Marine nutrient cycling - How will the ocean's capacity of biological carbon pumping change?

PRESENT | PAST

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The biological pump refers to a suite of biologically mediated processes that transport carbon from the ocean's surface layer to its interior. Its efficiency depends on the balance between the rates of carbon photo-assimilation, export and mineralization. Our knowledge of the biological carbon pump relies on our mechanistic understanding of factors structuring phytoplankton distributions and marine food webs, and the associated biogeochemical cycles. To assess the extent to which the strength and the efficiency of this pump will change in the future, we need to know how these factors – light, nutrients and temperature – might change in a warmer ocean. Models coupling an ecosystem module to a global circulation model provide important tools for understanding the dynamics of the carbon pump and its response to warming. But as pointed out by Sarmiento et al. (2004), existing tools are still not mature enough to allow this.

The last decade has seen increasing awareness of the relationship between key phytoplankton groups and their pivotal role in the functioning of the biological carbon pump (e.g. Boyd et al. 2010). Until recently, the picture was of a simple subdivision between efficient carbon export via a diatom-copepod-fish linear food chain in nutrient-rich waters and retention of surface carbon in nutrient-poor waters via a microbial network initiated by the ubiquitous pico/nano phytoplankton (Chisholm 2000). This picture has been recently complicated by the recognition of two additional small-sized players – the calcifying coccolithophores and the nitrogen-fixing cyanobacteria. These have a competitive advantage over diatoms in warm, well-illuminated surface waters supplied with imbalanced inorganic nitrogen and phosphorus nutrients. Their participation in carbon export is indirect and involves either aggregation with calcium carbonate liths acting as ballast particles or the release of nitrogen that sustains the growth of concomitant diatoms, thereby triggering carbon export (Chen et al. 2011)

Ocean warming affects the pelagic ecosystem both directly and indirectly, by increasing temperature and stratification. The latter tend to favor the dominance of small phytoplankton (e.g. Falkowski and Oliver 2007; Li et al. 2009) over large cells such as diatoms (Fig. 1). The resulting photo-assimilated carbon benefits the heterotrophic microbial food web, whose activity is stimulated by the warmer temperature (Sarmento et al. 2010), increasing the rate of carbon mineralization. Small phytoplankton cells also have low sinking rates. The spreading of such cells anticipated with increased ocean stratification will decrease the overall capacity of the biological pump, but the extent remains uncertain (Barber 2007).

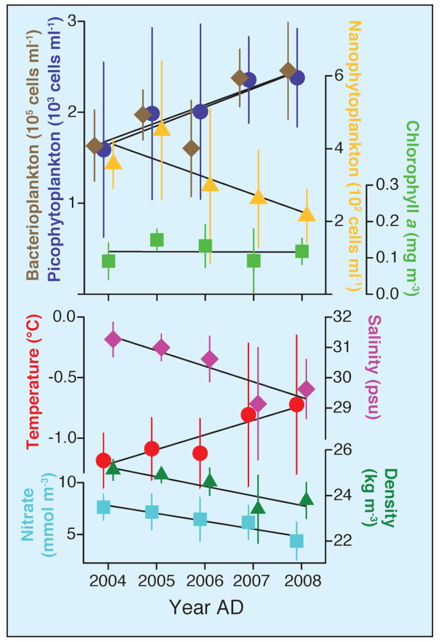

Figure 1: Summer conditions in the upper water layers of the Canadian Basin. The lower panel shows the physical water properties over the period 2004-2008, the upper the response of the plankton organisms during the same period. Figure modified from Li et al. 2009.

The prevalent view of the picophytoplankton carbon being totally remineralized in the surface waters has been recently challenged by data from the equatorial Pacific Ocean and Arabian Sea, which point to significant export of picophytoplankton-related carbon through indirect paths such as aggregation and fecal pellets (Richardson and Jackson 2007). The future latitudinal extent of coccolithophore blooms due to warming is unknown, however, particularly in light of the possible alteration of calcification rates by ocean acidification (Cermeño et al. 2008).

The potentially large ocean deoxygenation due to the increased temperature and stratification projected for a warmer ocean (Keeling et al. 2010) will have direct consequences for marine biota, but only an indirect effect on ocean productivity and nutrient and carbon cycling. An expansion of suboxic/anoxic conditions would increase the release of phosphate and iron from sediments while some reactive nitrogen would be eliminated by denitrification or anaerobic ammonia oxidation. The subsequent shift in the ocean nitrate-to-phosphate balance will affect the composition and productivity of marine organisms, notably diazotrophic cyanobacteria, with uncertain consequences for the efficiency of the biological pump.

On the whole, the smallest phytoplanktons seem to have a competitive advantage in a warmer ocean. In contrast, diatoms are at an advantage in surface waters with transient nutrient pulses. In particular, they will benefit in coastal regions from stronger wind-driven upwelling events that are expected from increased storm and frequency resulting from climate warming. Improving the evaluation of changes to the biological carbon pump via ocean models is currently hampered by several uncertainties on mechanisms controlling phytoplankton dominance and food-web structures. Also, many global circulation models remain coarse in resolution and don't serve high frequency forcing. To be able to predict better the efficiency of the pump under future conditions, the complexity in biology needs to be matched with an appropriate complexity in the representation of the physical and chemical environment in ocean models.

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Changes in oceanic carbon storage have been hypothesized to be the cause of the ~100 ppm variations in atmospheric CO_2 between glacials and interglacials (Sigman and Boyle 2000). If the ocean stores more carbon in colder than in warmer climates this implies that a positive feedback exists: as climate warms the ocean releases carbon, which increases atmospheric CO_2 and amplifies the original warming. However, presently we don't know how much ocean carbon storage changed in the past and why.

Carbon isotope data from Last Glacial Maximum (LGM, 19-22 ka) sediments indicate that more carbon was stored in the deepest ocean layers, particularly in the Atlantic (Fig. 1). $\delta^{13}C$ is fractionated during carbon uptake by phytoplankton, which favors the light isotope ^{12}C. Its distribution in the deep ocean is therefore determined to a large degree by the efficiency of the biological carbon pump, with lower values indicating more respired carbon.

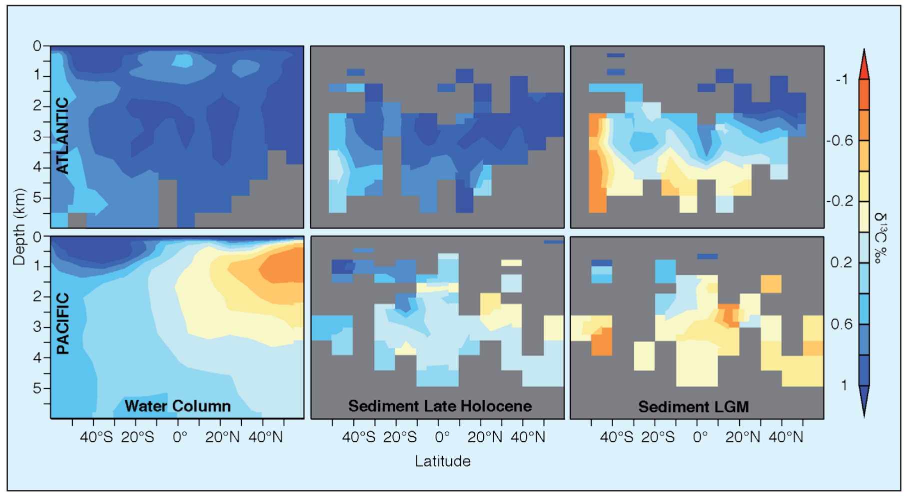

*Figure 1: Latitude-depth distribution of zonally averaged $\delta^{13}C$ in the Atlantic (**top**) and Pacific (**bottom**). **Left**: modern water-column measurements from WOCE and CLIVAR cruises (Schmittner et al., unpublished data). Middle: Late Holocene sediment data. **Right**: LGM sediment data. Sediment data from Hesse et al. (2011) and Matsumoto et al. (2002).*

In the modern ocean deep waters in the North Atlantic are high (+1‰) in $\delta^{13}C$ and quite homogenous, corresponding to sinking of low nutrient, high oxygen surface waters. $\delta^{13}C$ values are lower in the Southern Ocean (+0.5‰) and decrease towards the North Pacific (–0.6‰), reflecting today's ocean's oldest waters with high nutrients and carbon and low oxygen.

During the LGM, the deep Atlantic had much larger vertical gradients than today, $\delta^{13}C$ values measured on microfossil shells of benthic foraminifera were up to 1‰ lower below 2-3 km depth, but similar above 2 km in the north (e.g. Curry and Oppo 2005). The deep Pacific Ocean was also higher in $\delta^{13}C$ particularly in the south (Matsumoto et al. 2002). In contrast to the North Atlantic, however, the North Pacific did not exhibit larger vertical gradients. The coherent large-scale differences in $\delta^{13}C$ suggest that biological carbon storage in the glacial ocean was most likely higher than today. But how much higher was it and why?

Biological, physical and chemical processes determine biological carbon storage in the ocean. It is likely that more than one process must be invoked to explain glacial to interglacial changes (Köhler et al. 2005). Increased solubility of CO_2 in colder water explains less than 20 ppm of the full glacial-interglacial difference. The increase in the vertical gradient of $\delta^{13}C$ in the Atlantic may require changes in the circulation such as a shoaling of the southward flowing deep waters or a change in the rate of northward flowing bottom waters. Furthermore, in a simple model of Bouttes et al. (2009) increased brine rejection from sea ice around Antarctica and its effect on deep ocean stratification lowered atmospheric CO_2 by ~42 ppm, but the effect will need to be reproduced with more realistic models. Large changes in the contributions of northern versus southern sources to the global deep water can also change the efficiency of the biological pump (Martin 1990), but did not contribute much to the glacial-interglacial atmospheric CO_2 changes.

The efficiency of plankton to use nitrate and phosphate may have been enhanced by more iron input to the surface ocean by higher dust deposition (Brovkin et al. 2007, estimate 37 ppm). A glacial dust plume from Patagonia may partly explain lower $\delta^{13}C$ in South Atlantic bottom waters but the increase in aeolian iron input may have been counteracted by a decrease in sedimentary sources due to lower sea level (Moore and Braucher 2008). It is also possible that the biologically available (fixed) nitrogen inventory of the glacial ocean was overall higher than today because denitrification was lower due to higher dissolved oxygen concentrations in the colder glacial ocean and reduced continental shelf area due to the sea level drop. However, these processes have not been quantified yet with a realistic 3-dimensional model.

Some or all of the processes that controlled changes in the glacial-interglacial ocean carbon storage may also be important for our warming planet. Decreased CO_2 solubility in a warming ocean will certainly occur. However, how important some of the other processes will be in the future is more uncertain. Better understanding of how and why ocean carbon storage varied in the past and in the future may now be possible due to coordinated international modeling projects and efforts to synthesize and increase the spatial coverage of paleoclimate data.

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