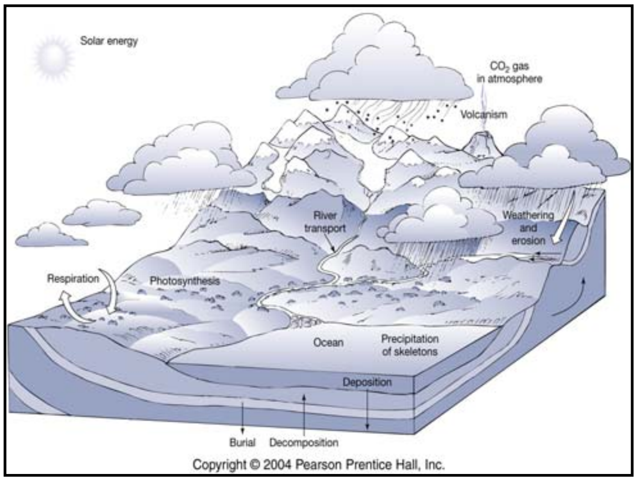
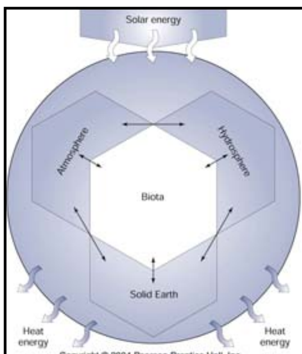
.. to support life over ~ 4 billion years, Earth must be <u>sustainable</u> system..

key to long-term sustainability is recycling..





"Earth System"

how are key elements needed for life (C, N, P) recycled on the Earth's surface through interactions between the biota, the atmosphere, the hydrosphere, and the lithosphere?

Chap 8 pp.147-152
Lecture notes (in power point presentation)

THE CARBON CYCLE

Carbon is important in the Earth system because:

1. ALL life on Earth is based on Carbon

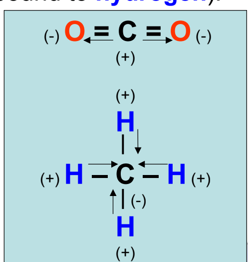
2. Carbon dioxide (CO_2) and methane (CH_4) are important greenhouse gases

3. The acidity of the oceans is determined by CO_2 dissolving in seawater

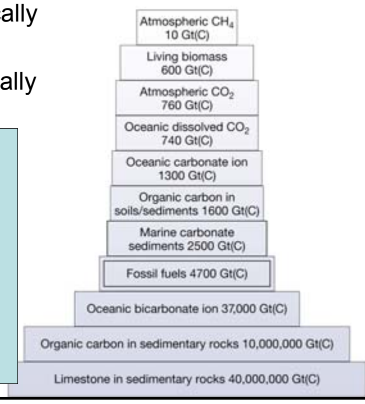
4. Atmospheric oxygen is a result of the carbon cycle

Carbon is found in two main forms:

- oxidized (i.e. chemically bound to **oxygen**)
- reduced (i.e. chemically bound to **hydrogen**).



G (giga) = 10^9 (billion)



Reduced (i.e. chemically bound to **hydrogen**)

G (giga) = 10^9 (billion)

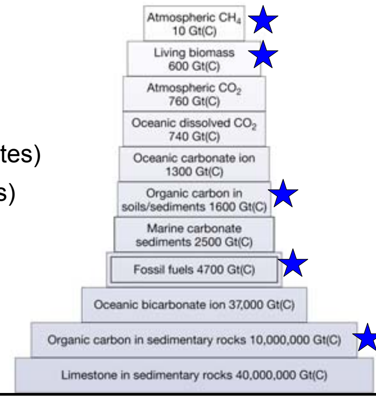
= **Organic Carbon**

e.g.:

CH_4 (methane)

$C_6H_{12}O_6$ (carbohydrates)

C_nH_{n+2} (hydrocarbons)



Oxidized (i.e. chemically bound to **oxygen**)

G (giga) = 10^9 (billion)

= **Inorganic Carbon**

e.g.:

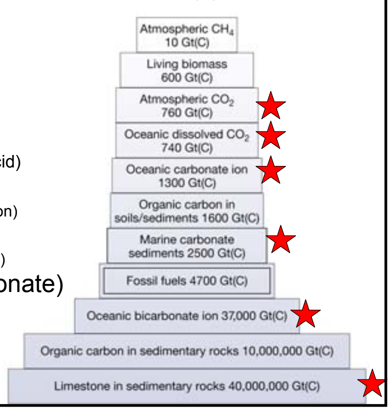
CO_2 (atm./water)

$CO_2 + H_2O \leftrightarrow H_2CO_3$
(carbonic acid)

$H_2CO_3 \leftrightarrow H^+ + HCO_3^-$
(bicarbonate ion)

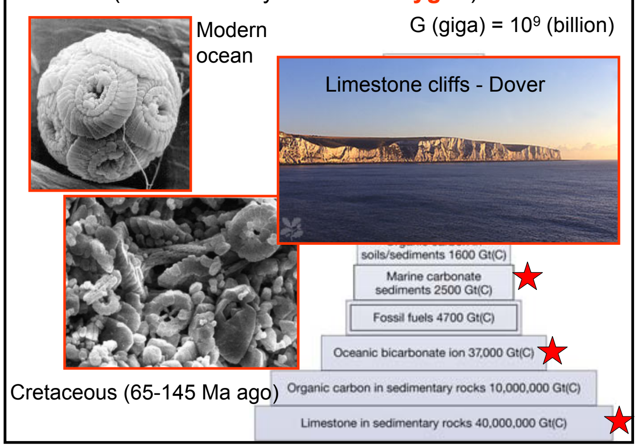
$HCO_3^- \leftrightarrow H^+ + CO_3^=$
(carbonate ion)

$CaCO_3$ (calcium carbonate)



Oxidized (i.e. chemically bound to **oxygen**)

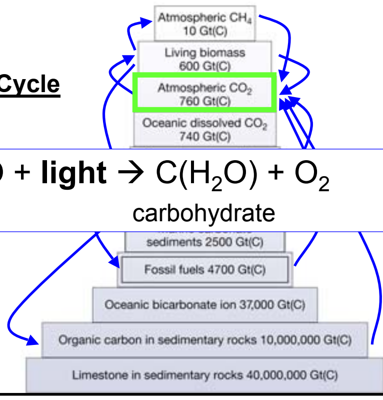
G (giga) = 10^9 (billion)



Cretaceous (65-145 Ma ago)

Carbon is continuously exchanged between reservoirs

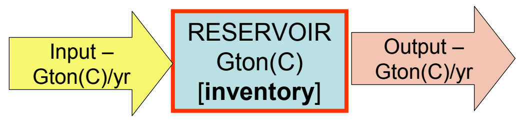
= **Organic Carbon Cycle**



CO_2 + H_2O + **light** → $C(H_2O)$ + O_2

carbohydrate

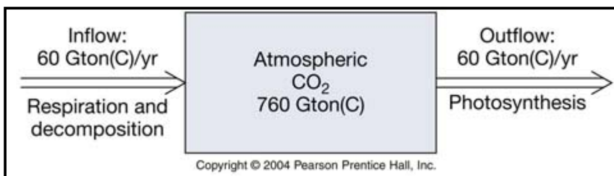
Reservoirs are temporary depositories as carbon flows through them



Input = Output

Size of reservoir stays constant

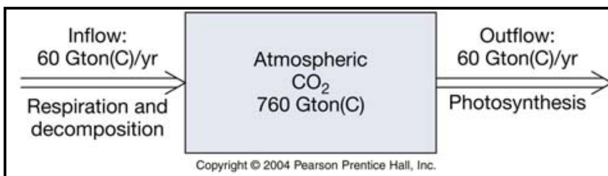
STEADY STATE



For instance:
if rate of addition of CO_2 to the atmosphere (respiration & decomposition)..
.. equals rate of removal of CO_2 (photosynthesis)

→ level of atmospheric CO_2 level does not change as a result of the exchange of carbon with the biota

At steady state → carbon has a fixed residence time in each reservoir

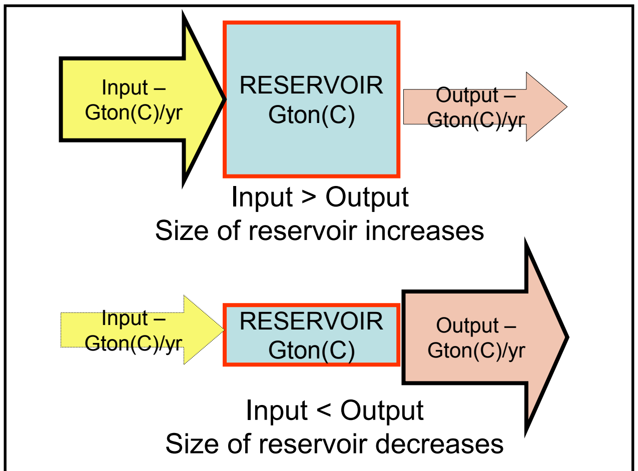


Residence Time (τ_{res})
=
Reservoir Size at Steady State

Inflow (Outflow) rate

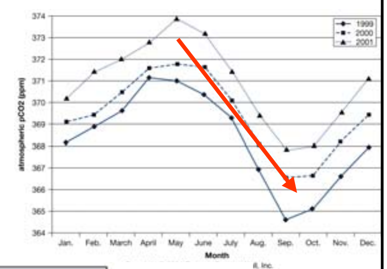
..average length of time spent by an atom of carbon in a reservoir

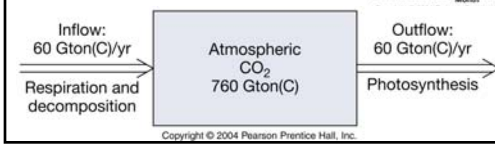
${}^{Atm}\tau_{res}(CO_2)$ = 760 Gt / 60 Gt.yr^{-1}
= 12.7 yr



Atmospheric CO_2 – Hawaii (N. hemisphere)

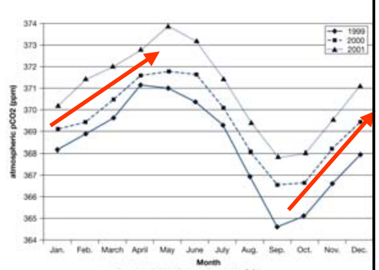
..decreases from May to September when photosynthesis > respiration

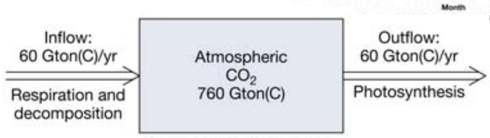




Atmospheric CO_2 – Hawaii (N. hemisphere)

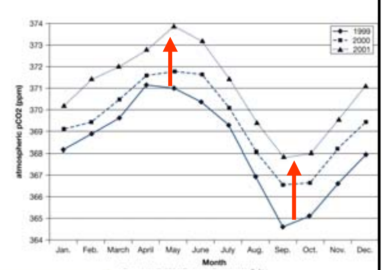
..increases from October to April when photosynthesis < respiration

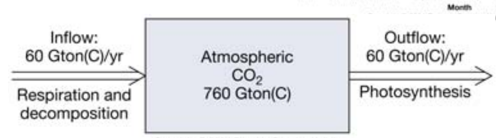




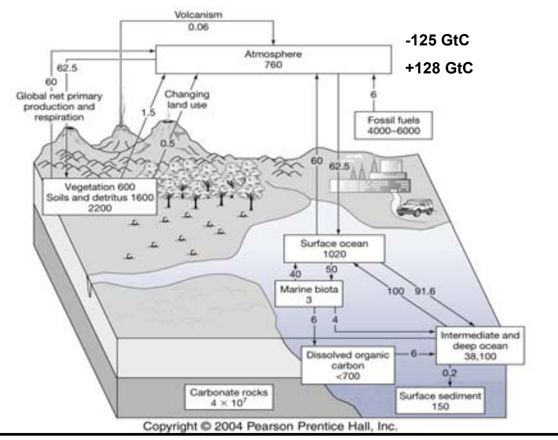
Atmospheric CO_2 – Hawaii (N. hemisphere)

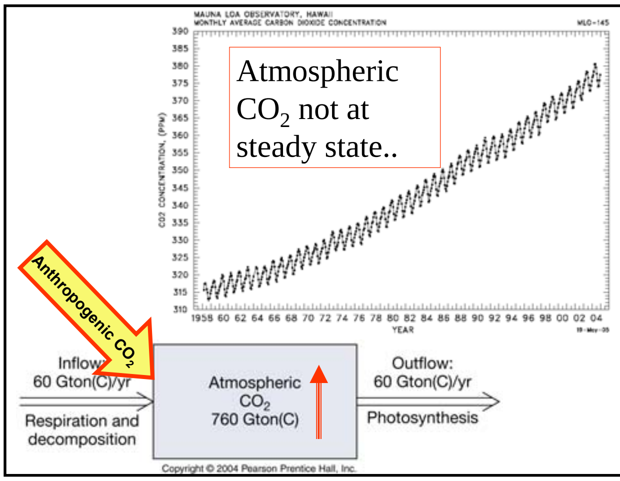
+ gradual increase in atmospheric CO_2





Global carbon cycle (x 10^{15}gC.year^{-1})

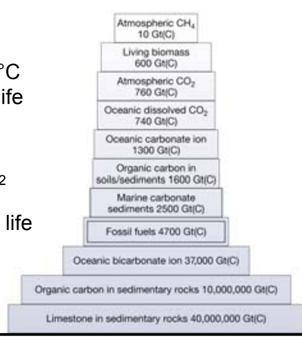




Although the size of reservoirs can vary, it is important that on the long run some sort of steady state be maintained, otherwise..

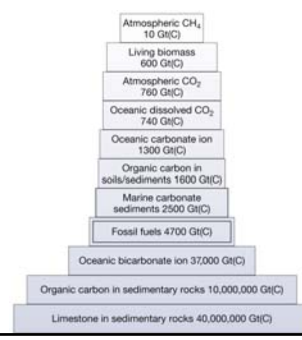
→**Venus:**
Atmospheric CO_2 very high
→ Surface temperature of 400°C
➔ too hot for liquid water and life

Mars:
has lost all its atmospheric CO_2
Locked in carbonate rocks
➔ too cold for liquid water and life



Although the size of reservoirs can vary, it is important that on the long run some sort of steady state be maintained..

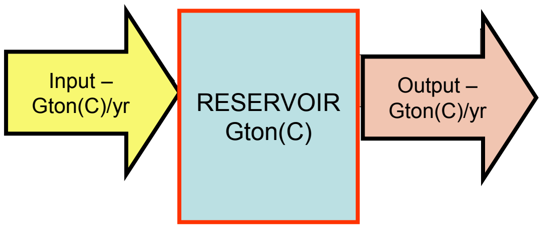
negative feedback loops

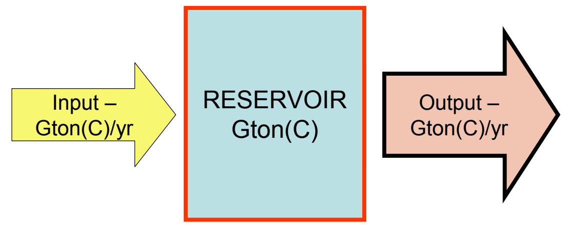


Negative feedback loop

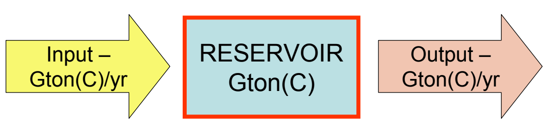
Input and/or output rates are sensitive to the size of the reservoir

[input rates decrease or output rates increase as the size of the reservoir increases]





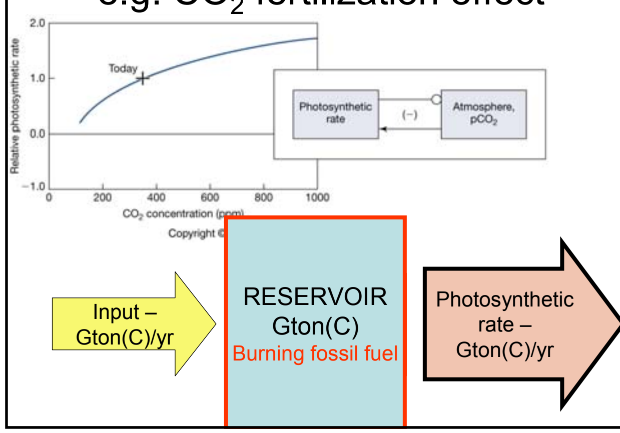
How fast does a system regain steady state after a perturbation?



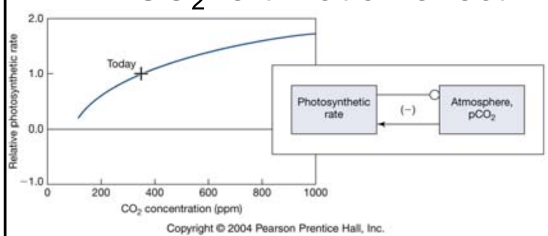
The ***characteristic response time*** is proportional to the residence time

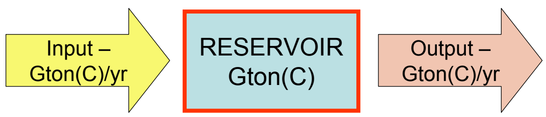
..smaller reservoirs regain steady state faster that large reservoirs

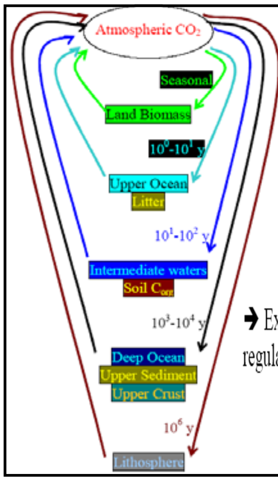
e.g. CO_2 fertilization effect



CO_2 fertilization effect

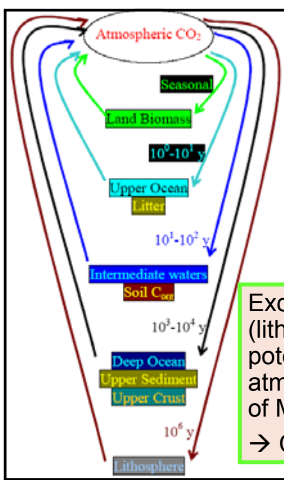


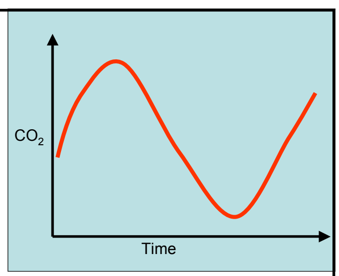




C cycle = series of nested loops recycling carbon through the atmosphere at different rates by exchanging carbon with reservoirs of different size

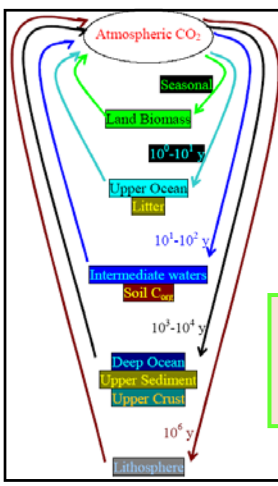
→ Exchanges between atmosphere and different C reservoirs regulate atmospheric CO_2 on different timescales.

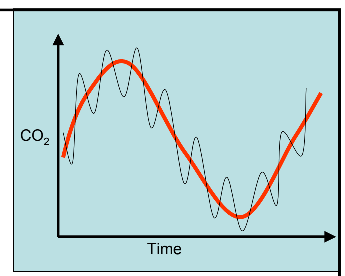




Exchanges with large reservoirs (lithosphere) result in slow (but potentially large) changes in atmospheric CO_2 over hundreds of Ma

→ Controlled by Plate Tectonics

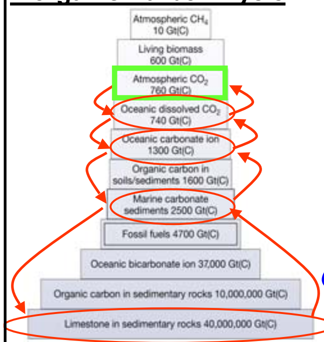


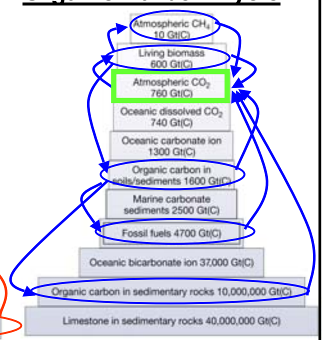


Superimposed on these long term changes, we find higher frequency variations…

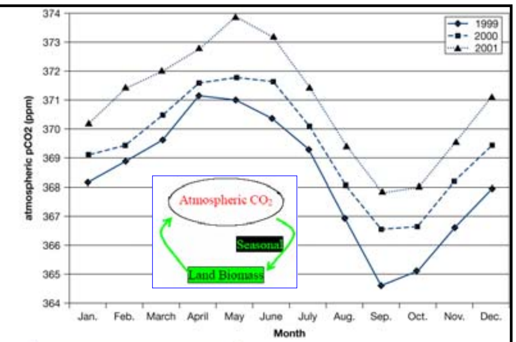
Inorganic Carbon Cycle

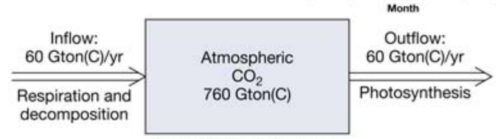
Organic Carbon Cycle

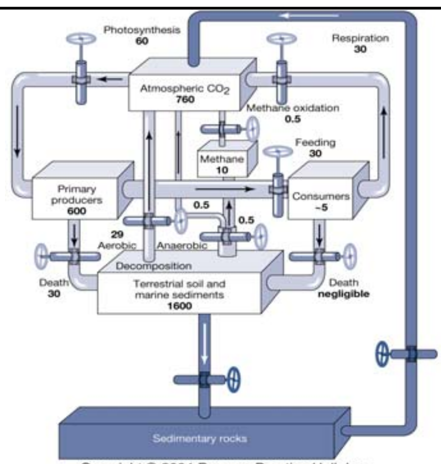




How do these cycles operate on different timescales?

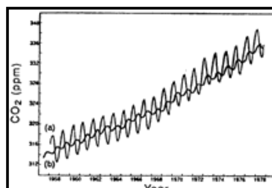




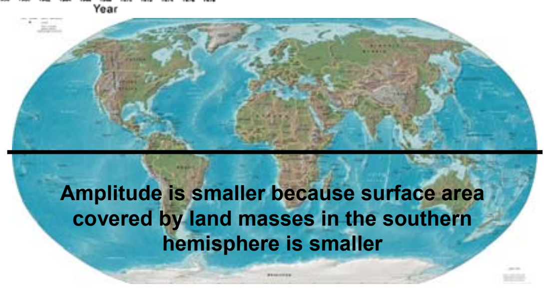


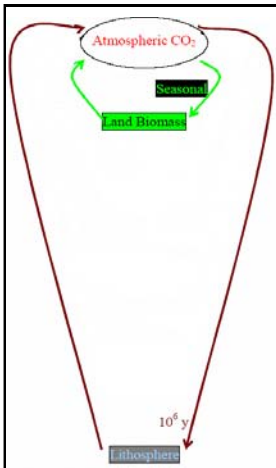
In the southern hemisphere:

→ opposite trends because the seasons are reversed



Atmospheric CO_2 concentration (a) at Mauna Loa Observatory, Hawaii, 19.5°N, 155.6°W and (b) at the South Pole.

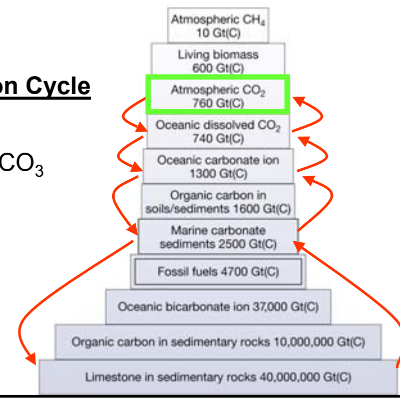




Carbon is continuously exchanged between reservoirs

= **Inorganic Carbon Cycle**

$Ca^{++} + CO_3^{=} \rightarrow CaCO_3$



Inorganic carbon cycle

The ocean is the largest reservoir of the Earth's active C (~ 40000 Gt):

- dissolved CO_2
- bicarbonate HCO_3^-
- carbonate CO_3^{2-}
- organic compounds

In the ocean, far more carbon stored in inorganic form (~ 97% of total) than in organic form

CO_2 is the most soluble of the major gases in sea water and the ocean thus has an enormous capacity to buffer changes in the atmospheric CO_2 content

The concentration of dissolved CO_2 in sea water is small (only ~ 1.5% of C atoms are in CO_2 form). This is because:

- biological uptake, resulting in losses of C in carbonate sediments (limestone);

Inorganic carbon cycle

- CO_2 reacts with water to form carbonic acid (H_2CO_3), which within milliseconds forms bicarbonate and carbonate ions

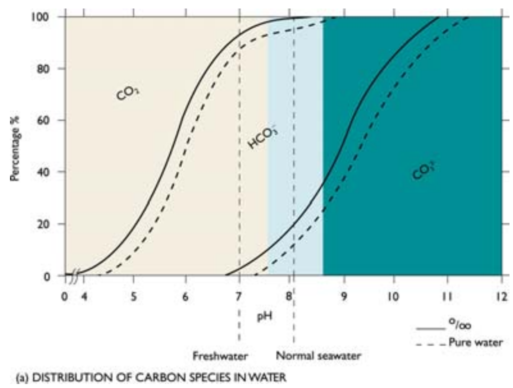
$$\mathbf{CO_2(gas) + H_2O \leftrightarrow H_2CO_3\,(aq) \leftrightarrow H^+(aq) + HCO_3^-(aq) \leftrightarrow 2H^+(aq) + CO_3^{2-}(aq)}$$

For every 20 molecules of CO_2 absorbed by the ocean, 19 are rapidly converted to bicarbonate and carbonate ions

Perturbation of the above equilibrium changes the pH of seawater. CO_2 dissolves → carbonic acid → bicarbonate & hydrogen ions → pH drops → hydrogen ion reacts with carbonate ion → bicarbonate ion → pH rises.

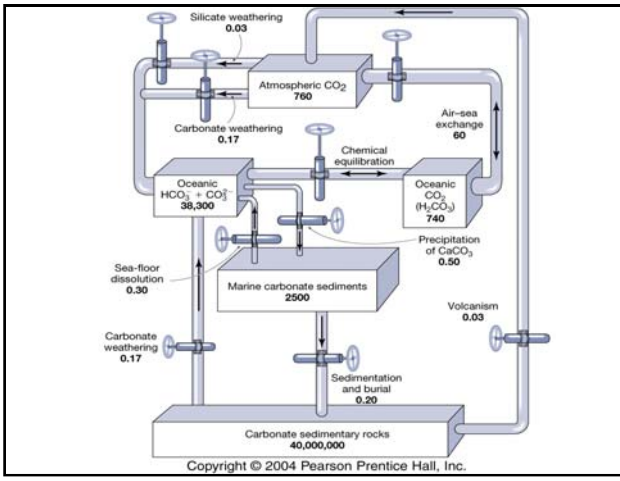
Overall chemical reaction: $\mathbf{CO_2 + CO_3^{2-} + H_2O \leftrightarrow 2H_2CO_3}$

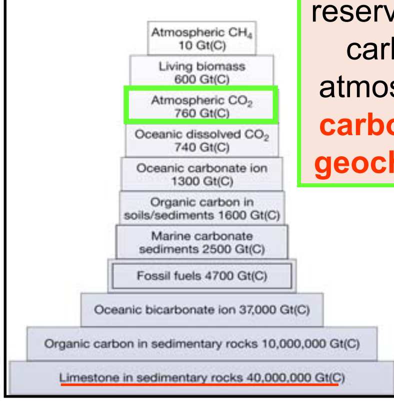
Ocean's capacity to absorb anthropogenic CO_2 is enormous as it converts it into other forms of inorganic C. Note: limited residence time, approximately 600-1000 years, of the deep ocean



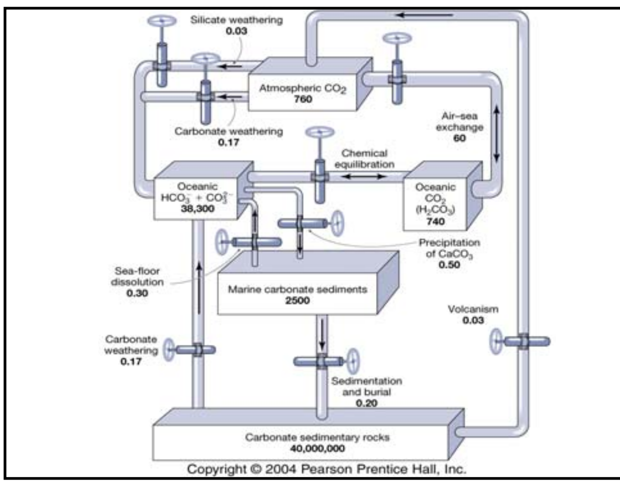
Water too acidic: $H_2CO_3 \leftarrow H^+ + \mathbf{HCO_3^-}$ pH rises

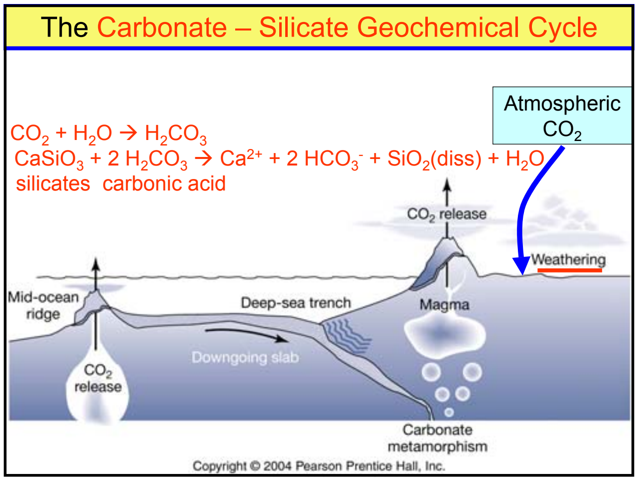
Water too basic: $\mathbf{HCO_3^-} \rightarrow H^+ + CO_3^{-2}$ pH drops





The large limestone reservoir exchanges carbon with the atmosphere via the **carbonate-silicate geochemical cycle**




The Carbonate – Silicate Geochemical Cycle
Atmospheric CO_2
$CO_2 + H_2O \rightarrow H_2CO_3$
$CaSiO_3 + 2\ H_2CO_3 \rightarrow Ca^{2+} + 2\ HCO_3^- + SiO_2(diss) + H_2O$
silicates carbonic acid
CO_2 release
Weathering
Mid-ocean ridge
Deep-sea trench
Magma
Downgoing slab
CO_2 release
Carbonate metamorphism
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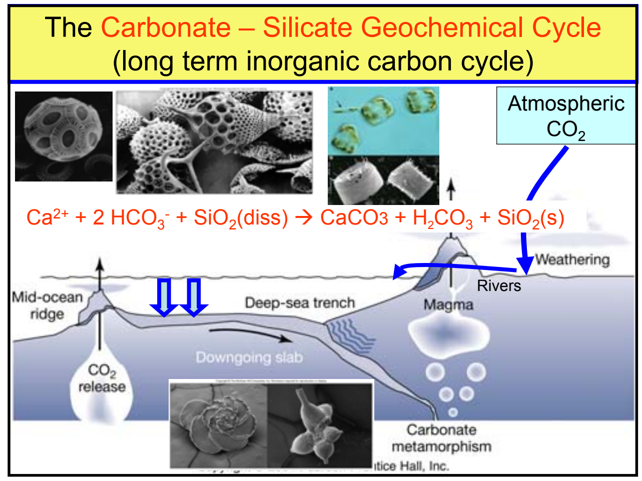
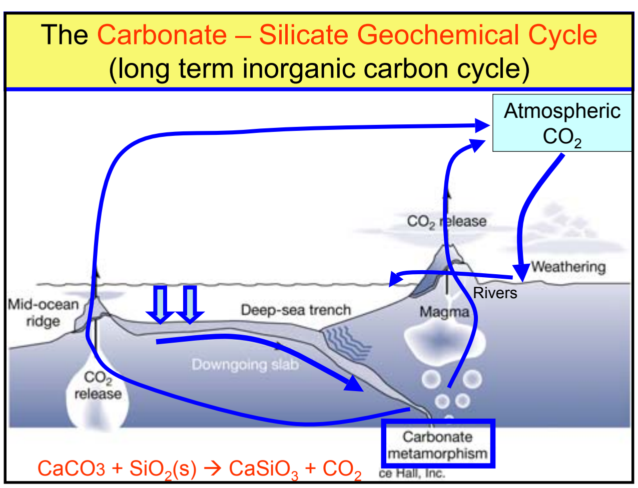
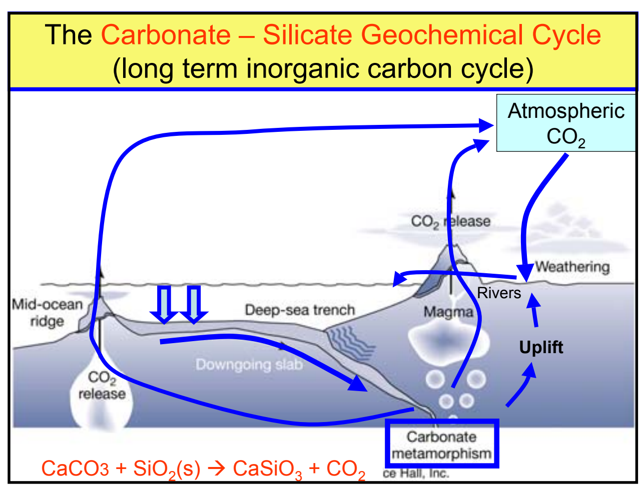
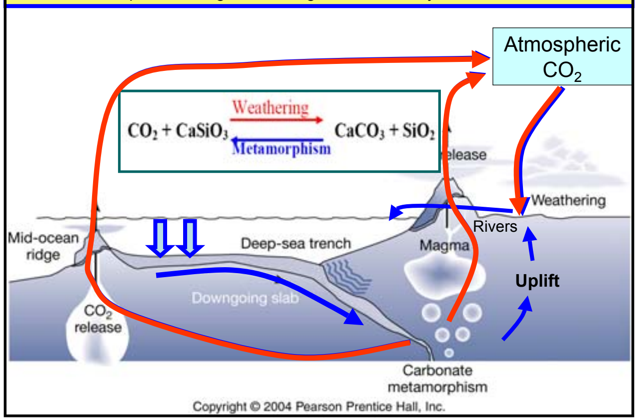

The Carbonate – Silicate Geochemical Cycle
(long term inorganic carbon cycle)
Atmospheric CO_2
$Ca^{2+} + 2\ HCO_3^- + SiO_2(diss) \rightarrow CaCO3 + H_2CO_3 + SiO_2(s)$
Weathering
Rivers
Mid-ocean ridge
Deep-sea trench
Magma
Downgoing slab
CO_2 release
Carbonate metamorphism


The Carbonate – Silicate Geochemical Cycle
(long term inorganic carbon cycle)
Atmospheric CO_2
CO_2 release
Weathering
Rivers
Mid-ocean ridge
Deep-sea trench
Magma
Downgoing slab
CO_2 release
Carbonate metamorphism
$CaCO3 + SiO_2(s) \rightarrow CaSiO_3 + CO_2$


The Carbonate – Silicate Geochemical Cycle
(long term inorganic carbon cycle)
Atmospheric CO_2
CO_2 release
Weathering
Rivers
Uplift
Mid-ocean ridge
Deep-sea trench
Magma
Downgoing slab
CO_2 release
Carbonate metamorphism
$CaCO3 + SiO_2(s) \rightarrow CaSiO_3 + CO_2$

.. long term evolution of atmospheric CO_2 is controlled by the balance between uptake during weathering and release by volcanism



Net removal of CO_2 from ocean and atmosphere

Carbonate weathering: $CaCO_3 + H_2CO_3 \rightarrow Ca^{2+} + 2HCO_3^-$

Carbonate precipitation: $Ca^{2+} + 2HCO_3^- \rightarrow CaCO_3 + H_2CO_3$

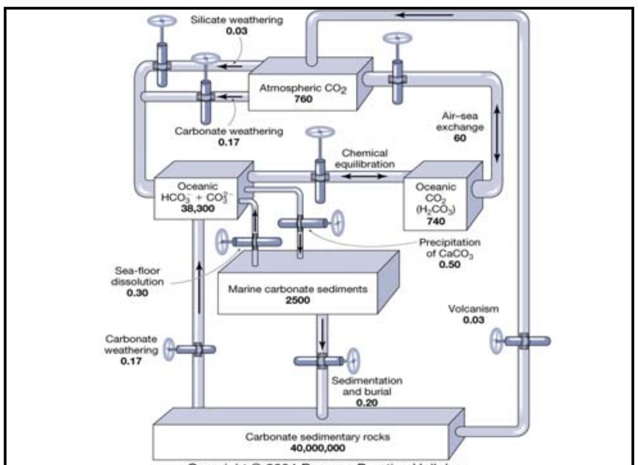
Net result: **0**

Silicate weathering: $CaSiO_3 + 2H_2CO_3 \rightarrow Ca^{2+} + 2HCO_3^- + SiO_2 + H_2O$

Carbonate precipitation: $Ca^{2+} + 2HCO_3^- \rightarrow CaCO_3 + H_2CO_3$

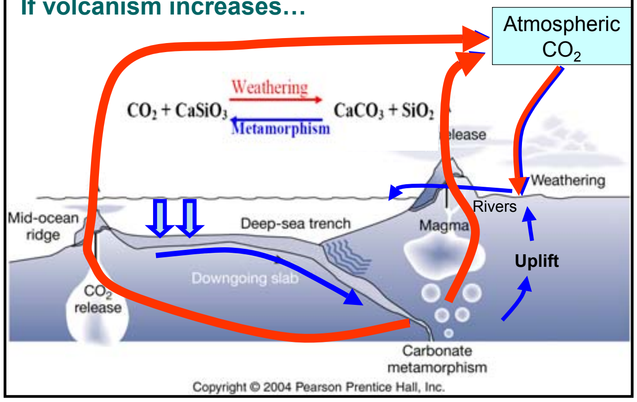
Ocean/atmosphere CO_2 exchange: $CO_2 + H_2O \rightarrow H_2CO_3$

Net result: $CaSiO_3 + CO_2 \rightarrow CaCO_3 + SiO_2$



..negative feedback loop, which tends to stabilize the level of atmospheric CO_2 in the long term

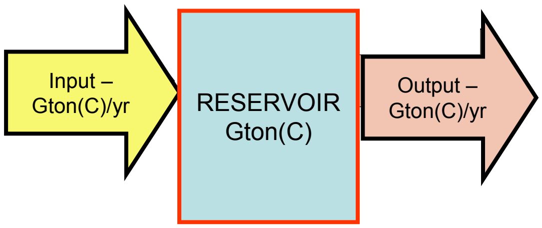
If volcanism increases…



Negative feedback loop

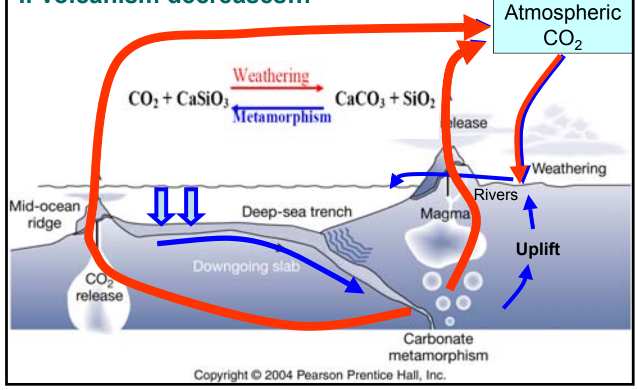
Input and/or output rates are sensitive to the size of the reservoir

[input rates decrease or output rates increase as the size of the reservoir increases]



..negative feedback loop, which tends to stabilize the level of atmospheric CO_2 in the long term

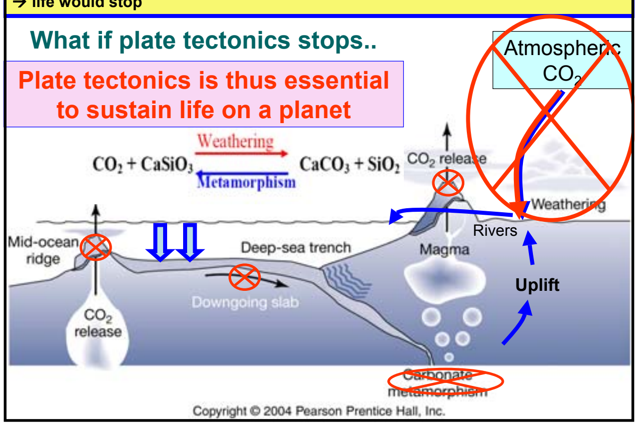
If volcanism decreases…



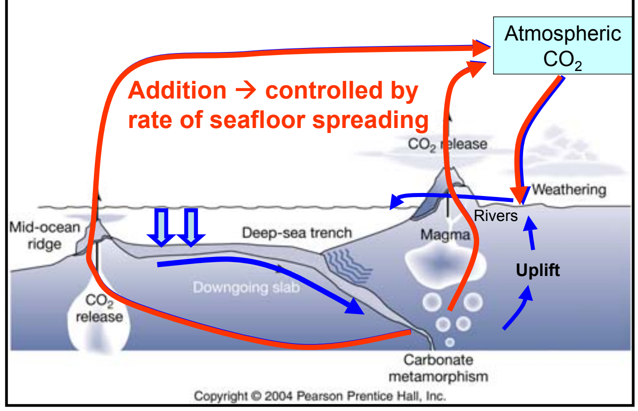
→ atm CO_2 would drop to zero as all the carbon becomes locked in carbonate
→ planet would freeze
→ life would stop

What if plate tectonics stops..

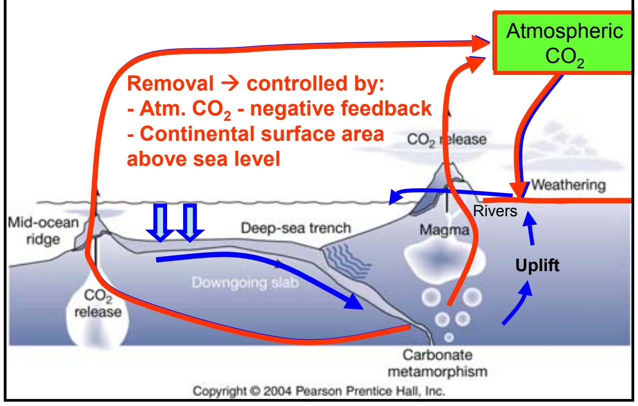
Plate tectonics is thus essential to sustain life on a planet



What factors control addition/removal of atm. CO_2?



What factors control addition/removal of atm. CO_2?



How can we change the surface area of continents above sea level?

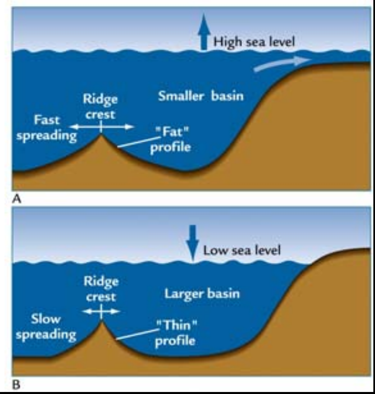
→ By changing the volume of the ocean

When spreading rates are fast

→younger (hotter) rocks are found at a greater distance from the ridge

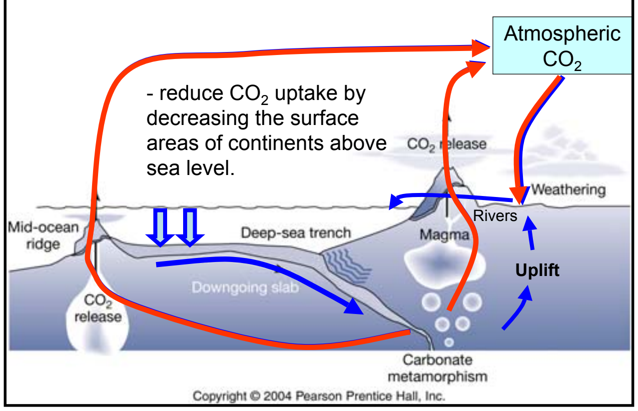
→ Mid ocean ridges are "fatter"

→ Displace water over the continental shelves



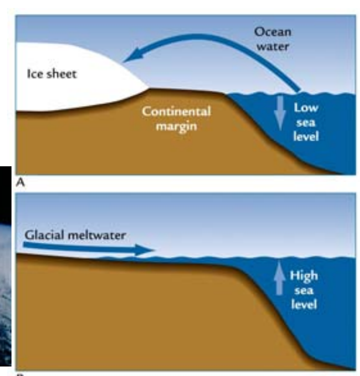
→ higher spreading rates:

- increase CO_2 outgassing to the atmosphere

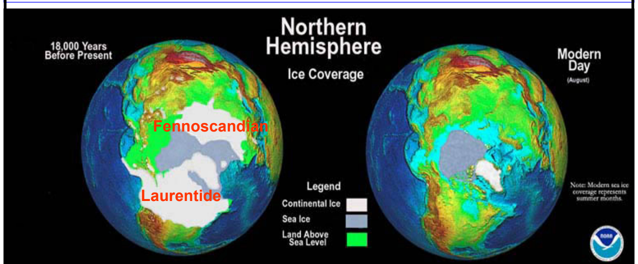


How can we change the surface area of continents above sea level?

→ By accumulating ice on continents



During ice ages → large ice sheets (Laurentide, Fennoscandian) covered northern continents



→ Sea level dropped by 130 m

Increasing continental ice affects atmospheric CO_2 in several opposing ways..

Larger ice sheets:

→Lower sea level

→Larger continental surface area

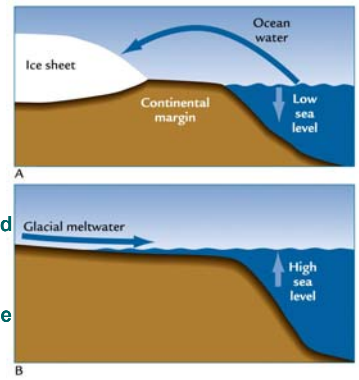
→Higher rate of CO_2 uptake

Larger ice sheets:

→Cover large areas of land

→Decrease continental surface area

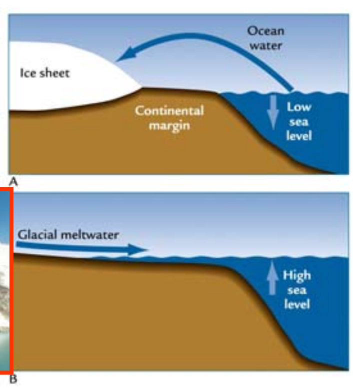
→Lower rate of CO_2 uptake



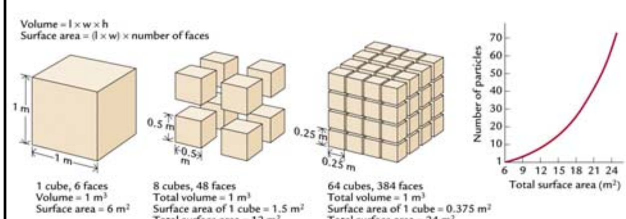
Increasing continental ice affects atmospheric CO_2 in several opposing ways..

+ Glaciers grind up rocks by abrasion:

→Higher rate of CO_2 uptake



As rocks are broken in smaller pieces, the total volume stays the same but surface area increases…



.. increases surface of contact between rocks and acidic rain water

→ accelerates the rate of chemical reactions (chemical weathering) between carbonic acid and rock minerals.

Increasing continental ice affects atmospheric CO_2 in several opposing ways..

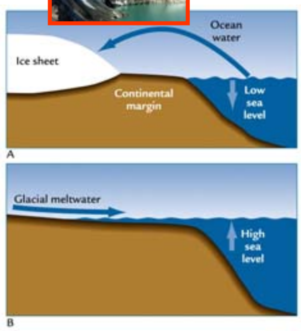
→ **Net effect is difficult to evaluate**

It depends on the balance between:

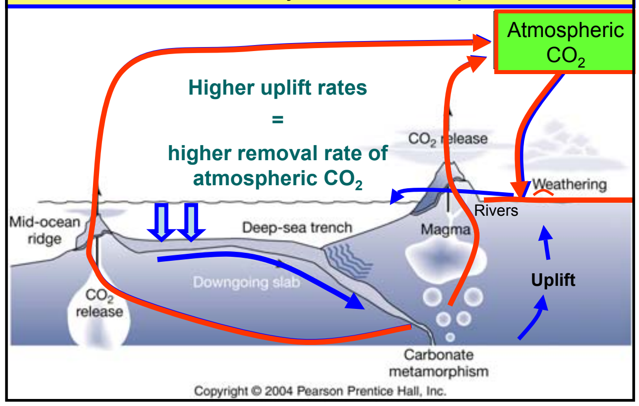
-Increase in CO_2 uptake due to lower sea level (increasing continental surface area)

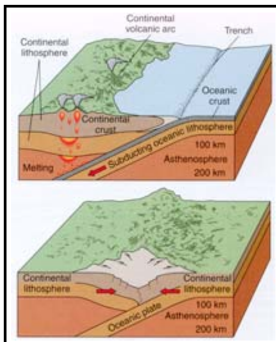
-Decrease in CO_2 uptake due to more ice cover (decreasing continental surface area)

-Increase CO_2 uptake due to rock abrasion (physical weathering)



CO_2 removal from the atmosphere is also controlled by the rate of uplift



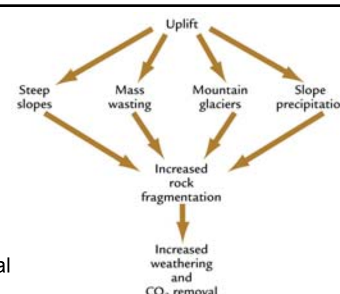


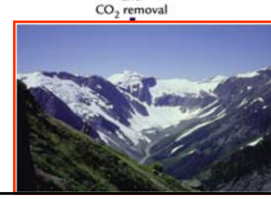
Uplift and orogeny (mountain building) occur mainly at converging plate boundaries, especially when two continents collide…

In periods of extensive orogeny:

- more mountain glaciers (abrasion)
- steeper slopes
- more precipitation (rain, snow)

→ Higher rates of mechanical breakdown of rocks
→ Higher rates of atmospheric CO_2 uptake





Control of atmospheric CO_2 on Ma timescale..
Summary

Atmospheric CO_2
- Controlled by balance between uptake rate by continental weathering & outgassing rate by tectonic activity
- Stabilized by negative feedback

Control of atmospheric CO_2 on Ma timescale..
Summary

Increasing seafloor spreading → increases CO_2 outgassing
→ decreases CO_2 uptake (continental flooding)

Increasing orogeny (mountain building) → increases CO_2 uptake

Increasing continental ice sheets → increased uptake (rock fragmentation, lower sea level)
→ decreased uptake (continental ice cover)
→ Net effect difficult to evaluate

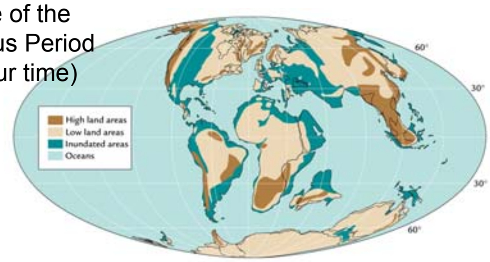
100 Million years ago…

Changes in seafloor spreading rate and orogeny have been invoked to suggest large changes in atmospheric CO_2 and climate during the last 600 Ma

..middle of the Cretaceous Period (Dinosaur time)

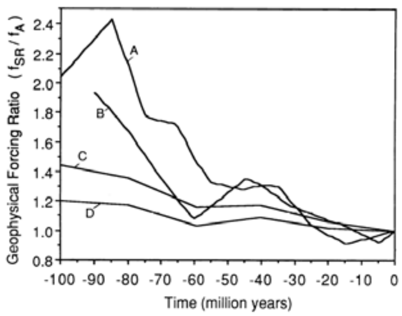
100 Million years ago…

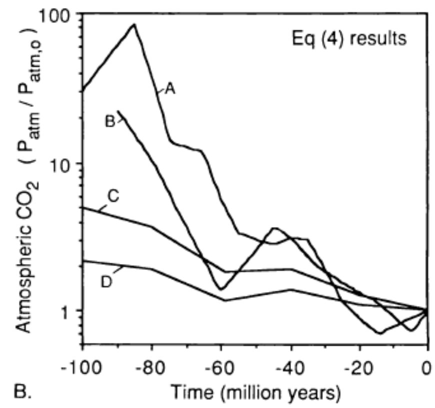
..middle of the Cretaceous Period (Dinosaur time)



→High seafloor spreading rate
→Continental flooding (dark green areas)
→Lower CO_2 uptake rate

Geophysical Forcing Ratio
=
(rate of seafloor spreading)/(continental area)

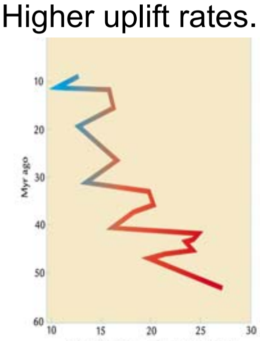


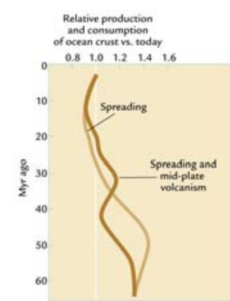


..high atmospheric CO_2 = **warm climate**
→ tropical plant & dinosaurs at the poles..

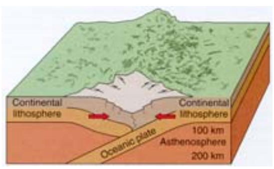
..65 Ma ago → temperature started to drop

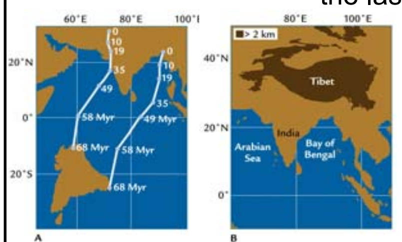
→Slower spreading rates..
→Higher uplift rates..





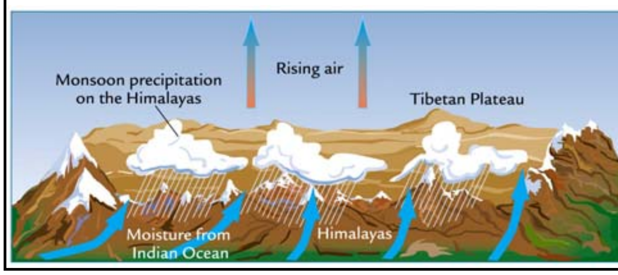
Collision between Indian and Asia created the Himalayas and the Tibetan plateau (gradual buildup over the last 60 Ma or so..)



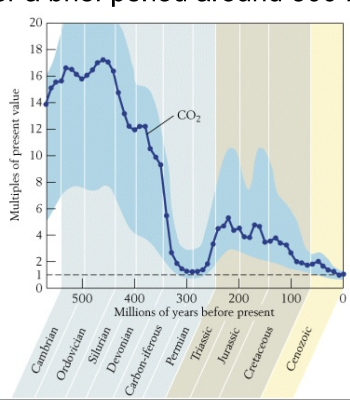


Rising air in summer over the Tibetan Plateau brings in warm moist air from the tropical Indian ocean (monsoon)

→ **rock fragmentation + steep slopes + heavy rain = accelerated CO_2 uptake**

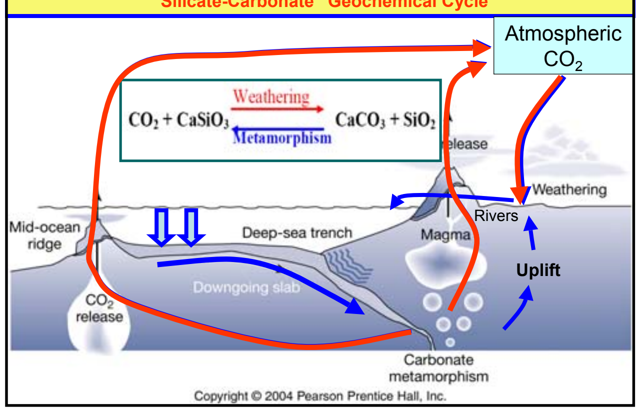


Atmospheric CO2 was higher in the past, except for a brief period around 300 Ma ago



.. long term evolution of atmospheric CO_2 is controlled by the balance between uptake during weathering and release by volcanism

"Silicate-Carbonate" Geochemical Cycle



Oxidized (i.e. chemically bound to **oxygen**)

= **Inorganic Carbon**

e.g.:

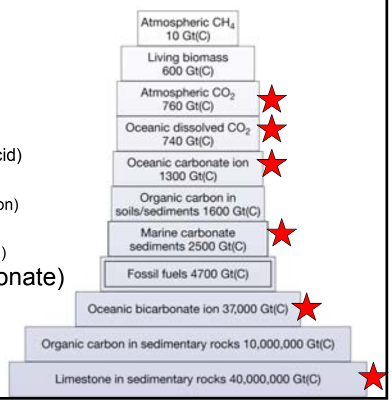
CO_2 (atm./water)

$CO_2 + H_2O \leftrightarrow H_2CO_3$ (carbonic acid)

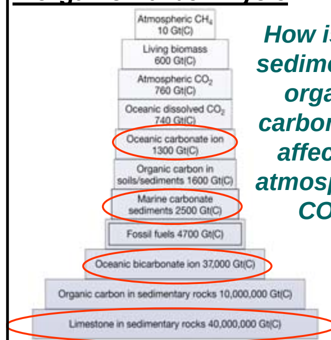
$H_2CO_3 \leftrightarrow H^+ + HCO_3^-$ (bicarbonate ion)

$HCO_3^- \leftrightarrow H^+ + CO_3^=$ (carbonate ion)

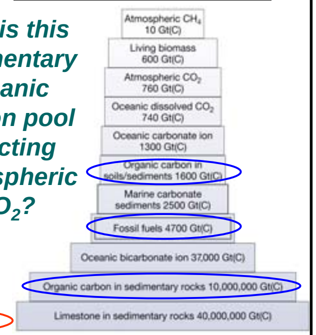
$CaCO_3$ (calcium carbonate)



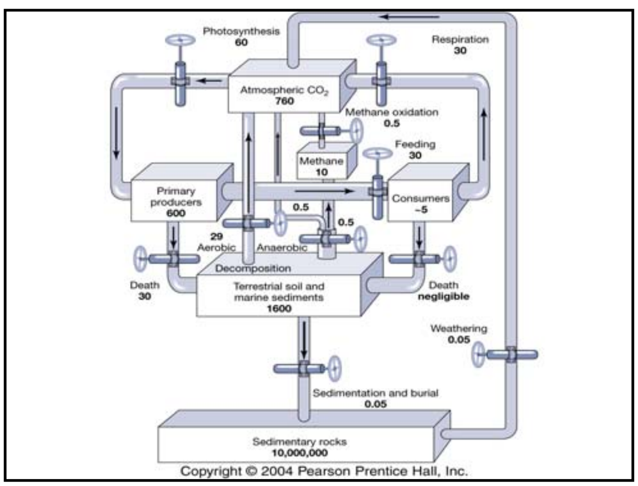
Inorganic Carbon Cycle | **Organic Carbon Cycle**



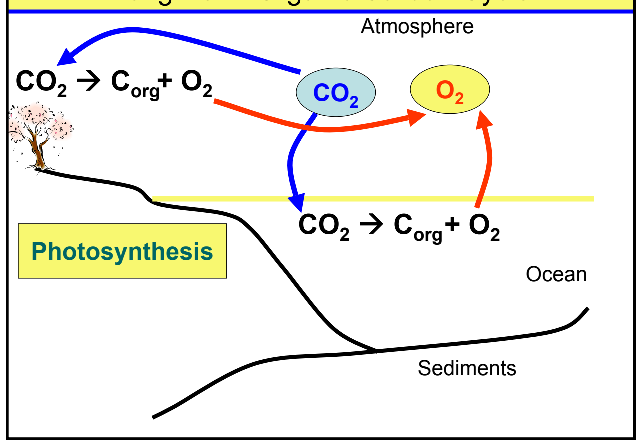
How is this sedimentary organic carbon pool affecting atmospheric CO_2?



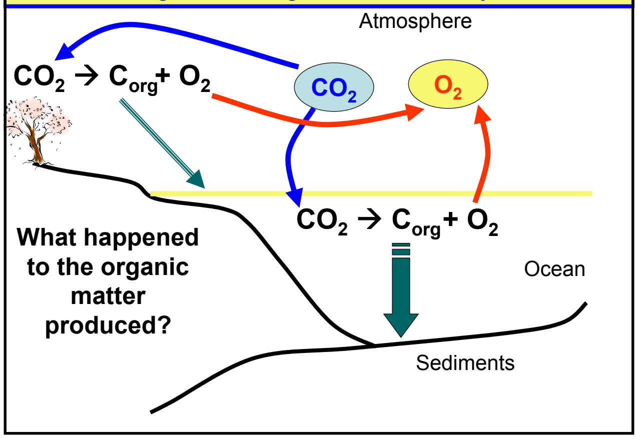
Most of the organic carbon in the crust is found at very low concentration in sedimentary rocks (<<1%).
Fossil fuels are concentrated in localized regions of the crust that can be mined or drilled to)



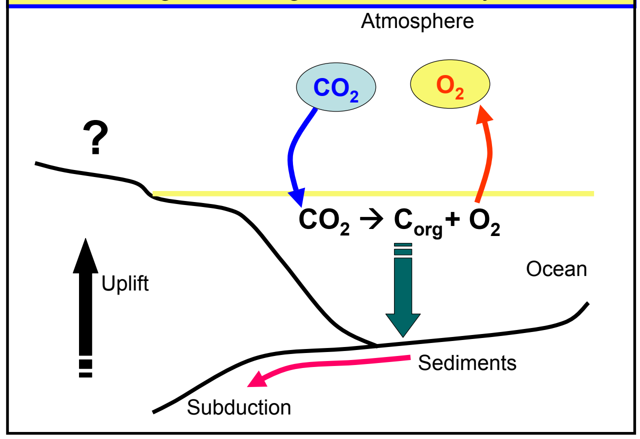
Long-Term Organic Carbon Cycle



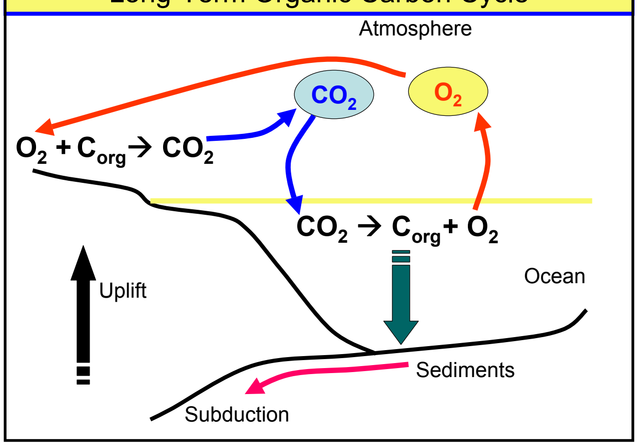
Long-Term Organic Carbon Cycle



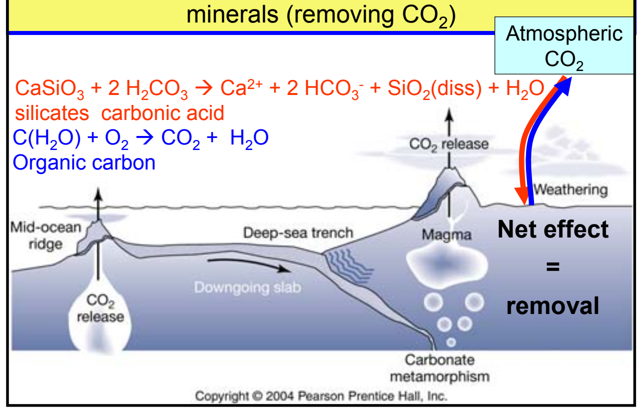
Long-Term Organic Carbon Cycle



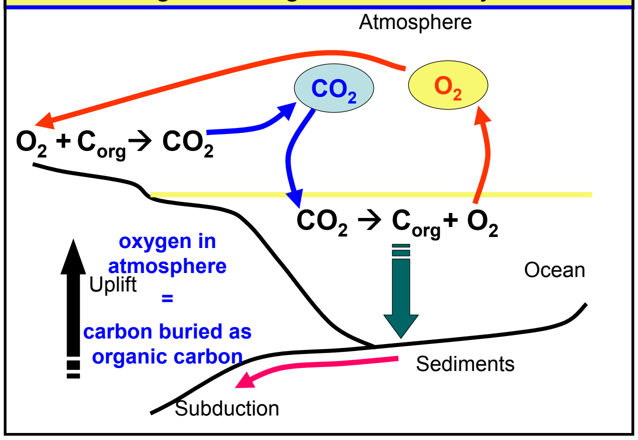
Long-Term Organic Carbon Cycle



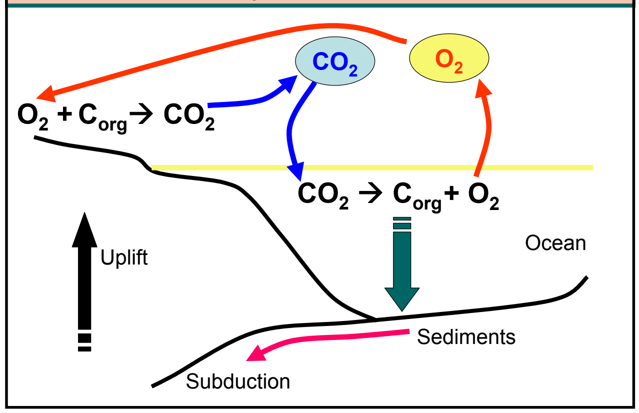
Weathering of sedimentary organic carbon (producing CO_2) partially cancel the effect of weathering of silicate minerals (removing CO_2)



Long-Term Organic Carbon Cycle

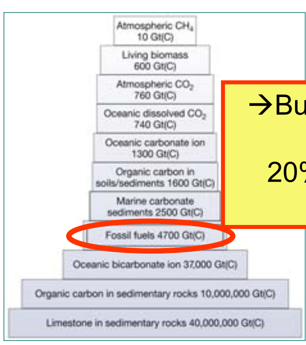


What would happen if we oxidize all organic carbon locked in sedimentary rocks, sediments and biomass?



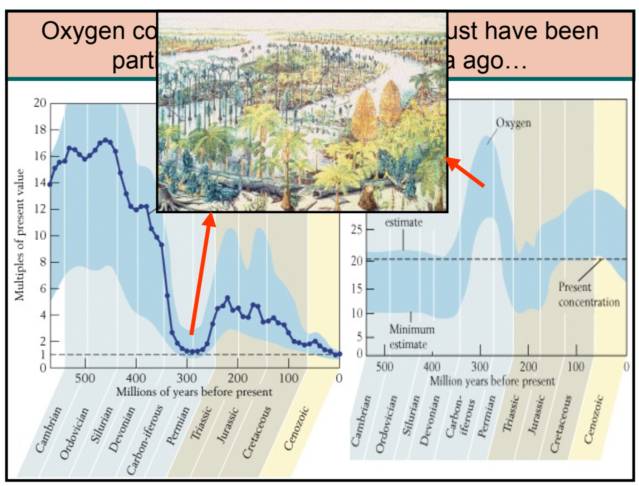
What would happen if we burn all the fossil fuel?

20% O2 = 10,006,900 Gt(C)



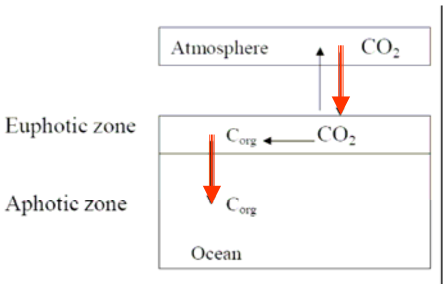
→Burning 4,700 Gt(C) would remove:
20% x 4,700 / 10,006,900
= 0.0094% O_2

Oxygen co… part… …ust have been …a ago…

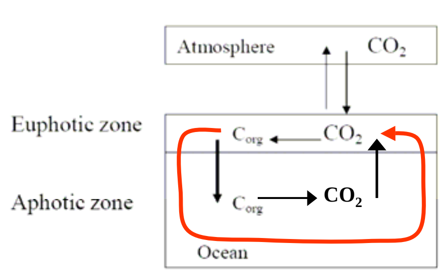


Processes affecting the exchange of CO_2 between the ocean and atmosphere

THE BIOLOGICAL PUMP

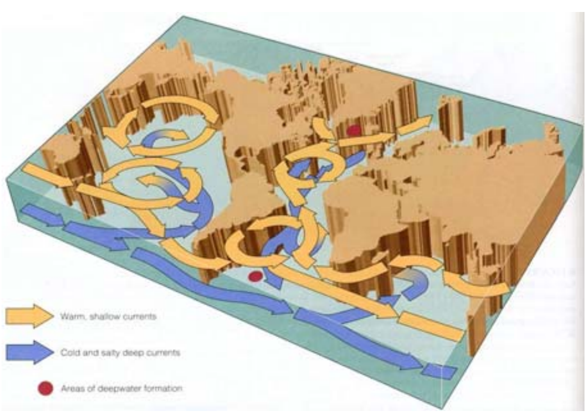


THE BIOLOGICAL PUMP



THE THERMOHALINE CIRCULATION

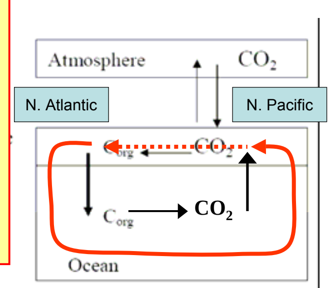
Thermohaline circulation



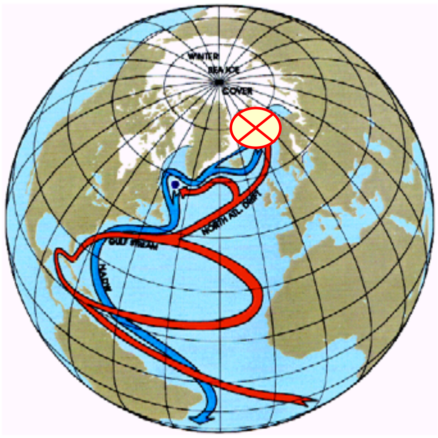
THE BIOLOGICAL PUMP

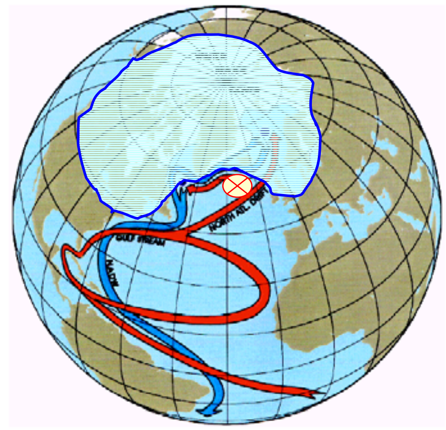
Atmospheric CO_2 can be lowered by:
- increasing the biological pump
- decreasing the thermohaline circulation

Atmospheric CO_2 can be increased by:
- decreasing the biological pump
- increasing the thermohaline circulation

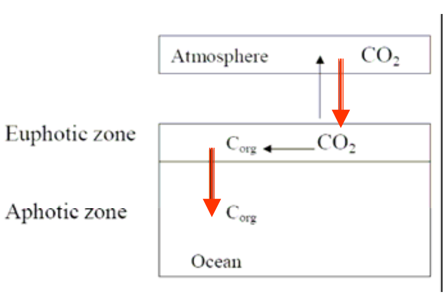


THE THERMOHALINE CIRCULATION

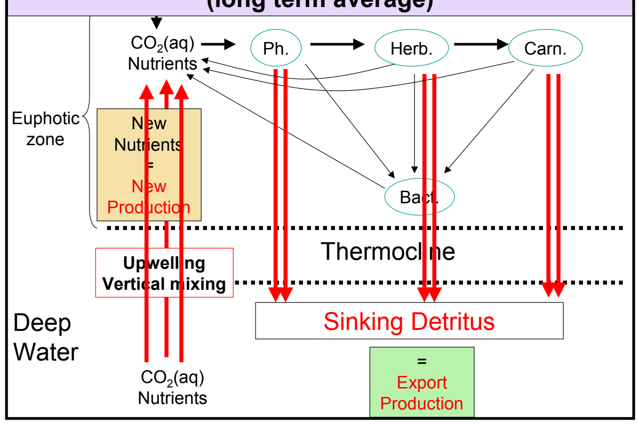




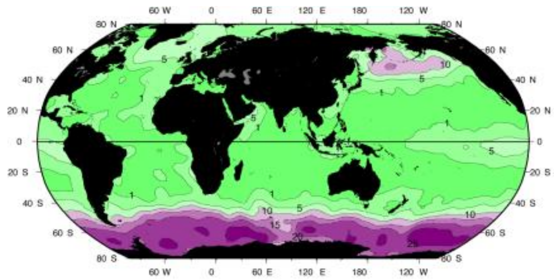
THE BIOLOGICAL PUMP



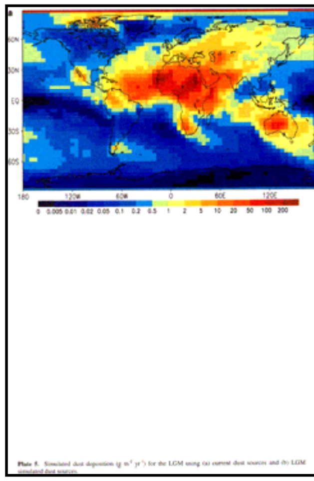
New Production = Export Production (long term average)



THE BIOLOGICAL PUMP could be increased by using all the (new) nutrients supplied to surface waters that do not get utilized by phytoplankton

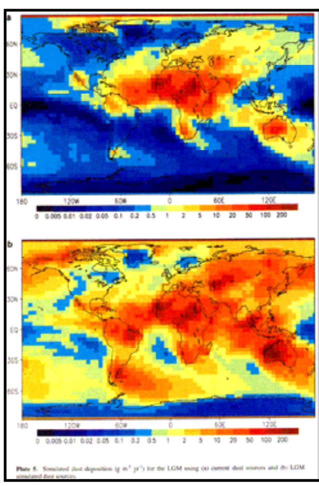


Purple = High Nutrients Low Chlorophyll (HNLC) regions



Today

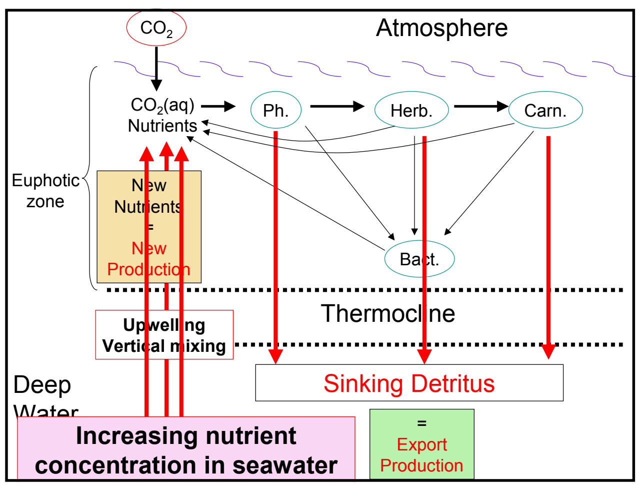
Supply of dust to the ocean surface



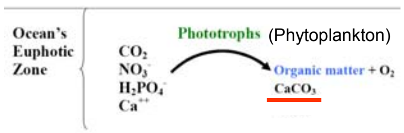
Today

Supply of dust to the ocean surface

Ice Age



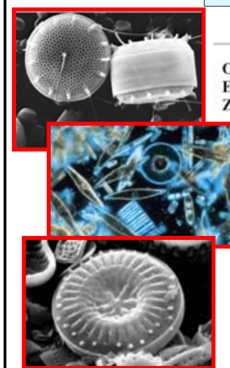
At sea…



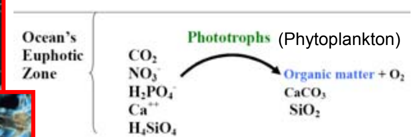
Coccolithophorid

$Ca^{2+} + 2\ HCO_3^- \rightarrow CaCO_3 + H_2O + CO_2$

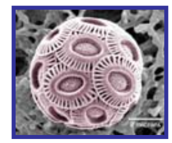
Diatoms



Coccolithophorids are less efficient than diatoms at exporting CO2 to the deep sea…



Coccolithophorid



we can decrease atmospheric CO_2 by sequestering it in the ocean.
(1,000 to 10,000 years needed to reach a new steady state)

This could be achieved by several means:

- Slowing down the overturning of the ocean (thermohaline circulation)
- Increasing the biological pump by:
  - increasing nutrient utilization in the HNLC regions
  - by favoring diatoms
  - by increasing the nitrate and phosphate seawater concentrations in the deep sea

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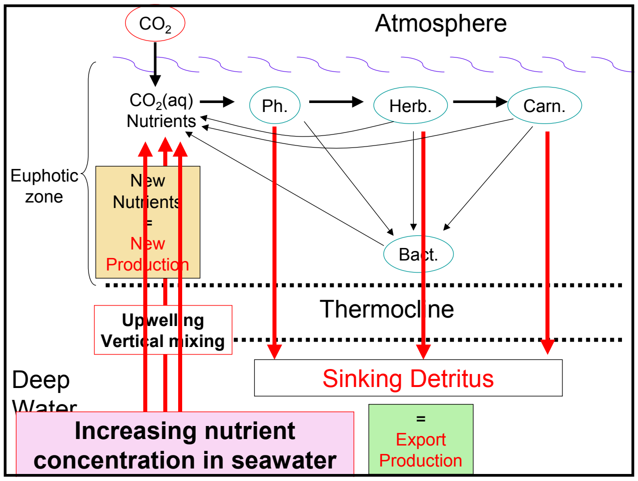
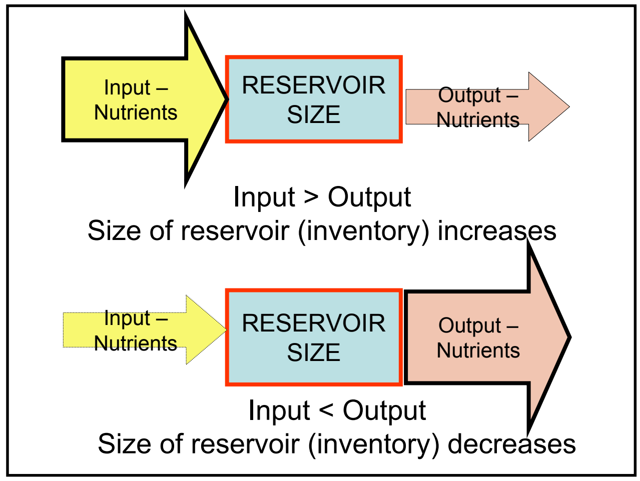
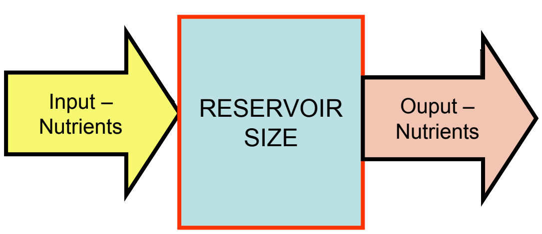




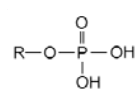
Negative feedback loop

Output rate is proportional to the size of the reservoir



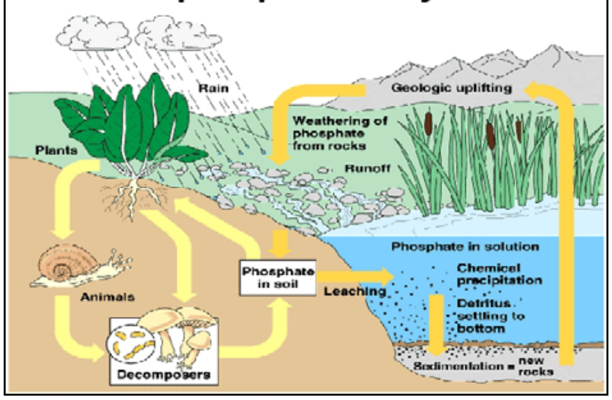
Phosphorous

- Simple environmental chemistry
- Always in its oxidized form (phosphate)

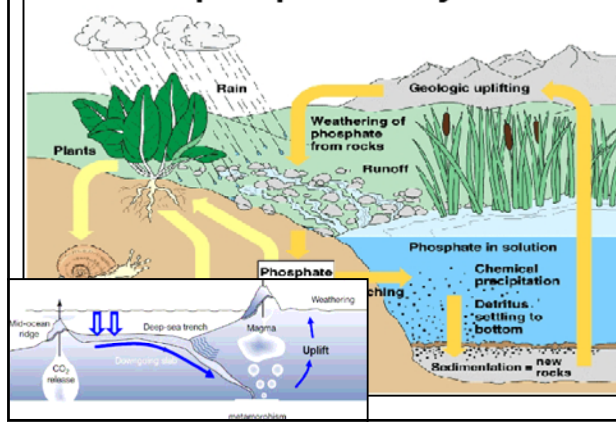
(PO_4^{3-}) 

- No P-containing gases in atmosphere (but phosphate present in dust carried by winds)

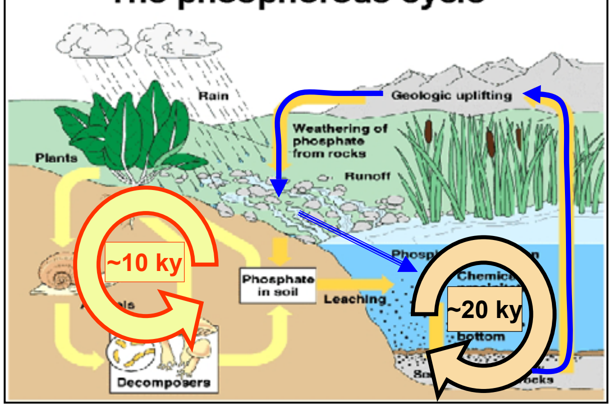
The phosphorous cycle



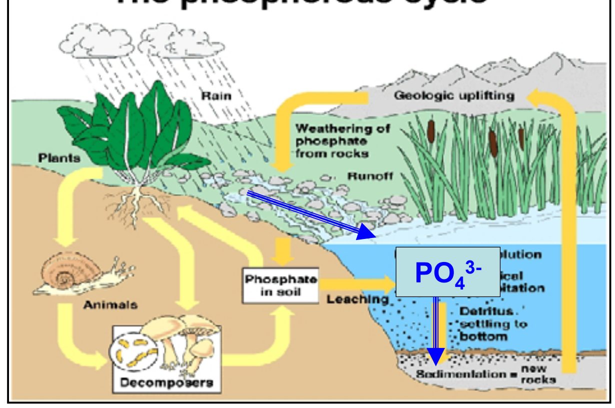
The phosphorous cycle



The phosphorous cycle



The phosphorous cycle



Nitrogen

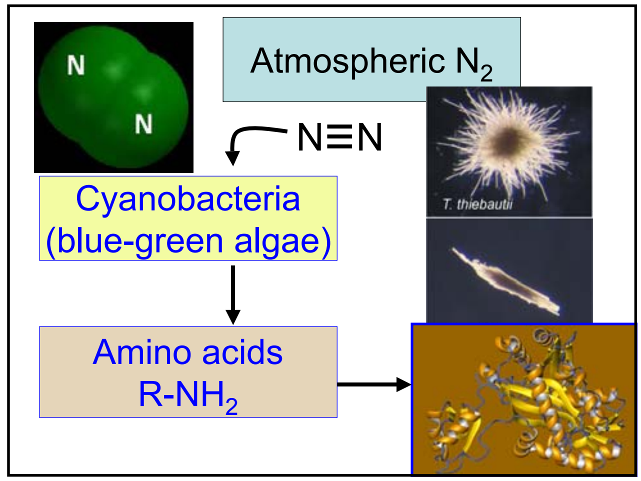
- Complex environmental chemistry
- Many oxidation states:

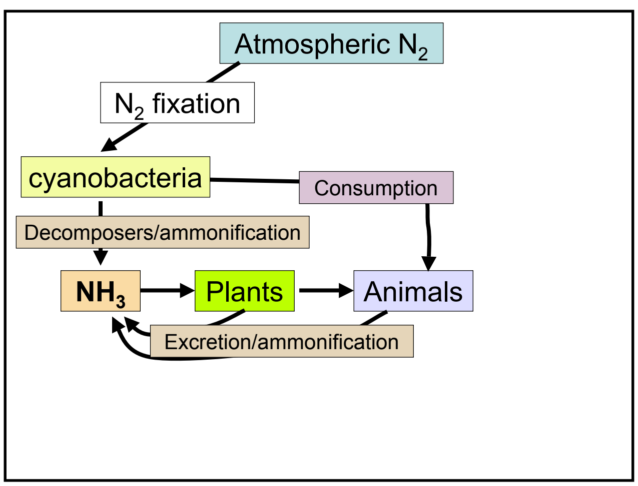
e.g. - 3 (NH_3; $R\text{-}NH_2$)

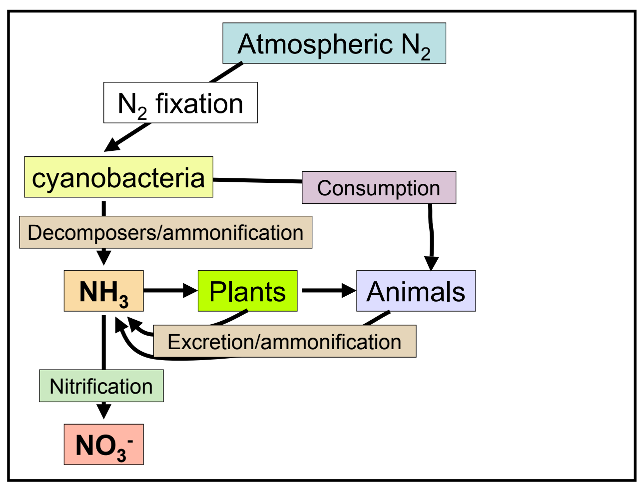
0 (N_2)

+ 5 (NO_3^-)

- Includes atmospheric gases (e.g. N_2), dissolved salts (e.g. NO_3^-) and solids (e.g. $R\text{-}NH_2$)





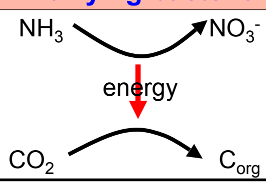


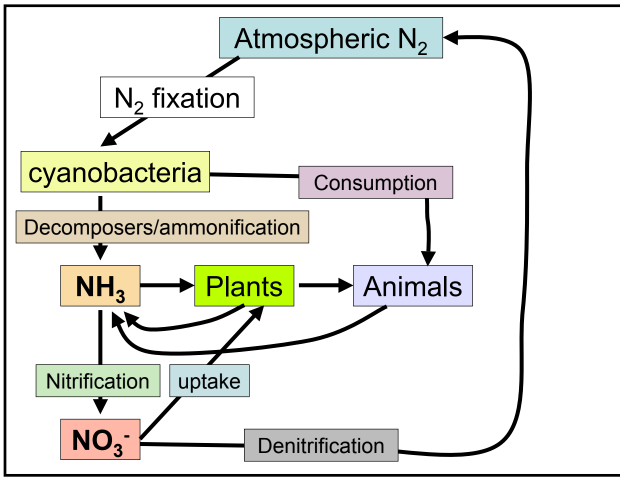
Nitrification

$2\ NH_3 + 3\ O_2 \rightarrow 2\ NO_2^- + 2\ H^+ + 2\ H_2O + \text{energy}$

$2\ NO_2^- + O_2 \rightarrow 2\ NO_3^- + \text{energy}$

Reactions catalyzed by
nitrifying bacteria

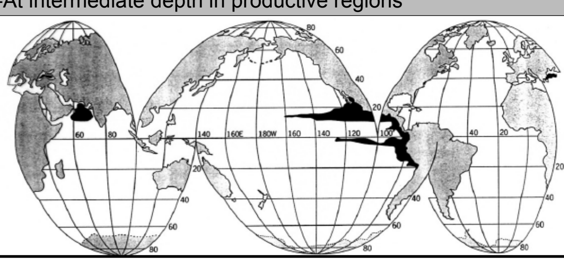


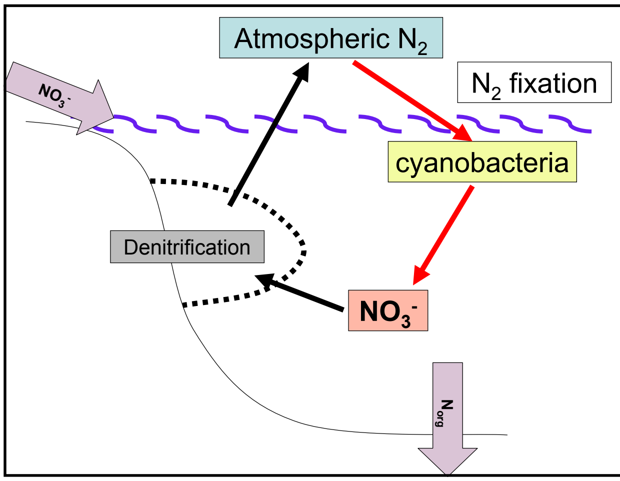


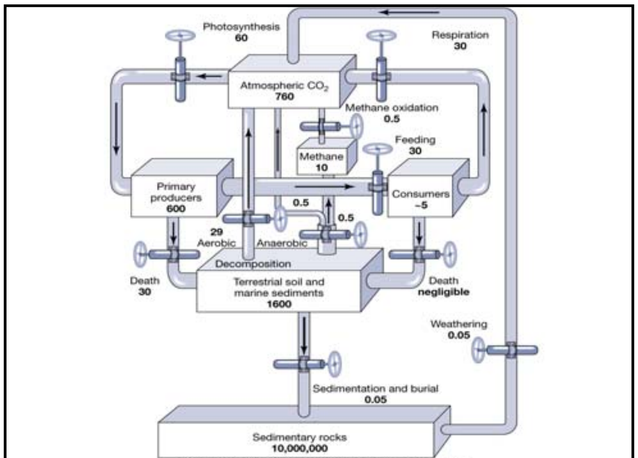
Denitrification

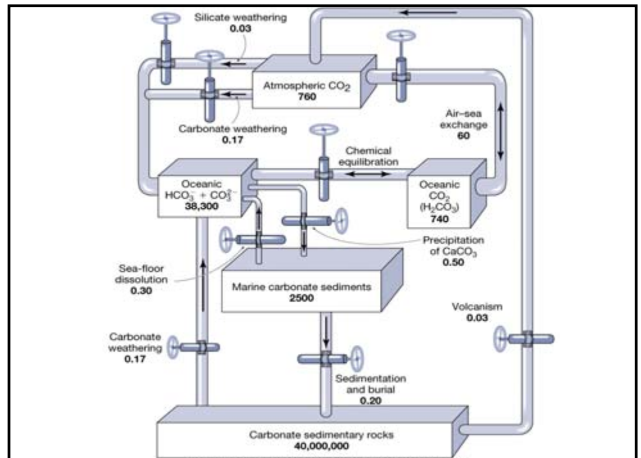
$CH_2O + 0.8\ NO_3^- + 0.8\ H^+ \rightarrow CO_2 + 0.4\ N_2 + 1.4\ H_2O + 505\ kJ$

In areas depleted in oxygen:
- Anoxic basins (Black Sea, some Fjords)
- Some sediments
- At intermediate depth in productive regions






Photosynthesis
60
Respiration
30
Atmospheric CO_2
760
Methane oxidation
0.5
Methane
10
Feeding
30
Primary
producers
600
Consumers
~5
0.5
0.5
29
Aerobic
Anaerobic
Decomposition
Death
30
Terrestrial soil and
marine sediments
1600
Death
negligible
Weathering
0.05
Sedimentation and burial
0.05
Sedimentary rocks
10,000,000


Silicate weathering
0.03
Atmospheric CO_2
760
Carbonate weathering
0.17
Air–sea
exchange
60
Chemical
equilibration
Oceanic
$HCO_3^- + CO_3^{2-}$
38,300
Oceanic
CO_2
(H_2CO_3)
740
Precipitation
of $CaCO_3$
0.50
Sea-floor
dissolution
0.30
Marine carbonate sediments
2500
Volcanism
0.03
Carbonate
weathering
0.17
Sedimentation
and burial
0.20
Carbonate sedimentary rocks
40,000,000