

Some main processes of organic matter mineralization and nutrient fluxes at the sediment-water interface in the Great Astrolabe lagoon (Fiji)

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Abstract

The fluxes of dissolved inorganic N, P and Si from the sediments were calculated using pore water gradient concentration measured using the peeper technique at 2 stations in the Great Astrolabe Reef lagoon (Fiji). The nutrient concentrations of pore water reached maximum values of $130 \mu\text{M NH}_4$, $8 \mu\text{M PO}_4$ and $90 \mu\text{M SiO}_2$. Fluxes calculated from concentration gradients were from the sediment to the water column. NH_4 , PO_4 and SiO_2 fluxes were respectively in the range $18\text{-}64$, $0\text{-}2$, $2\text{-}50 \mu\text{mol m}^{-2} \text{d}^{-1}$.

1. Introduction

The Great Astrolabe Reef (Fiji) belongs to the remote places in the world which can be considered as free of anthropic impacts. Then, it can represent a « reference » level for environmental studies.

Since the small population uses natural resources for survival where fishing is the main local food supply, it is necessary to estimate the ability of the lagoon to recycle its natural potentiality which allows the people to live here, at the back of beyond.

One of the keys to solve this problem is the estimation of the primary production which is sustained by the nutrient pool of C, N, P and Si.

The inputs of Si and P can be accounted on the weathering of basaltic islands which emerge in many places inside of the lagoon : 13 islands which represent a total surface of 34 km^2 while the lagoon surface is 210 km^2 (16 %). The input of dry deposition is the other main source for this very large lagoon surrounded by oligotrophic waters of the Pacific Ocean.

From these rough observations it can be supposed that the main limiting factor for primary production is Nitrogen. This is probably the reason why the phytoplankton is submitted to the dominance of nitrogen-fixing species (cyanobacteria).

An important compartment involved into the recycling of the organic matter is the sediment and its associated pore water. The aim of this contribution is an attempt to quantify the fluxes of the nutrients which come from the sediment-water-

interface (SWI). A further step will be the evaluation of the contribution of these fluxes compared to the other lagoonal sources and sinks.

2. Material and methods

The field work has been focused on stations 4 and 13 where the depth is 40 and 36 m respectively Figure 1.

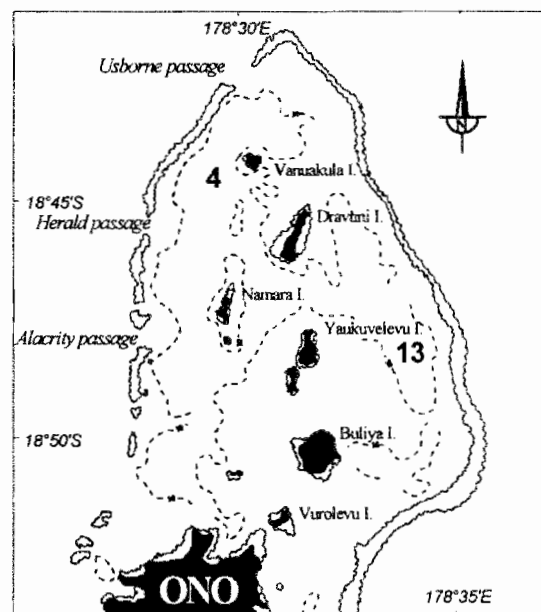


Figure 1: Station location

The pore water sampling has been done with « peepers » (Hesselin 1976) embedded into the sediment during 18 days to insure chemical equilibration.

Conventional analytical methods have been used on the field or in the laboratory to carry out the analysis of nutrients and major elements.

The pH was determined at room temperature ($T = 22 \text{ }^\circ\text{C}$) on submicro-samples ($500 \mu\text{l}$) using an Ingold micro-electrode and the Hansson's calibration method described by Almgren et al. (1975) with a precision of ± 0.01 pH unit. Gran's potentiometric titration was used for alkalinity measurements

(Stumm and Morgan 1981). The accuracy was $\pm 0.5\%$. Standard colorimetric procedures were used for nutrient analysis (SiO_2 , NH_4 and PO_4) and total dissolved sulfide, adapted for submicro samples (Merck Spectroquant methods; sample volume: 0.5 or 1 ml, precision $\pm 4\%$).

3. Results

Concentrations profiles obtained at stations 4 and 13 are shown on Figures 2 to 9. For both stations large concentration gradients occur for all the analyzed species.

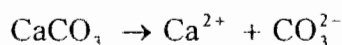
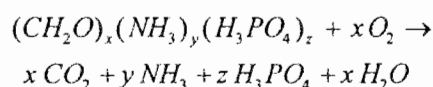
3.1 Concentration profiles

Data are available only for station 13 but the observed profiles are these usually observed in such marine environments (Charpy et al. 1996).

Both pH and alkalinity show a sharp variation at the SWI :

- pH decreases from 8.40 in the water column, just above the interface to 8.00 at 2.5 cm below it.
- Alkalinity, on the reverse, exhibits increasing values from 2.25 to 2.40 mM within the same depth interval.

These opposite variations show that active organic matter mineralization occurs in the sediment. Within the first centimeter below the interface we can reasonably suppose that oxidation of organic matter is carried out by using dissolved oxygen. Since this reaction increases the total DIC ($\Sigma\text{CO}_2 = [\text{H}_2\text{CO}_3] + [\text{HCO}_3^-] + [\text{CO}_3^{2-}]$) without any significant variation of the alkalinity, we can infer that the observed increase of alkalinity is due to the dissolution of the sediment i.e. CaCO_3 (probably under the aragonite form, which is the main component of coral sand). Therefore, within the 0 to 2 cm layer below the SWI the main OM mineralization processes can be represented by the two reactions :



From which we can deduce :

$$\frac{\Delta\Sigma\text{CO}_2}{\Delta\text{O}_2} = 1 \text{ and } \frac{\Delta\text{Alk}}{\Delta\text{O}_2} = \frac{y-z}{x}$$

$$\text{then: } \frac{\Delta\Sigma\text{CO}_2}{\Delta\text{Alk}} = \frac{x}{y-z}$$

As the stoichiometric coefficients are always as $y > z$ and $x \gg (y-z)$ we can infer that $\Delta\Sigma\text{CO}_2 > \Delta\text{Alk}$ and the pH *must* decrease, as it can be observed. However, this pH variation must be buffered by the

dissolution of calcium carbonate which leads to an increase of the alkalinity by the amount of :

$$\Delta\text{Alk} = 2\Delta[\text{Ca}^{2+}]$$

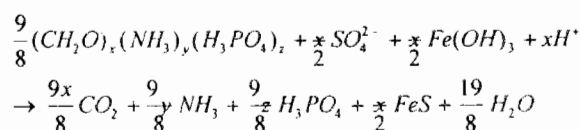
Within the 2cm upper layer the alkalinity increase is within the range 30 to 70 μM . then the variation of the dissolved calcium concentration is only within the range 15 to 35 μM and this small variation is not recorded on the calcium profile.

As pH and alkalinity keep changing below the oxygenated layer of the sediment, this show that the mineralization of the organic matter is still going on but the terminal electron acceptor cannot be the oxygen in account of the anoxic environment. Usually, in marine coastal environments, the sulfate ion SO_4^{2-} replaces oxygen and is reduced into S(-II) by the sulfate-reducing bacterial strains. This leads to a sharp increase in dissolved sulfide concentration into the pore water.

At both stations no dissolved $\Sigma\text{H}_2\text{S}$ (with $\Sigma\text{H}_2\text{S} = [\text{H}_2\text{S}] + [\text{HS}^-]$) occurs into the pore water. A possible explanation can be found in the fact that lots of detrital iron-rich particles are bring into the sediment by the basalt weathering which is a quite different situation as it is observed in other atolls where the iron is only a trace element.

If Fe(III) and SO_4^{2-} are both oxidants with respect to organic matter, then soluble Fe(II) and soluble sulfide S(-II) are released into the pore water medium. Subsequently, Fe(II) reacts very quickly on sulfide and leads to FeS precipitation. This reaction pathway is able to explain why neither Fe^{2+} nor $\Sigma\text{H}_2\text{S}$ can be detected into the pore water.

The sum of the three reactions can be written as follow :



and the ratio

$$\frac{\Delta\text{Alk}}{\Delta\text{SO}_4} = \frac{\frac{9}{8}(y-z) + x}{\frac{x}{2}} \quad (1)$$

$$\text{and : } \frac{\Delta\Sigma\text{CO}_2}{\Delta\text{SO}_4} = \frac{9x}{x} \approx 2.4 \quad (2)$$

since the ratio (1) is always close to 2 in account of the usual values accepted for x , y and z (with $y-z \ll x$) we can deduce that the ratio $\frac{\Delta\Sigma\text{CO}_2}{\Delta\text{Alk}}$ is always

greater than 1. Hence the total DIC increases faster than the alkalinity and the pH keeps decreasing. This trend is well observed for the peeper C at station 13 (Figure 4). Unfortunately, NH_4 and SiO_2 have not been measured on peeper C. Only the PO_4 increase can yield about the mineralization processes.

Nutrients profiles observed at station 4 (40 m depth) show very sharp concentration gradients at the SWI in account of the same processes observed at station 13. A striking fact is an increase of the concentrations up to 7 to 10 cm below the SWI for NH_4 and SiO_2 while, for PO_4 , the concentration increases between the SWI and 15 cm below it. The

same pattern is observed for this nutrient at station 13 (Figure 7).

These observations suggest that a sediment layer located between 7 and 10 cm below the SWI acts as a nutrients source allowing diffusion to process either upward or downward this layer. Since the N:P molar ratio is close to 30 within this layer, the organic matter which undergoes the mineralization is probably not constituted of phytoplankton alone. This ratio represents more likely the degradation of dead meiofauna. The decrease observed downward is probably due to several synergetic processes such as bacterial uptake and chemisorption onto carbonate crystals surfaces.

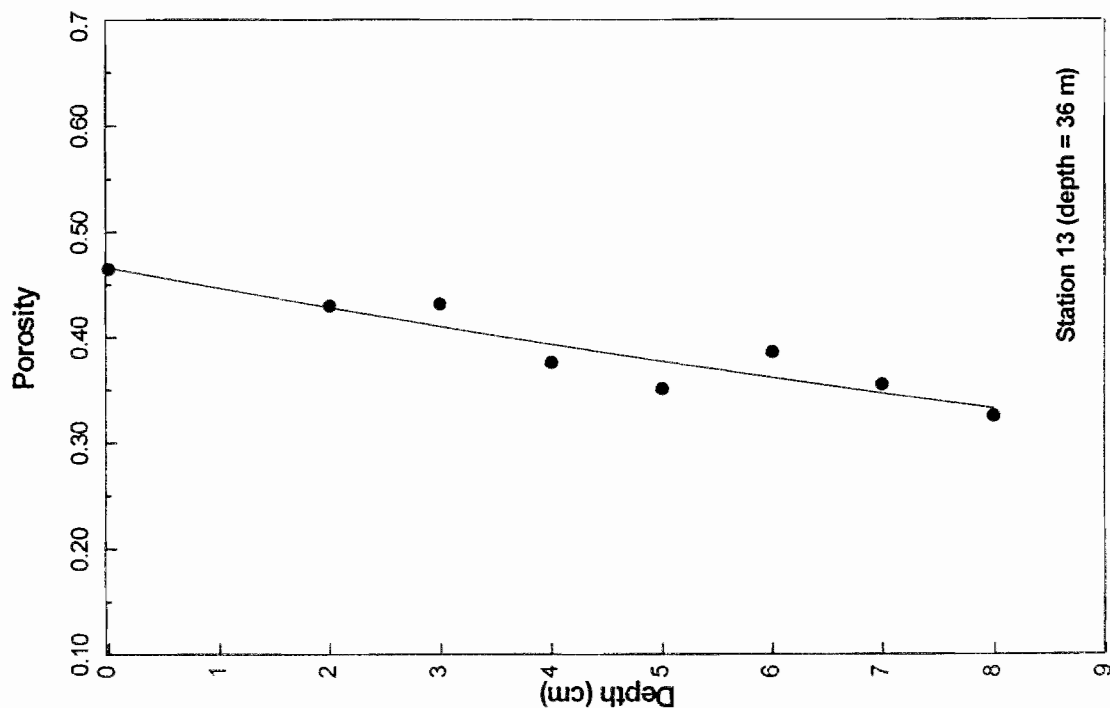


Figure 2: Porosity profile at station 13

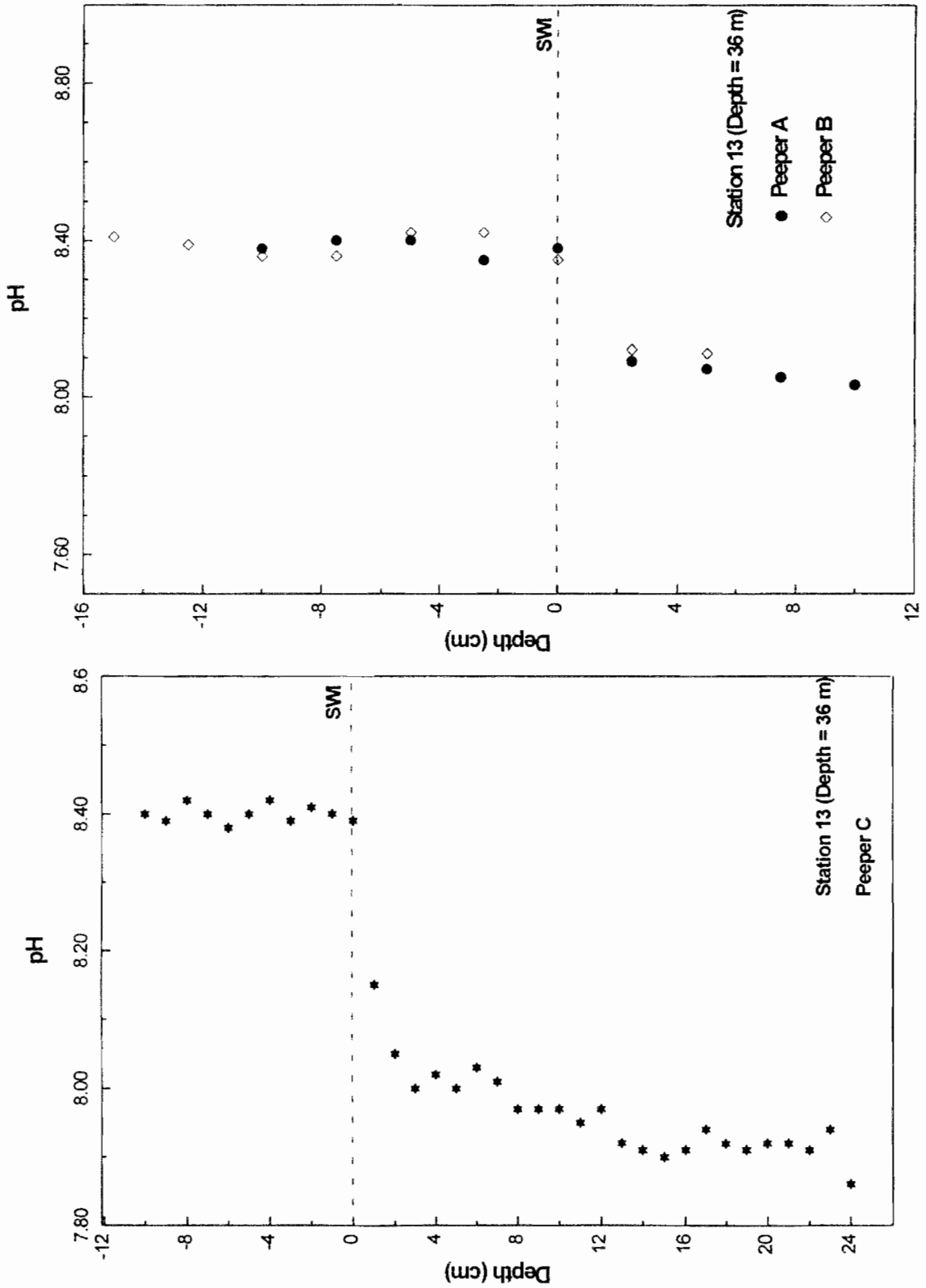


Figure 3: pH profiles at station 13

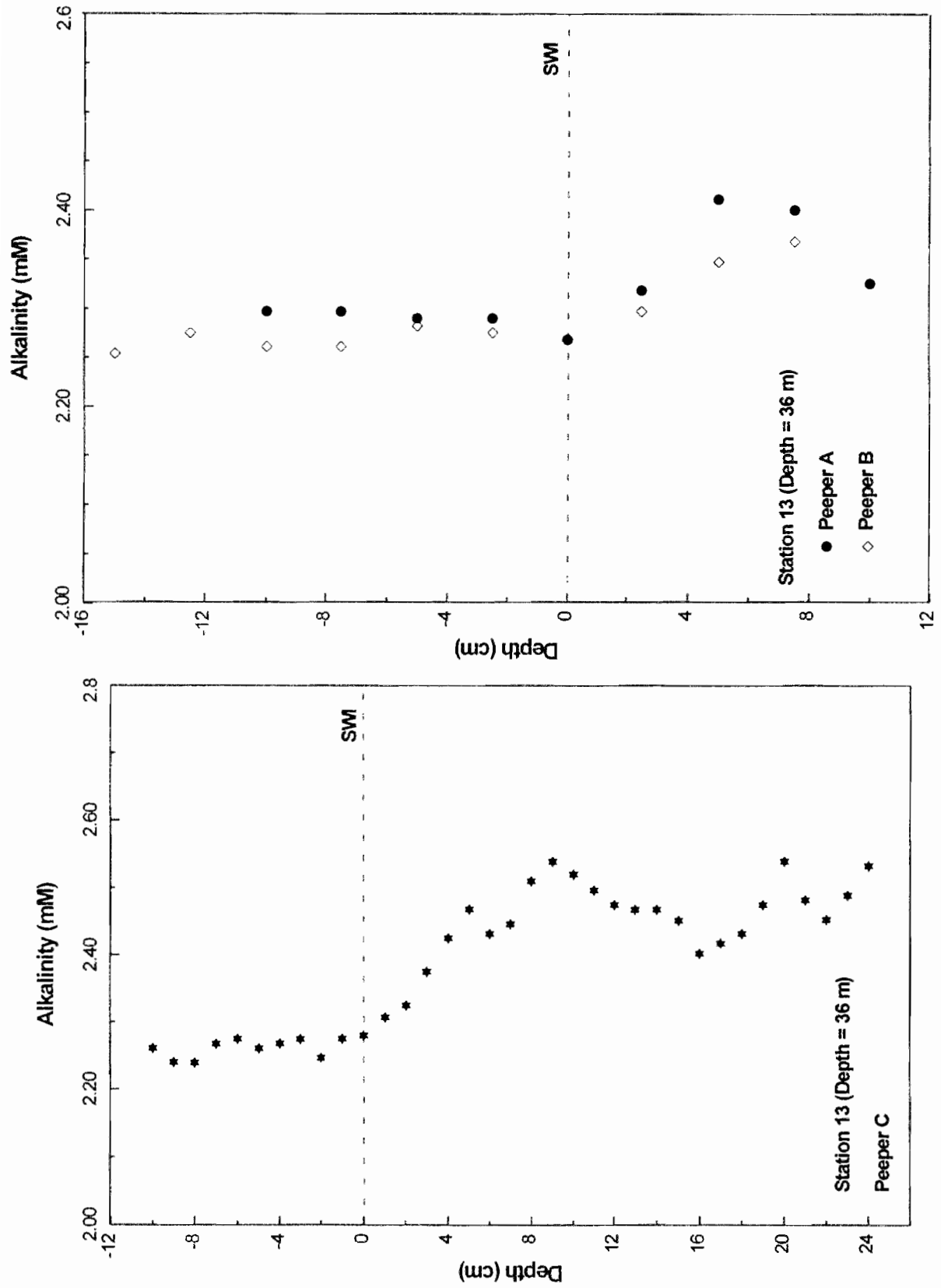


Figure 4: Alkalinity profiles at station 13

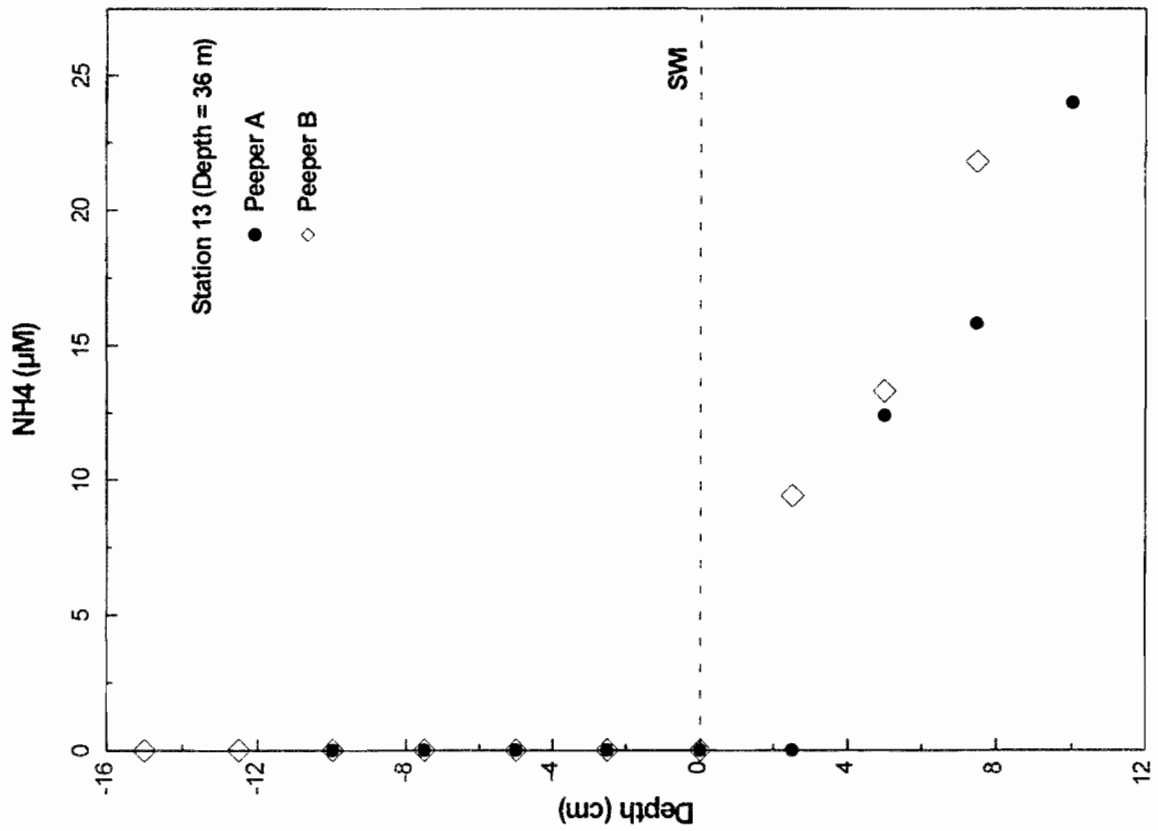


Figure 5: NH₄ profiles at station 13

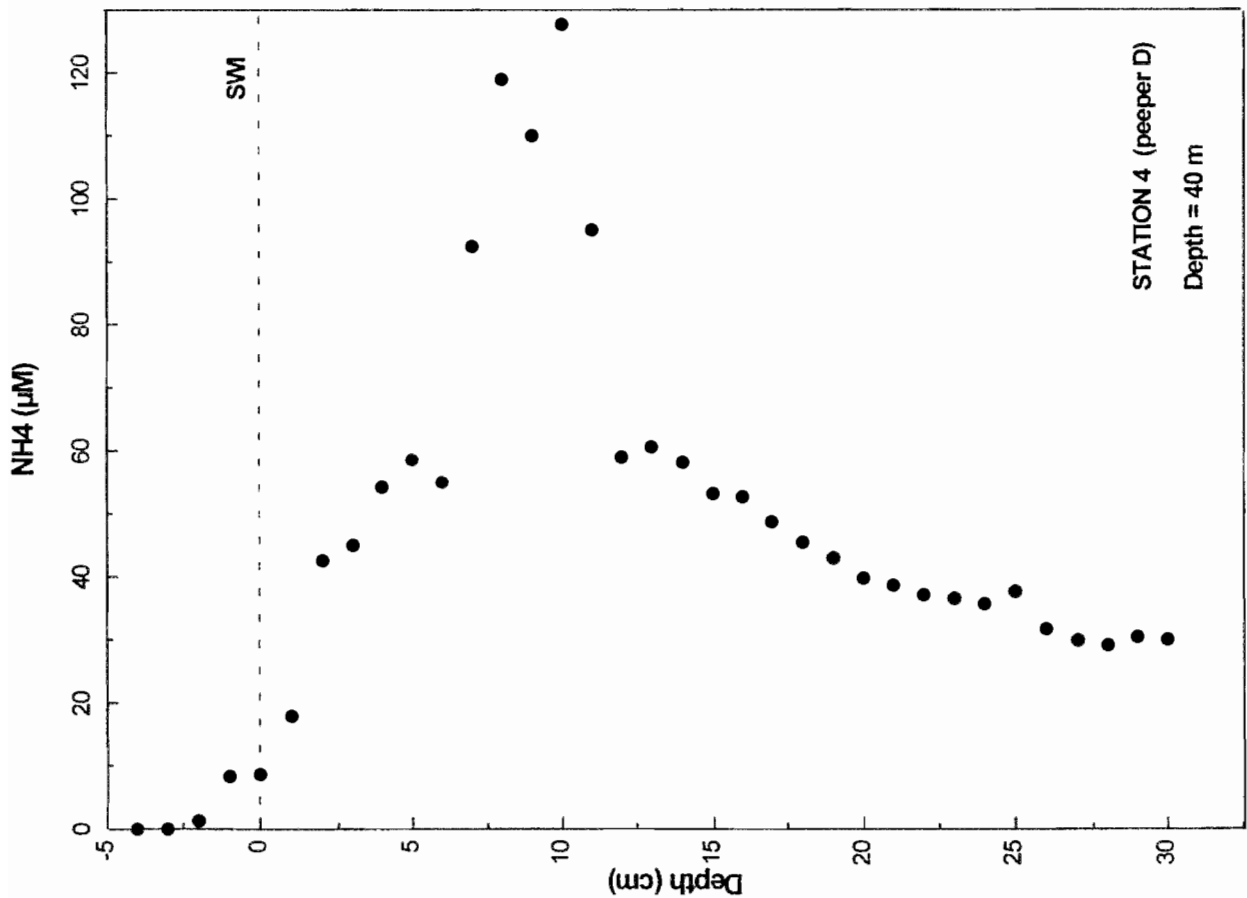


Figure 6: NH₄ profiles at station 4

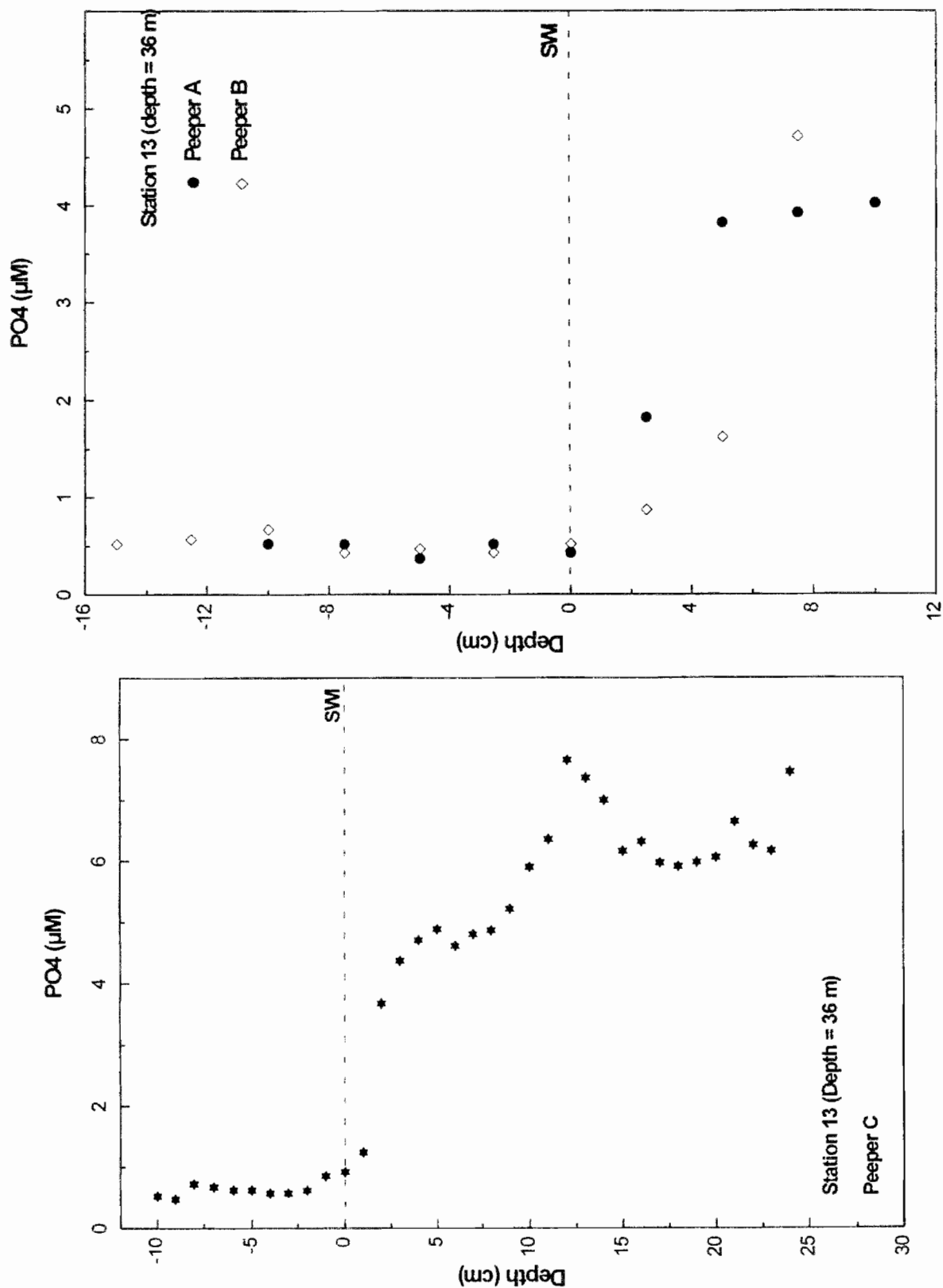


Figure 7: PO₄ profiles at station 13

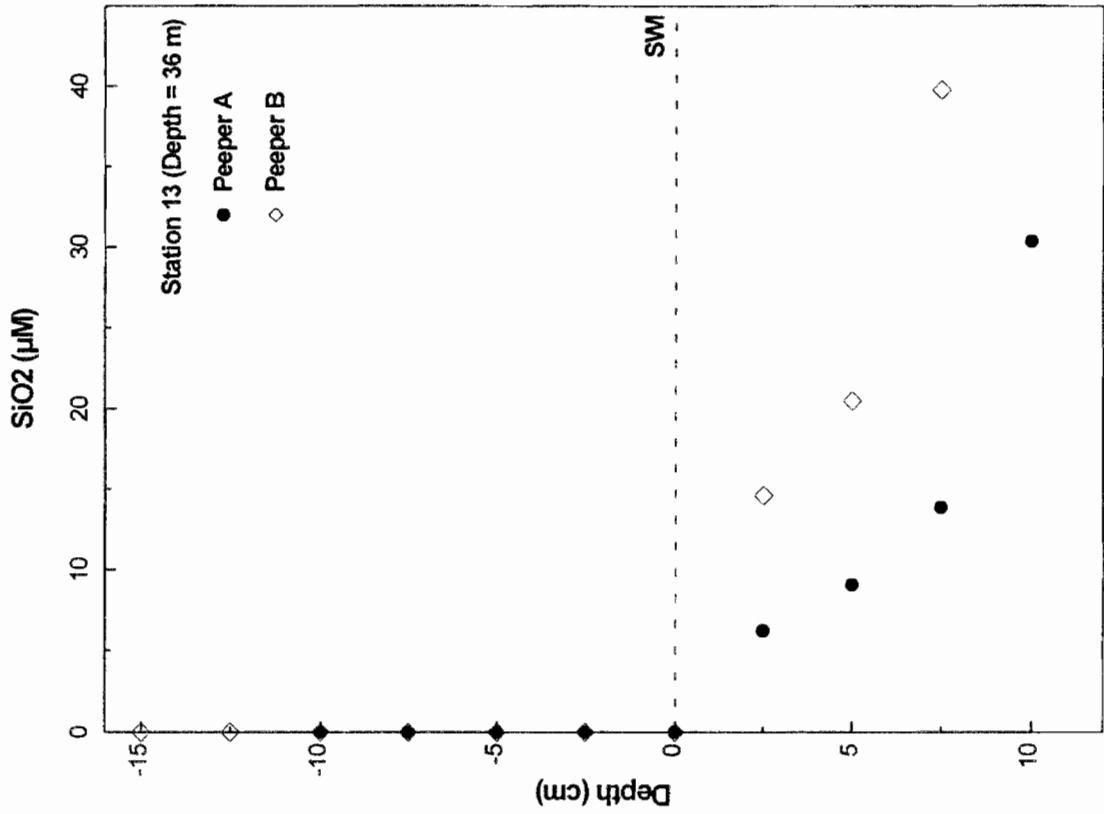


Figure 8: SiO₂ profiles at station 13

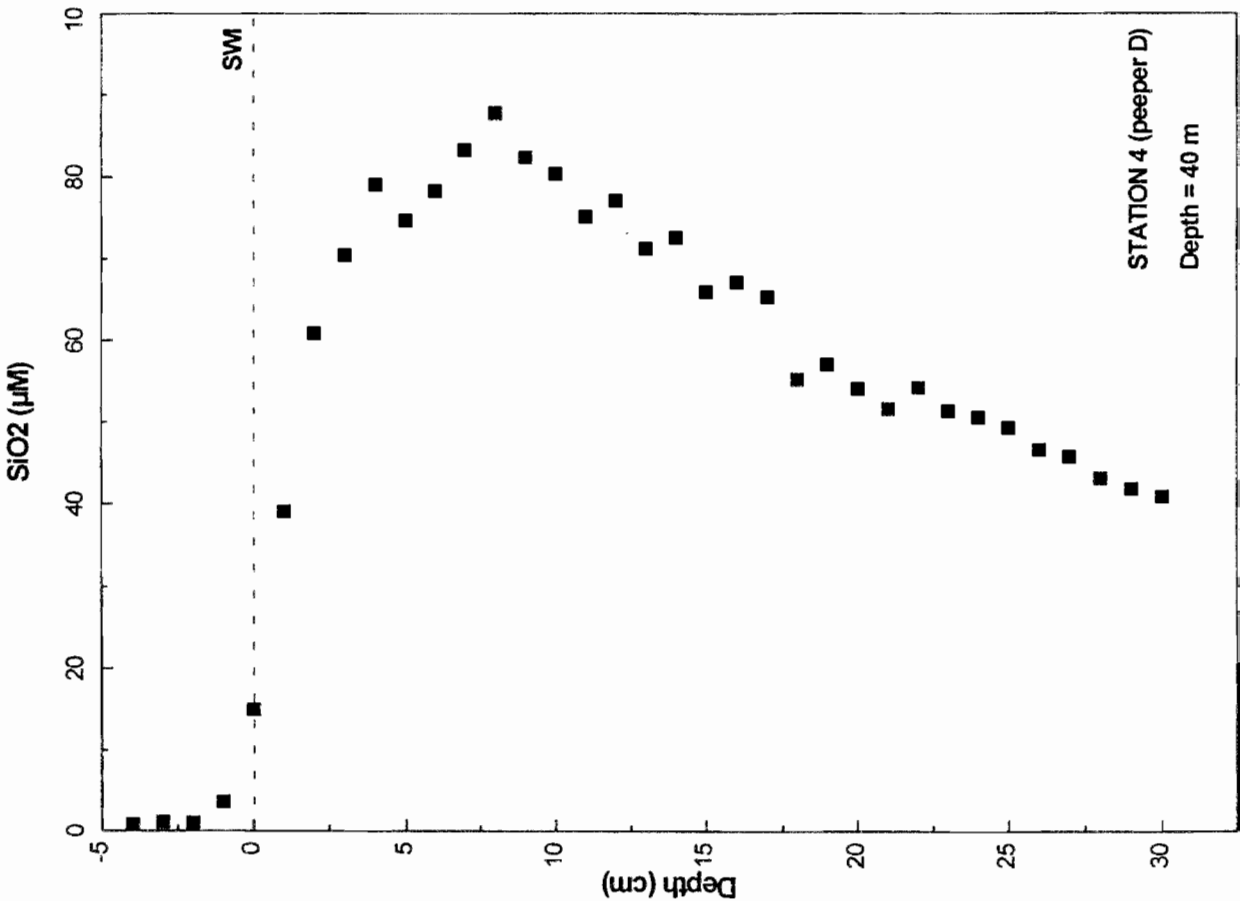


Figure 9: SiO₂ profiles at station 4

3.2 Nutrients fluxes at the sediment-water interface

Fluxes of nutrients have been calculated after the estimation of the concentration gradient at the SWI for NH_4 , PO_4 and dissolved SiO_2 .

The concentration gradient is related to the diffusional flux through the first Fick's law :

$$F = - \Phi \cdot D_s \cdot \text{grad}C \quad \text{where :}$$

Φ is the sediment porosity at the SWI : $\Phi = 0.46$

D_s is the *in situ* diffusion coefficient is calculated from the molecular diffusion coefficient corrected from the tortuosity coefficient and from the sea water viscosity :

$$D_s = \frac{\Phi}{k} \cdot D^0 \quad \text{where } k \text{ is the ratio of the viscosity}$$

of sea water to fresh water ($k = 1.08$) and D^0 the molecular diffusion coefficient.

$$\text{grad}C = \lim_{z \rightarrow 0} \left| \frac{\partial C}{\partial z} \right| \quad \text{where } z \text{ is the space}$$

coordinate positively oriented downward. Thus a negative flux means that the species diffuses upward, from the sediment to the water column.

The results are summarized in the Table 1.

Table 1: Nutrient fluxes at the WSI

STATION	PEEPER	FLUX ($\mu\text{mol m}^{-2} \text{d}^{-1}$)		
		NH_4	PO_4	SiO_2
4	D	-64.2	-1.9	-50.3
13	A	-18	-0.51	-2.4
«	B	-21	-0.13	-1.6
«	C	-	-0.98	-

As generally observed the calculated fluxes vary greatly from one station to another. This can be accounted on particular sedimentation conditions which include bottom currents, local inputs from the shore and so on.. Since the water column is highly depleted in metabolizable nitrogen (NO_3^- , NH_4^+) the SWI appears to be able to satisfy a part of the requirements in this element for the benthic communities. High flux of silica together with nitrogen at station 4 can provide a correct nutritional environment to sustain the production of benthic diatoms, while directly metabolizable N under the NH_4^+ form can be used to develop cyanobacteria mats.

Some meteorological conditions (high wind speed or storm) can disturb the superficial sediment, enhance the dispersion flux at the SWI and bring into the water column an unusual amount of nutrients

If such fluxes can be developed in many parts of the lagoon they could be partly at the origin of random algal bloom which have been reported in the past and for which no clear explanation was found (D.R. Green).

4. References

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