

ATMOSPHERIC CO₂, OCEAN CHEMISTRY, AND MECHANISMS OF CLIMATE CHANGE

- I. Recall: Ice core evidence for changes in atmospheric CO₂
- A. Pre-anthropogenic p_{CO₂} was about 280 ppmV
 - B. Glacial p_{CO₂} was about 190 ppmV. Four (soon to be five) Antarctic ice cores give same number. Greenland cores agree tolerably well, but their high dust loading and reaction with CO₂ leads to some problems.
 - C. In the early 1980' s, it seemed that the Dye 3 Greenland ice core showed relatively large, rapid (few hundred year) fluctuations in CO₂ during glacial stadial/interstadial fluctuations. These were not observed in the high resolution Byrd Antarctic ice core, and it is now thought that these apparent high CO₂ events were artifacts due to melt layers or interactions with dust.
 - D. So why did atmospheric CO₂ change with the glacial/interglacial cycles? Is the CO₂ change a chicken or an egg in the progression of climate change?
 1. Many ideas have been proposed; almost an equal number have been disposed (or is it deposited?). Speaking informally, it is as if theories on glacial/interglacial carbon dioxide are radioactive with two year half-lives.
 2. Despite this situation, there is much to be learned about CO₂ in the ocean from those ideas, so a historical examination of them is still worthwhile.
- II. CO₂ digression
- A. Two useful conservative quantities (properties that mix linearly) are:
 1. $\Sigma\text{CO}_2 = [\text{CO}_2(\text{aq})] + [\text{HCO}_3^-] + [\text{CO}_3^{=}]$
 2. **Alkalinity** = $[\text{HCO}_3^-] + 2[\text{CO}_3^{=}]$
 $+ [\text{B}(\text{OH})_4^-] + [\text{OH}^-] - [\text{H}^+]$
 $+ (\text{etc...})$
 - a. "Alkalinity" is a device for employing a special form of the charge balance equation which divides ions into those that have acid-base reactions and those which don't: e.g., in a system consisting of a solution of NaCl, MgSO₄, NaCO₃, and NaHCO₃:

$$[\text{Na}^+] + 2[\text{Mg}^{++}] - 2[\text{SO}_4^{=}] = [\text{HCO}_3^-] + 2[\text{CO}_3^{=}] + [\text{OH}^-] - [\text{H}^+]$$

$$= \text{Alkalinity}$$
 - b. Adding or removing CO₂ from a water sample does not change the alkalinity (convince yourself of this!)
 - C. A common simplification is that of Broecker:

$$[\text{CO}_3^{=}] = \text{Alk} - \Sigma\text{CO}_2$$

This approximation is conceptually simple but not accurate because of the neglect of significant contributions from borate and aqueous CO₂. For example, for values of Alk and ΣCO₂ typical of pre-anthropogenic warm surface waters (Alk=2275, ΣCO₂=1900; pH=8.3; pCO₂=283) and cold deep waters (Alk=2375, ΣCO₂=2260; pH=7.9):

units: (μmol/kg)

	[CO ₂ (aq)]	[HCO ₃ ⁻]	[CO ₃ ⁼]	Alk-ΣCO ₂	[B(OH) ₃]	[B(OH) ₄ ⁻]
ws	8	1628	264	375	288	119
cd	27	2150	85	115	349	58

D. By expressing all species in terms of [H⁺], then combining into a single equation in [H⁺], it can be shown that

$$pH = f\left(\frac{Alk}{\Sigma CO_2}, (T, S, P)\right)$$

Assuming deviations within the range of values likely to be found in the ocean, the relationship can be expressed as a simple cubic equation which has an exact solution. Since the relative amount of the various CO₂ species is a function of the pH, we can compute the absolute amount of each species as f(pH)*ΣCO₂; or directly as f(Alk, ΣCO₂).

The most correct formulation for CO₂ system calculations is to solve the cubic equation for the hydrogen ion where Alk, ΣCO₂, and borate are taken into account:

1. Let **A = Alk/ΣCO₂** and **B = ΣB/ΣCO₂**, let **[H⁺]** = the activity of hydrogen ion, and let K₁', K₂', and K_B' be the apparent constants in seawater at the appropriate temperature, pressure, and salinity; then:

$$\begin{aligned}
 2. \quad 0 &= [H^+]^3 A \\
 &+ [H^+]^2 [K_1'(A-1) + K_B'(A-B)] \\
 &+ [H^+] [K_1'K_B'(A-B-1) + K_1'K_2'(A-2)] \\
 &+ [K_1'K_2'K_B'(A-B-2)]
 \end{aligned}$$

E. A quick approximate solution to the carbonate system can be obtained by combining the equations for K₂' and K_B':

$$[B(OH)_4^-] / [CO_3^{=}] = (K_B'/K_2') ([B(OH)_3]/[HCO_3^-])$$

and noting that the ratio [B(OH)₃]/[HCO₃⁻] • 1/6 does not vary much throughout the ocean, (and (K_B'/K₂') • 3), so:

$$[B(OH)_4^-] \bullet [CO_3^{=}] / 2$$

and hence (assuming noting that CO₂(aq) is small and so are its variations (17±10):

$$CO_3^{=} \bullet (Alk - \Sigma CO_2 + 17) / 1.5 \quad (17 \text{ is the approximate aqueous } CO_2)$$

$$HCO_3^- \bullet \Sigma CO_2 - CO_3^{=} - 17 \quad \text{“} \quad \text{“} \quad \text{“}$$

$$[H^+] \bullet K_2' [HCO_3^-] / [CO_3^{=}]$$

$$[CO_2(aq)] \bullet [H^+] [HCO_3^-] / K_1'$$

$$p_{\text{CO}_2} \cdot [\text{CO}_2(\text{aq})] / \alpha_s'$$

F. Carbon Isotope Fractionation: equilibrium carbon isotope fractionation between CO_2 system species is a function of temperature. Kinetic lags are important.

$\delta^{13}\text{C}$ and gas exchange

A. Temperature-dependent equilibrium: fixing the carbon isotope composition of the atmosphere and allowing seawater to equilibrate:

	25° C	0° C
gaseous CO_2	-6.4	-6.4
~~~~~		
aqueous $\text{CO}_2$	-7.6	-7.7
$\text{HCO}_3^-$	+1.5	+4.4
$\text{CO}_3^{=}$	-0.5	+0.8
( $\text{CaCO}_3$ )	+3.0	+0.0
<hr/> $\Sigma\text{CO}_2$	+1.3	+4.1

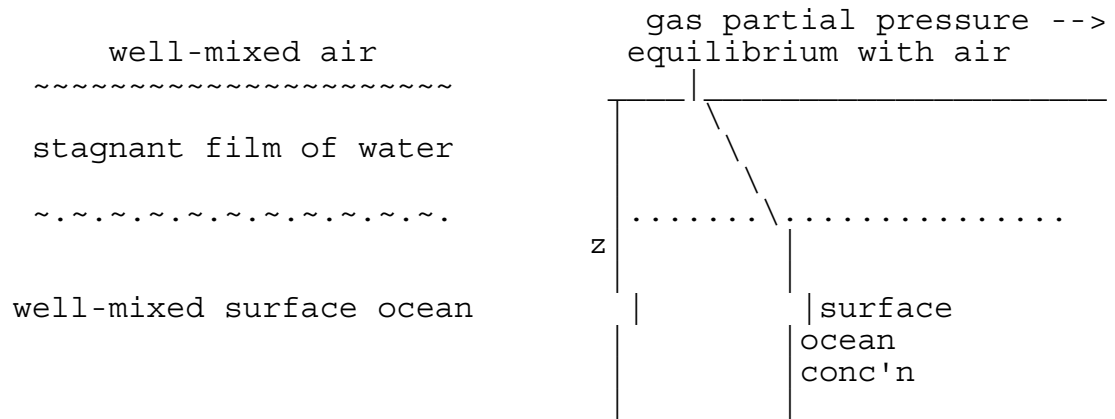
These isotope fractionation factors can be incorporated into the apparent equilibrium constants and then you can treat  $^{13}\text{CO}_2$  as a completely independent chemical system (apart from the pH control which is set by  $^{12}\text{CO}_2$ ).

# I. CO₂ gas exchange

## CO₂ exchange between the ocean and atmosphere

Stagnant film model:

D (~10⁻⁵cm²/sec; depends on gas)  
 z_{film} (~30 μm; depends on wind conditions).



$$\text{Flux} = D \frac{C_m - C_o}{z_{\text{film}}}$$

where C_m = concentration of dissolved gas in (interior) mixed layer of ocean  
 C_o = concentration of dissolved gas at surface of ocean (equilibrium with atm)  
 z_{film} = thickness of stagnant film  
 D = diffusion coefficient of dissolved gas

1. Piston Velocity concept. If we look at the dimensions of the above equation, it has dimensions of distance per time. One can think of this process as *if* imaginary pistons were moving through the water column and simultaneously pushing gas in and out of the ocean. This "piston" has a velocity magnitude of about 2000 m/yr!
2. For gases like oxygen and nitrogen, which equilibrate entirely between the gas phase and the dissolved aqueous form, gas exchange is very effective and surface waters are almost at equilibrium. For carbon dioxide however, where the dissolved aqueous gas equilibrates with the bicarbonate and carbonate ions which occur at much higher concentrations, gas exchange takes much longer and surface waters are usually out of equilibrium with the atmosphere - sometimes by up to a factor of two.
3. Rate of exchange of CO₂ between ocean and atmosphere. Speaking in round terms, we can calculate the average rate at which CO₂ moves across the sea surface:

$$2000 \text{ m/yr} * 10^{-5} \text{ moles/kg} * 1000 \text{ kg/m}^3 = 20 \text{ moles/m}^2/\text{yr}$$

piston vel.    aq. CO₂ conc.    conversion factor

Given that the upper 75 m of the water column (mixed layer) underneath a square meter contains 150 moles of carbon, full equilibration for carbon isotopes (e.g. ¹³C, ¹⁴C) can take some time (years). pCO₂ equilibration takes less time however, because the pH shift induced by gas exchange shifts the water towards equilibrium (e.g., if water is supersaturated with respect to atmosphere, CO₂ is lost and pH of seawater rises and

some aqueous  $\text{CO}_2$  is lost to the bicarbonate pool. Hence  $p_{\text{CO}_2}$  is moved towards equilibrium both by loss of aqueous  $\text{CO}_2$  to atmosphere and by loss of aqueous  $\text{CO}_2$  to bicarbonate pool (this is related to the Revelle Factor).

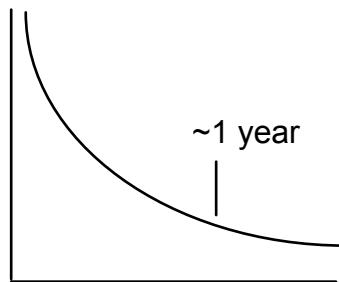
4. One square meter of the the upper 100m of the ocean ( $10^5\text{kg}$ ) contains 200 moles of carbon. 0.5% of this carbon is as dissolved  $\text{CO}_2$ , and the Revelle Factor is  $\sim 10$ , so about  $200 \times 0.005 \times 10 = 10$  moles of  $\text{CO}_2$  needs to be transferred to equilibrate the mixed layer with the atmosphere. Since the gas exchange rate is  $20 \text{ moles/m}^2/\text{yr}$ , the mixed layer of the ocean can equilibrate with the atmosphere on a time scale of about a half-year (note however that  $^{13}\text{C}$  and  $^{14}\text{C}$  will take longer to equilibrate because the total carbon dioxide must exchange to fully equilibrate the isotopes, i.e.,  $200/20 = 10$  years). Hence the "average" anthropogenic  $\text{CO}_2$  molecule has plenty of time to equilibrate with the mixed-layer.

Simple Gas Exchange  
(e.g.  $\text{O}_2$ , Ar)



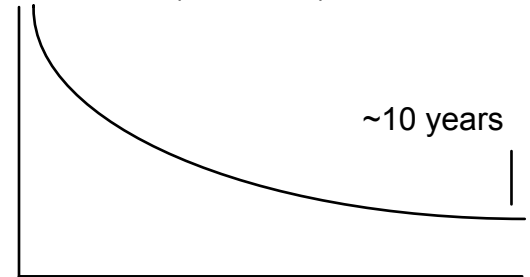
(depth of mixed layer divided by piston velocity - i.e. total gas content divided by gas flux)

$p_{\text{CO}_2}$  equilibration



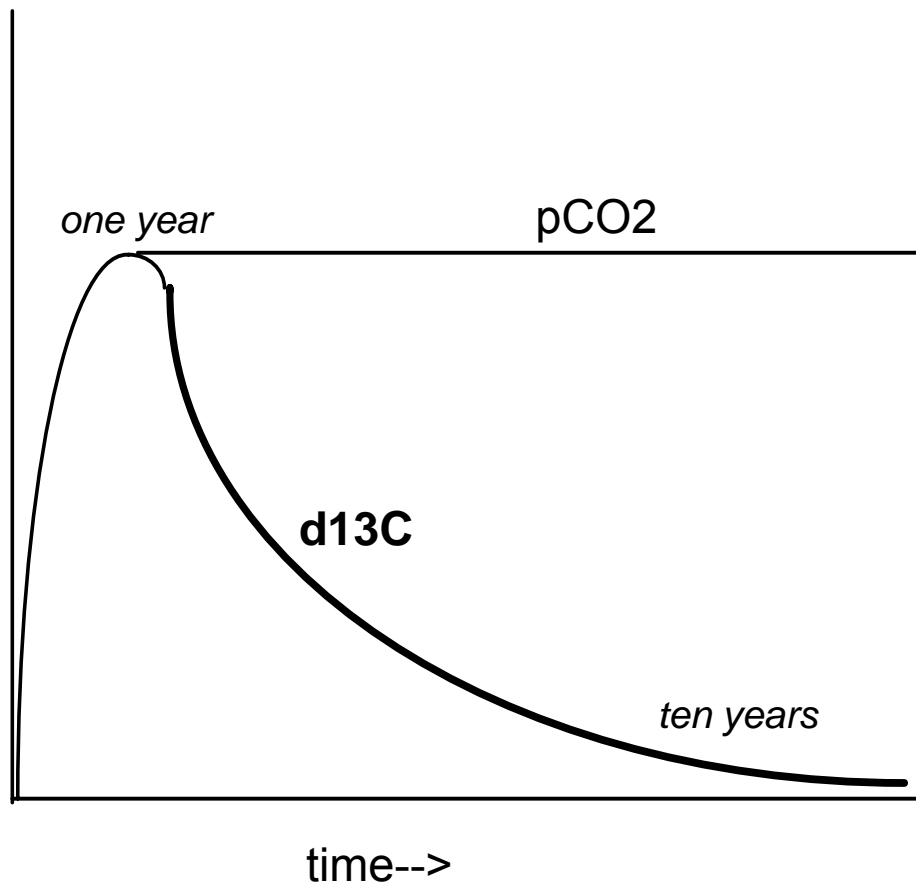
(change in  $\text{TCO}_2$  required to change  $p_{\text{CO}_2}$  in seawater is divided by gas flux: e.g. for a 3% increase in  $p_{\text{CO}_2}$ ,  $\text{CO}_2(\text{aq})$  rises by 3% and  $\text{TCO}_2$  rises by 0.3%; but because  $\text{TCO}_2$  is  $\sim 200\times \text{CO}_2(\text{aq})$ , it then takes  $200 \times 0.3/3 = 20\times$  longer

carbon isotope equilibration  
( $\text{C}^{13}$ ,  $\text{C}^{14}$ )

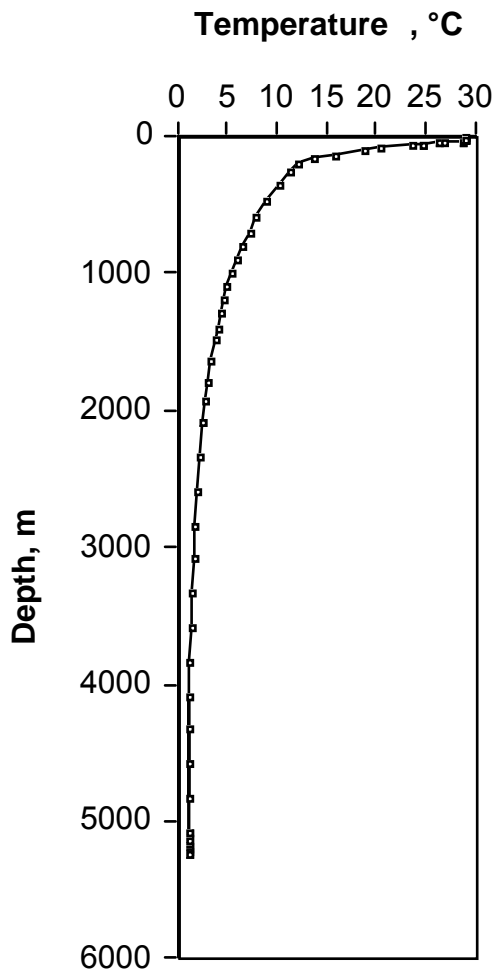


(total dissolved carbon dioxide divided by  $\text{CO}_2$  gas flux,  $\sim 20 \text{ moles/m}^2/\text{yr}$ )

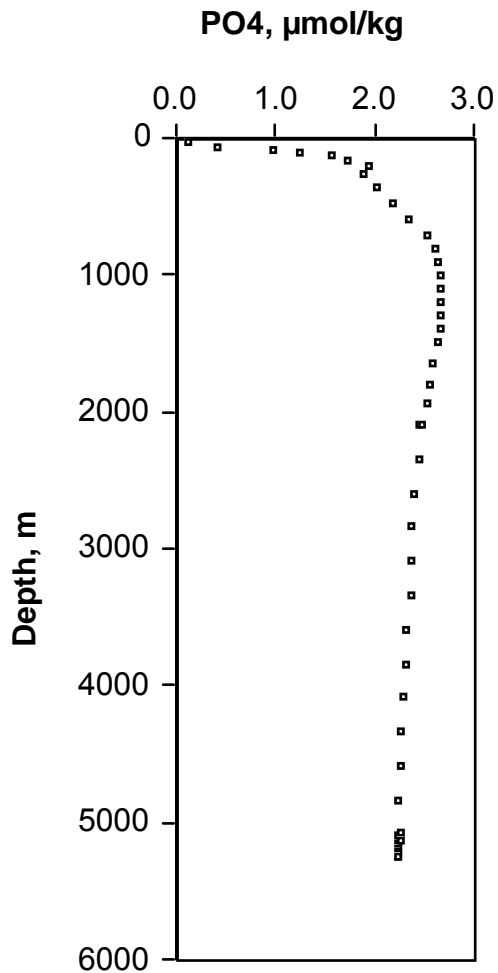
Kinetic-isotope disequilibrium (Lynch-Stieglitz et al., 1995): Imagine a mixed layer system which is in equilibrium for carbon isotope ratio, but below equilibrium for  $p_{\text{CO}_2}$ . Gas exchange will drive the mixed layer into equilibrium in one year's time; but in doing so, it will shift the mixed layer towards the lighter isotope (because it is the light  $\text{CO}_2(\text{aq})$  species which is involved in gas exchange). After about 10 years, the system comes back into isotopic equilibrium.



H. CO₂ and the oceans

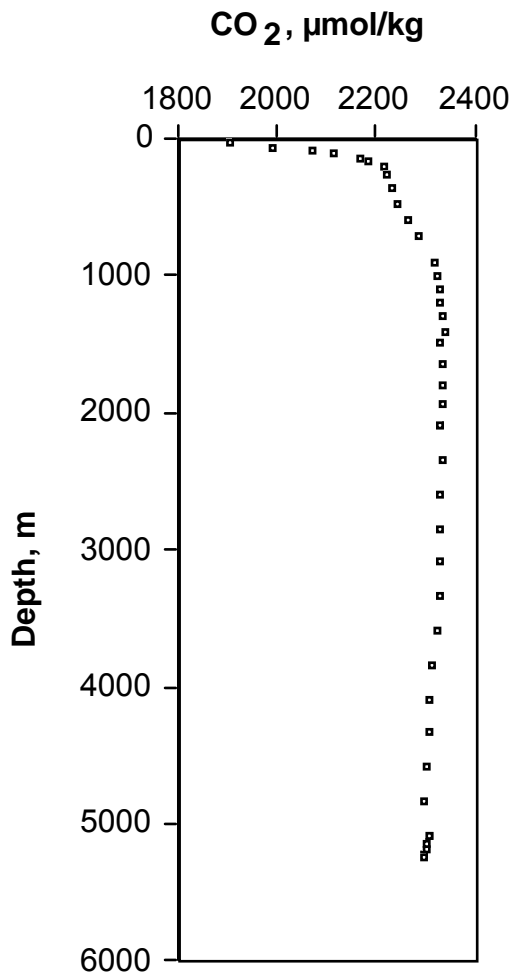


1. **The solubility pump:** CO₂ is more soluble in cold waters than in warm waters. If alkalinity were uniform throughout the ocean and if both cold and warm surface waters equilibrated their pCO₂ with the atmosphere, then cold surface waters would have a higher dissolved carbon dioxide content than warm surface waters. As these cold surface waters circulate into the deep interior of the ocean, then deep waters will have more CO₂ than warm surface waters.
2. **The biological pump:** organisms remove carbon and nutrient elements from the surface ocean (which is equilibrated with atmospheric oxygen; note oxygen solubility is a function of temperature); the debris from these organisms sinks and decomposes, releasing carbon and nutrient elements into the deep water and consuming oxygen. Redfield Ratio: -138 O₂: 106 CH₂O: 16 NH₃: 1 H₃PO₄. Example: phosphorus-depth profile from the Northern Indian Ocean



3. **The Salt Pump:** Evaporation-precipitation cycle will change the  $p_{\text{CO}_2}$  of a surface water significantly, because:
  - a.  $\text{pH} \approx \text{constant}$  (so  $\text{CO}_2(\text{aq})/\Sigma\text{CO}_2 = \text{constant}$ )
  - b.  $\Sigma\text{CO}_2$  goes up (linearly with salinity)
  - c. Revelle factor of 10 for change in  $p_{\text{CO}_2}$  relative to  $\Sigma\text{CO}_2$
4. Total dissolved carbon dioxide distribution (combination of all pumps, but mainly due to solubility pump and biological pump)





III. Broecker (1982) interpretation of G/I pCO₂ and oceanic δ¹³C evidence

- A. Considers various mechanisms which might change pCO₂. Must involve the ocean, because it is dominant reservoir with which atmosphere communicates on ~20,000 year time scale. Salinity change (due to ice buildup) makes small difference, as does temperature change (if you accept the CLIMAP T estimates); these latter cancel each other out.

B. Consider surface water as deep water which has warmed up, had its P removed by organisms  
 (and accompanying removal of 106 CH₂O : 21 CaCO₃ : 16 NH₃ : 1 P)

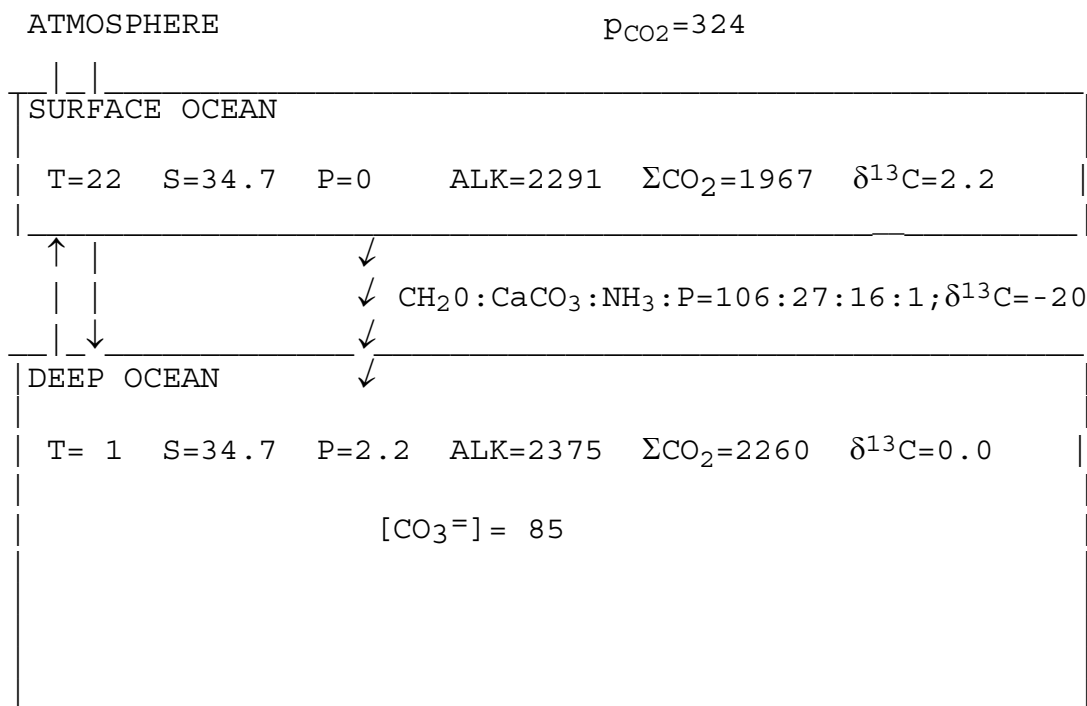
$$\text{CO}_2(\text{surface}) = \text{CO}_2(\text{deep}) - (\text{106} + \text{21}) * \text{P}(\text{deep})$$

$$\text{Alk}(\text{surface}) = \text{Alk}(\text{deep}) - (2*21 - 16) * \text{P}(\text{deep})$$

Surface ΣCO₂ = 2260 - 2.2*(106+27) = 1967

Surface Alk = 2375 - 2.2*(54-16) = 2291

Interglacial two-box ocean:



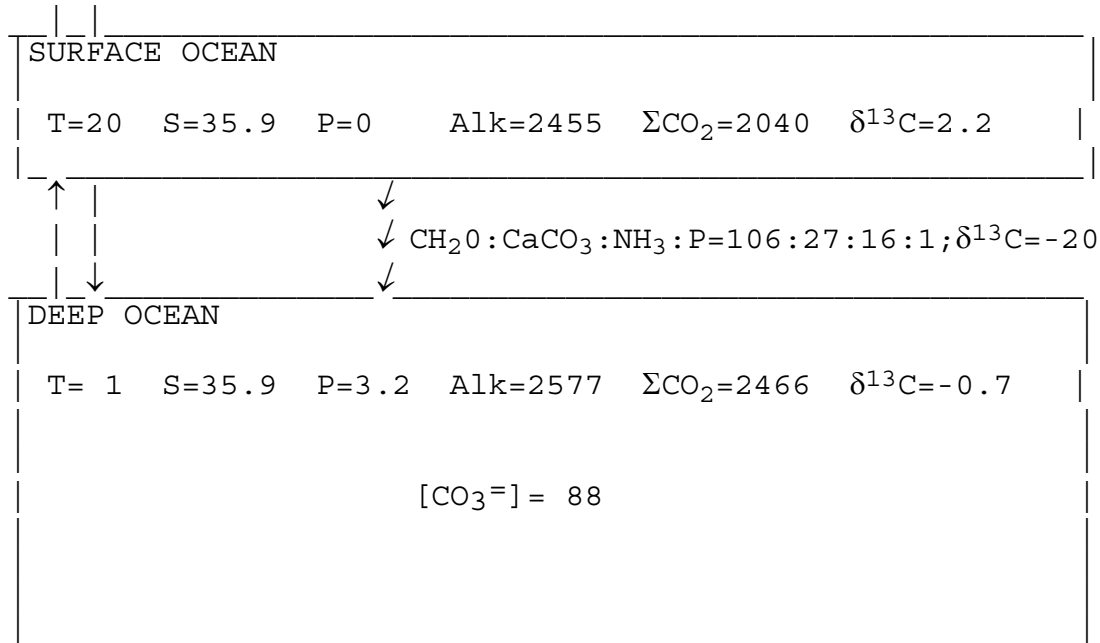
Proposed Glacial two-box ocean (add "Redfield" carbon from continental shelves):

Surface ΣCO₂ = 2466 - 3.2*(106+27) = 2040

Surface Alk = 2577 - 3.2*(54-16) = 2455

ATMOSPHERE

$P_{CO_2}=241$



• (Assumes that marine organic matter from continental shelf is oxidized and put into ocean, and that sedimentary  $CaCO_3$  dissolves to keep  $[CO_3^{=}]$  of deep ocean is approximately constant)

### C. Comparison to data

1. At the time, the calculated deep-sea  $\delta^{13}C$  change was similar to the -0.7 permil suggested by Shackleton (1977) [although we now know that the change is less, about -0.3 to -0.4 ‰]
- 2. Planktonic (surface)  $\delta^{13}C$  does not change in model; this result was consistent with evidence available at the time [but data for planktonics had a relatively large scatter which raises some concerns].
3. One major problem with this model is that it would deplete the deep-sea of oxygen. This outcome clearly did not happen (as proven by the existence of glacial-age benthic foraminifera, which are aerobic organisms).

### IV. Shackleton et al. (1983) planktonic-benthic $\delta^{13}C$ record

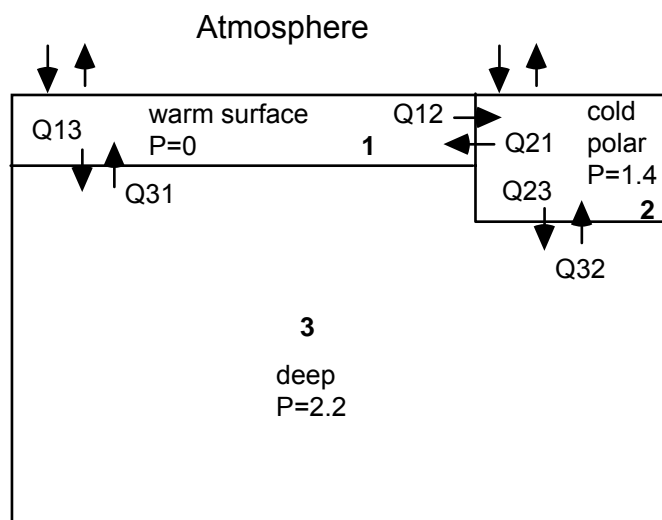
- A. Did detailed analysis of N. dutertrei and Uvigerina  $\delta^{13}C$  in the same core (V19-30). Subtracted the two to get surface-deep  $\delta^{13}C$  contrast, which determines atmospheric  $CO_2$ , according to Broecker model. Appears to confirm reduced  $CO_2$  during glacials; however, it appears that  $CO_2$  change occurs before that of sea level (as indicated by  $\delta^{18}O$ )

### V. Cd evidence:

- A. It does not appear that deep ocean Cd did not increase; best current evidence is that it decreased by -15% with a margin of error comparable to the change (Broecker theory requires +40% increase in P).
- B. Paired Cd- $\delta^{13}C$  evidence suggests that much of the global  $\delta^{13}C$  change is due to P-free carbon (trees and soils).

### V. Pre-formed phosphate scenario (Toggweiler and Sarmiento; Wenk and Siegnethaler; Ennever and McElroy)

- A. One unrealistic feature of the two-box ocean is that not all of the phosphorus is removed before water that upwells at high latitudes is cooled and returned to the deep sea as bottom. About  $\frac{1}{3}$  of all deep ocean phosphorus arrives advectively rather than by particulate transfer. In effect, this makes the deep ocean "leaky" with respect to  $\text{CO}_2$ , so that atmospheric  $\text{CO}_2$  levels are higher than they would be in the absence of this "leak". Factors that change the "leakiness" of the deep ocean carbon dioxide can change atmospheric  $\text{CO}_2$
- B. Another unrealistic feature of the two-box ocean is that warm surface waters do not fill up most of the deep ocean; instead, cold surface waters at high latitudes sink to fill up most of the ocean. This aspect brings more oxygen into the deep sea.
- C. So it is more realistic to have the deep ocean in communication with a cold surface ocean which contains residual phosphorus.
- D. Consider a three-box ocean with a cold polar outcrop



1. Definition of pre-formed phosphate:  $\text{PFP} = P - r \text{AOU}(T)$   
 where  $r = \Delta\text{O}_2:\Delta\text{P}$  "Redfield ratio"  
 and  $\text{AOU} = \text{Apparent oxygen utilization (equil. sol. - observed O}_2)$
  2. Calculation of magnitude of  $\text{CO}_2$  changes for a given PFP change.
  3. Ways to change pre-formed phosphate:
    - a. Change relative exchange rates of warm surface and deep ocean with cold polar.
    - b. More biological productivity in cold polar surface waters (?why? Martin iron hypothesis).
    - c. Change nutrient content of thermocline and hence upwell water of different nutrient content at high latitudes.
- D. Advantages/Disadvantages of this theory:
1. Advantage: can produce rapid  $\text{CO}_2$  changes
  2. Advantage: consistent with benthic-planktonic  $\delta^{13}\text{C}$  evidence
  3. Disadvantage: doesn't solve oxygen problem
  4. Disadvantage: there is no evidence that PFP changes! (high latitude planktonic Cd does not change, and planktonic  $\delta^{13}\text{C}$  actually gets more negative)
- E. Multi-box steady-state ocean modeling using linear matrix methods: Ocean Box Modeling.

## VIII. Coral Reef hypotheses

- A. Basic premise is that  $\text{CaCO}_3$  precipitation leads to release of  $\text{CO}_2$ :  $\text{Ca}^{++} + 2\text{HCO}_3^- = \text{CO}_2 + \text{CaCO}_3$ , and that a rise in sea-level will increase the growth opportunities for corals.
- B. Keir-Berger w/"Worthington's Lid": this effect is enhanced by formation of a meltwater lid that keeps much of the released  $\text{CO}_2$  in the mixed layer/atmosphere (hence coral growth rates don't have to be especially high). Snag: no evidence that deep ocean ventilation rates were cut off during deglaciation (recall: benthic-planktonic  $^{14}\text{C}$  difference).
- C. Opdyke version (Holocene coral growth rates greater than input from weathering). Coral growth rates are not likely to be this high, and the continuation of this phenomenon for 10,000 years would drive the deep ocean into a state of intense carbonate undersaturation.

## VI. Nutrient deepening/alkalinity response model (Boyle, 1988).

- A. Recall that one of the results of the paleo-deep water studies is that glacial P and  $\delta^{13}\text{C}$  distributions change to move nutrients from the upper ocean into the deepest parts of the ocean. Since  $\text{CO}_2$  is a weak acid, this movement of  $\text{CO}_2$  into the deep ocean will make the deep ocean more corrosive to  $\text{CaCO}_3$  (i.e., deep  $[\text{CO}_3^-]$  will go down).
- B. As Broecker (1982) pointed out, if weathering input and deep-sea biological carbonate production remains constant, then in order to maintain the CCD/lysocline at the same position, the deep  $[\text{CO}_3^-]$  must remain approximately constant. If there is more  $\text{CO}_2$  in deep waters, then to keep  $[\text{CO}_3^-]$  constant, then the alkalinity will have to rise. This is achieved by dissolving some of the  $\text{CaCO}_3$  on the seafloor:  $\text{CO}_2 + \text{CaCO}_3 = \text{Ca}^{++} + 2\text{HCO}_3^-$ . Note that this reaction increases the alkalinity of the whole ocean by two  $\mu\text{eq/kg}$  for every  $\mu\text{mol/kg}$  of  $\text{CO}_2$  added from dissolution. Hence the ratio of  $\text{Alk}/\Sigma\text{CO}_2$  in the whole ocean increases, pH rises, and  $p\text{CO}_2$  drops.
- C. A complication of this model is that  $\text{CO}_2$  may not be transferred vertically independently of alkalinity; in some scenarios, the  $[\text{CO}_3^-]$  response is small (e.g. versions that increase vertical mixing between the surface and intermediate depth ocean) while in others it is much larger (e.g., an increase in the fraction of organic matter that decomposes in the deep ocean as opposed to the intermediate depth ocean).
- D. The state of this idea: (1) it doesn't appear that the whole ocean nutrient-deepening is quantitatively large enough to account for more than a fraction of  $p\text{CO}_2$  drop, even given the most sensitive scenario (Boyle, 1992), and (2) the  $[\text{CO}_3^-] = \text{constant}$  constraint may not be the best way to simulate the lysocline/CCD response; using a more explicit model of carbonate dissolution on the seafloor reduces the effectiveness of this mechanism at changing  $p\text{CO}_2$  (Emerson and Archer, 1993).

## VII. Rain Ratio Model

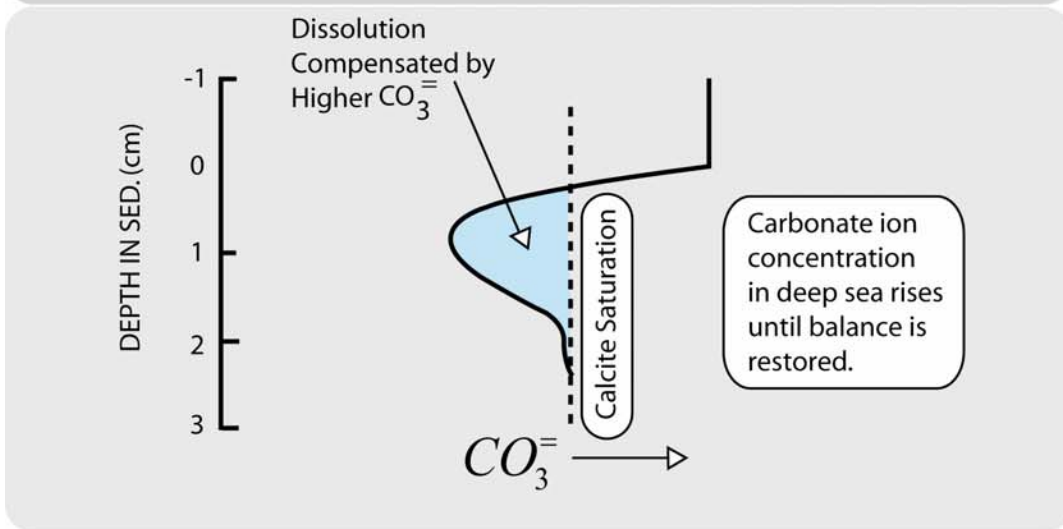
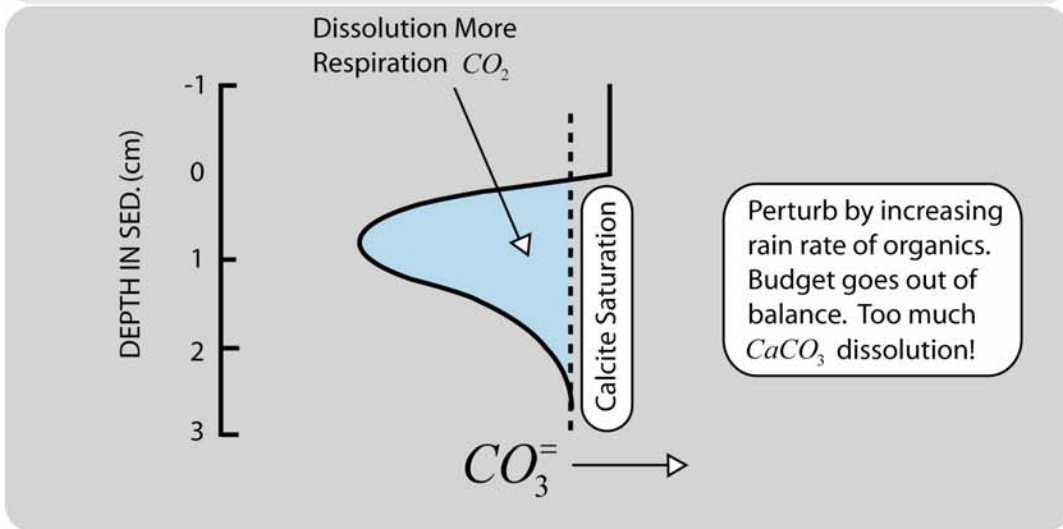
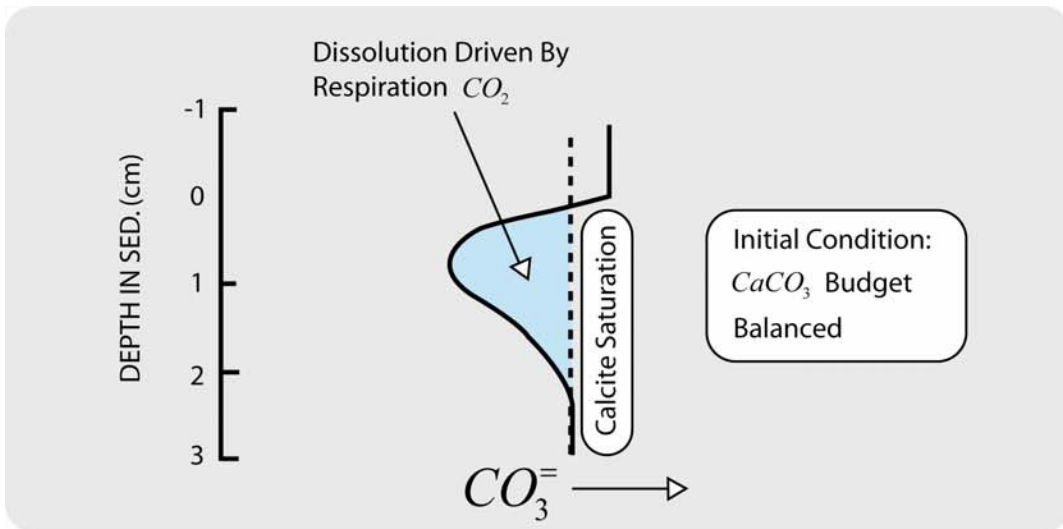
- A. Exists in various incarnations. What the ideas have in common is noting that there is a "rain ratio" of  $\text{CH}_2\text{O}:\text{CaCO}_3$ , and that this ratio may have changed between glacial and interglacial times (e.g. Keir and Berger, 1985).
- B. Early versions of this idea sprang from the idea that "diatoms are more efficient at reducing surface water  $p\text{CO}_2$  than coccolithophorids". At face value, this notion seems subtly obvious: diatoms remove only  $\text{CO}_2$  (as  $\text{CH}_2\text{O}$ ) while coccolithophorids remove both  $\text{CO}_2$  and  $\text{CaCO}_3$ . But because coccolithophorids remove two units of alkalinity for every  $\text{CaCO}_3$  removed, they lower the pH and hence create a higher  $p\text{CO}_2$  for a given  $\Sigma\text{CO}_2$  level. However, as we saw earlier, reducing the low-latitude  $\text{CaCO}_3:\text{CH}_2\text{O}$  rain ratio by a factor of

in the Toggweiler-Sarmiento 3-box ocean model had a significant response (279->254 ppmV), but not enough to account for glacial  $p_{\text{CO}_2}$  lowering. Even lowering the ratio by a factor of 4 in that model - placing the rain rate precariously close to the point at which carbonate isn't being deposited as fast as it is being weathered - being only gives an additional 12 ppmV. The reason for this lack of response is that a mass-conserving ocean will often take  $\text{CO}_2$  that is removed in one part of the ocean pop back up in another part of the ocean, cancelling the apparent effectiveness of the original change. In this case, the upwelling of deep water to the high latitude ocean counters the direct effect of the low-latitude rain ratio change. Let this serve as a warning about purely qualitative ideas about mechanisms controlling  $p_{\text{CO}_2}$ .

C. In a more subtle variation on this theme, it is noted that currently about 5 times more calcium carbonate is being formed by organisms than is coming in from rivers, hence 80% of the biogenic flux is dissolved. If the rain ratio is reduced by a factor of 50% (biogenic =  $2.5 \times \text{input}$ ), only 60% of the biogenic flux needs to be dissolved. Hence the  $[\text{CO}_3^{2-}]$  concentration of the deep ocean can be higher, leading to higher oceanic alkalinity and hence lower  $p_{\text{CO}_2}$ .

D. Archer/Maier-Reimer O-GCM w/ $\text{CO}_2$  chem:

1. Model is an ocean GCM including carbon cycle chemistry and a  $\text{CaCO}_3$  dissolution model for the seafloor, holding ocean circulation constant. Key characteristic of dissolution model is a major role for *in-situ* organic carbon degradation in  $\text{CaCO}_3$  dissolution. In the model, the  $\text{CH}_2\text{O}:\text{CaCO}_3$  rain ratio is varied slightly, and it is found that the degree of *in-situ* dissolution increases greatly - so much so, that most of the deep ocean is supersaturated with respect to  $\text{CaCO}_3$ ! Hence, the balance between input and output of  $\text{CaCO}_3$  is mediated by a balance between the extent of supersaturation and the extent of *in-situ* organic carbon degradation (higher supersaturation leads to lower  $\text{CaCO}_3$  dissolution, higher *in-situ* organic degradation leads to more  $\text{CaCO}_3$  dissolution; for a given organic degradation the deep ocean supersaturation moves to the point where dissolution is sufficient to compensate for excess of production over input. Notable in model is the relatively small change in rain ratio necessary to create large  $p_{\text{CO}_2}$  changes.



The Archer-Maier-Reimer scenario for the cause of the low atmospheric  $CO_2$  content during glacial time (Archer and Maier-Reimer, in press). An increased rain rate of organic matter raises the amount of respiration  $CO_2$  released to sediment pore waters. This increase causes the rate of  $CaCO_3$  dissolution to exceed that required for material balance. As a result, the  $CO_3^{2-}$  ion content of the sea rises, eventually compensating for the excess dissolution. This rise in ocean  $CO_3^{2-}$  content drives down the atmosphere's  $CO_2$  content.

2. Model predicts large pH increase in deep ocean. This may be verifiable by boron isotope paleo-pH method.

VIII. R. Keeling "sea ice/gas exchange mechanism: proposes that LGM sea ice cover over the Antarctic prevented the CO₂ "leak" in the Southern Ocean. Problem: sea ice cover must be extremely efficient (>95%) for this mechanism to work.

IX. Nitrate as a limiting nutrient: the idea and it's development

- A. McElroy et al. proposed that instead of changes in oceanic phosphorus, nitrate changes could control glacial/interglacial CO₂ cycles. The reason for an increase in nitrogen in the glacial ocean would be a decrease in denitrification (perhaps due to lower sealevel, although other causes may also produce this effect). Because nitrate has a residence time of about 10⁴ years (compared to 10⁵ years for P), it is easier to change the oceanic N value on G/I timescales.
- B. This idea was initially discounted on two grounds:
  1. Redfield argued that P, not N, is the limiting nutrient in seawater because it is only supplied and lost through geological interactions (dissolution from rocks, loss to sedimentation). Nitrogen, on the other hand, is dominantly fixed biologically (nitrogen fixers), and if the supply of N is not sufficient, the ecological balance would be shifted towards nitrogen fixers, and the nitrate supply would "catch up" with the available phosphorus.
  2. Altabet and Curry showed that it did not appear that the isotopic composition of oceanic nitrogen changed on G/I timescales. This was taken as evidence against a large decrease in denitrification during glacial times, because denitrification is accompanied by a large isotopic fractionation.
- C. This idea has been recently revived (in somewhat modified form) given the following arguments:
  1. Falkowski has argued that the Redfield argument is wrong because it does not take into account another limiting nutrient: iron. Redfield derived his arguments from terrestrial studies (Hutchinson), where Fe is not limiting. However, in the ocean in many environments, iron is limiting – in some cases to marine life, period, but perhaps almost everywhere with respect to nitrogen fixation. Nitrogen fixation is an iron-intensive biochemical process. Hence nitrate, not phosphate might be the limiting nutrient.
  2. It is not clear that a reduction in denitrification will necessarily be reflected in a change in the isotopic composition of oceanic nitrate. Recent studies by Devol show that continental margin sediments – where nitrate is completely depleted at some depth in the sediment due to denitrification – do not evidence any isotopic fractionation in nitrate because all of the nitrate – both N¹⁴ and N¹⁵ forms – are completely consumed within the sediments. Although a very small isotope fractionation may remain – due to the difference in the diffusion of ¹⁵NO₃⁻ (63 amu) compared to ¹⁴NO₃⁻ (62 amu) – the large isotope fractionation due to denitrification has little effect.
- D. Studies of the nitrogen isotope composition of continental margin sediments off of Mexico (Ganeshram et al., 1995) and in the Arabian Sea (Altabet et al., 1995) suggest less efficient glacial denitrification in these environments.
- E. Denitrification today appears to occur roughly equally on continental shelves and in the low-oxygen zones of the Eastern Tropical Pacific. If both of these sinks was diminished, and the supply of fixed nitrogen enhanced by the greater iron supply in the dusty glacial atmosphere, then the nitrate content of the ocean may well have been higher. The fly in this



ointment: how will we ever establish that this happened (paleo-nitrate indicators being in limited supply).

- F. Studies of the nitrogen isotopic composition of Antarctic sediments and diatoms (suggesting somewhat more efficient nitrate utilization) have also revived the polar nutrient hypothesis to some extent.
- X. Toggweiler "change in mode of Southern Ocean deepwater formation" model (personal communication, 2002). Premise is that in the ocean today, AABW formation occurs on the continental shelf after nearly complete equilibration of the surface water with atmosphere (hence CO₂ is "lost" from ocean to atmosphere). He proposes that during the last glacial maximum, deepwater formation proceeded by deep ocean convection (short-lived uniform column of high density water from surface to bottom) where gas exchange equilibration cannot occur - hence sealing off the ocean CO₂ "leak" and reducing atmospheric CO₂.

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