Box 2.4. The Marine Carbon Cycle

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Carbon in the oceans resides in dissolved organic carbon (DOC), dissolved inorganic carbon (DIC), particulate organic carbon (POC) and in marine biota. Most of the carbon in all the ocean layers is in the form of DIC but less than 1% of this carbon is gaseous carbon dioxide (CO_2). Around 90% of the inorganic form is bicarbonate ions and the rest is in the form of carbonate ions. These three forms are in a pH-dependent equilibrium. The capacity of the ocean's bicarbonate system to buffer changes in CO_2 is limited by the addition of calcium and, to a lesser extent, magnesium from the slow weathering of rocks. Thus, the ability of oceans to absorb excess CO_2 on short time scales is constrained.

Marine biomass makes up less than about 2 Pg C in the oceans. Despite this fact, marine biota are globally as productive as terrestrial systems (Falkowski et al. 2000). However, the biota do not accumulate biomass in carbon-rich structures in the way that land plants do. Primary production in the photic zone, approximately 50 Pg C per year globally, is the major input of organic carbon to the marine ecosystem. Over the past decade much has been learned about primary producers as well as what controls their distribution, rate of carbon uptake and turnover areas.

The carbon fixed at the surface is transported only slowly to deep ocean layers. Most of the carbon in the oceans is in the intermediate and deep waters. Once there, estimated lifetimes are approximately decades to centuries. The higher concentrations of inorganic carbon in these deeper layers in all basins result from the combined influence of two fundamental processes, known as the solubility and the biological pumps (Fasham et al. 2001), which operate at different rates temporally and spatially in diverse regions of the oceans. Thus, the strength and magnitude of these two pumps gives rise to regions of natural sources and sinks of carbon in the oceans.

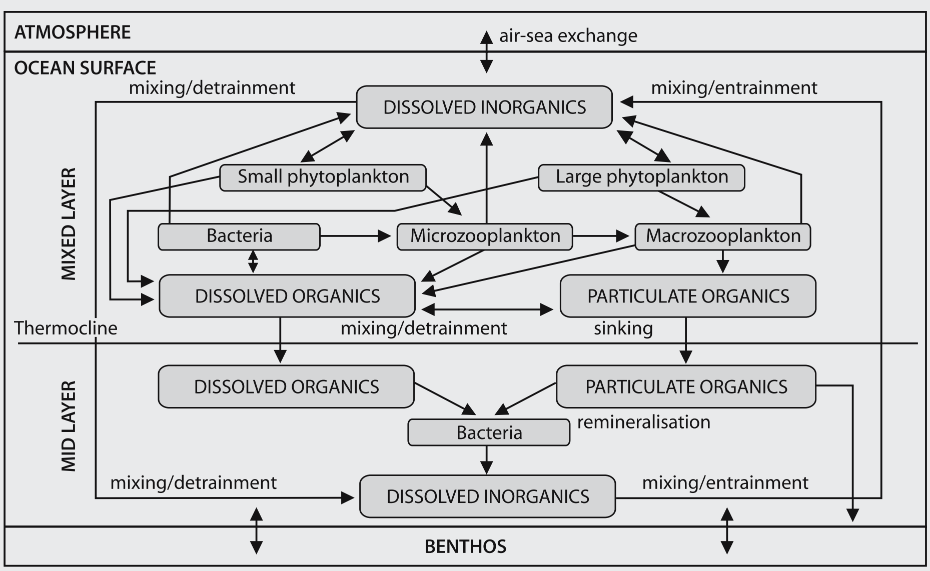
The solubility pump operates as a result of thermohaline circulation and latitudinal and seasonal changes in ocean ventilation. Being more soluble in cold, more saline waters, CO_2 is sequestered to deeper waters by the formation of cold, dense water at high latitudes. The most intense natural sink region in the world is the North Atlantic, where the Gulf Stream and North Atlantic Drift transport warm water northward. There it cools and sinks, increasing the solubility of CO_2 from the atmosphere and removing it from the surface to deeper waters. Off Scandinavia and Canada, the dissolved CO_2 and fixed carbon in these waters sinks to great deeps over 500 m and is subsequently transported south along the bottom.

The biological pump (Fig. 2.29) operates through the biological assimilation of dissolved CO_2, which enhances the air-sea gas exchange and contributes to the vertical transport of carbon via physical processes. It functions through a complex food web of small plankton, which primarily recycle CO_2 within the photic zone, and larger plankton, which generate most of the particulate and dissolved forms that sink to the deep ocean. Approximately 25% of the carbon fixed by phytoplankton globally in the upper ocean layer sinks to the interior (Falkowski et al. 1998), where it is oxidised and recycled by bacteria and other heterotrophs into dissolved inorganic and organic carbon forms. Only a small percentage (1–2%) reaches the ocean floor. The deep benthos consumes and recycles most of what falls from the interior while the rest is buried in ocean sediments.

The magnitude of vertical flux from the surface layer to the deep ocean depends on the pathway and food web. Food webs dominated by large phytoplankton and macrozooplankton produce the largest vertical export flux to the deeper parts of the ocean (Fasham 2003). Phytoplankton species, which form silicate (opal) or carbonate shells, contribute to what is sometimes referred to as the silicon or carbonate pumps. There is no best way to estimate the magnitude of the biological pumps. However, several global estimates agree quite well and set the transfer rate at about 10 Pg C yr^{-1} (Falkowski et al. 2000, Laws et al. 2000, Schlitzer 2002). Recent findings suggest that the phytoplankton species involved in this process may hold a key to the rate and potential for the flux to change.

Once carbon sinks to the ocean interior via the biological and solubility pumps and then is transported laterally, dissolved CO_2 in the water is effectively prevented from re-equilibrating with the atmosphere until transported back to the surface later. Along the equator, vigorous upwelling occurs, which warms the deep water as it rises to the surface, decreases the solubility of CO_2 and thus releases it to the atmosphere. Along many diverse ocean margins, intense seasonal upwelling occurs and supports a strong phytoplankton and heterotrophic foodweb, referred to as a continental carbon pump. Recent evaluation of the magnitude of this pump indicates that these food webs absorb upwards of 0.2–1 Pg C yr^{-1}, a significant component of the global carbon cycle (Fasham 2003). In many regions, eastern and western ocean boundary currents sweep carbon deep offshore, while in high latitude margins, cold water sinks into the intermediate layer of the open ocean. The magnitude of the lateral and the vertical carbon flux at the seafloor, which is less than 0.2 Pg C yr^{-1} globally, agrees well with benthic respiration, sediment trap and primary production data (Fasham 2003).

Fig. 2.29.
The biological pump (conceptual diagram from the Joint Global Ocean Flux Study (JGOFS) International Project Office, Bergen, Norway)



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