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The oceanic carbon sink - processes, timescales, and impacts

REDUCED VERSION OF TALK AS GIVEN AT GIFT WORKSHOP

Christoph Heinze
christoph.heinze@gfi.uib.no
University of Bergen

Norwegian Centre of Excellence

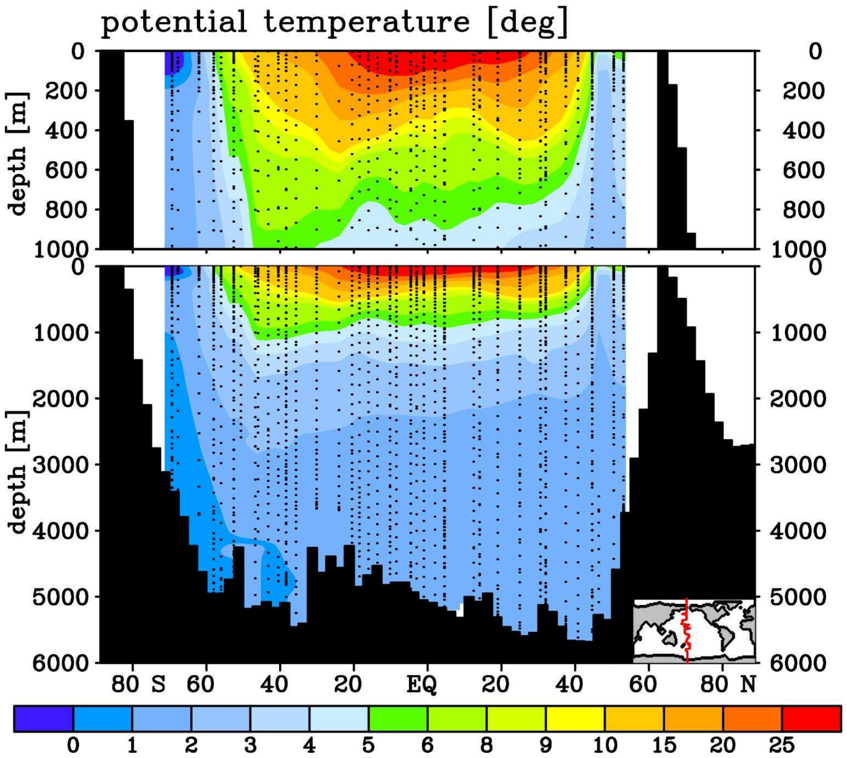
CARBOOCEAN

CarboSchools

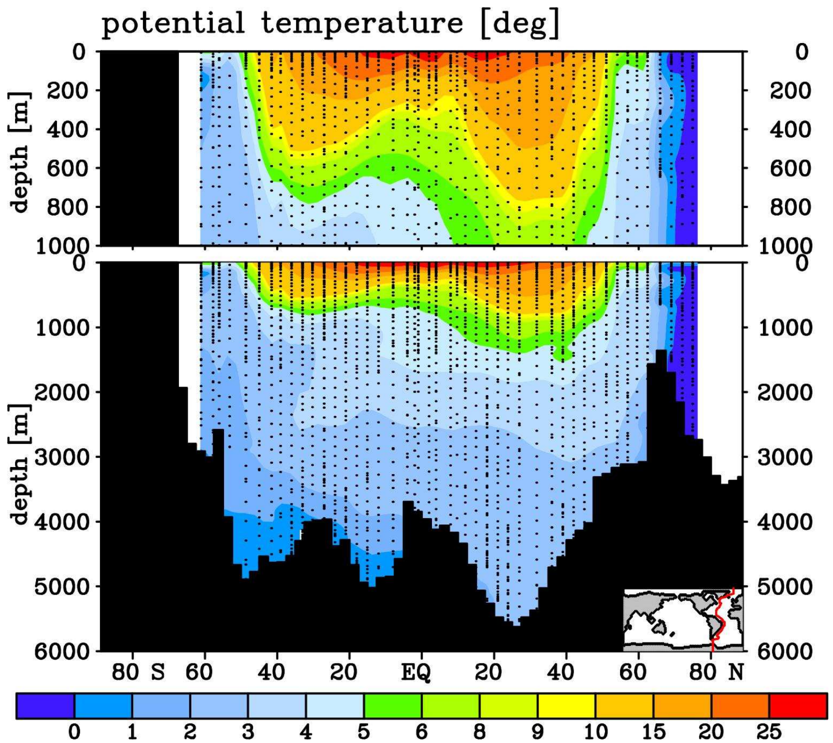
NERSC

Nansen Center, Bergen

Most of the oceanic water column is cold (below 4 degrees Celsius) and deep (3-4 km).



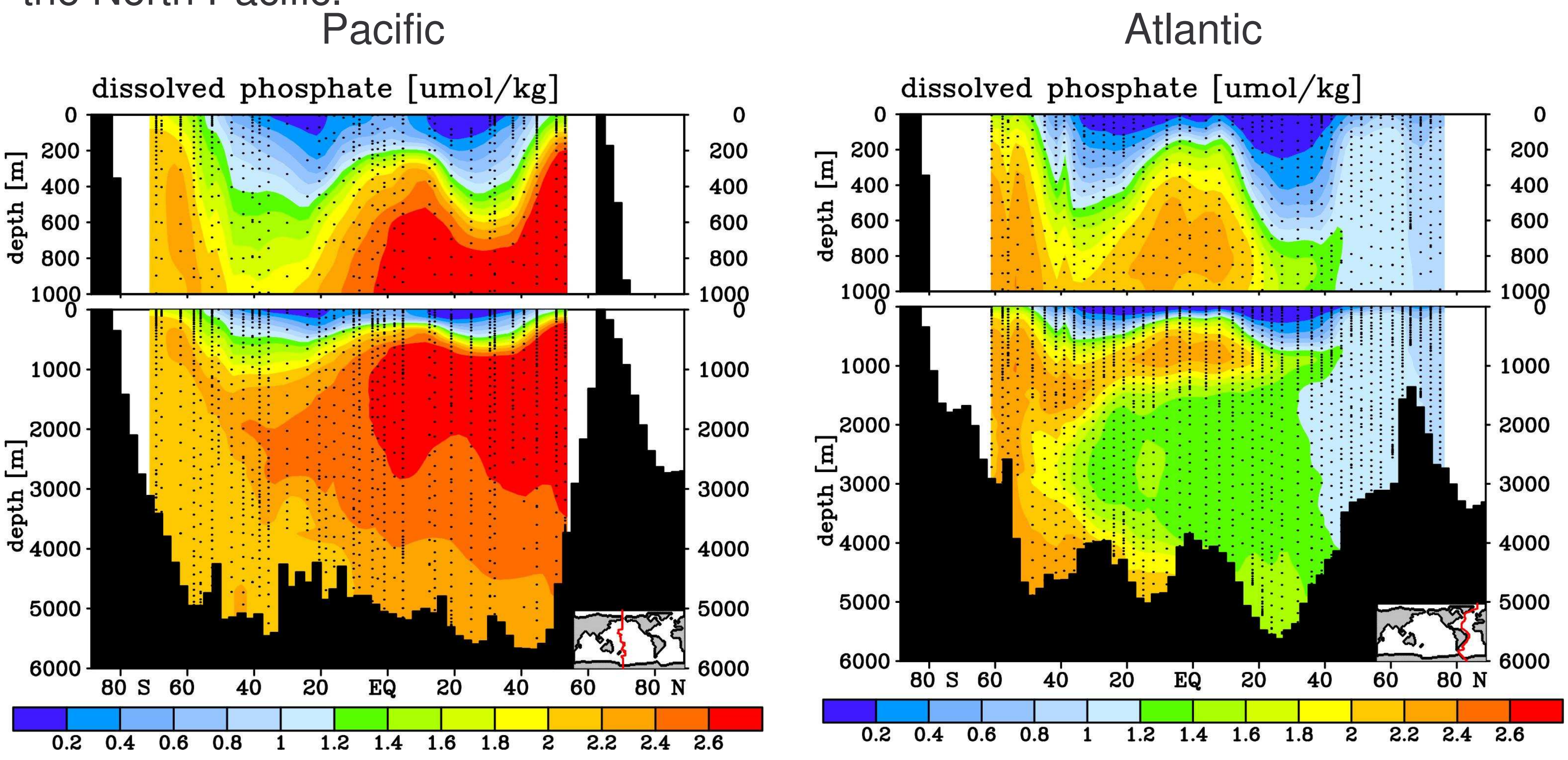
Pacific



Atlantic

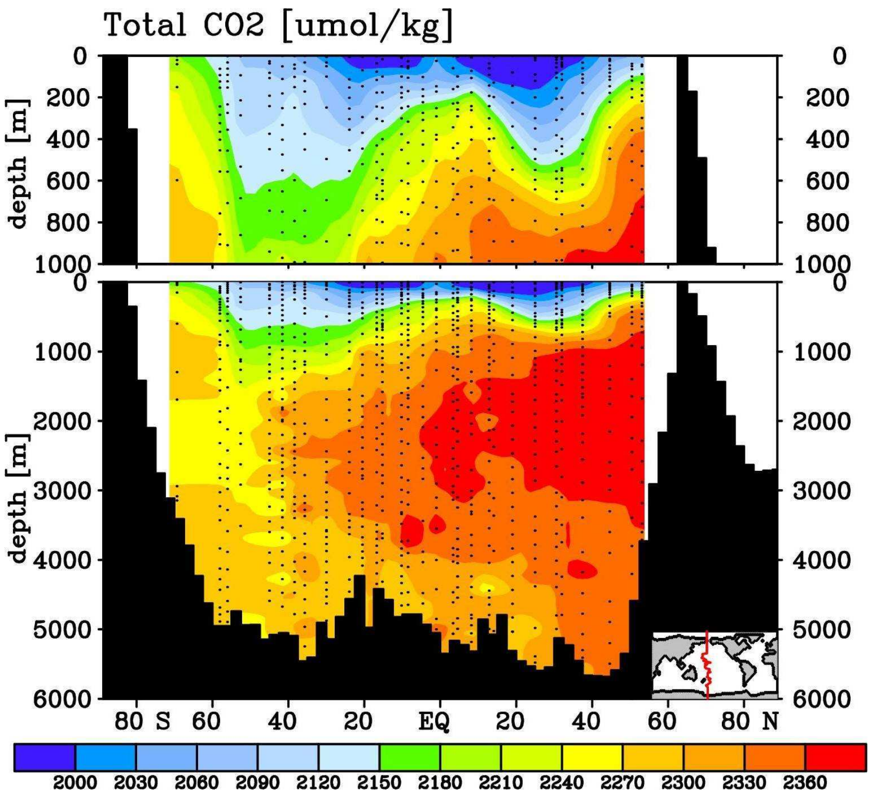
GEOSECS observations (Bainbridge et al., 1981, Broecker et al., 1982)

The ocean large scale circulation cannot directly be seen from the temperature distribution. However, dissolved phosphate (a nutrient) shows how the ocean works. The phosphate distribution results from a superposition of circulation, biological surface production, particle rain through the water column, and remineralisation). The ”older” the water is, the higher the nutrient concentration. The turnover time of the ocean is ca. 1500 years. Deep water is produced in the North Atlantic and the Southern Ocean. The oldest water is found in mid-depth in the North Pacific.

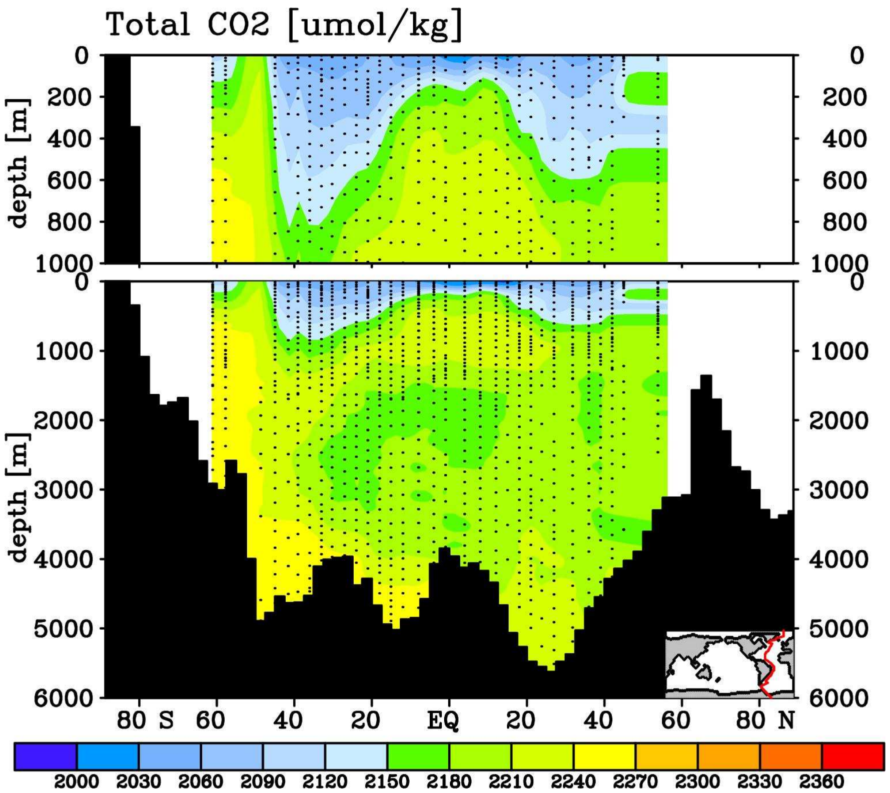


GEOSECS observations (Bainbridge et al., 1981, Broecker et al., 1982)

The dissolved inorganic carbon in the ocean resembles the pattern of the nutrient phosphate. Note however, that the carbon concentration never goes to zero at the surface.



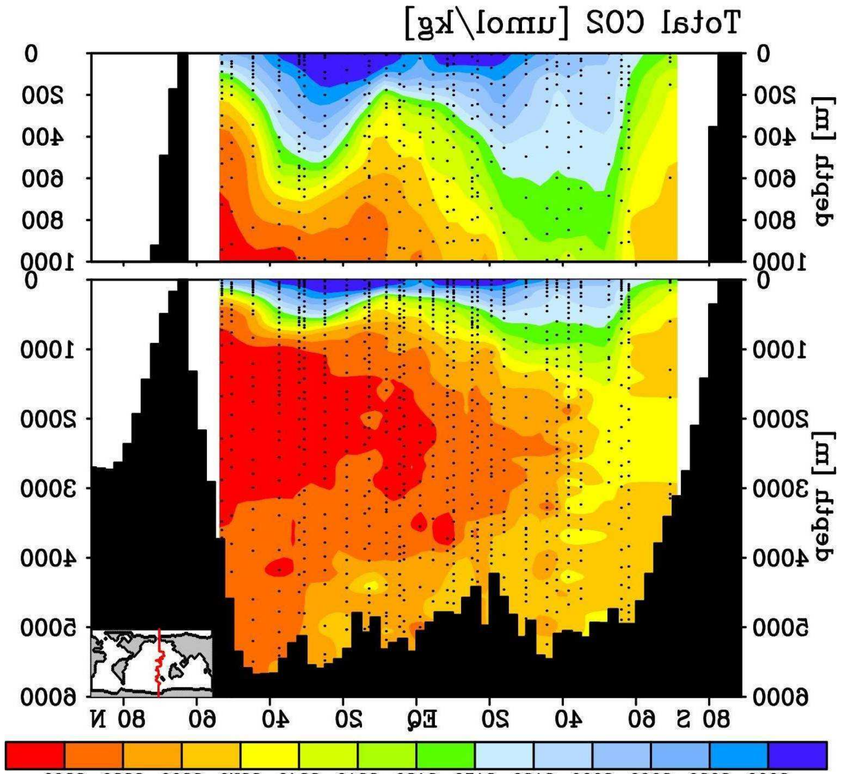
Pacific



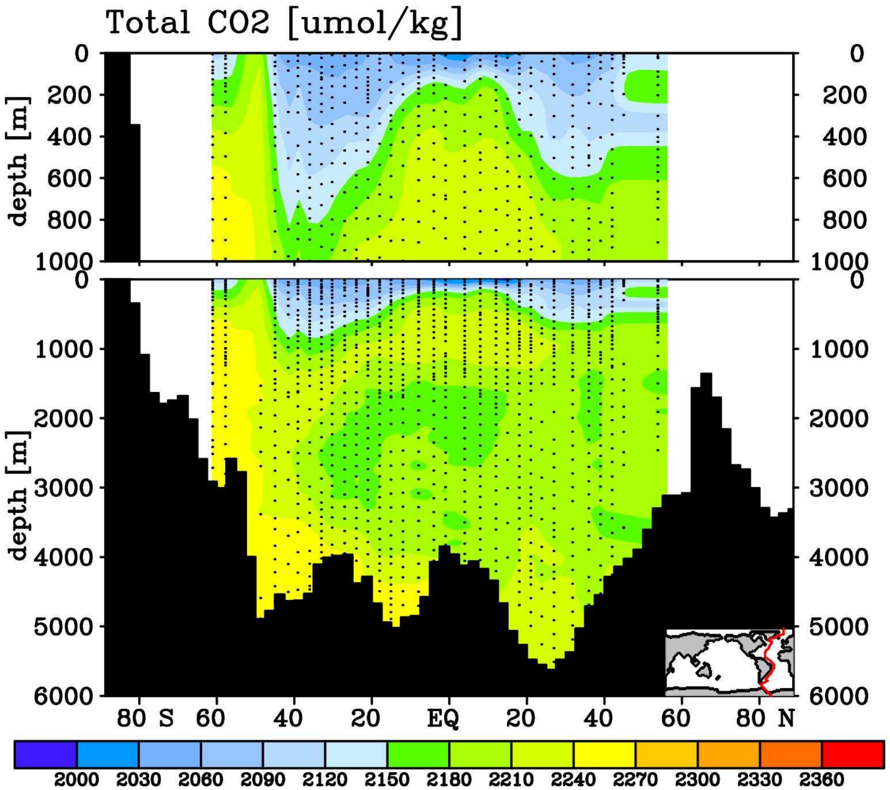
Atlantic

GEOSECS observations (Bainbridge et al., 1981, Broecker et al., 1982)

The oceanic conveyor belt circulation can be seen even better when we flip the Pacific cross section around, so that the section from Atlantic and Pacific ”meet” at the Southern Ocean.



Pacific



Atlantic

GEOSECS observations (Bainbridge et al., 1981, Broecker et al., 1982)

CO_2 in seawater is reactive in contrast to atmosphere, definitions of carbon species (free CO_2, bicarbonate, carbonate, DIC, Alk)

Difference!

Atmosphere:

CO_2 is
a fairly inert gas

Low storage capacity

Ocean:

CO_2 reactive $CO_2:HCO_3^-:CO_3^{2-}$

1 : 100 : 10

Large storage capacity

Reason – seawater alkalinity
"ability to dissociate weak acids"

main contributor to alkalinity changes:

$Ca^{2+}CO_3^{2-}$ shells

$CO_2+H_2O+CO_3^{2-} \rightarrow 2HCO_3^-$

Carbon in seawater

CO$_2$ solution → hydration → 1st dissociation step → 2nd dissociation step

$$CO_2^{gas} + H_2O \leftrightarrow H_2CO_3^{aq} \leftrightarrow HCO_3^- + H^+ \leftrightarrow CO_3^{2-} + 2H^+$$

carbonic acid ($H_2CO_3^{aq}$), bicarbonate ion (HCO_3^-), proton (H^+), carbonate ion (CO_3^{2-})

total alkalinity, seawater property determining dissociation of weak acids

$$TA = [HCO_3^-] + 2[CO_3^{2-}] + [B(OH)_4^-] + [OH^-] - [H^+] + \text{small terms}$$

$$DIC = [CO_2 + H_2CO_3] + [HCO_3^-] + [CO_3^{2-}]$$

dissolved inorganic carbon

If 2 of the "green" variables are known, all the others can be computed.

Partial pressure of CO_2:

pCO_2 x solubility = CO_2

Carbon in seawater: CO_2 added

Principal effect is formation of bicarbonate using up carbonate:

$$CO_2^{gas} + H_2O + CO_3^{2-} \leftrightarrow 2HCO_3^-$$

$$\textit{some} \nwarrow\!\!\searrow \; H^+ + HCO_3^- + CO_3^{2-}$$

Some CO_2 reacts with water:

$$CO_2^{gas} + H_2O \leftrightarrow H^+ + HCO_3^-$$

Net effect: less CO_3^{2-} and more H^+

More H^+ means less alkaline (”more acid”) conditions and negative pH value shift:

$$pH = -\log_{10}([H^+])$$

Carbon in seawater: CO_2 added, carbonate saturation

The major factor for changing TA is precipitation/dissolution of $CaCO_3$:

calcium carbonate (calcite, aragonite)

$$CaCO_3^{solid} \leftrightarrow Ca^{2+} + CO_3^{2-}$$

So in principle one can neutralise CO_2^{gas} by dissolving $CaCO_3$:

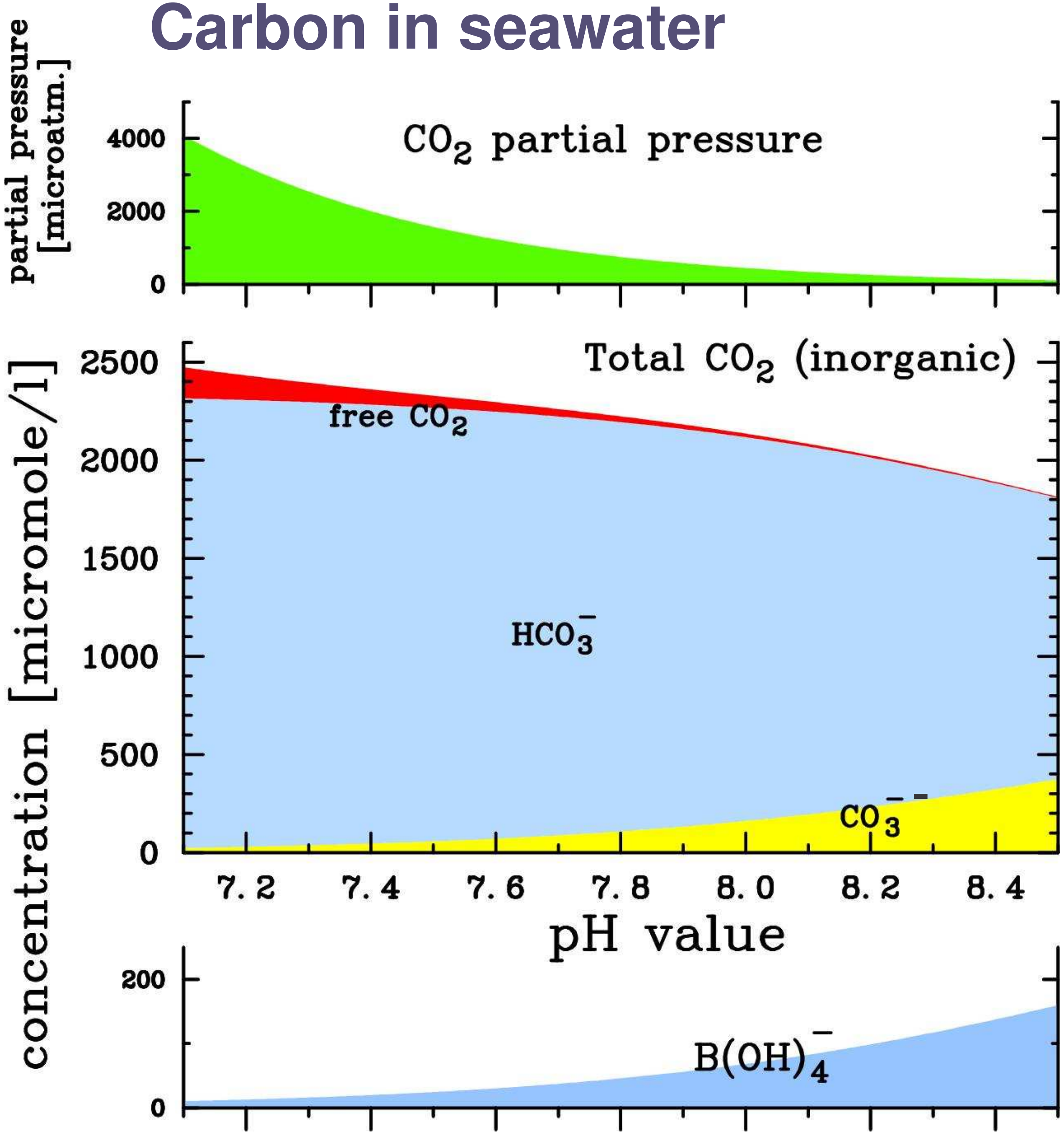
$$CaCO_3^{solid} + CO_2^{gas} + H_2O \leftrightarrow Ca^{2+} + 2HCO_3^-$$

Over-/undersaturation with respect to $CaCO_3$ is determined by the solubility product:

$$K_{sp} = [Ca^{2+}] \times [CO_3^{2-}] = const. \times [CO_3^{2-}]$$

Therefore: By adding CO_2 we decrease the carbonate saturation.

Carbon in seawater



Inevitable marine carbon cycle equilibrium:

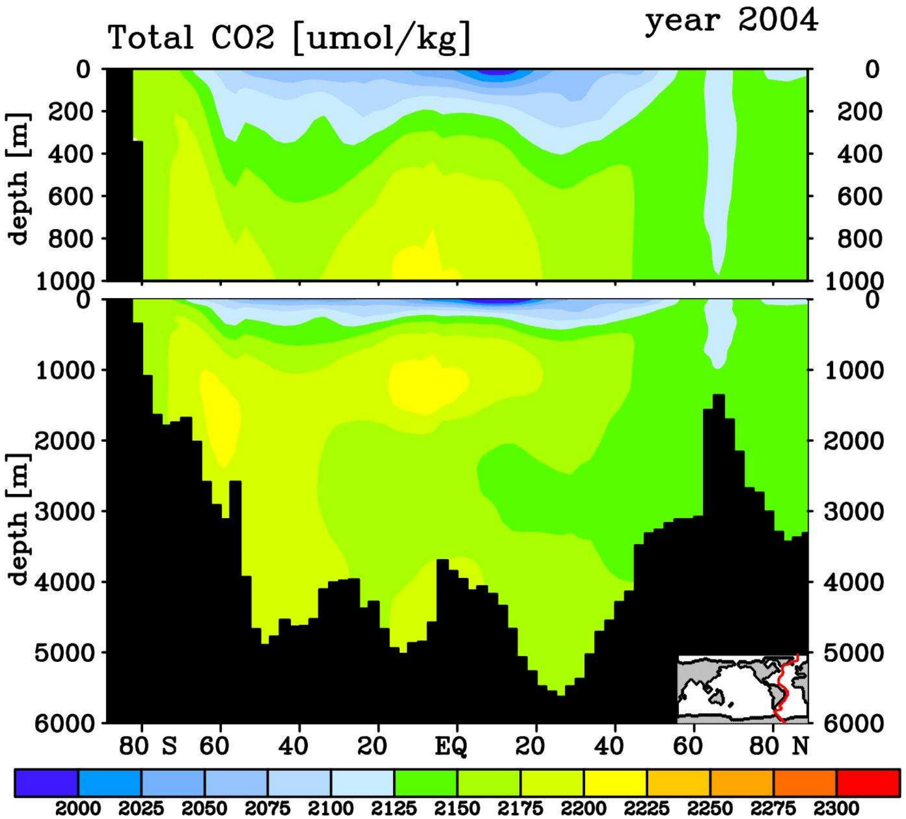
rising pCO_2

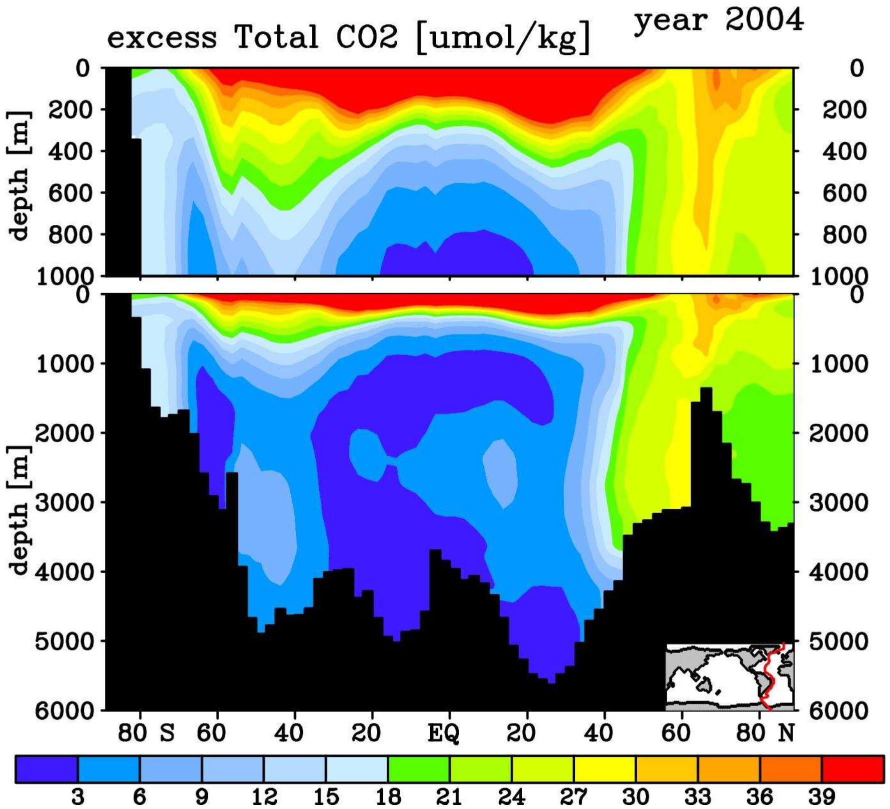
↓

pH value sinks

relative buffer capacity decreases

The actual TCO_2 and the anthropogenic TCO_2 have completely different patterns.
Only the "actual" can readily be observed. On the right figure one can see how the anthropogenic carbon slowly mixes into the interior of the ocean. Due to this slow mixing, more and more human-made CO_2 will temporarily accumulate in the atmosphere.





Heinze, model simulation using the HAMOCC4 model as developed by Ernst Maier-Reimer, MPI Hamburg.

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The ocean will be the major ultimate sink for anthropogenic carbon (after final equilibrium).

Its ultimate uptake capacity is about 90% of any external CO_2 addition to the atmosphere (*Bolin & Eriksson* 1959, Rossby Memorial Volume).

Neutralise CO_2^{gas} through inorganic buffering in seawater:

$CO_2^{gas} + H_2O + CO_3^{2-} \leftrightarrow 2HCO_3^-$ *(some $H^+ + HCO_3^- + CO_3^{2-}$)*

Also: $CO_2^{gas} + H_2O \leftrightarrow H^+ + HCO_3^-$

Neutralise CO_2^{gas} by dissolving $CaCO_3$ sediment:

$$CaCO_3^{solid} + CO_2^{gas} + H_2O \leftrightarrow Ca^{2+} + CO_2^{gas} + H_2O + CO_3^{2-} \leftrightarrow$$
$$\leftrightarrow Ca^{2+} + 2HCO_3^-$$

Warming will decrease CO_2 solubility in ocean but enhance buffering, high CO_2 will decrease buffering.

However!

…this overall buffering requires a long time

…temporary large build up of CO_2 in the atmosphere

Key question:

What are the <u>oceanic uptake kinetics</u> for anthropogenic CO_2?

A good summary is given in:

Maier-Reimer, E., and K. Hasselmann, 1987, Transport and storage of CO_2 in the ocean - an inorganic ocean-circulation carbon cycle model, *Climate Dynamics*, 2, 63-90.

Summary conclusion:

Saving CO_2 emissions helps the ocean to more efficiently take up carbon.

Too fast emissions are not synchronous with the oceans CO_2 uptake kinetics and temporarily a large build-up of CO_2 in the atmosphere will occur.

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