

Parflux E site east of Barbados¹⁵; however, the short duration of this trapping experiment limits its usefulness in establishing ocean fluxes.

The predicted carbon fluxes (Table 1) are within 12% of the measured values. Thus, the correlation seems to apply to the Atlantic as well as the Pacific Ocean.

The one-dimensional model discussed thus far considers only ²¹⁰Pb derived from atmospheric deposition or ²²⁶Ra decay in the water directly above the trap (our term *P*). Another source of ²¹⁰Pb is water moving through the area that carries ²¹⁰Pb produced elsewhere. A complete ²¹⁰Pb balance in a defined volume above each trap is described as follows:

$$\text{input of } ^{210}\text{Pb} = P + V_{\text{in}}C_{\text{in}}$$

$$\text{loss of } ^{210}\text{Pb} = F + \lambda C_{210}Ah + V_{\text{out}}C_{210}$$

where V_{in} = the volume of water moving into a defined area above the trap and equals V_{out} , the volume moving out; C_{in} = the ²¹⁰Pb activity in V_{in} ; and C_{210} = the ²¹⁰Pb activity in the defined volume above the trap, λ is the decay constant of ²¹⁰Pb, A is the defined area and h is the depth of the trap. F and P (including atmospheric input) have been defined earlier.

At steady state the input and loss terms are equal, so we may simplify the ²¹⁰Pb balance to:

$$P - F = V_{\text{in}}(C_{210} - C_{\text{in}}) + \lambda C_{210}Ah$$

If the activities of ²¹⁰Pb are constant in the water passing into and out of the area over the trap, this reduces to the one-dimensional case:

$$P - F = \lambda C_{210}Ah$$

If we neglect atmospheric input, P is a function of ²²⁶Ra

$$P = \lambda C_{226}Ah$$

where C_{226} is the activity of ²²⁶Ra. Thus:

$$F/P = 1 - \frac{C_{210}}{C_{226}}$$

and we should be able to derive F/P from the activity ratio of ²¹⁰Pb/²²⁶Ra in the water above the trap. This offers another way to deduce organic C fluxes.

These simplified relationships will not apply when waters flow across gradients of production that cause changes in the flux of organic C. In this case the water entering and leaving an area would be expected to have different activities of dissolved ²¹⁰Pb.

One area where this has been well documented is the Santa Barbara Basin off southern California where ²¹⁰Pb sediment inventories¹⁶ and the fluxes of ²¹⁰Pb recorded in sediment traps¹⁷ are many times greater than the production rate (including atmospheric deposition) in the basin waters. Clearly the advection of water through the basin carries dissolved ²¹⁰Pb into the basin where ²¹⁰Pb is stripped, probably in response to high productivity in the coastal waters. It is interesting to note that about seven times more ²¹⁰Pb is deposited in the basin than is produced there and that the organic carbon flux is about 10 times greater than in the open ocean¹⁷. Another region where ²¹⁰Pb fluxes are much higher than *in situ* production plus fallout is the continental shelf of Washington state¹⁸. Similar transport must occur there.

The model outlined above assumes steady state, an assumption that may not be valid. There is abundant evidence for significant changes in productivity over a range of time scales. The residence time of ²¹⁰Pb varies from a year or two in near surface waters to tens or hundreds of years in the deep sea⁶. Achieving a steady state in such a system is unlikely. Our data show that comparisons of annual mean fluxes reduce the scatter significantly from the seasonal samples. However, we have not considered longer time-scale variability. Because the sediment inventory of ²¹⁰Pb integrates the flux over a 100-year period, this approach to defining the organic C flux to the bottom should

be less subject to breakdown of the steady-state assumption.

If future studies confirm the interpretations we have made, excess ²¹⁰Pb inventories in surficial deep-sea sediments can be used to calculate F/P and to estimate the rain rate of organic carbon to the sediment. The rain rate coupled with measurements of organic carbon burial rates defines the rate of recycling of organic carbon at the sediment-water interface. Both rates could be inferred by measurements in surficial sediments without the complexities of deploying sediment traps in the deep sea. But the technique must first be tested by comparing ²¹⁰Pb and organic carbon fluxes from other regions.

This research was supported by the NSF as part of the MANOP project. We thank M. Bacon, M. Lyle, D. Kadko, C.-A. Huh and R. Thunell for their comments on the manuscript.

Received 22 June; accepted 20 November 1987

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Iron deficiency limits phytoplankton growth in the north-east Pacific subarctic

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An interesting oceanographic problem concerns the excess major plant nutrients (PO_4 , NO_3 , SiO_3) occurring in offshore surface waters of the Antarctic¹⁻³ and north-east Pacific subarctic Oceans⁴. In a previous study⁵, we presented indirect evidence suggesting that inadequate Fe input was responsible for this limitation of growth; recently we had the opportunity to seek direct evidence for this hypothesis in the north-east Pacific subarctic. We report here that the addition of nmol amounts of dissolved iron resulted in the nearly complete utilization of excess NO_3 , whereas in the controls—without added Fe—only 25% of the available NO_3 was used. We also observed that the amounts of chlorophyll in the phytoplankton increased in proportion to the Fe added. We conclude that Fe deficiency is limiting phytoplankton growth in these major-nutrient-rich waters.

Water, with its resident phytoplankton, was collected from a depth of 20 m at Ocean Station 'PAPA' (50.0° N; 145.0° W), using ultra-clean trace-element sampling techniques (30-litre Go-Flo bottles on Kevlar line; samples processed inside a positive pressure portable laboratory)^{6,7}. Aliquots of the raw water were placed in 2-l acid-washed polycarbonate bottles, to which various amounts of unchelated Fe (1,000 p.p.m. Fe in HNO_3 , diluted appropriately) were added (Table 1, Fig. 1). In the oxygenated seawater (pH 8), the Fe was presumed to be in the ferric (III) state in the form of hydroxy complexes⁸. Additional methodological details will be published elsewhere.

Chlorophyll concentrations remained fairly constant for all treatments on days 1-3 (Table 1, Fig. 1). On day 4, however,

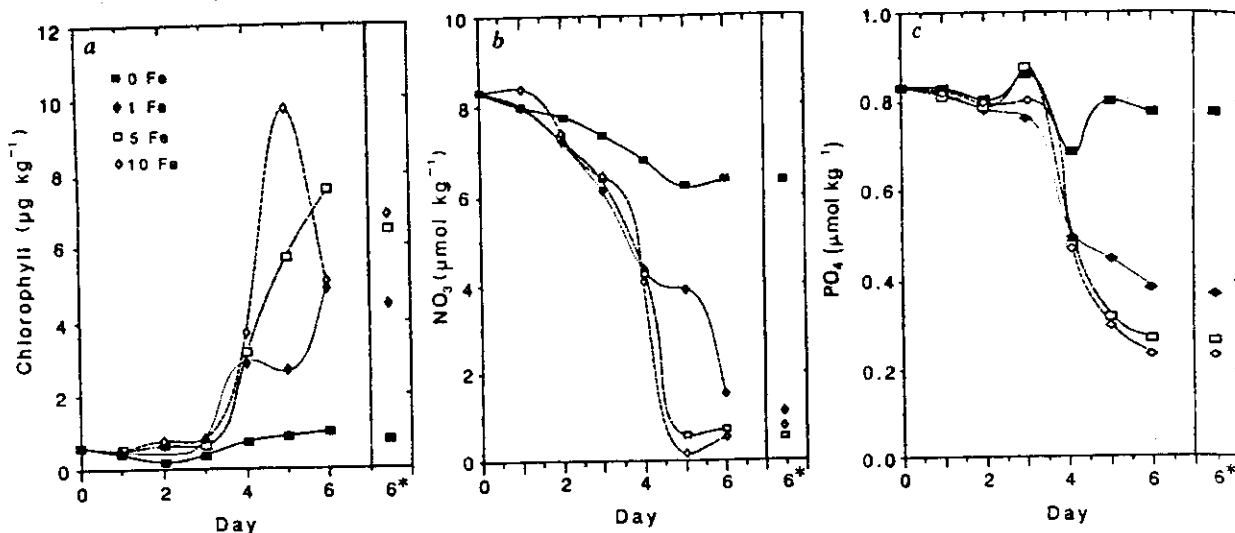


Fig. 1 Chlorophyll concentrations versus experimental day (a) in control (no added Fe) and experimental bottles with 1 nmol Fe, 5 nmol Fe and 10 nmol Fe added per kg. The data shown for day 6* represent a second set of replicates that were opened only upon the completion of the experiment. Nitrate (b) and phosphate (c) concentrations versus day are also shown. Because of the unintentional addition of $3 \mu\text{mol NO}_3 \text{ kg}^{-1}$, the data for the experimental bottles with 10 nmol Fe plus 1.0 nmol Mn and 0.1 nmol Co per kg are not included in this figure (see Table 1). Bottles were placed in three plastic bags and maintained in an all plastic deck-top incubator. Light levels were $\sim 20\%$ of those at the sea surface; temperatures were kept at $\sim 14^\circ\text{C}$ via running seawater.

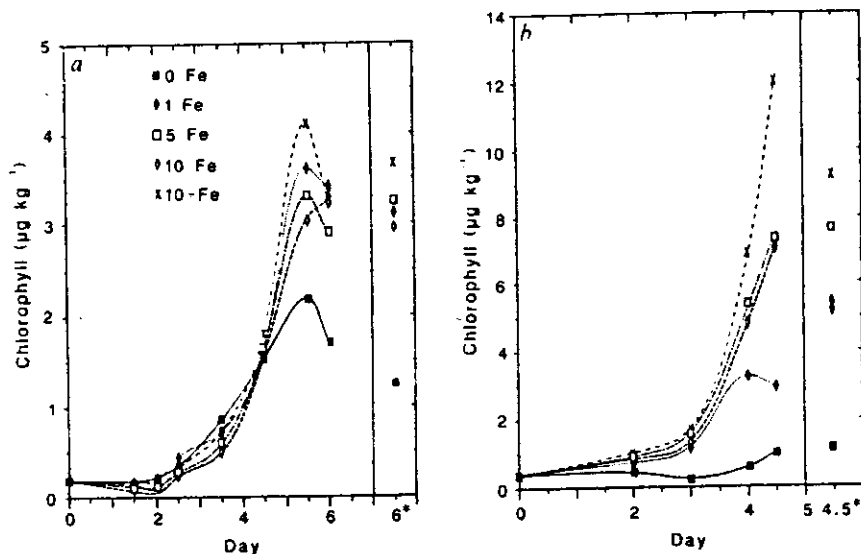


Fig. 2 Chlorophyll concentrations versus experimental day at 45.5°N ; 142.89°W (a) and at 55.5°N ; 147.5°W (b) in control (no added Fe) and experimental bottles with 1 nmol Fe, 5 nmol Fe, 10 nmol Fe, and 10 nmol Fe plus 1.0 nmol Mn and 0.1 nmol Co per kg. Day 6* and day 4.5* data represent second sets of replicates that were opened only at the end of the experiment. Initial conditions in a (15 m depth, 3 August 1987) = $0.18 \mu\text{g chlorophyll kg}^{-1}$, 6.43, 0.70 and $10.2 \mu\text{mol kg}^{-1}$ of NO_3 , PO_4 and SiO_3 , respectively. Initial conditions in b (30 m depth, 11 August 1987) = $0.35 \mu\text{g chlorophyll kg}^{-1}$, and 7.53, 0.88 and $23.06 \mu\text{mol kg}^{-1}$ of NO_3 , PO_4 and SiO_3 , respectively.

rapid growth began and substantial chlorophyll increases were observed in all of the bottles with added Fe. Accelerated growth continued into day 5 and appeared to peak in the 10 and 10+ (10 nmol Fe with 1.0 nmol Mn and 0.1 nmol Co kg^{-1} added) experimental bottles. Similar chlorophyll concentrations were found in replicate experimental bottles opened for the first time on day 6. In contrast, both control bottles had very low chlorophyll concentrations (1.07 and $0.85 \mu\text{g kg}^{-1}$) on day 6 that were scarcely higher than the initial day 0 values ($0.69 \mu\text{g kg}^{-1}$).

Nearly identical NO_3 uptake was observed in the 1, 5 and 10 nmol added Fe kg^{-1} bottles on days 1-4. On day 5 substantial NO_3 depletion continued in the 5 and 10 nmol Fe bottles, although it appeared to slow in the 1 nmol Fe bottle (Fig. 1, Table 1). Once again, the control bottles with no added Fe were completely different. The bottle sampled each day showed very slow NO_3 utilization and at the end of the experiment on day 6, 75% of the NO_3 originally present remained. The concentration in the other replicate initially opened on day 6 was nearly identical (6.38 versus $6.34 \mu\text{mol NO}_3 \text{ kg}^{-1}$). Much the same uptake pattern was also observed for PO_4 (Fig. 1). In the case

of SiO_3 , however, hardly any differences were observed between bottles (Table 1). This was surprising because scanning electron microscope examinations showed substantial diatom (*Nitzschia* spp.) growth in the experimental bottles, in contrast to the controls, which had abundant coccoliths but few diatoms.

Similar experiments were also performed 625 km north-west (55.5° N; 147.5° W) and 625 km south-east (45.0° N; 142.89° W) of Ocean Station 'PAPA' (50.0° N; 145.0° W). Essentially the same results were found; chlorophyll data are shown in Fig. 2. These data, together with those in Fig. 1, provide consistent evidence that Fe limits phytoplankton growth in the north-east Pacific subarctic. With appropriate independent confirmation, this area could become a classic marine example of Liebig's law of the minimum⁹. This research also exemplifies the necessity of using ultraclean techniques, without which the control bottles would have had enough contaminate Fe to stimulate growth and make the control results indistinguishable from those with intentionally added Fe.

The essential roles of Fe in living systems are well known¹⁰ and it is not surprising that chlorophyll levels increase with

Table 1 Nutrient uptake and chlorophyll production by phytoplankton from 20 m depth

Day	Fe added (nmol kg ⁻¹)				
	0	1	5	10	10+*
	Chlorophyll (µg kg ⁻¹)				
0	0.69 (initial value)				
1	0.49	0.46	0.52	0.46	0.49
2	0.35	0.68	0.67	0.77	0.67
3	0.41	0.84	0.67	0.84	0.92
4	0.77	2.98	3.31	3.85	3.65
5	0.92	2.77	5.97	10.16	9.09
6	1.07	5.21	7.91	5.26	6.95
6‡	0.85	4.63	6.81	7.17	14.70
	NO ₃ (µmol kg ⁻¹)				
0	8.37 (initial value)				
1	8.04	7.94	8.03	8.37	11.43†
2	7.72	7.19	7.22	7.34	10.58
3	7.34	6.10	6.40	6.40	9.75
4	6.80	4.32	4.24	4.04	7.57
5	6.25	3.84	0.60	0.15	2.69
6	6.34	1.51	0.75	0.50	0.52
6‡	6.38	1.14	0.58	0.77	0.56
	PO ₄ (µmol kg ⁻¹)				
0	0.83 (initial value)				
1	0.83	0.82	0.83	0.83	0.83
2	0.80	0.78	0.79	0.79	0.79
3	0.86	0.76	0.86	0.80	0.80
4	0.68	0.50	0.48	0.47	0.48
5	0.80	0.44	0.32	0.30	0.33
6	0.78	0.38	0.27	0.23	0.09
6‡	0.74	0.42	0.22	0.23	0.07
	SiO ₃ (µmol kg ⁻¹)				
0	16.13 (initial value)				
1	16.94	16.96	16.88	17.01	17.04
2	17.24	17.22	17.19	16.86	17.14
3	19.49	18.51	18.53	18.08	17.96
4	18.40	17.65	17.21	17.40	16.65
5	18.80	—	16.90	15.82	15.81
6	—	—	17.31	15.82	14.43
6‡	—	18.60	—	15.34	11.69

The daily chlorophyll and nutrient data are from control and experimental bottles with 0, 1, 5 and 10 nmol added Fe kg⁻¹ at Ocean Station PAPA (50.0° N; 145.0° W): water was collected on 6 August 1987 from depth of 20 m. Data from a second set of replicates opened only at end of experiment are also shown.

* 10+ = 10 nmol Fe + 1 nmol Mn + 0.1 nmol Co kg⁻¹.

† About 3 µmol NO₃ kg⁻¹ inadvertently added with Co.

‡ The second replicate that was not opened until day 6.

increased Fe, as this element is required for the synthesis of this pigment¹¹. Iron is also required for the synthesis of several photosynthetic electron transport proteins and for the reduction of CO₂, SO₄ and NO₃ during the photosynthetic production of organic compounds¹².

In addition to the necessity for Fe at the cellular level, it is possible that oceanic Fe availability may be important in determining global atmospheric CO₂ levels and, hence, affecting global climate. The major role of the Antarctic Ocean in the removal of CO₂ from the atmosphere has been noted in several studies¹³⁻¹⁷. It is argued that the lower ice age atmospheric CO₂ levels (200 p.p.m.v.) measured in ancient air¹⁸⁻²¹ trapped in ice resulted from the increased biological productivity and CO₂ removal at high latitudes. When high-latitude surface nutrients are completely used up by the phytoplankton, the biological removal mechanism operates at maximum efficiency and CO₂ is withdrawn from the atmosphere^{14,15}. In contrast, when surplus nutrients are left unutilized, the biological pump is inefficient and CO₂ levels increase. The range in these two extremes results in a factor of three difference in the atmospheric CO₂ content¹⁶.

We believe that a situation similar to that in the Gulf of Alaska also exists in offshore Southern Ocean waters, where surface concentrations of major nutrients such as NO₃ (25.2 µmol kg⁻¹), PO₄ (1.68 µmol kg⁻¹) and SiO₃ (54.9 µmol kg⁻¹) are 2-3 times higher (examples from GEOSECS Station 287; 2 m depth, 69°S; 173°30' W)²². We postulate⁵ that Fe supplies from local Antarctic continental margin sources are rapidly used up in the highly productive neritic waters of this region and little, if any, Fe mixes out into the open Southern Ocean. Also, atmospheric Fe input is limited, as Southern Ocean dust fluxes between 40° S and the Antarctic continent are among the lowest in the world because of the absence of land masses; that is, only the tip of South America and the southern island of New Zealand are in this region of strong prevailing westerly winds²³. This low dust input is exemplified by the small amounts of Al found in present-day interglacial Antarctic Vostok ice core samples. De Angelis *et al.*²⁴ report that mean Al concentrations are 2.5 ± 0.9 ng g⁻¹ during the present interglacial (Holocene) and are not significantly different from those of the previous interglacial (3.0 ± 1.5 ng Al g⁻¹, 127 kyr BP). This would be ~0.03-0.04 nmol of Fe g⁻¹ based upon Taylor's²⁵ crustal abundance estimates (8.23% Al and 5.63% Fe; Fe:Al ratio = 0.68). Levels of CO₂ in the same Vostok ice core, were highest during these interglacials, on the order of 270-280 p.p.m.v.²¹.

In contrast to interglacials, glacial periods had increased amounts of atmospheric dust Al and Fe and low CO₂ concentrations; for example, during the last glacial maximum, when Al levels were 50 times higher (~125 ng Al g⁻¹)²⁴, CO₂ concentrations were <200 p.p.m.v.²¹. The larger glacial period Al amounts are expected as tropical arid zones were five times larger, wind speeds 1.3-1.6 times higher, and atmospheric dust loads 10-20 times greater during the ice ages²⁶. We postulate that the enhanced supply of Fe from the atmosphere stimulated photosynthesis, which led to the drawdown in atmospheric CO₂ levels during glacial maxima.

We thank Mike Gordon, Sara Tanner and Ginger Elrod for their help with the field work. The VERTEX nutrient data were supplied by Sandy Moore and Larry Small of Oregon State University. This research was supported by grants from the NSF Marine Chemistry Programme and the US Office of Naval Research Oceanic Chemistry Programme.

Received 9 September; accepted 11 December 1987.

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