Tracing the present and past ocean carbon cycle

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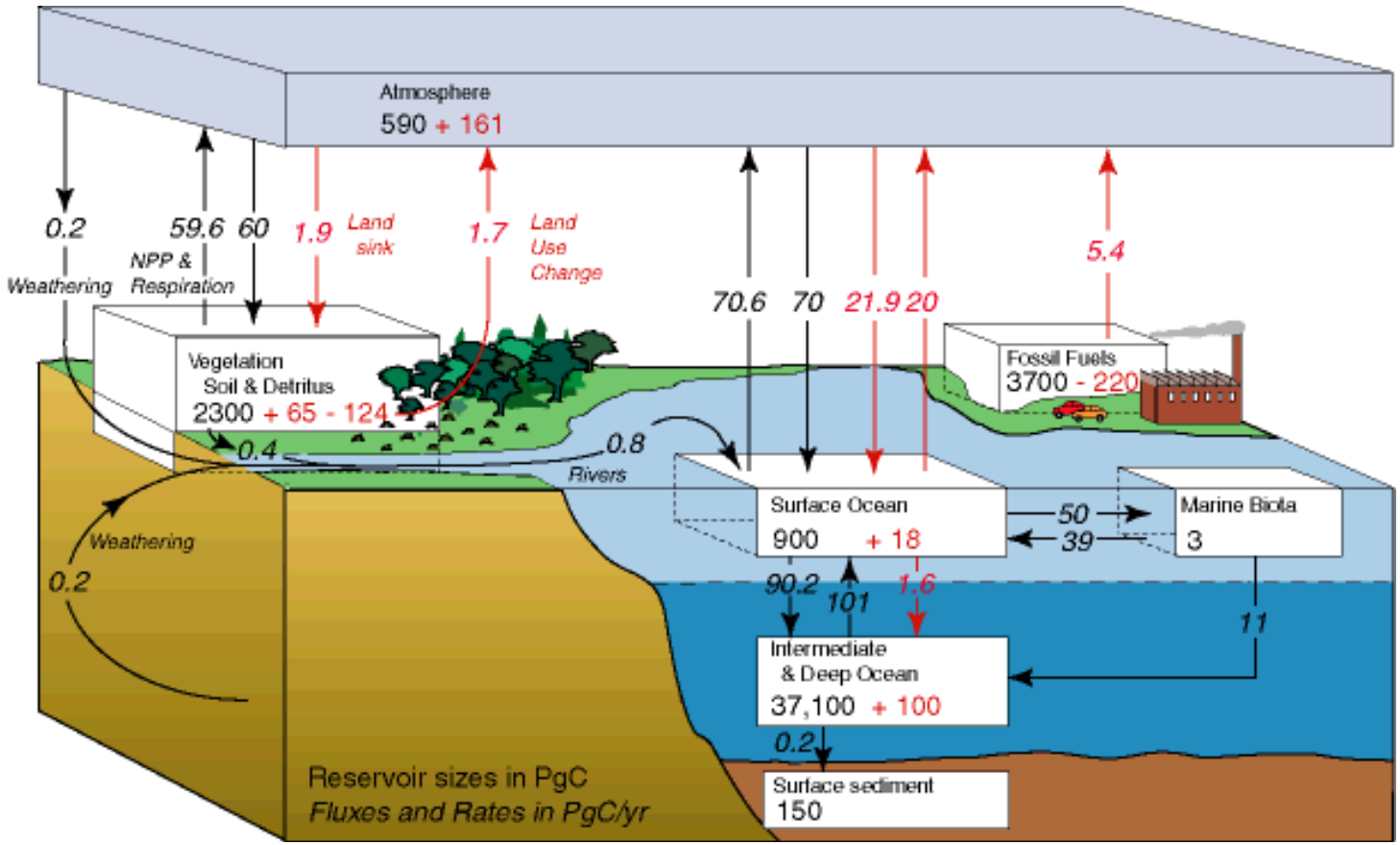
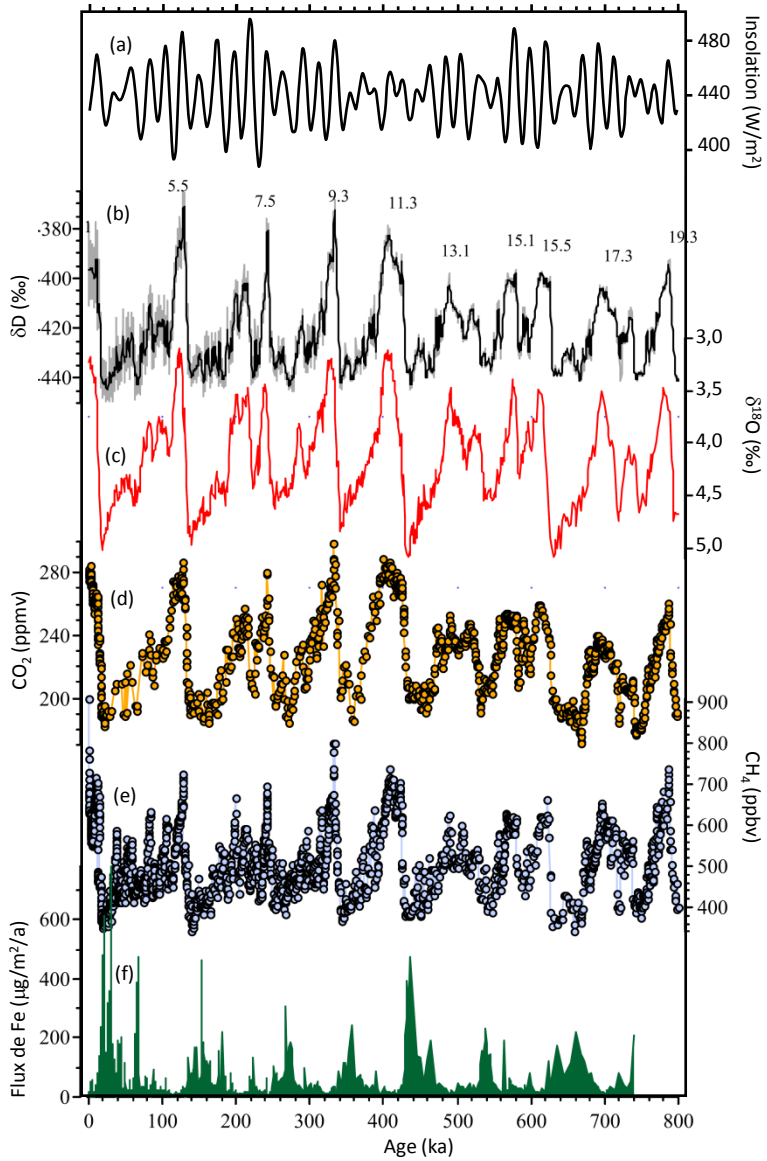
Global carbon cycle



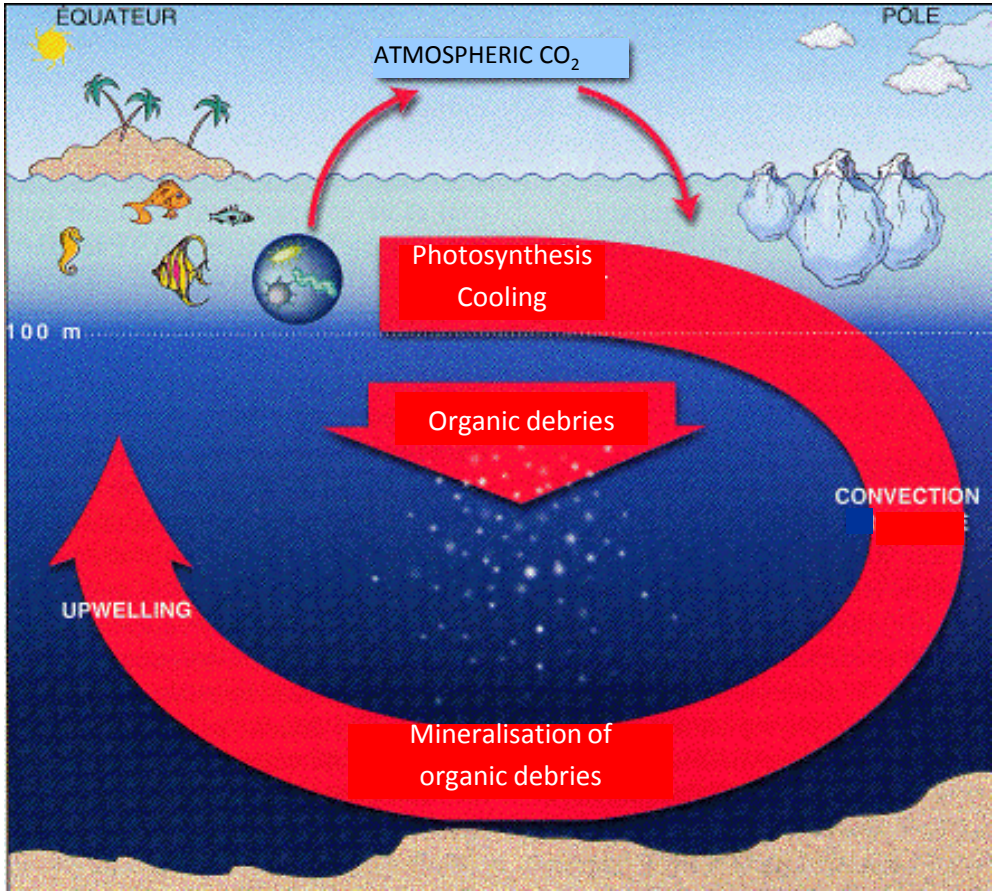
→ Ocean is the main C reservoir

Past carbon variability



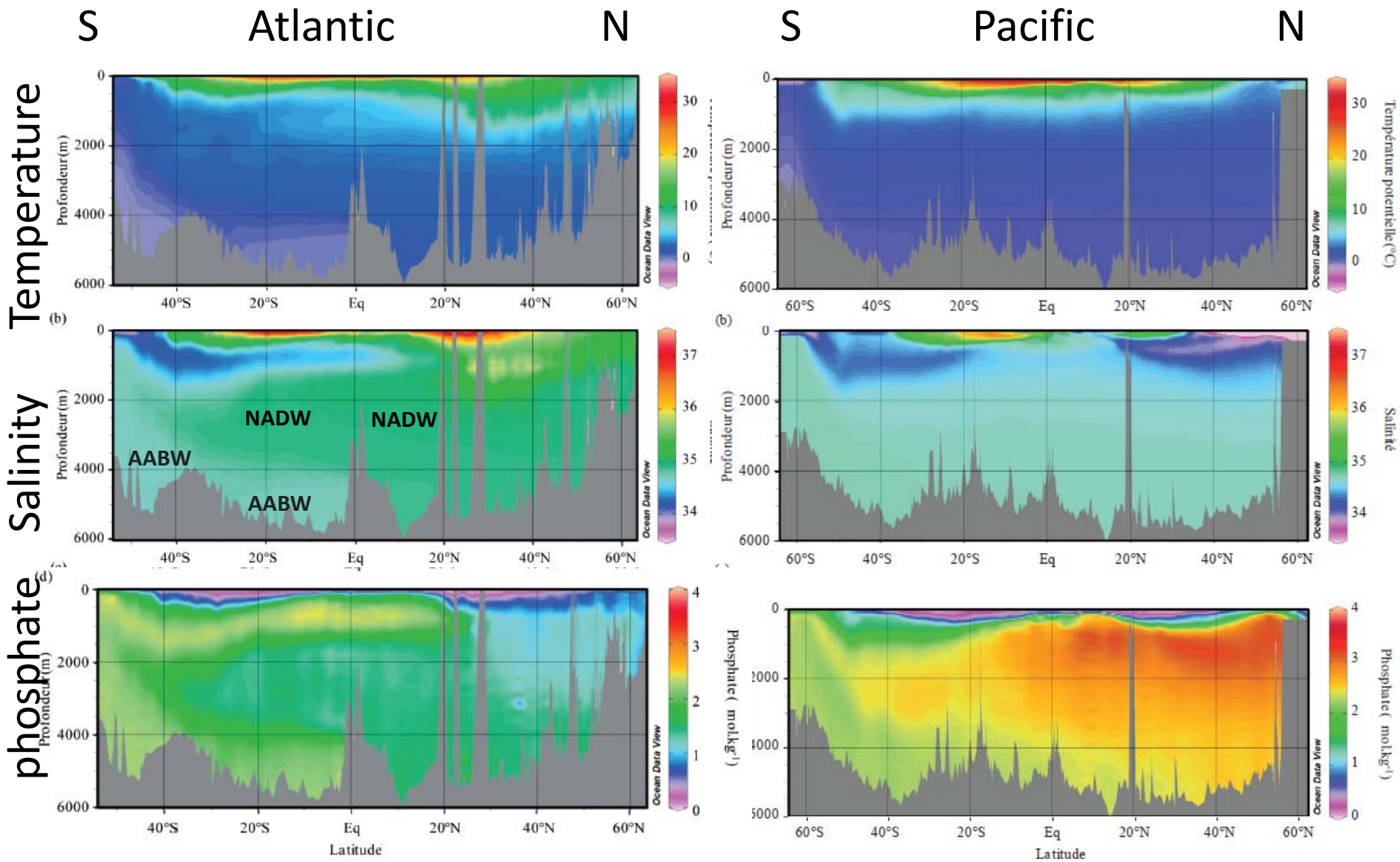
Tracing the present and past ocean carbon cycle

The oceanic carbon cycle

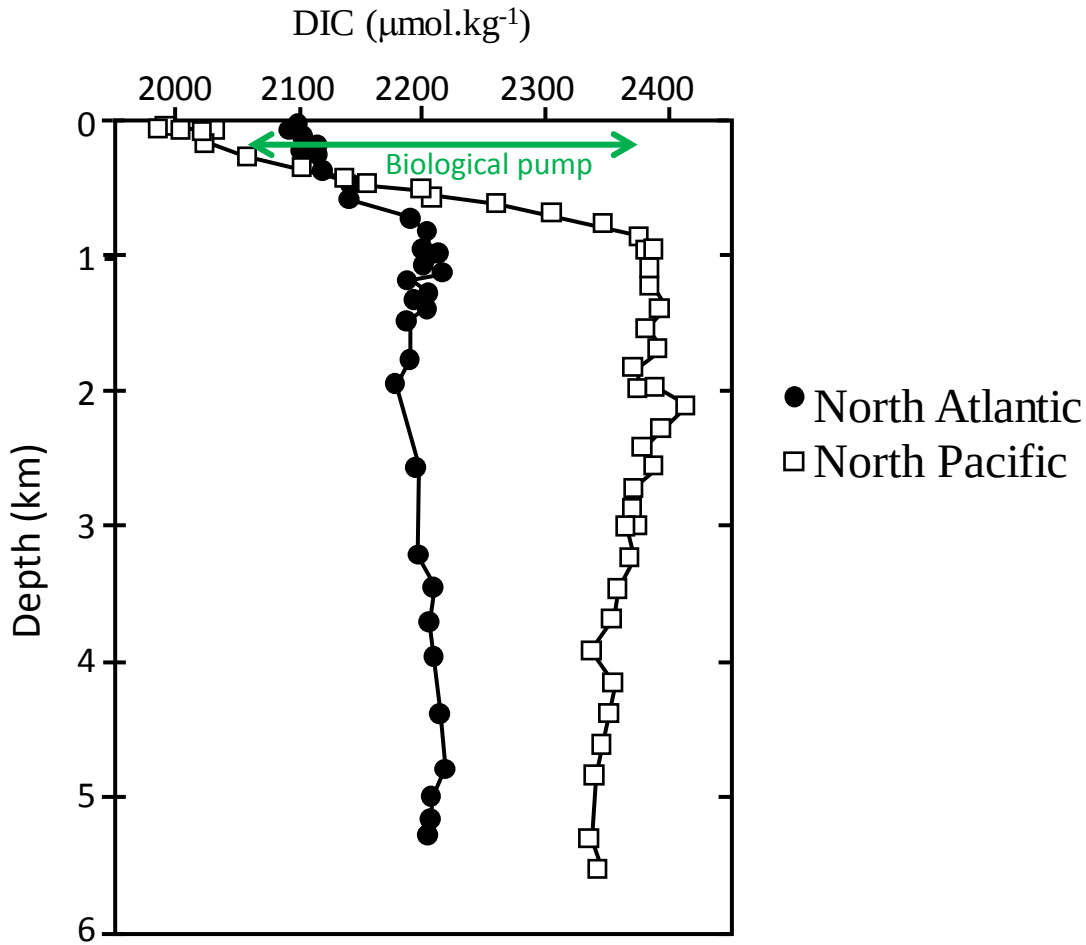


→ Requires constrains on dissolved and particulate carbon fluxes

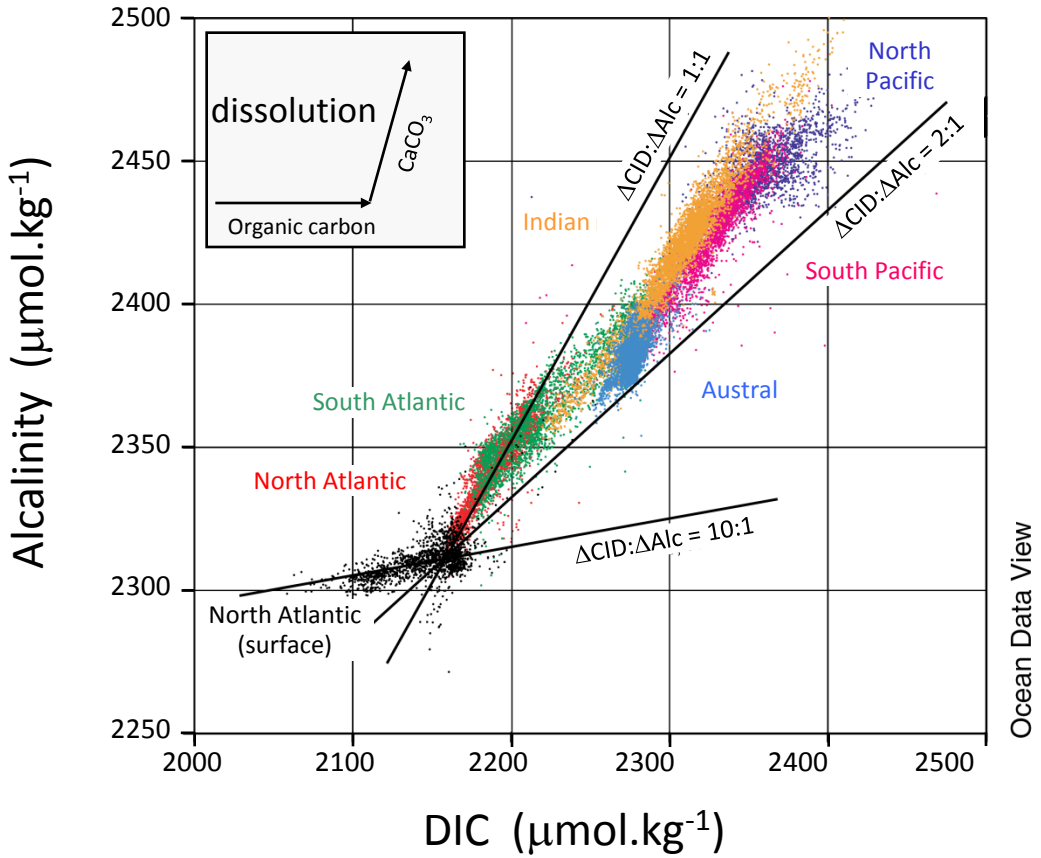
Hydrographic sections



Dissolved inorganic carbon

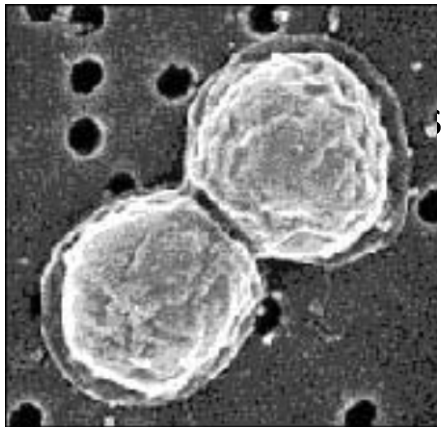


→ Requires constrains on dissolved and particulate carbon fluxes

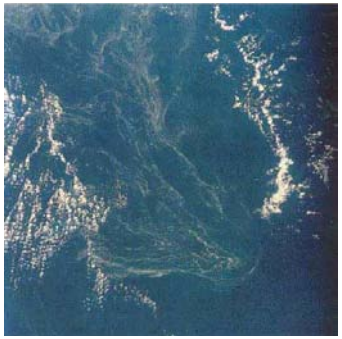

dissolution
$CaCO_3$
Organic carbon
ΔCID:ΔAlc = 1:1
ΔCID:ΔAlc = 2:1
ΔCID:ΔAlc = 10:1
North Pacific
Indian
South Pacific
Austral
South Atlantic
North Atlantic
North Atlantic (surface)
Alcalinity (μmol.kg^{-1})
DIC (μmol.kg^{-1})
2500
2450
2400
2350
2300
2250
2000
2100
2200
2300
2400
2500
Ocean Data View

Photosynthetic bacteria

- Prochlorococcus (most abundant living being on earth)

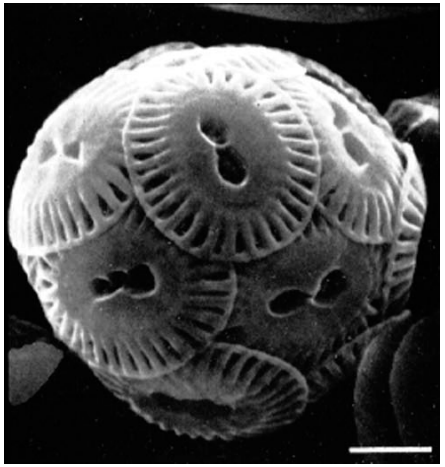


- Trichodesmium (N2 fixing bacteria).

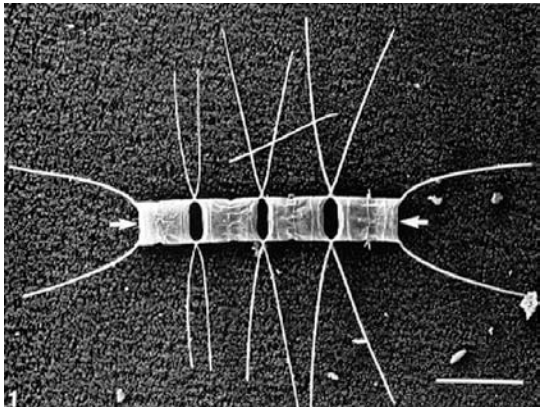


Photosynthetic eucariots

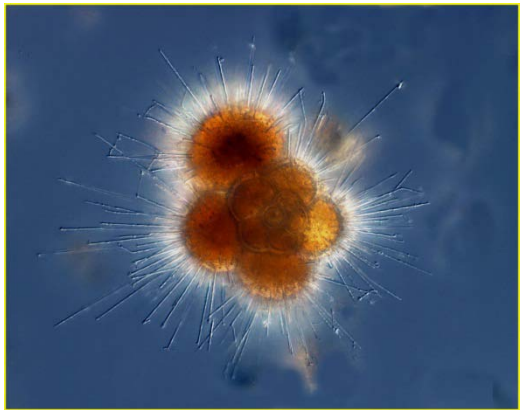
Coccolitophorids (calcium carbonate test)



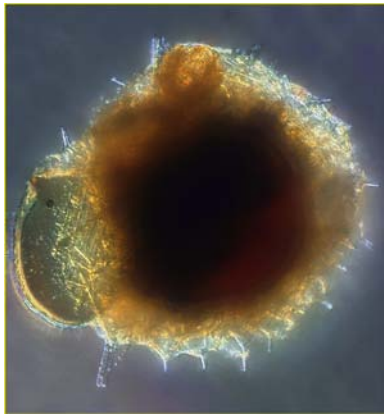
Diatoms (siliieous test)



Consummers

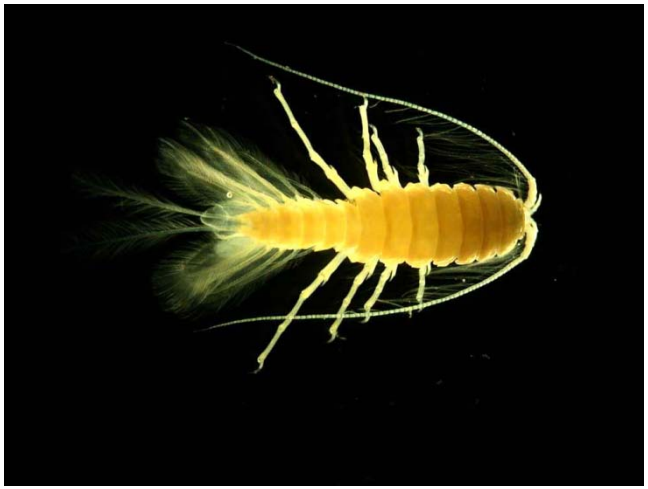


Foraminifera



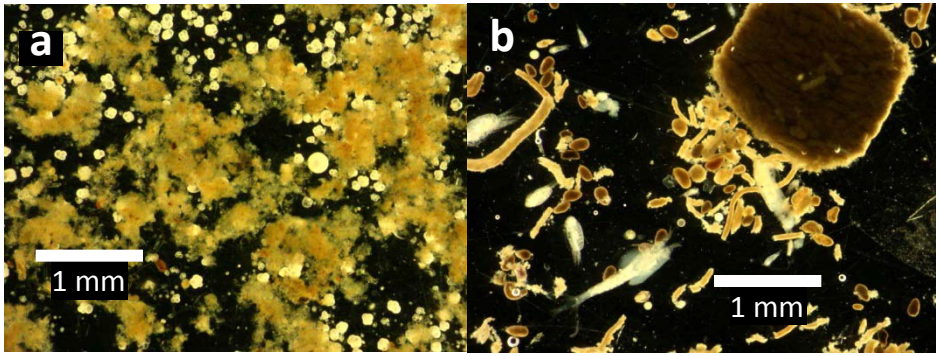
radiolarian eating a tintimid

Seaweed

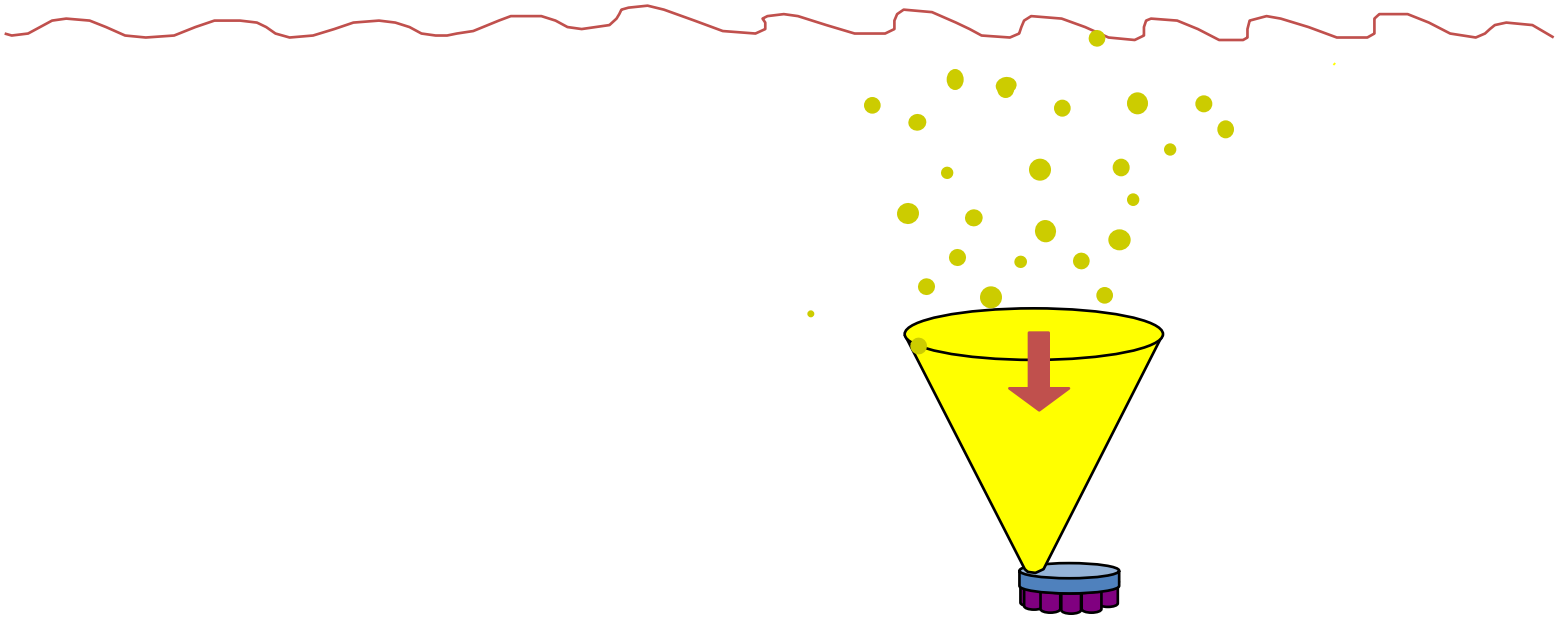


copepod

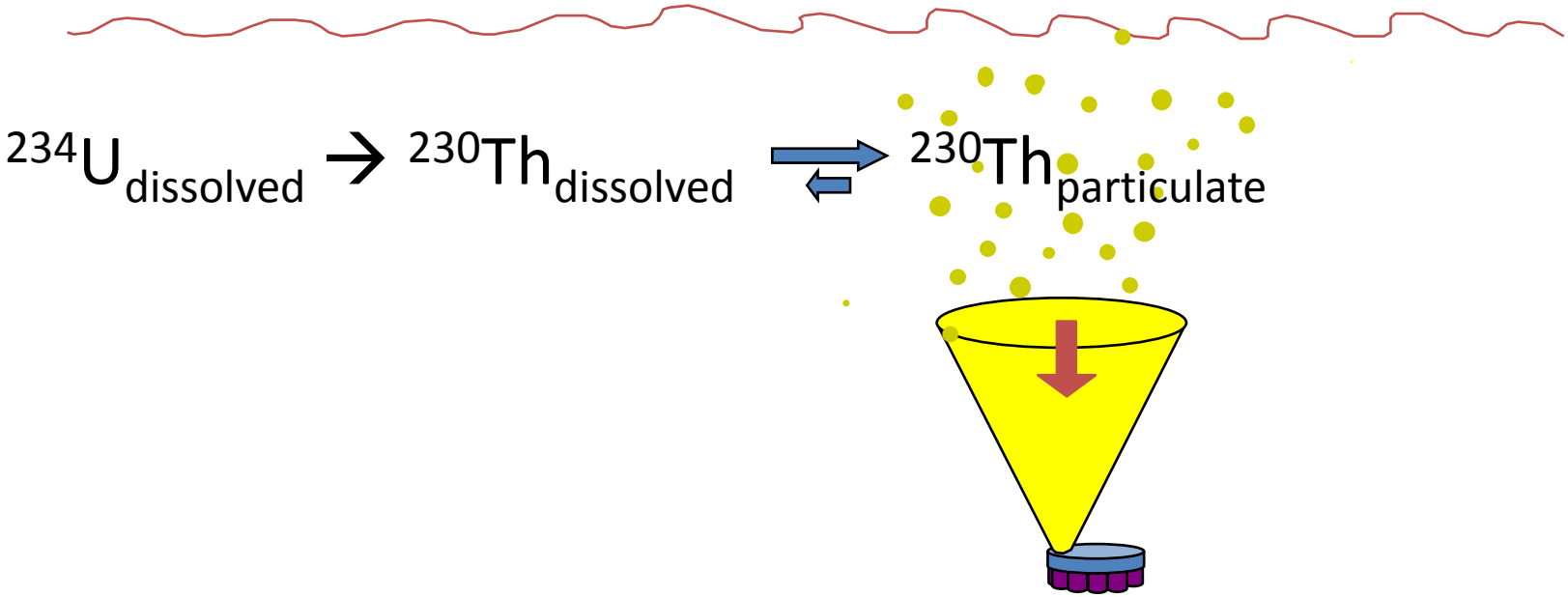
Measuring the biological pump



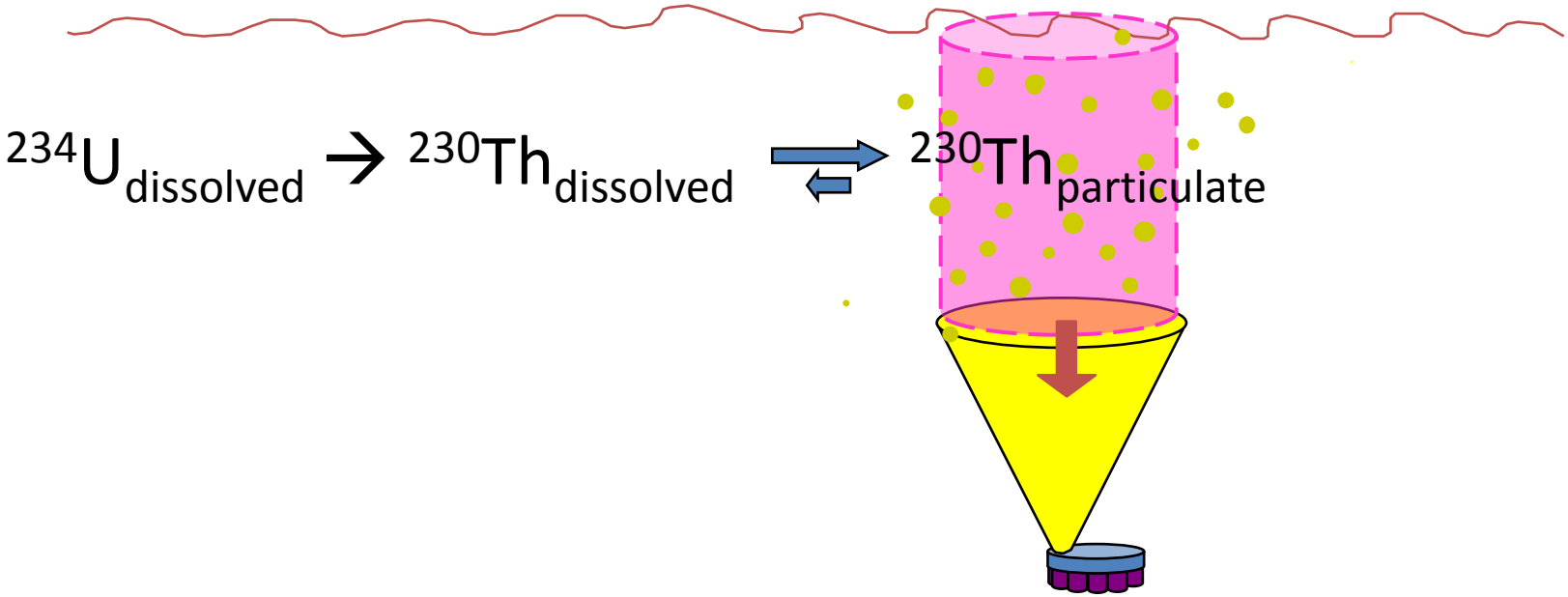
Estimation of particulate fluxes



Estimation of particulate fluxes

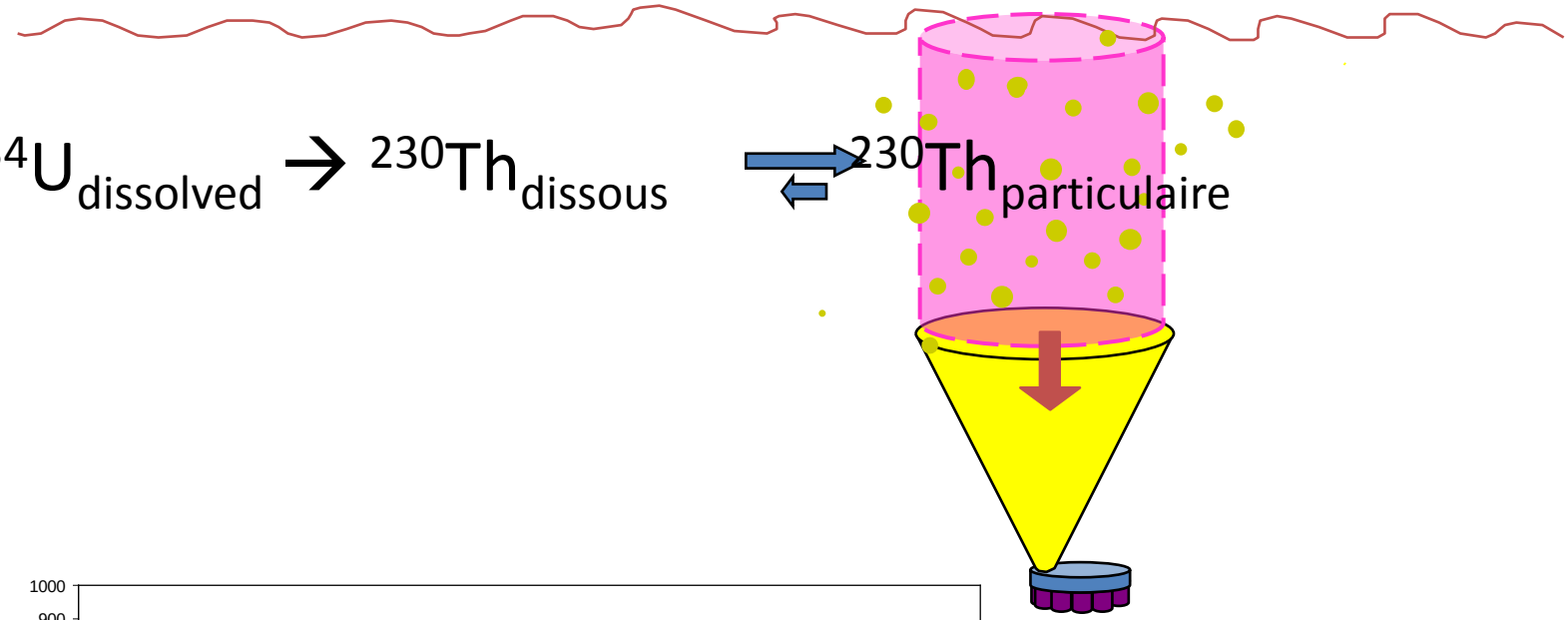


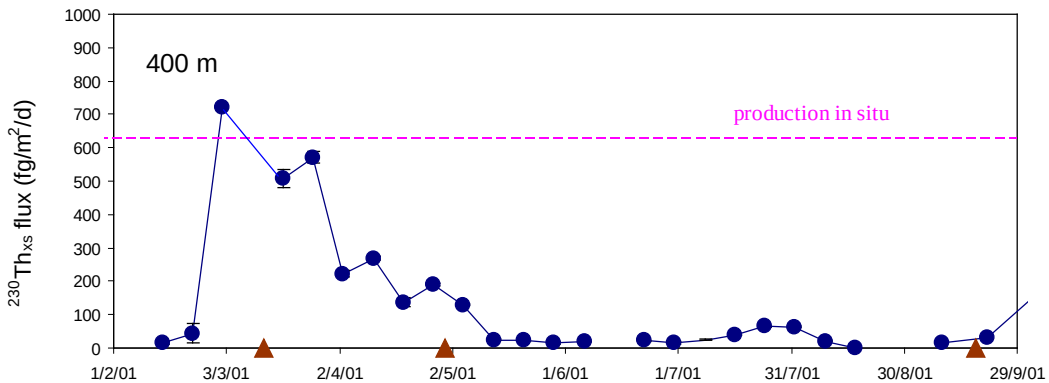
Estimation of particulate fluxes



Estimation of particulate fluxes

$^{234}U_{dissolved} \rightarrow {}^{230}Th_{dissous} \rightleftarrows {}^{230}Th_{particulaire}$





N.E. Atlantic ocean (Guieu et al., 2005)

Focussing/winnowing factor

$$\Psi_{i,j} = \mathrm{MAR}_{i,j} \frac{{}^{0}[{}^{xs}\mathrm{Th}]_{i,j}}{P_{\mathrm{Th}}(t_j - t_i)}$$

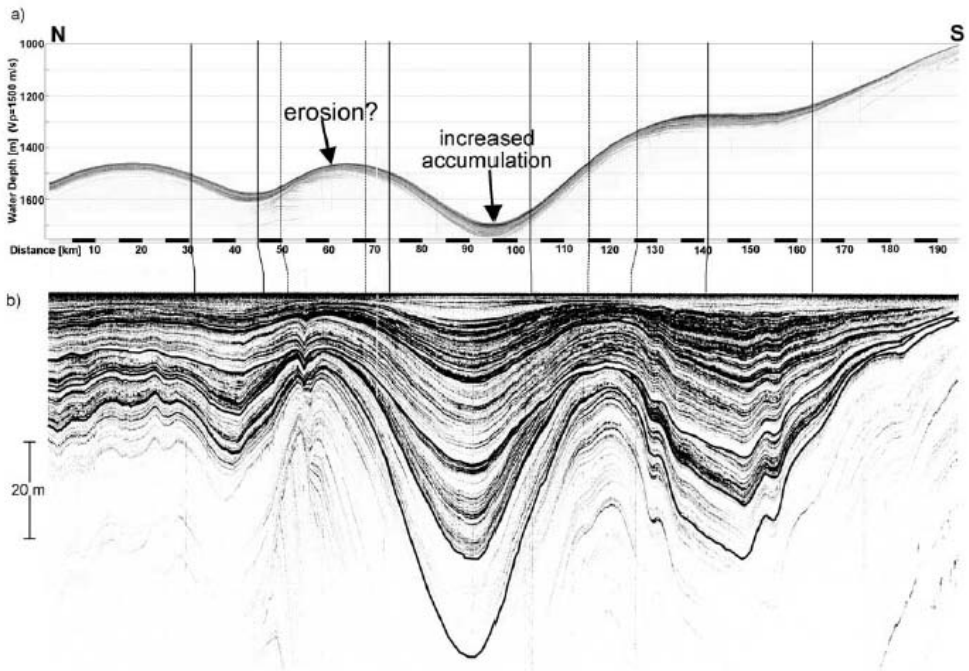
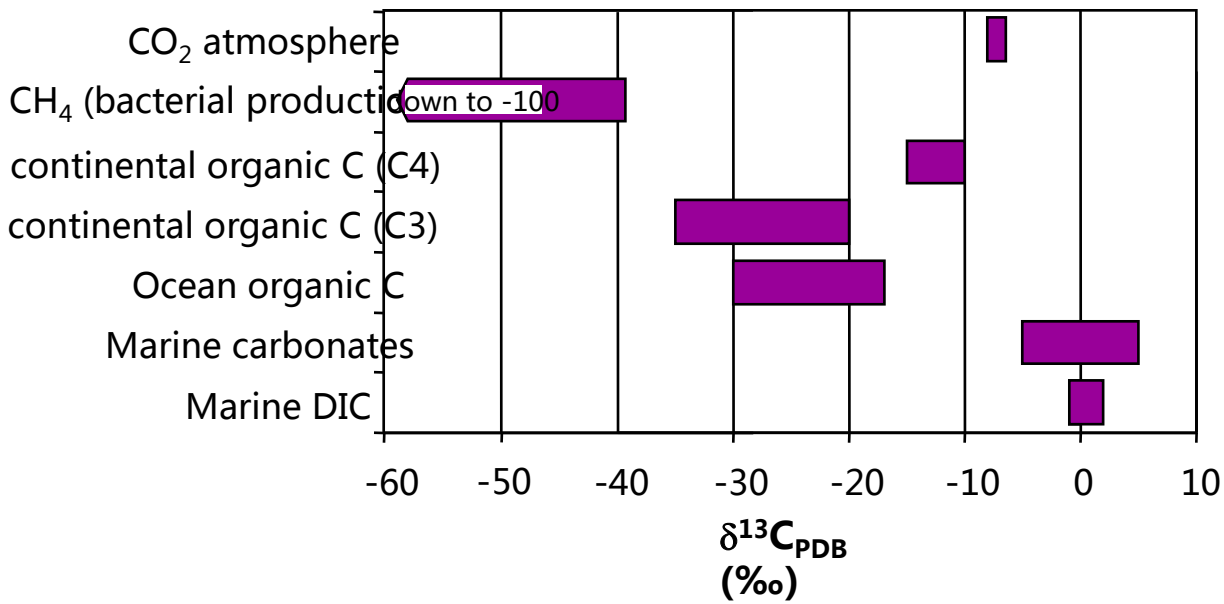


Figure 1. Echo sounder profile on a section parallel to the coast of Namibia showing evidence for sediment winnowing and pounding in mesoscale troughs [*Mollenhauer et al.*, 2002].

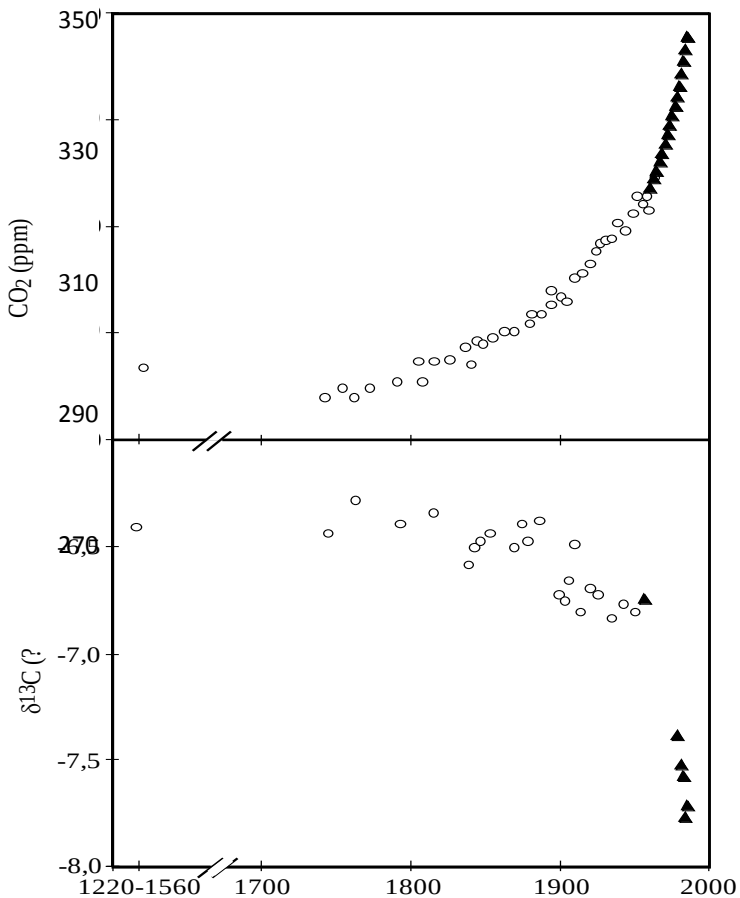
Normalisation to ^{230}Th allows quantifying sediment redistribution

carbon isotope fractionation

$$\delta^{13}C_{\text{sample/reference}} = \left\{ \frac{\left({}^{13}C/{}^{12}C\right)_{\text{sample}}}{\left({}^{13}C/{}^{12}C\right)_{\text{reference}}} - 1 \right\} \times 1000$$

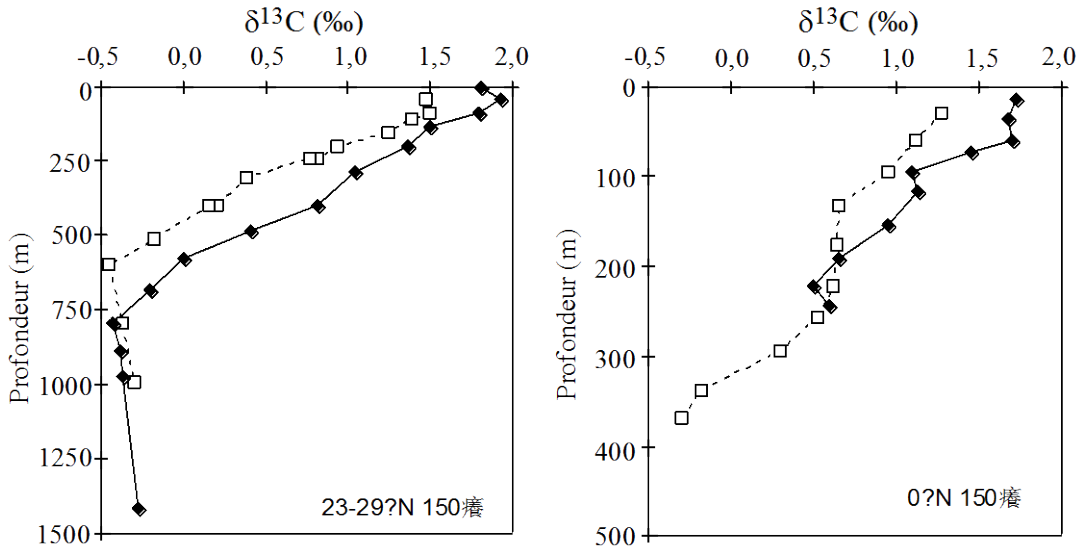


carbon isotope anthropic signature (Suess effect)



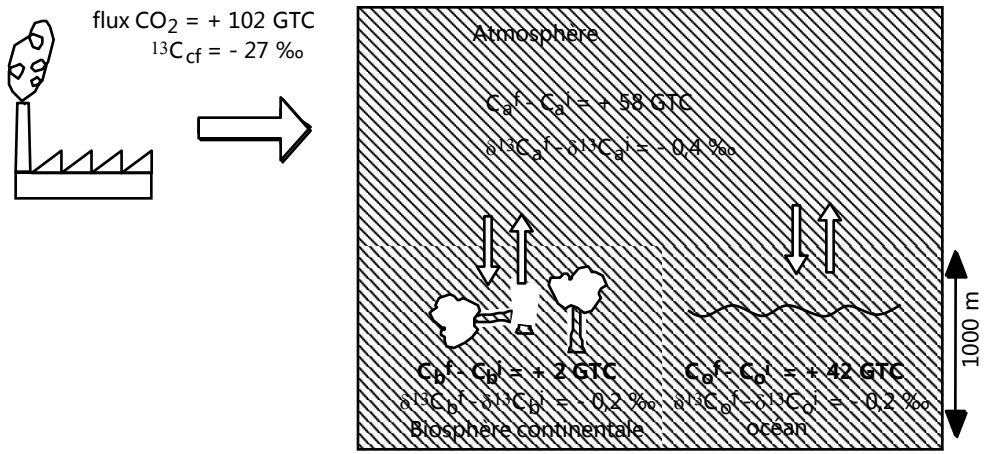
anthropic signature invading the ocean in 20 years

$\delta^{13}C$ measured 20 years appart



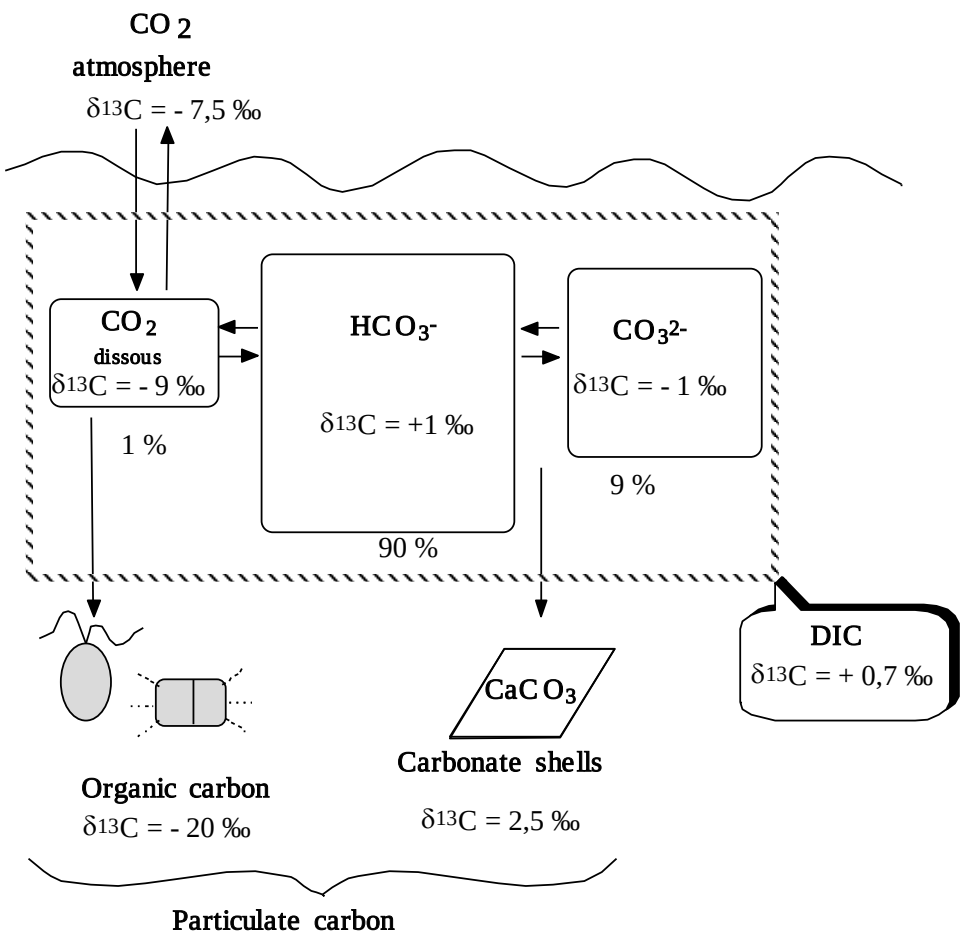
The first 500 m of is in contact with the atmosphere

Global 13C budger

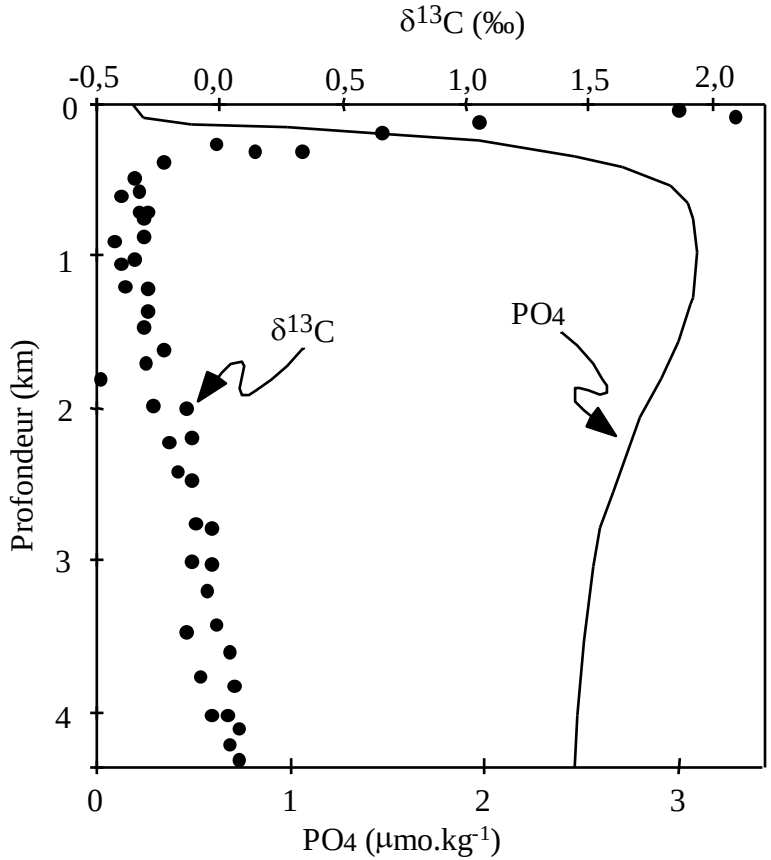


Ocean uptake > continental biosphere uptake

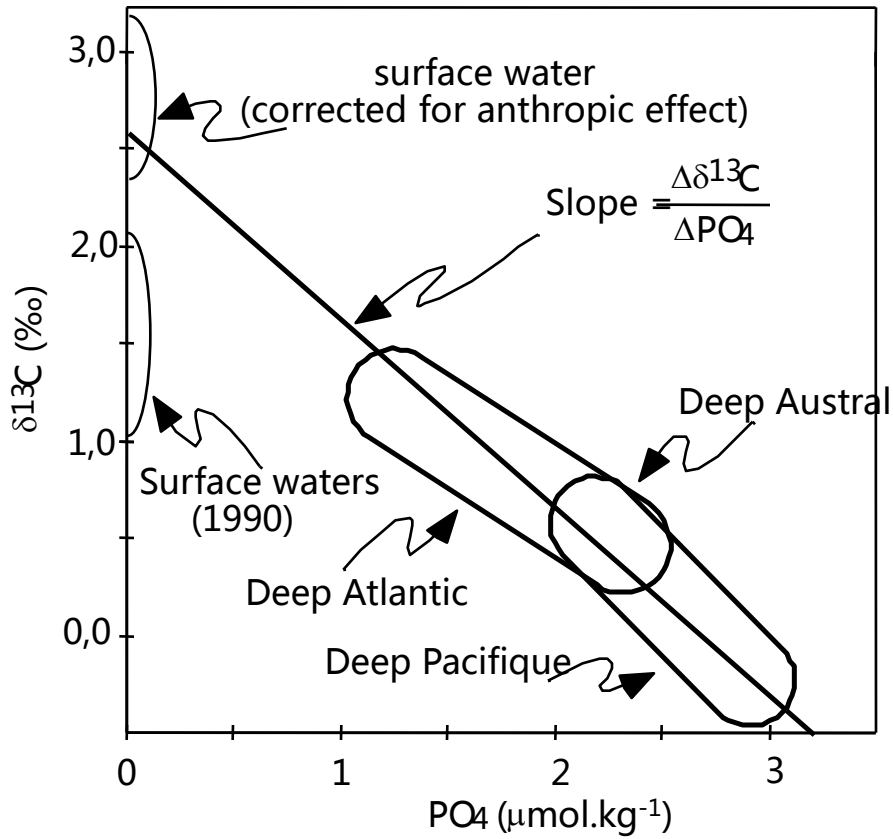
Carbon isotope fractionnation in the ocean



Carbon isotope fractionation in the ocean

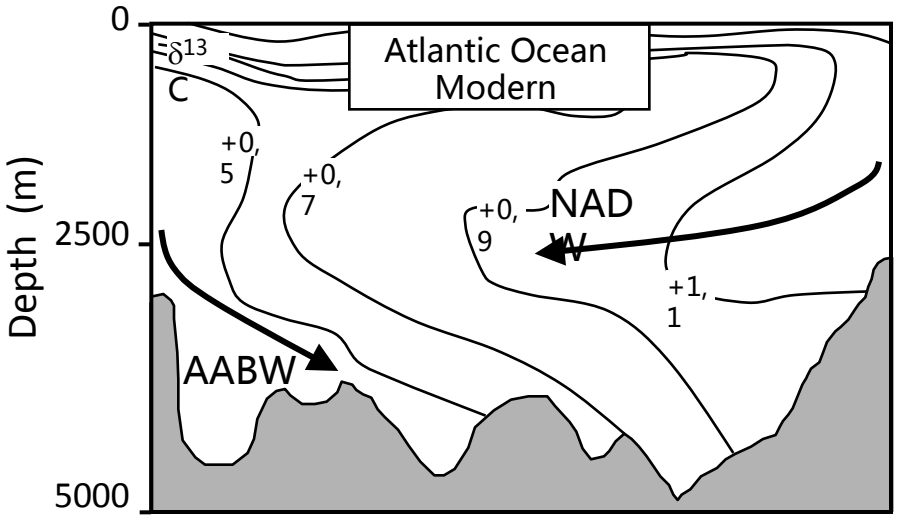


$\delta^{13}C$ for tracing watermasses

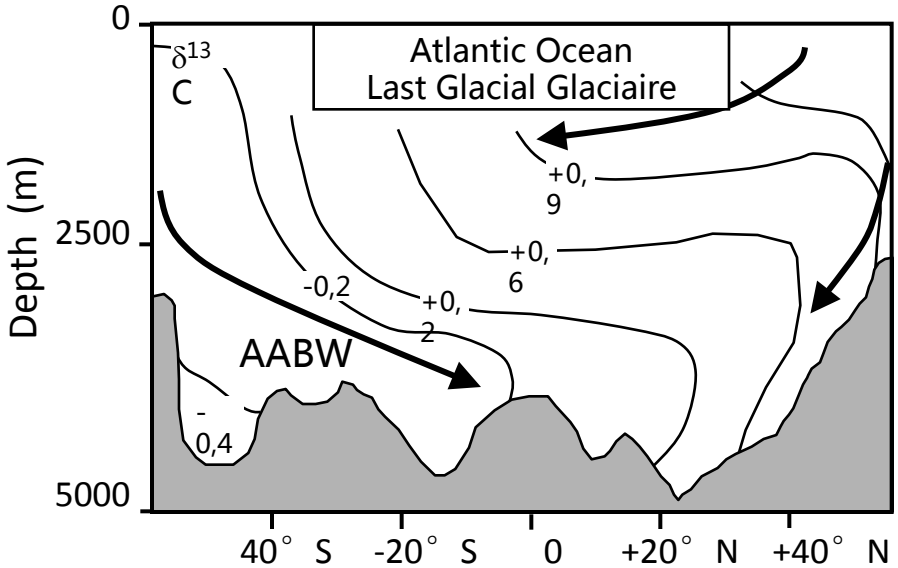


$\delta^{13}C$ in foraminifera

today

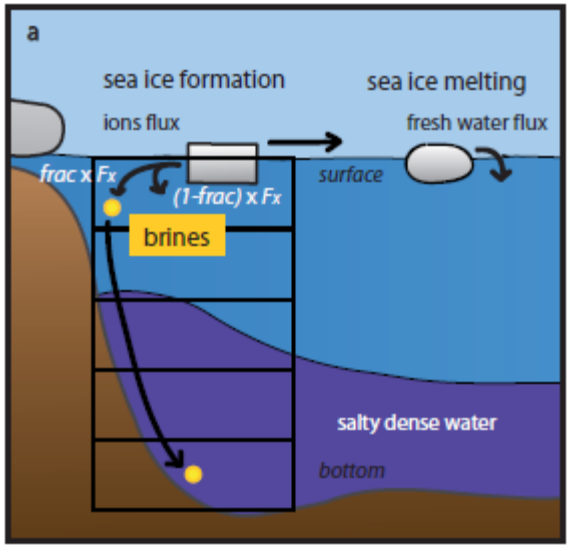


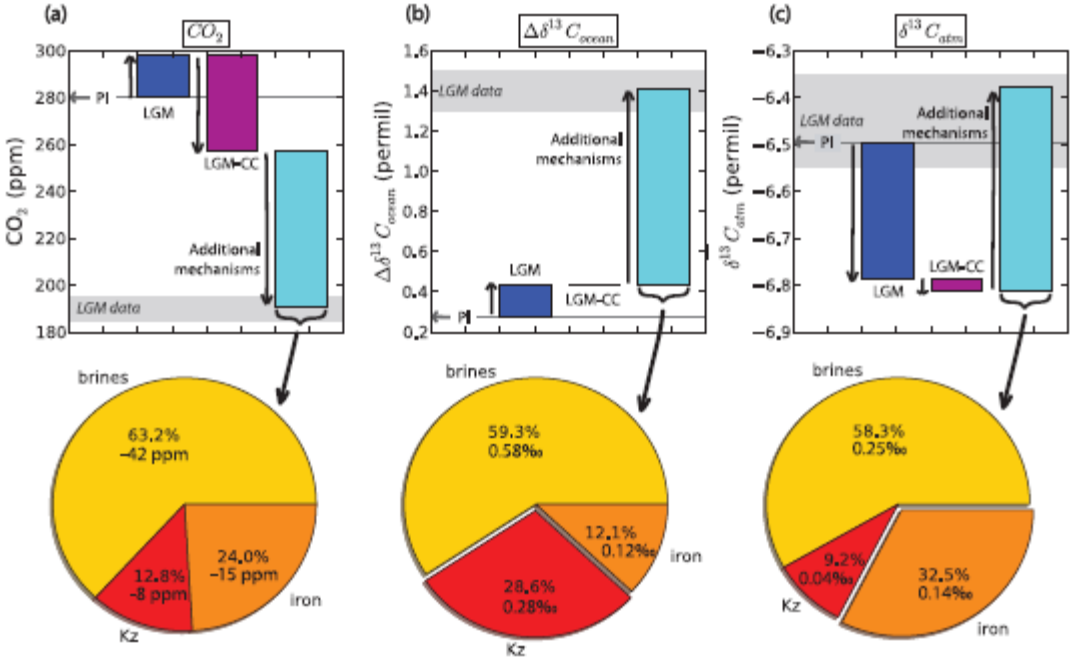
LGM



During LGM storage of carbon on the deep ocean.

Balancing LGM carbon budget





storage of carbon in the deep ocean explained by brine formation + iron fertilisation

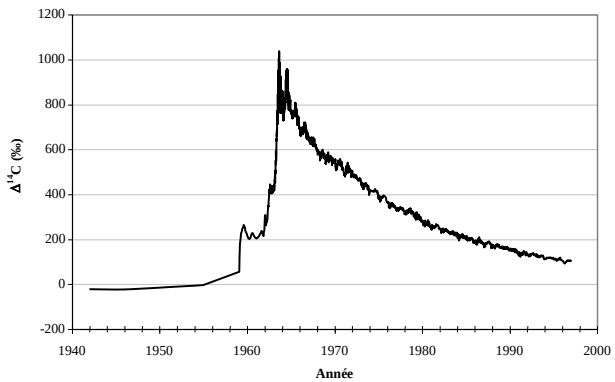
Bouttes et al. 2011

^{14}C production and fates

Natural production in the atmosphere

$$^{14}_{7}N + n \rightarrow {}^{14}_{6}C + p$$

anthropic production in the atmosphere



Radioactive decay in superficial enveloppes

$$^{14}_{6}C \rightarrow {}^{14}_{7}N + e^{-} + \bar{\nu}$$

^{14}C half life = 5700 y → ~12‰ lost every century

Clock for the century-milenium timescale

^{14}C notations

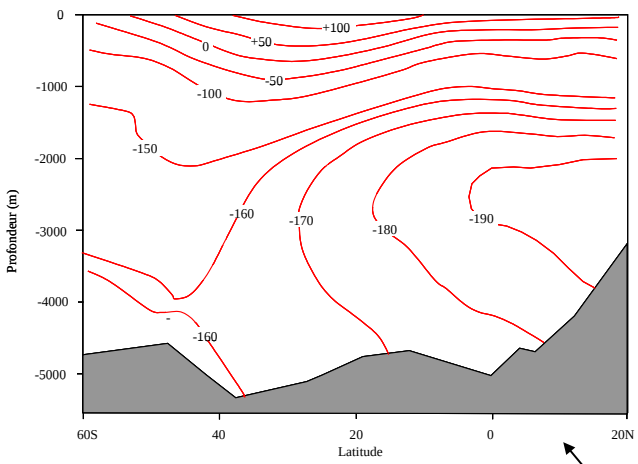
$$\delta^{14}C_{sample} = \left\{ \frac{\left(^{14}C/^{12}C\right)_{sample}}{\left(^{14}C/^{12}C\right)_{reference}} - 1 \right\} \times 1000$$

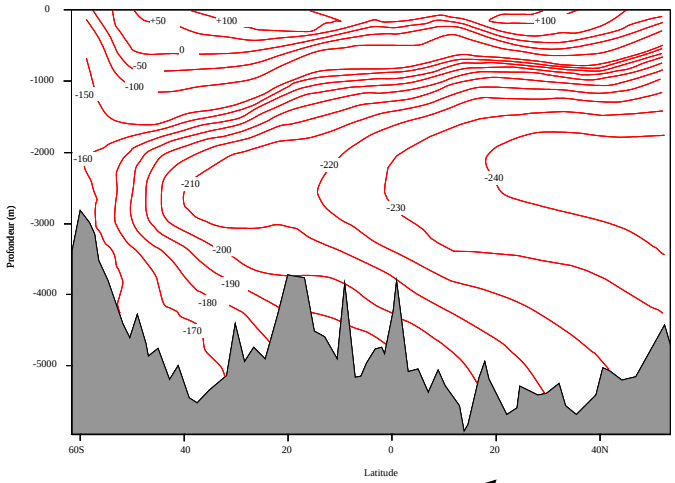
avec $(^{14}C/^{12}C)_{référence} = 1{,}2 \times 10^{-12}$

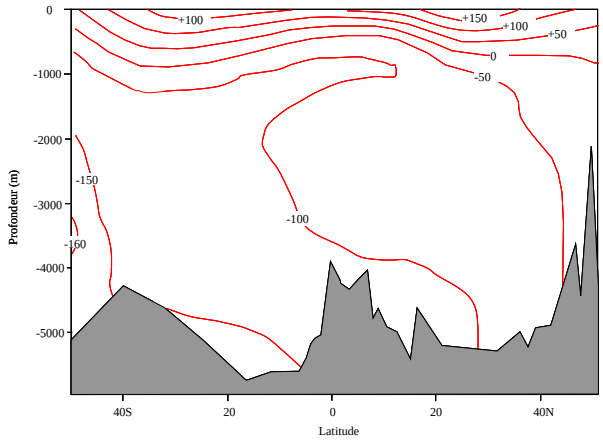
$$\Delta^{14}C_{sample} = \left(\left(1 + \frac{\delta^{14}C_{sample}}{1000} \right) \times \left(1 + 2 \times \frac{25 - \delta^{13}C_{sample}}{1000} \right) - 1 \right) \times 1000$$

Correction of isotopic fractionnation with $\delta^{13}C$

$\Delta^{14}C$ sections of the ocean







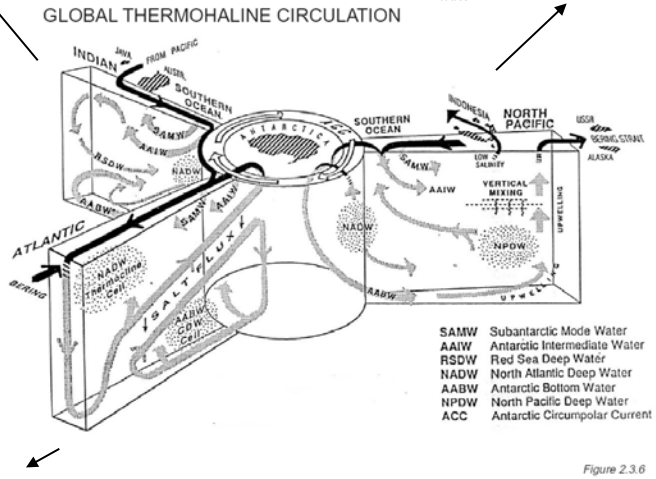


Figure 2.3.6

- → Thermocline « invaded » by anthropic 14C
- → Aging of the deep water masses

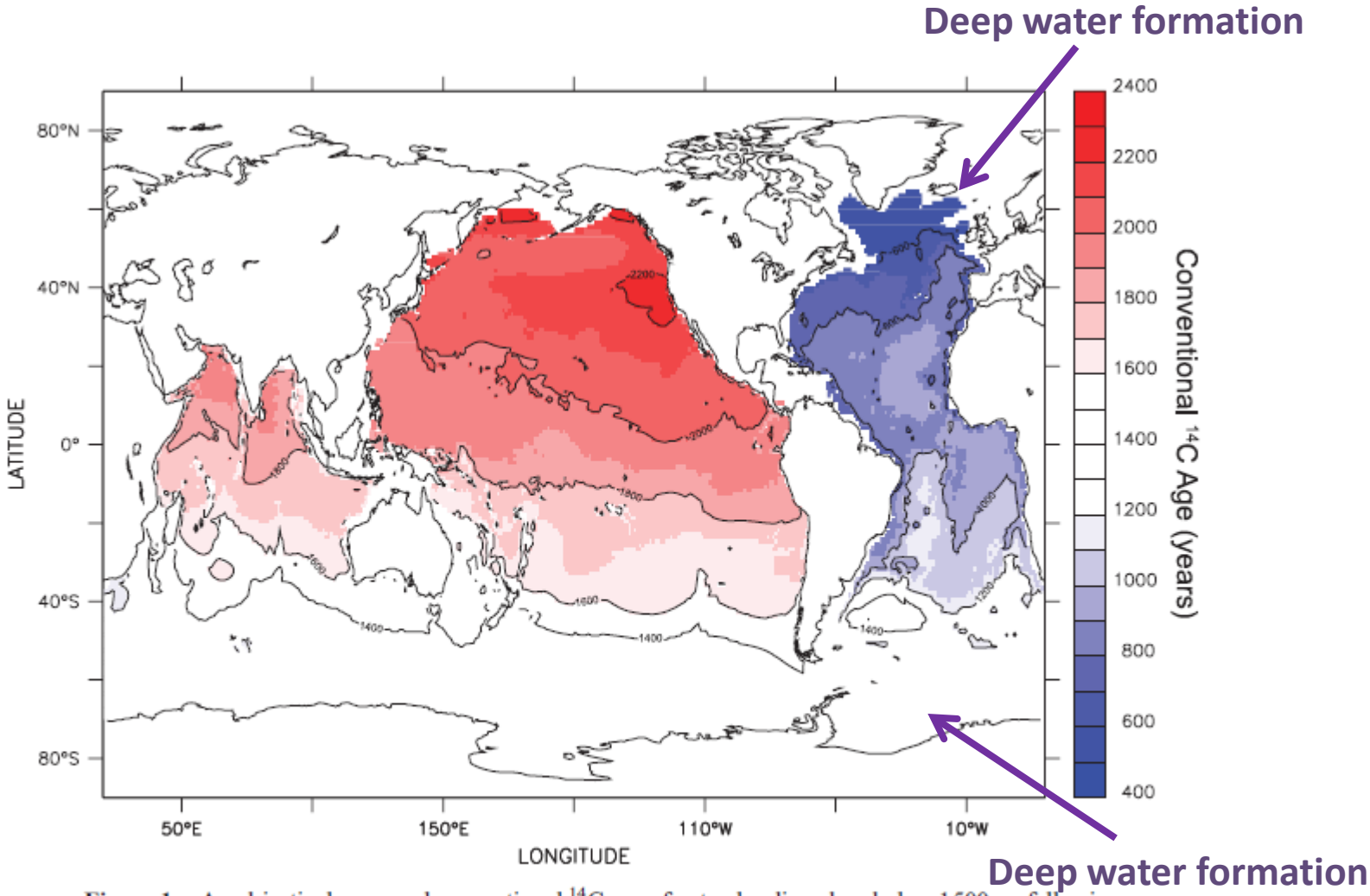
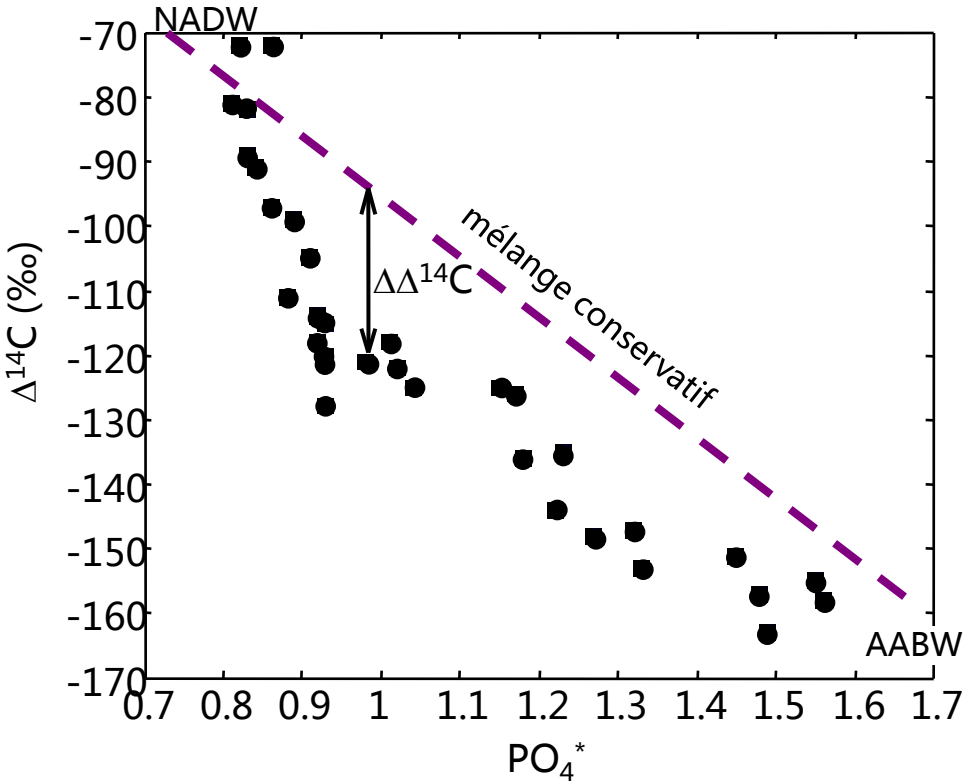


Figure 1. An objectively mapped conventional ^{14}C age of natural radiocarbon below 1500 m, following *Matsumoto and Key* [2004]. Unit is years.

Matsumoto, 2007



After Broecker, 1997

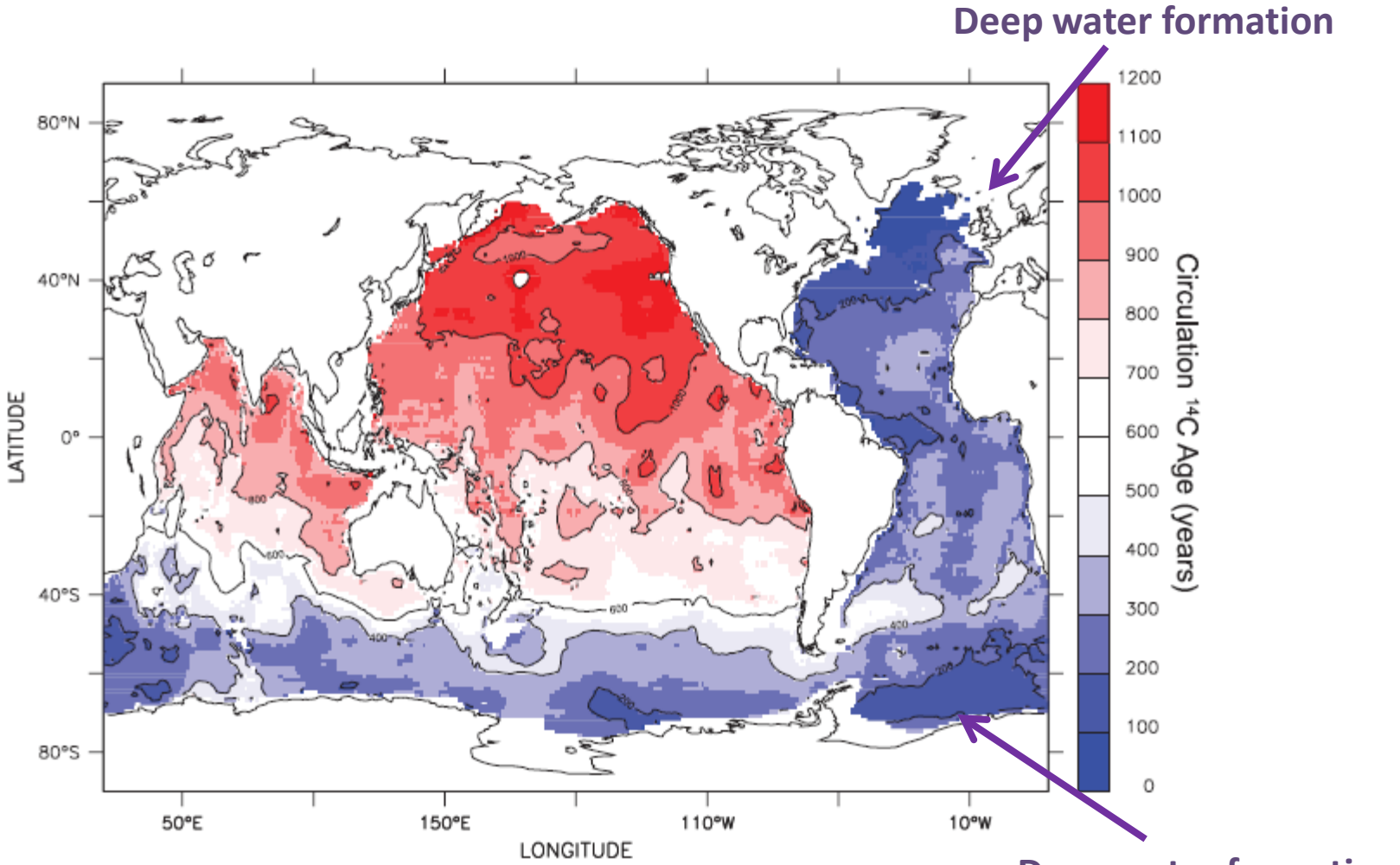
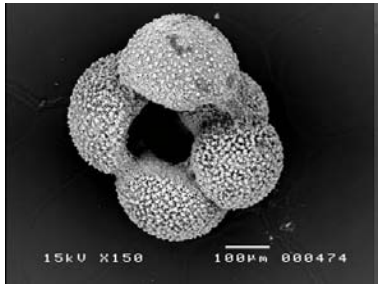


Figure 4. A map of circulation ^{14}C age below 1500 m. This is equivalent to conventional ^{14}C age (Figure 1) but accounts for surface ocean ^{14}C reservoir age and the different sources of deep water. Unit is years.

Matsumoto, 2007

Comparing $\Delta^{14}C$ in surface and deep foraminifera



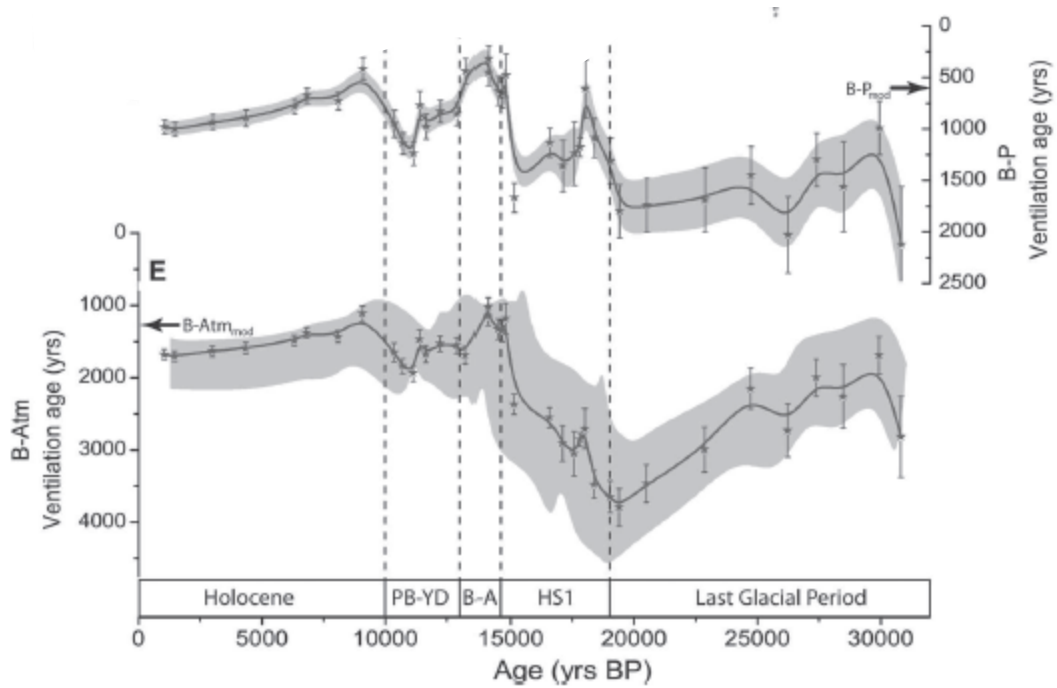


Fig. 2. Deep-water ventilation changes in the Atlantic sector of the Southern Ocean. **(A)** Atmospheric CO_2

→ Younger « ventilation » age in the southern Atlantic today and at LGM

Neodymium isotope rationale

Neodymium is an insoluble element
(residence time = 500 y)

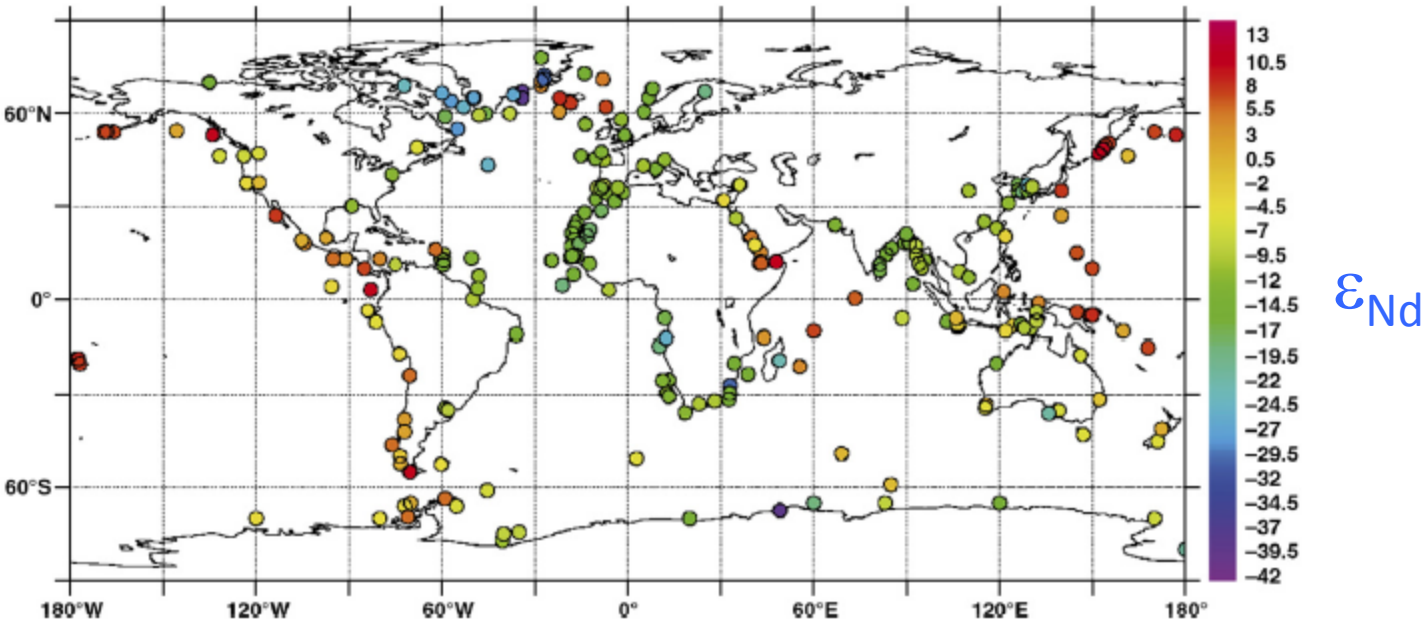
$^{147}Sm \rightarrow {}^{143}Nd + \alpha$

^{147}Sm half life = 106 Gy

→ very small changes of $^{143}Nd/^{144}Nd$

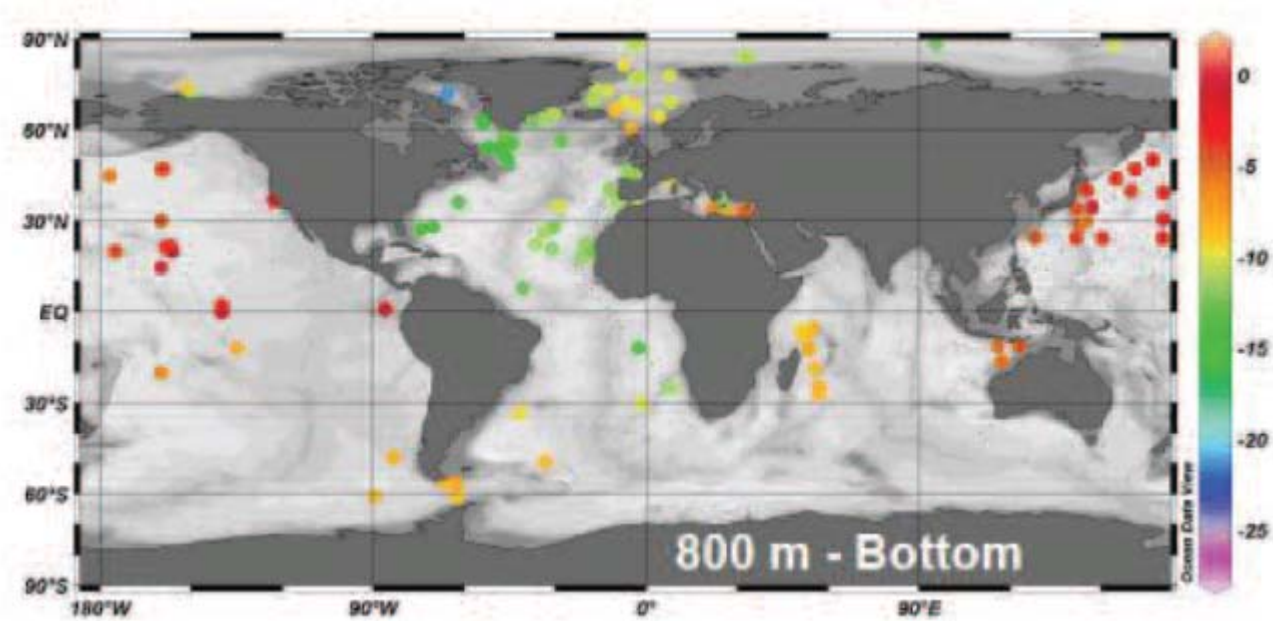
$$\varepsilon_{Nd} = \left(\frac{\left(\frac{^{143}Nd}{^{144}Nd} \right)_{sample}}{\left(\frac{^{143}Nd}{^{144}Nd} \right)_{CHUR}} - 1 \right) \times 10^4$$

Rocks around the ocean



Jeandel et al., 2007

Water mass signatures



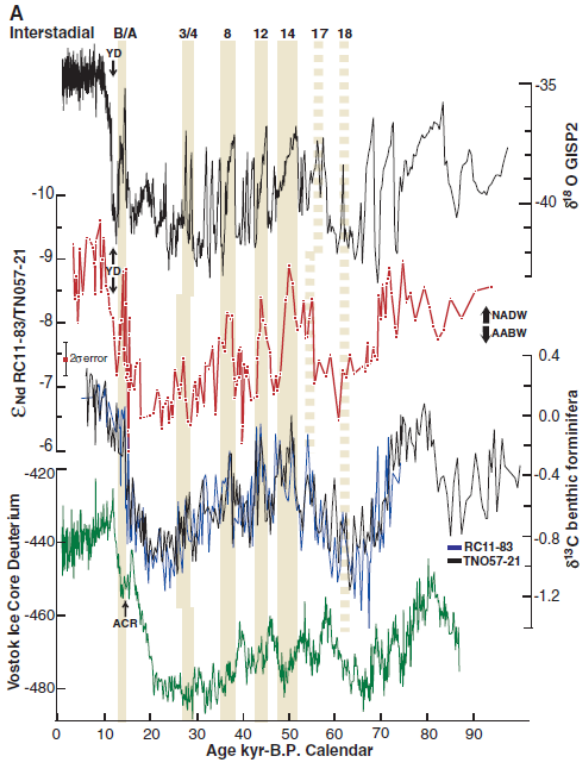
ε_{Nd}

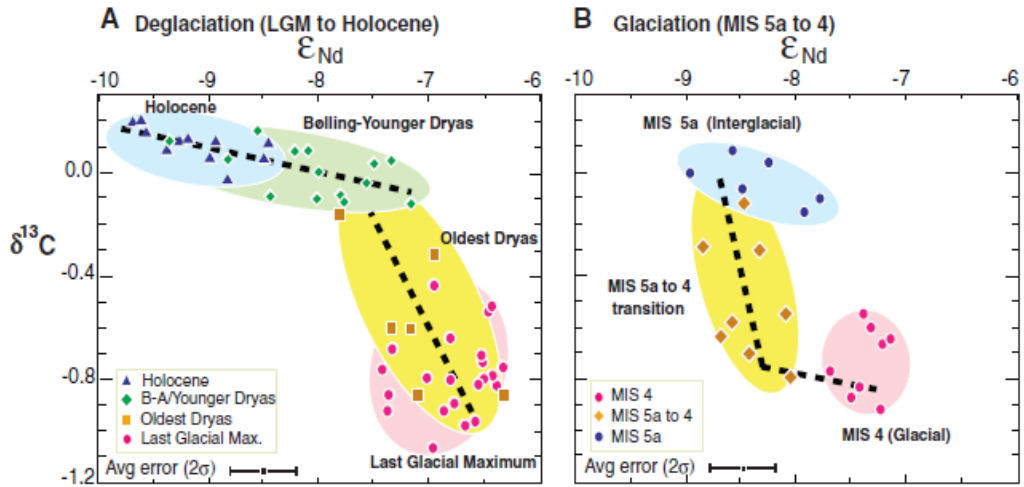
Jeandel et al., 2012

Neodymium isotopes -13C comparison

NADW

AABW

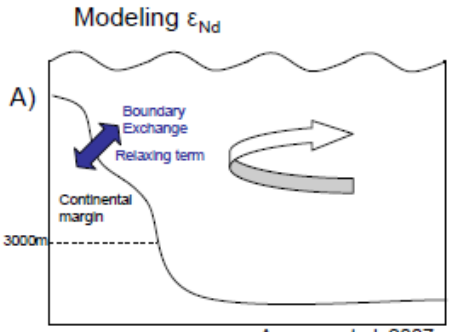




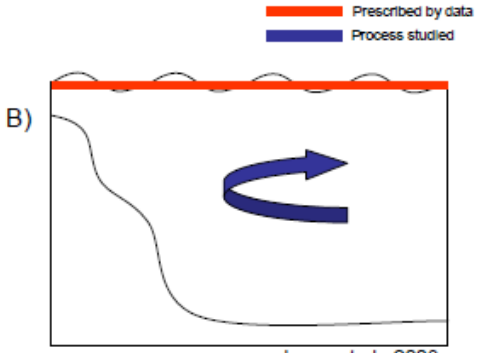
Piotrovski et al., 2005

Nd modeling: the role of particles

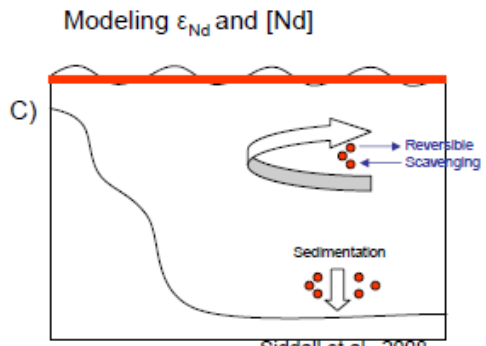
No particles



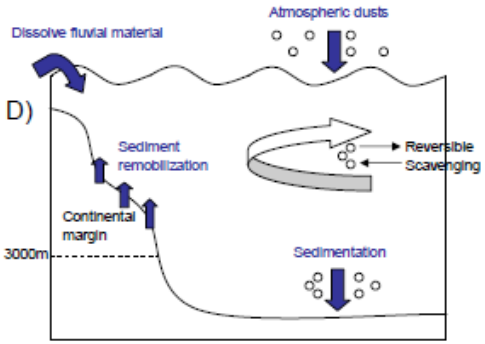
Arsouze et al. 2007



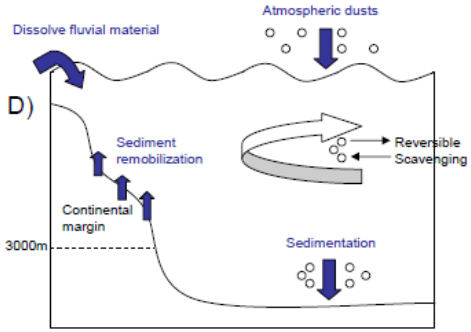
Jones et al., 2008



Siddall et al., 2008



Arsouze et al., 2009



Rempfer et al., 2011

Oka et al., 2009

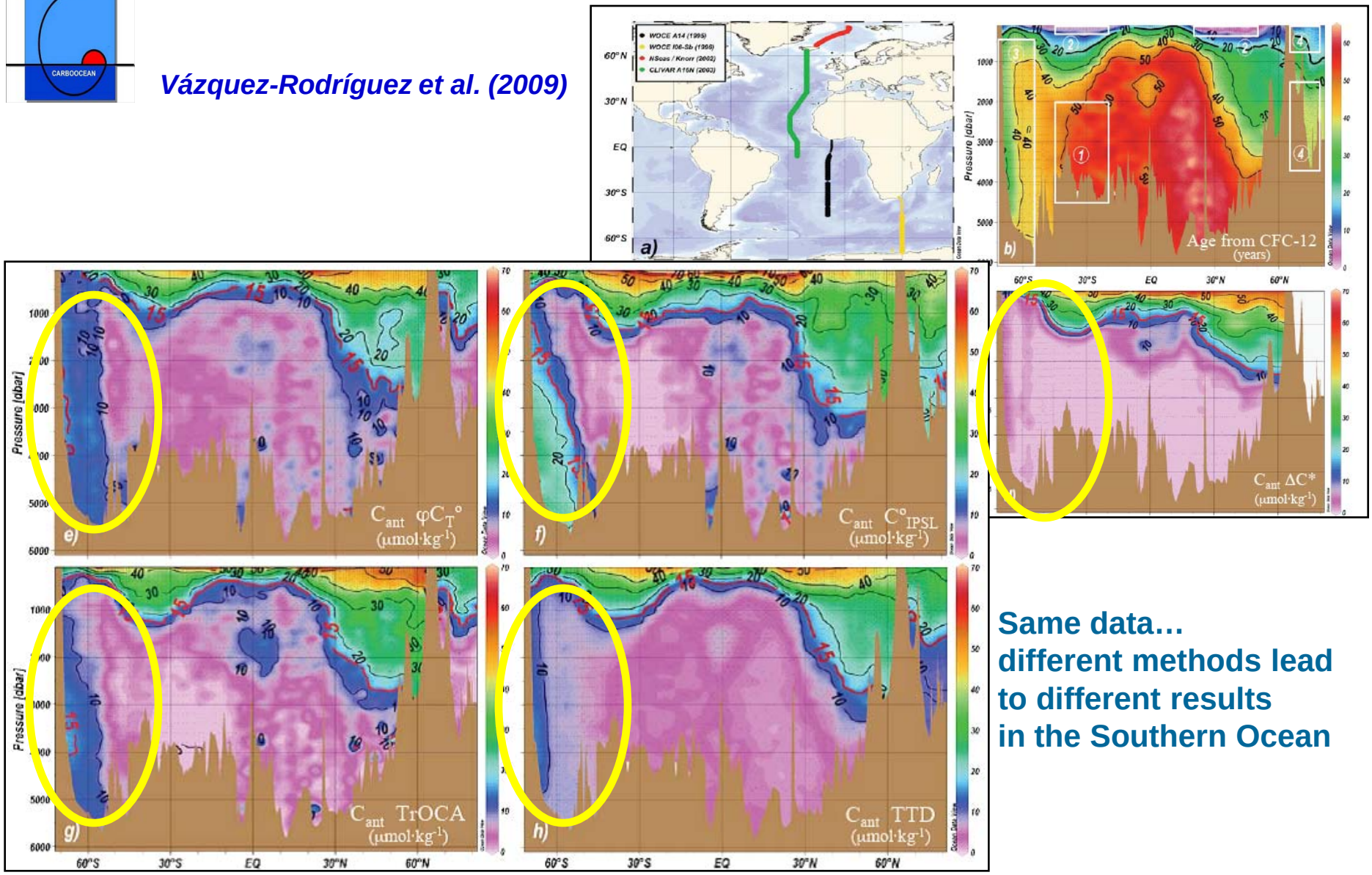
Particles composition	bSi/$CaCO_3$/litho	OM/litho	no
Boundary input	No	yes	yes

5 different models claim to reproduce « very well » the ε_{Nd} in the ocean

updapted from Arsouze et al., 2009

Atlantic: from the Artic to the Southern Ocean

Vázquez-Rodríguez et al. (2009)



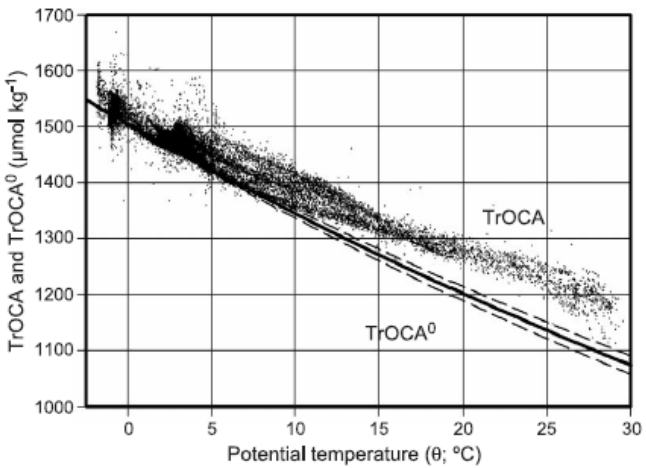
Same data…
different methods lead
to different results
in the Southern Ocean

Troca Cook-book

$$\mathrm{TrOCA} = \mathrm{O_2} + 1.2\,\mathrm{TCO_2} - 0.6\mathrm{TA}$$

$$\mathrm{TrOCA}^0 = \mathrm{O_2^0} + 1.2\,\mathrm{TCO_2^0} - 0.6\mathrm{TA}^0$$

$$\mathrm{C_{ANT}} = \mathrm{TCO_2} - \mathrm{TCO_2^0} = \frac{\mathrm{TrOCA} - \mathrm{TrOCA}^0}{1.2}$$



Touratier et al., 2004

Take home message

Isotopic tracer are important at all scales:

* source tracer
* clock

trace elements are important:

* specific informations