

Insight into contributions of different iron sources to the ocean from a model of the Fe stable isotopes

Motivation

It has been widely accepted that iron is an essential micronutrient for ocean biota, and its distribution strongly affects the magnitude of phytoplankton primary productivity and thus the carbon uptake in the ocean. Yet there is still no consensus on the mechanisms behind the distribution of iron and especially on the relative role of different external iron sources to the ocean. In recent years, the analysis of the stable isotopic composition of dissolved iron in sea water has been increasingly used to constrain the relative role of different iron sources. In a global biogeochemical model, we aim to take into account processes in the ocean interior to fractionate between iron isotopes and physical processes to mix water masses with different isotopic compositions.

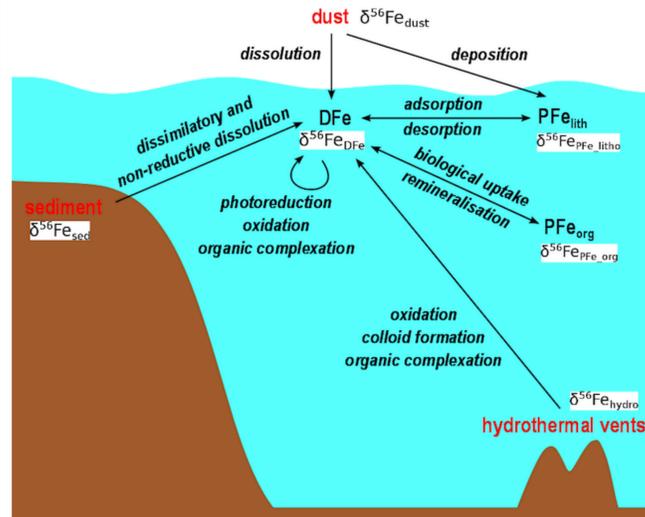


Fig. 1: Concept of the Fe isotope model REcoM-Felso.

REcoM-Felso Model

A global biogeochemical model including iron (REcoM, Hauck et al. 2013) has been extended with an explicit representation of isotopic effects (REcoM-Felso). The current model version considers (Fig. 1): ⁵⁶Fe in dissolved iron (DFe), in all organic iron pools (phytoplankton, zooplankton and detritus), and in scavenged iron. Three external iron sources, i.e. dust, hydrothermal vents and sediment, bring DFe into the ocean with different isotopic composition. ⁵⁶Fe from different sources is transported in the dissolved form, mixed between water masses, taken up by phytoplankton and goes through the biological cycle. To compare with observations, we calculate δ⁵⁶Fe (‰) with:

$$\delta^{56}\text{Fe} = \left[\frac{(^{56}\text{Fe}/^{54}\text{Fe})_{\text{model}}}{(^{56}\text{Fe}/^{54}\text{Fe})_{\text{IRMM-014}}} - 1 \right] \times 1000$$

Sensitivity runs with varying source signals

At the first step we wanted to understand the effect of physical mixing on the isotopic composition of DFe and conducted first simulations without chemical and biological fractionation. δ⁵⁶Fe (‰) values for dust, hydrothermal and sedimentary source were from literature and a set of sensitivity runs was conducted in order to illustrate how variations in source signals affect the end-member isotopic composition of DFe (Tab. 1).

Run	dust	hydrothermalism	sediment
R ₀	+0.10	-0.50	-3.00
R _{non-red}	+0.10	-0.50	+0.30
R _{high_sed}	+0.10	-0.50	-1.00
R _{low_sed}	+0.10	-0.50	-5.00
R _{high_hydro}	+0.10	-0.30	-3.00
R _{high_dust}	+0.30	-0.50	-3.00

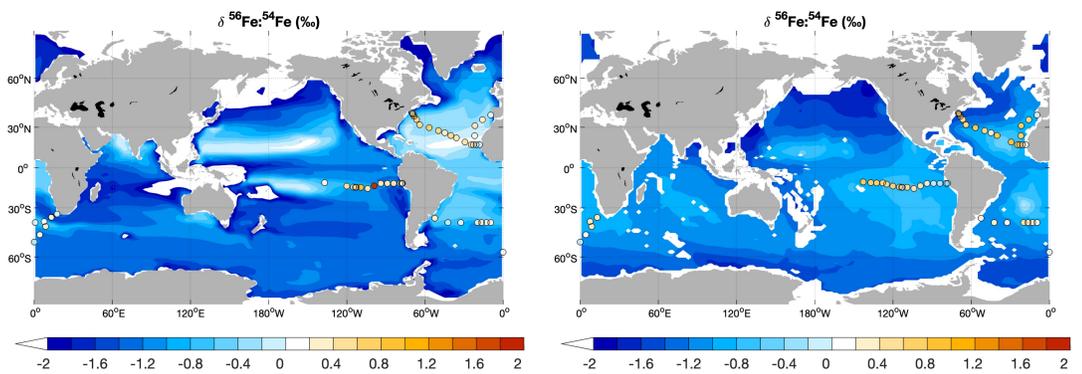


Fig. 2: δ⁵⁶Fe (‰) in R₀, averaged for the upper 200 m (left) and between 2000 and 3000 m (right). Colour dots are GEOTRACES measurements.

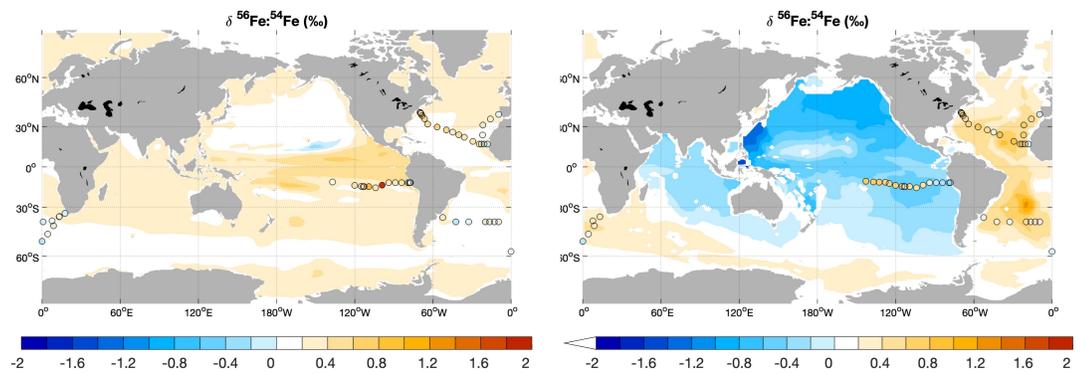


Fig. 3: δ⁵⁶Fe (‰) in R_{non-red}, averaged for the upper 200 m (left) and between 2000 and 3000 m (right). Colour dots are GEOTRACES measurements.

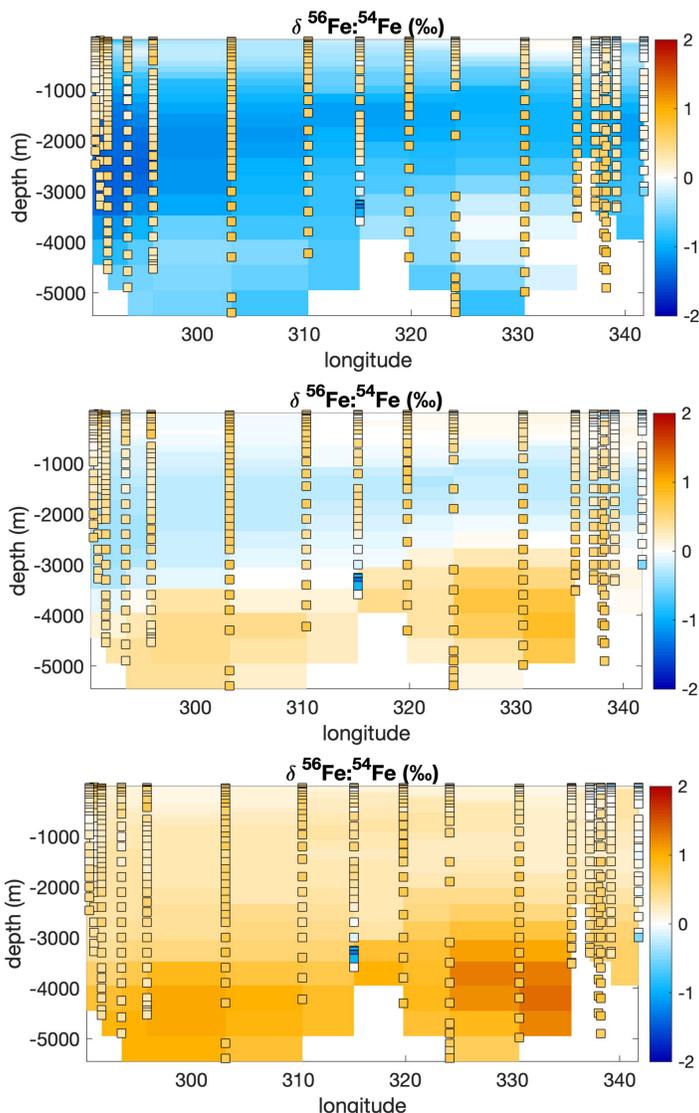


Fig. 4: δ⁵⁶Fe (‰) in R₀ (top), R_{non-red} (middle) and R_{high_sed} (bottom) along GA03. Colour dots are GA03 measurements from GEOTRACES IDP 2017.

Effect of the source signals on global distribution of δ⁵⁶Fe in DFe

The pattern of δ⁵⁶Fe distribution in the model is determined by physical transport and mixing of three sources through circulation, and vertical transport of iron from the surface to the ocean interior by biological uptake, sinking and remineralisation of organic particles.

Slightly positive values of δ⁵⁶Fe in R₀ (Fig.2) are found in regions receiving much dust input of iron, e.g. the subtropical North Atlantic, eastern North Pacific affected by the Asian dust events and Indian Ocean. In HNLC regions, the main source of DFe is sediment input and upwelling. Both are isotopically light and result in more negative δ⁵⁶Fe. In the subtropical Pacific gyres, δ⁵⁶Fe is close to 0, due to lack of any external source and intensive biological activities. In all ocean basins and throughout the water column (Fig.4 top), values in R₀ produced are much lower than observed.

With enhanced δ⁵⁶Fe in the individual sources, R_{high_hydro} and R_{high_dust} show just slightly higher δ⁵⁶Fe in the deep and surface ocean, respectively. From R_{low_sed} via R₀ to R_{high_sed}, sediment supplies DFe with an increasing but still negative δ⁵⁶Fe. Modelled δ⁵⁶Fe in DFe increased correspondingly, but even in R_{high_sed}, δ⁵⁶Fe in the surface and mesopelagic waters is still lower than observation (Fig.4 middle). The positive signals in the deep ocean which is in agreement with observation, can not be explained by the negative signal from sediments but by transport of iron deposited at the surface.

R_{non-red} assumes that all sedimentary sources supply iron through the non-reductive release. This significantly increases δ⁵⁶Fe in DFe (Fig.3 and Fig.4 bottom), particularly in the global surface ocean and in the deep Atlantic and Southern Ocean. Below the surface water however, the model underestimates δ⁵⁶Fe in the open Pacific ocean where hydrothermal source might be the most important iron source. This simulation very likely overestimates the isotopic composition in the sediment source. But its overall effect on both surface and deep distribution of δ⁵⁶Fe strongly suggest for future studies:

- that variations in the source signals are crucial for determination of contribution by the single sources, particularly because of their effect in remote regions through biological cycling and ocean circulation;
- that these variations could be explained by processes quickly fractionating iron isotopes after iron released from the sources. More comprehensive studies on how iron from different sources undergoes fractionation are needed to understand the cycle of iron isotopes.