Carbon pumps in the Southern Ocean are impacting global climate

Ulrich Bathmann and Uta Passow

Introduction

The Earth's climate is changing rapidly. Data series available for over 35 years have documented a response to this climate change in the terrestrial biosphere, the ice (cryosphere) and the world's oceans. Climate models show that the changes are more pronounced in Polar regions. However, because of the high sensitivity of Polar regions to climatic changes and insufficient data coverage, the different global climate models also vary most at the Poles,

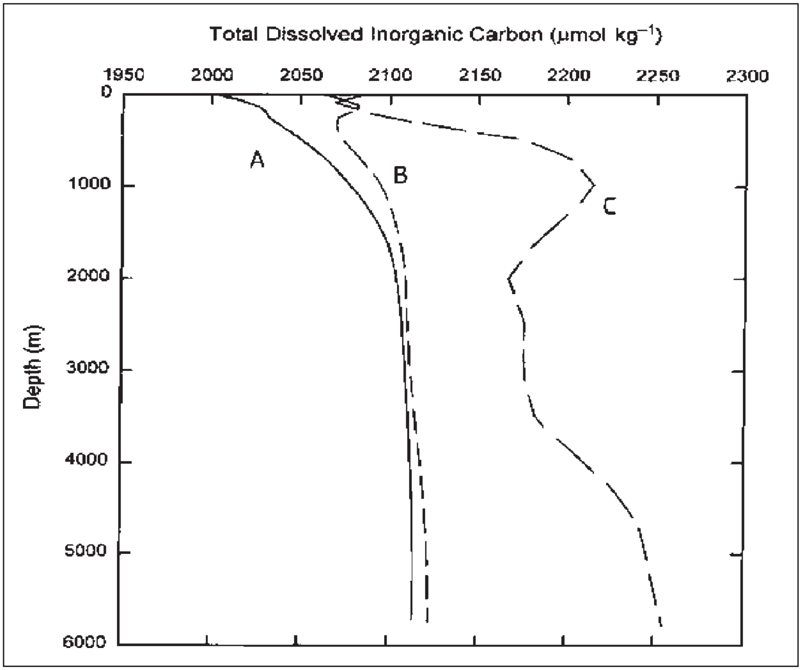


Fig. 1:
Vertical profiles of dissolved inorganic carbon in the ocean showing increasing concentrations with depth. Line A (pre-industrial level) represents the theoretical Dissolved Inorganic Carbon (DIC) distribution in the ocean at 280 ppmv atmospheric CO_2. Line B represents the theoretical distribution given 380 ppmv, representative for the year 1955 without considering biological pumps. The difference between lines A and B represents the ocean uptake of anthropogenic CO_2 without biological reactions. Line C reflects the actual present state and is created based on measurements from the Pacific Ocean. The difference between B and C is an indication of the high efficiency of biological transport of carbon in the ocean. (ppmv = Parts per million in volume). Figure modified after Raven and Falkowski (1999).

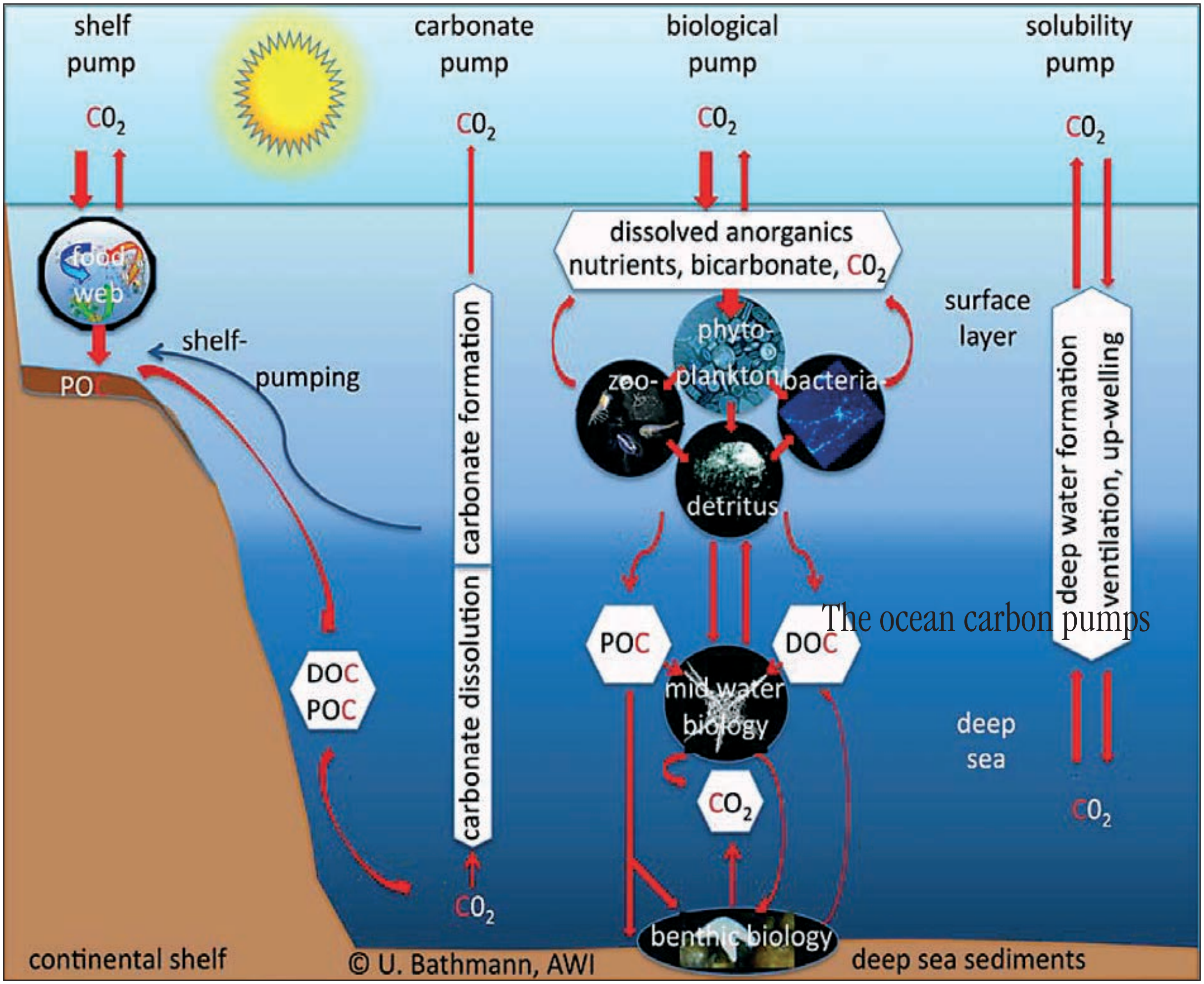


Fig 2:
The carbon pumps in the ocean. From right to left: The solubility pump is based solely on chemical-physical principles as CO_2 is taken up by water when it cools. Down- and upwelling processes ventilate the ocean interior. Biological production, consumption and sedimentation are the major drivers for the biological pump. In the surface ocean, CO_2 is fixed by photo-synthesis and only partially remineralised by heterotrophic processes. The "left-overs" sink to ocean depths. Only the fraction of carbon that is not consumed by mid-water or deep-sea benthic organisms gets buried in the ocean sediments. The carbonate pump describes the biological production of $CaCO_3$ and the sedimentation of these $CaCO_3$ shells. During the production of calcium carbonate CO_2 is released into the ocean, in contrast to the effect of the soft tissue pump where CO_2 is utilised. Carbonate dissolution e.g. in the deep ocean removes CO_2 from the water. The shelf pump describes the physical processes that move organic carbon sedimented in shallow shelf areas down slope into the deep sea, where heterotrophic processes (e.g. bacterial activity) releases CO_2. POC – particular organic carbon; DOC – dissolved organic carbon; CO_2 – carbon dioxide.

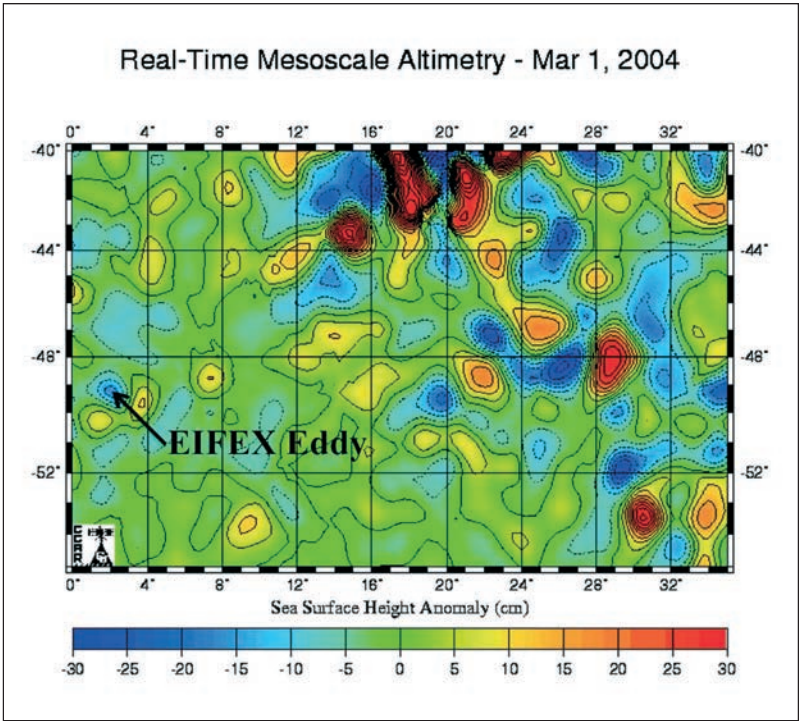


Fig. 3:
Mesoscale patchiness in a segment of the Southern Ocean as seen from a satellite. Sea surface height is influenced by ocean eddies. Clockwise rotating eddies create dimples (blue, downwelling) and anti-clockwise eddies result in surface pimples (red, upwelling). The height is given in centimetre against the mean sea level (green). This information was used during the European Iron Fertilization EXperiment EIFEX in 2004.

making predictions more difficult. Changes in sea surface temperature, sea-ice coverage of the polar oceans, atmospheric CO_2 and the oceans carbonate system (including pH) strongly effect ocean biology in many ways (see Gradinger, this book). This paper addresses processes that are responsible for the vertical transport of carbon from the surface layer of the ocean, which is in close contact with the atmosphere, down to the deep-sea sediments where carbon is stored on geological time scales.

We will describe the role of the different carbon pumps in the ocean by selecting examples from our own studies carried out in polar seas over the past two decades.

Role of ocean in the global carbon cycle

Atmospheric CO_2 is in equilibrium with CO_2 in the ocean surface and is constantly exchanged between the atmosphere and the upper ocean. Molecular diffusion and CO_2 dissolution in water are the major processes by which CO_2 is transported into the water when the pressure of CO_2 is higher in the atmosphere than in the surface ocean. As CO_2 dissolves in the ocean, it reacts with water to form other products (carbonic acid and its dissociation products, bicarbonate and carbonate ions). The sum of these different inorganic carbon compounds is called dissolved inorganic carbon (DIC).

As a consequence of these chemical reactions, the ocean can take up large amounts of CO_2 and store it as DIC. The global ocean stores 50 times more carbon (37500 Pg (1 Petagram = 10^{15} Gramm)) compared to the atmosphere (750 Pg), and small changes in the oceanic CO_2-reservoir result in significant changes in the atmosphere. A decrease in the pH of the ocean – termed ocean acidification

– is another consequence of these changes (see Auel *et al.*, this book).

The atmospheric CO_2 has increased drastically since the industrial revolution (about 200 years ago). The world's oceans have taken up a large fraction (about 30%) of this newly added anthropogenic CO_2.

In contrast to the surface water, deep water (approximately below 500 m) is removed from the atmosphere for, on average, 1000 years. This is the time span for deep water to be recirculated back to the surface layer via the ocean-wide circulation system called the conveyer belt (see Fahrbach *et al.,* this book).

Therefore, any carbon that leaves the upper mixed ocean layer (0-500 m) downward is "stored" for several hundred years in the deep ocean. There are different pathways by which carbon may be transferred from the surface ocean to the deeper ocean, called the carbon pumps.

The ocean carbon pumps

Four different processes mediate the exchange of carbon between the ocean surfaces and the deep-sea, where DIC generally is higher compared to the surface layer (Figs. 1, 2). The downward transfer of dissolved and particulate carbon is facilitated by various types of pumps, both physical and biological, mostly interacting with each other. The solubility pump transports the dissolved inorganic carbon (DIC) that sinks with the cold, saline surface water of the poles to extreme depths (2000-5000 m). This so-called downwelling happens when the surface water gets denser by cooling and by enrichment in sea-salts (due to surface water evaporation or formation of sea ice) (see Fahrbach *et al.,* this book). Cold polar water can carry more than twice as much CO_2 gas than warmer tropical water and thus, the carrying capacity for CO_2 in polar oceans and in the cold deep-sea exceed the capacity of tropical oceans.

This downwelling of cold, saline water in polar regions is compensated by warmer water flowing from the tropics polewards along the surface. It takes a water particle about 500 to 1000 years to complete the journey of the general large-scale ocean circulation. The solubility pump carries about one quarter to half of the DIC downwards into the deep ocean.

The transport of organic carbon to depths is called the biological soft tissue pump (Bathmann 1996). Particulate organic carbon (POC) is produced from DIC during photosynthesis. This primary production can only happen near the sunlit sea surface (see Noethig *et al.*, this book). The POC generated mostly by algae provides the basis for bacterial and animal (heterotrophic) life in the sea. Zooplankton consume the phytoplankton, forming their own organic carbon. Bacteria utilise and modify POC. Some of this organic carbon remains particulate, some is released as dissolved organic carbon (DOC) and most is respired back to CO_2.

The fraction of organic carbon that is respired in the surface ocean remains in equilibrium with the atmosphere on timescales of less than a year. But about 30% of the organic carbon is transported to depths below 500 m. Bacteria and heterotrophs living there utilize this organic carbon during its transit and CO_2 is released at depths (see Assmy *et al.*, this book), explaining the high DIC concentrations in deep waters. Only a small fraction, about 0.1 percent, of the originally produced organic material reaches the seafloor in the deep open ocean.

Organic carbon may be transported to great depths by different pathways. After feeding near the surface, some zooplankton migrate downward where they respire and defecate, thus releasing carbon at depths. A quantitatively more important mechanism is the direct sinking of POC to deep waters. Sinking velocity of particles depends largely on their size, and partially on their excess density. The sinking velocity of the tiny phytoplankton cells and faeces is too small to sink deep enough, for their carbon to be removed from the atmosphere for years. However, many cells release a substance called TEP (transparent exopolymer particles) which acts as glue (Passow *et al.* 1994). TEP bind many cells together forming aggregates more than 0.5 mm big, called marine snow (so-called because when diving in the ocean in the presence of these large aggregates, it looks like being in a snow storm). Marine snow is large enough to sink at significant velocities, reaching waters below 500 m. This is the main pathway by which the POC pump (also called soft tissue pump) transports carbon to depths.

The pumping efficiency e.g. the fraction of POC that reaches great depths depends on the sinking velocity of marine snow and other large particles, because the faster the organic carbon sinks, the less it is utilised by bacteria or eaten by animals (De La Rocha and Passow 2007). The presence of calcium carbonate or opal-rich shells produced by marine organisms in marine snow is thought to increase the sinking velocity.

The carbonate pump cycles carbon when organisms produce calcium carbonate shells. Its effect on the ocean carbonate system is opposite to that of photosynthesis.

Whereas photosynthesis removes CO_2 from the water (by fixing it into organic carbon), the carbonate pump generates CO_2 during the production of calcium carbonate by organisms such as corals, coccolithophorids or foraminifera. The release of CO_2 during calcification is counterintuitive, but can be explained with the equilibrium between CO_2 and the other forms of DIC e.g. with the carbonate system of the ocean. The shelf pump describes the physical movement of organic matter present in the shallow sediments of the highly productive shelf areas to the open ocean. The shelf sediments are rich in organic carbon. They may be resuspended by strong ocean bottom currents and organic carbon transported down-shelf by internal waves or other processes. On average, the shelf pump transports about 1 Pg carbon per year into the deep sea.

Programmes in the Southern Ocean

Since 1989, we carried out 15 expeditions into polar seas working on details of the carbon pumps. One of us (UB) was at sea for almost 800 days to perform experiments and carry out field work for these studies. One major focus was the functioning of the biological pump under different environmental conditions.

We concluded that the efficiency of the biological pump to transport carbon to the deep ocean depends largely on the organisms involved, or in other words, on the biodiversity and the structure of the biological systems.

Several cruises in different seasons were organised under the umbrella of the Joint Global Ocean Flux Study (JGOFS), an international programme focusing on the biological pump in different parts of the world ocean.

During one RV *Polarstern* cruise in austral summer 1995/96, we investigated the physical, chemical and biological patterns across the Antarctic Polar Front between South Africa and the Antarctic continent (Strass *et al.* 2002).

Discrete water samples were taken along north-south transects and two ocean "boxes" at the edge of the Polar Front. Along this front, which separates subpolar from polar surface waters, we found meanders with lengths of about 50 km and distinct patches rich in phytoplankton biomass.

These patches merged to a larger "green" band where the frontal structures aligned with a cold core ocean eddy. Complex hydrographical processes resulted in mesoscale upwelling and downwelling several tens of kilometres apart. Fig. 3 shows an example of such variability in the environmental conditions of open ocean water on spatial scales of 50 km. Stratified areas were side by side to areas characterized by strong vertical mixing. This is of great importance for phytoplankton growth, as phytoplankton can only grow effectively in the stratified areas where cells remain near the light at the surface of the ocean. Deep vertical mixing or downwelling, will result in an average low light climate for each cell, therefore prohibiting the build up of phytoplankton biomass. This cruise taught us that even the open Southern Ocean is intensely structured at temporal and spatial scales of typical ocean eddies, resulting in the concomitant variability in the efficiency of the pumps.

Large spiny diatoms (see Assmy *et al.*, this book) that dominated the biomass in some of these areas sink in large numbers, and their presence suggests an efficient biological pump. They are protected against grazing by their heavily silicified cell frustules, which are hard to break by their predators (copepods). The long spines, the elongated shape of the cells, and the formation of long chains further make these algae difficult to handle by the grazing zooplankton. As a result, these cells do not get eaten up, but form marine snow and sink down to depths. The underlying sediment in 5000 m water depth is composed up to 95% of such diatoms.

We have mentioned the light regime, and thus mixing, and grazing as important factors determining how much organic carbon can sink to great depths. High grazing rates imply that a large fraction of the organic carbon will be respired and recycled near the surface. Low light levels for phytoplankton result in low photosynthetic rates, and thus low production rates of organic carbon, and hence a weak biological pump.

Besides light and macronutrients, phytoplankton also require micronutrients for growth, and in large areas of the otherwise nutrient-rich Southern Polar Ocean, iron is the key nutrient that limits growth, especially of the large diatoms mentioned above (see Assmy *et al.,* this book).

Potential of the biological pump

Iron fertilisation in the Southern Ocean has been proposed to boost phytoplankton production and consequently the export of organic material by means of the Biological Pump (Martin *et al.,* 1990, Assmy *et al.,* this book). Ocean iron fertilisation (OIF) however, is a contro-versial issue (Browman and Boyd 2008, and 12 papers therein) – partly for fear that it might be misused in the political discussions about the need for rapid reduction of CO_2 emissions.

On the other hand, increased primary production triggered by artificial iron fertilisation in the Southern Ocean has been thought to have a positive effect on the zooplankton which might in turn benefit the populations of whales, seals and penguins (Smetacek and Naqvi 2008).

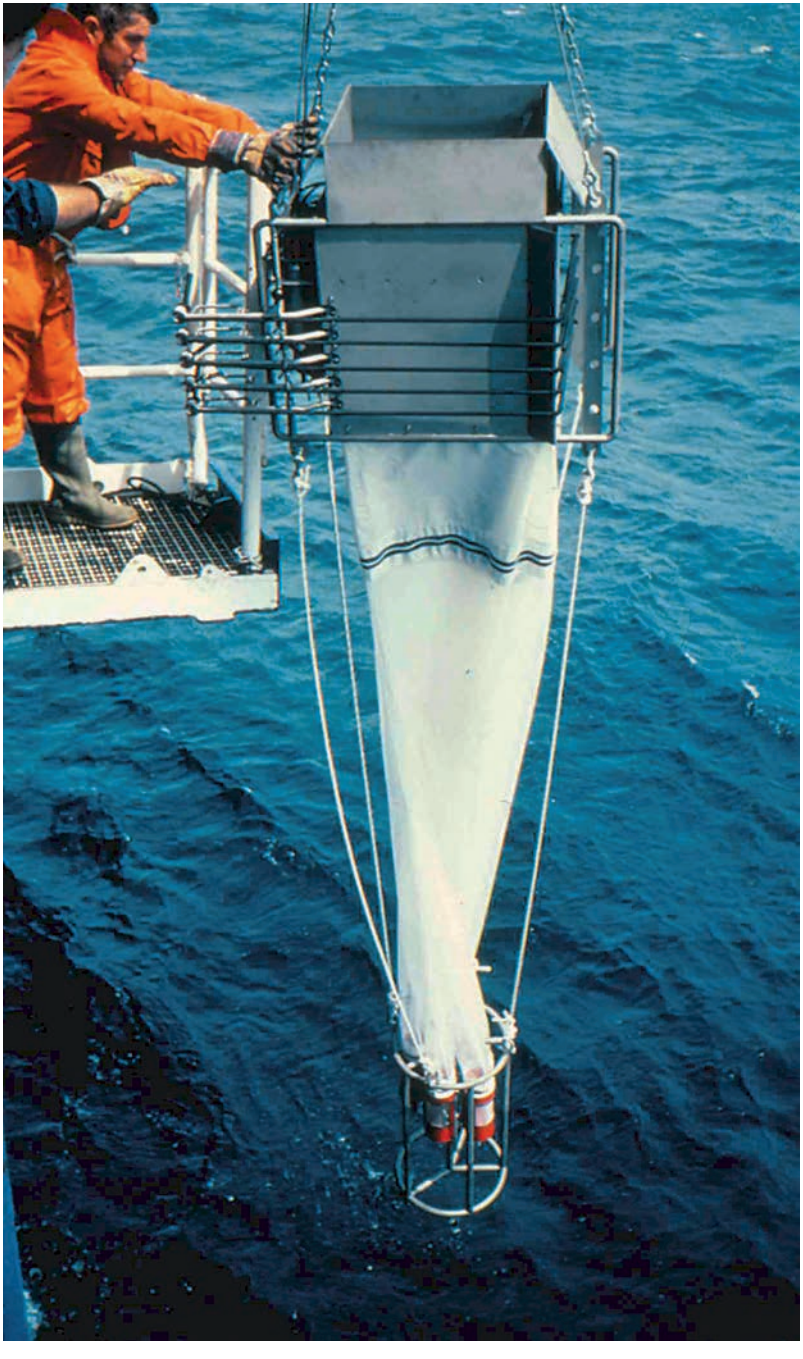
In the Southern Ocean, several iron fertilisation experiments have been conducted. But until 2009, only EIFEX (Bathmann 2005) (Fig. 3) was sufficiently long in the field to record the imprint of newly produced organic matter from the surface ocean to the deep-sea floor (Sachs 2008). The results of a further experiment in January to March 2009 (LOHA-FEX) are not yet available.

Preliminary data suggest that the additionally produced phytoplankton seems to have been grazed down by zooplankton and the carbon has instead been transferred into the higher trophic levels of the marine pelagic food web.

From these experiments, we concluded that iron fertilisation (artificial or natural) in the Southern Ocean might stimulate new carbon production in areas of stable hydrographical conditions that are also rich in macronutrients (Assmy *et al.,* this book).

Depending on phytoplankton species composition at the onset of the experiment, the physicochemical conditions of the upper mixed layer and the grazing pressure, the fate of the newly fixed carbon might be quite different (Bathmann 2005): POC might sink in reasonable quantities to the deep ocean or POC might end up in the pelagic food web.

The appropriate hypotheses to be tested *in situ* in the future are summarised by Smetacek and Naqvi (2008): i) there was alternating desertification of land and ocean mediated by rainfall and dust in earth geological history and the information is stored in sedimentary proxies; ii) the outcome of sustained fertilisation of nutrient-rich iron-poor waters will be determined by plankton species composition; iii) neritic diatom species draw down more carbon (associated with less silica sinking) compared to oceanic species; and iv) OIF will



The Multinet consists of five or more nets which can be opened and closed sequently to sample subsequent depth layers while the gear is vertically hauled. Photo:W Hagen produce more food for grazers and, thus, can reverse the ongoing krill decline.

Industrial up-scaling of OIF as a measure for CO_2 mitigation implies serious scientific problems. They are well addressed by Buesseler et al. (2008). The effects are unknown when performing OIF under different nutrient regimes or in areas of different populations of phytoplankton species or zooplankton grazers. The ecological impacts of OIF to fish, seabirds, and marine mammals are not well investigated. Oxygen concentration in the ocean might be altered by OIF as well as distributions, feedbacks, and cycling of non-CO_2 greenhouse gases, such as methane, nitrous oxide and dimethylsulfide. Ocean acidification may be hastened. Thus, the long-term effects of OIF need to be monitored and mathematical models being used to extrapolate effects beyond the study area and observation period have to be reviewed and improved.

In summary, the carbon pumps in the ocean link atmospheric and marine processes and are important for the global climate. The closer researchers investigate the mechanisms of the different pumps, the more complex the picture gets. Currently, the ocean is buffering anthropogenic stimulated climate change to a great extent, but we cannot precisely estimate for how long and what will be the developments in the next decades.