



Biogenic and nonliving labile particulate iron in the subtropical North Pacific Ocean

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Abstract. Biogenic and authigenic particulate iron (pFe) are key components of the marine iron cycle, influencing the fate of Fe in the upper ocean. However, their relative contributions to the total pFe pool are challenging to quantify. The chemical leach commonly used to operationally define ‘labile’ pFe is thought to extract both the biogenic and authigenic phases. To independently determine biogenic pFe, we conducted Fe uptake experiments in the surface mixed layer on 12 cruises in the
10 subtropical North Pacific Ocean. Bulk Fe uptake rates varied ~2.5-fold throughout the year, increasing with increasing *Prochlorococcus* and picoeukaryotes abundances. We used particulate carbon and phosphorus as biomass estimates in combination with Fe:C uptake ratios, finding that both led to overestimations of the biogenic pFe pool (>200% of labile pFe in the surface mixed layer). Using the nucleotide adenosine-5'-triphosphate (ATP) as an alternate estimate of living biomass instead suggested that biogenic pFe comprised ~60% of labile pFe in the mixed layer. The remainder of the labile pFe pool,
15 defined as ‘nonliving’, dominated the labile pFe pool below the euphotic zone, capturing contributions from detrital organic matter, authigenic minerals, and dust. A comparison of Fe phases between Station ALOHA and the subtropical North Atlantic revealed similar concentrations of dissolved Fe and biogenic pFe, but higher nonliving labile pFe concentrations in the North Atlantic, likely reflecting greater dust deposition. The greater role of biogenic pFe in the North Pacific may enable high efficiency Fe recycling, which rivals that observed in Fe-limited ecosystems.

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1 Introduction

The micronutrient iron (Fe) limits phytoplankton growth in ~30% of the global surface ocean, regulating microbial carbon uptake in these regions (Moore et al., 2013; Moore et al., 2001). Therefore, improving our understanding and modeling of marine Fe cycling is crucial for accurately representing drivers of marine carbon cycling and projecting impacts
25 of future environmental change. Global marine Fe models have historically struggled to reproduce observations of dissolved iron (dFe) (Tagliabue et al., 2016), in part due to uncertainties in the internal cycling of Fe in the surface ocean (Boyd et al., 2017; Tagliabue et al., 2019). Particulate Fe (pFe) phases act as sinks and sources of dFe, exerting a major influence on dFe distributions (Bressac et al., 2019; Hurst et al., 2010; Ohnemus et al., 2019). Chemically labile pFe undergoes both biotic and

abiotic transformations, leading to longer retention in the upper ocean than recalcitrant Fe phases (Bates et al., 2025).

30 However, quantifying the biotic and abiotic components of the labile pFe pool remains a challenge.

Biotic iron recycling is driven by the uptake and subsequent remineralization of Fe by microorganisms. Fe is an essential micronutrient for marine organisms, necessary for many microbial functions including photosynthesis, respiration, nitrogen fixation, and more (Raven et al., 1999; Sunda, 1988). Fe quotas, or biomass-normalized cellular Fe content, have been shown to vary for both cultured and natural phytoplankton communities across taxonomic groups, iron availability, and other environmental conditions such as light availability (Twining et al., 2021; Twining & Baines, 2013 and references therein). Boyd et al. (2015) proposed that biogenic pFe pools are often set by the balance of these Fe quotas with rates of Fe recycling, after observing similar biogenic pFe concentrations across ecosystems with dFe differing by an order of magnitude. Biogenic pFe pools can be challenging to estimate, due to variability in determined Fe quotas based on the methods used (King et al., 2012) and difficulties in quantifying C in living biomass. For the latter, previous studies have used measurements of particulate P (e.g., Sofen et al., 2023) or estimates deriving from biovolume (e.g., Boyd et al., 2015). Alternatively, adenosine-5'-triphosphate (ATP) has been employed as a proxy for living biomass, with studies at Station ALOHA finding just 26-42% of upper ocean particulate carbon is living (Henderikx-Freitas et al., 2021; Karl et al., 2022). Concurrent measurement of Fe quotas and ATP thus could provide reasonable estimates of Fe contained within living cells, or biogenic pFe.

45 Authigenic particulate Fe is thought to form via the aggregation of colloidal size (0.02-0.2 μm) Fe oxyhydroxides, by the adsorption of unchelated Fe onto existing particles, or by the precipitation of Fe oxide coatings from solution (Goldberg, 1954; Gunnars et al., 2002; Honeyman & Santschi, 1989; Landing & Bruland, 1987), all of which arise from the low solubility of Fe(III) in seawater (~ 0.2 nM; Kuma et al., 1996; Liu & Millero, 2002). When Fe supply is high, the dFe pool often contains a large fraction of colloidal Fe (Bergquist et al., 2007; Wu et al., 2001), which can be scavenged by both lithogenic material (Honeyman & Santschi, 1991; Ye & Völker, 2017) as well as phytoplankton surfaces (Hudson & Morel, 1989; Tang & Morel, 2006; Tovar-Sanchez et al., 2003). The quantification of the authigenic fraction in particulate samples has been difficult as there are currently no methods for direct measurement of the entire authigenic Fe pool. Previous studies have suggested large contributions of authigenic pFe to the labile pFe pool without direct measurement (Black et al., 2020; Marsay et al., 2017; Ohnemus et al., 2019). Recently, estimations of the biogenic pFe fraction have been used in combination with a short acetic acid and hydroxylamine leach accessing the labile pFe pool (Berger et al., 2008), allowing for the calculation of the authigenic pFe fraction by difference (Sofen et al., 2023). Sofen et al. (2023) found that the biogenic pFe pool was often much smaller than the labile pFe pool, and therefore that authigenic pFe was a significant contributor to labile pFe in several ocean basins under both high and low Fe concentrations. The addition of colloidal production of authigenic pFe as a sink of dFe from the surface ocean also improved the ability of a global-scale biogeochemical model to reproduce both temporal Fe observations in the North Atlantic gyre and global dFe observations (Tagliabue et al., 2023), highlighting the importance of this phase in driving upper ocean Fe cycling particularly in regions with high Fe inputs.



Here, we utilize bulk Fe uptake rates and concurrent particulate Fe measurements to explore the distributions and behaviors of the biogenic and nonliving labile pFe pools in the North Pacific Subtropical Gyre. Building off of a recently designed low level double spike technique (Hawco et al., 2022), we performed stable Fe isotope uptake experiments in the surface mixed layer on 12 cruises over three years at Station ALOHA, site of the Hawaii Ocean Time-series (HOT) program. We then compare three potential approaches to quantifying biogenic pFe concentrations in the upper 300 m based on the bulk Fe uptake rates. In combination with suspended particulate Fe measurements, we estimate the nonliving labile pFe fraction and assess differences between the two pools over time and with depth.

2 Methods

2.1 Sample collection and processing

Samples were collected on 12 Hawaii Ocean Time-series (HOT) cruises to Station ALOHA (22.75°N, 158.0°W) aboard the R/V *Kilo Moana* between May 2021 and September 2023. Collection and processing of water column dissolved and particulate Fe profiles have been previously described (Bates et al., 2025; Bates & Hawco, 2025). Briefly, water column particulate samples were filtered at 0.2 μm and first digested using an acetic acid-hydroxylamine hydrochloride leach to dissolve the chemically labile fraction (pFe_{Labile}; Berger et al., 2008). Lithogenic particulate iron (pFe_{Lith}) was operationally defined by an HF/HNO₃ digest performed sequentially following the labile digest (Planquette & Sherrell, 2012). The sum of the lithogenic and labile fractions represents the total particulate Fe (pFe_{Total}; see Bates et al., 2025).

2.2 Iron uptake incubations

Iron uptake incubations were performed following the low-level double spike procedure described in Hawco et al. (2022), with minor modifications. Four “pre-incubations” were prepared by adding 100 μL of a ~ 1 μM Fe double spike solution (Fe_{DS}) containing ⁵⁷Fe and ⁵⁸Fe to 100 mL of 0.2 μm filtered trace metal clean seawater, recently collected from the surface mixed layer. The pre-incubations were then left in the dark at room temperature for approximately 24 hours to allow the Fe spike to equilibrate with natural Fe-binding ligands. Seawater for the incubations was collected after sunset from the mixed layer (sampled at 25 m depth), following trace metal clean procedures. The incubations were set up in a “clean laboratory van” under red light to avoid light shock to the phytoplankton. The 100 mL pre-equilibrated solution was added to a 2 L acid-cleaned polycarbonate bottle pre-rinsed with filtered seawater, which was then filled with unfiltered trace metal clean seawater. The incubations were performed in triplicate. A control incubation was conducted using 0.2 μm filtered seawater to account for abiotic precipitation of pFe or adsorption of dFe onto the filter. The bottles were placed in an on-deck flow-through incubator, shaded to light levels equivalent to a depth of 25 m and left for 24 hours. The incubations were then harvested via vacuum filtration onto acid-cleaned 0.2 μm Supor filters (47 mm), with the filtrate collected in acid-cleaned 1 L bottles. The filtrate was acidified to pH 1.8 using hydrochloric acid (Fisher Scientific OPTIMA grade) and left



for > 2 months. Laboratory purification of samples and measurement via MC-ICPMS followed the methods described in
Hawco et al. (2022), with measurements performed using a Thermo Scientific Neptune at the University of Southern
95 California, using the IRMM-014 standard to define the natural abundance ratio of Fe.

The mean Fe_{DS} recovered from both the dissolved and particulate phases was 55 ± 21 pM, which represents ~12%
of the mean mixed layer dFe concentration for these cruises (0.43 ± 0.23 nM). Dissolved Fe turnover times (τ_{FeDS}) and
uptake rates (ρ_{dFe}) were calculated following the equations in Hawco et al. (2022):

$$\tau_{FeDS} = \left(\frac{pFe_{DS} + dFe_{DS}}{pFe_{DS}} \right) \times \Delta t, \quad (1)$$

100 based on the concentrations of the Fe double spike in the particulate and dissolved phases after 1 day. The median pFe_{DS} of
all control samples (1 pM, n=12) was subtracted from the pFe_{DS} for each sample. The uptake rate was calculated as:

$$\rho_{dFe} = \frac{[dFe]}{\tau_{FeDS}}, \quad (2)$$

where [dFe] reflects the dFe concentration at 25 m measured by isotope dilution after extraction with Nobias PA-1 resin
(Bates & Hawco, 2025).

105 2.3 Hawaii Ocean Time-series data

Primary production (*in situ* ^{14}C -radiotracer method; Steemann Nielsen, 1952), particulate adenosine-5'-triphosphate
(firefly bioluminescence assay; Karl, 1993), particulate carbon, particulate phosphorus, and flow cytometry (Rii et al., 2016)
data at Station ALOHA were obtained from the Hawaii Ocean Time-series Data Organization and Graphical System (HOT-
DOGS: <https://hahana.soest.hawaii.edu/hot/hot-dogs/>). Primary productivity measurements were not available for all cruises;
110 for March and August of 2021 and 2022, the 5-year mean (2019-2023) of March and August measurements were used. For
pFe sampling at 15 m and 300 m depths, no HOT corresponding samples were available and HOT parameters for these
depths were linearly interpolated from samples immediately above and below these depths.

3 Results and Discussion

3.1 Seasonal variability in Fe uptake

115 We measured iron uptake rates at 25 m on 12 cruises to Station ALOHA in 2021-2023, with good seasonal
coverage for spring (MAM, n=4), summer (JJA, n=5), and fall (SON, n=3), but with no experiments conducted during
winter. Concurrent measurements of dissolved, suspended particulate, and sinking particulate Fe were also conducted on
these cruises, as well as 9 additional cruises between December 2020-November 2023 (Bates et al., 2025; Bates & Hawco,
2025). Seasonal variability was observed for all Fe phases measured, with euphotic zone dFe, labile pFe, and lithogenic pFe
120 higher in winter and spring, driven by local island inputs in winter and Asian dust deposition in spring (Bates & Hawco,
2025). Significant interannual variability was also evident, with large dust-driven peaks in particulate Fe export observed in



May 2021 and 2022, but not in 2023, likely reflecting lower dust deposition in 2023 (Bates et al., 2025). Additionally, anomalously high dFe was found in September 2022, coinciding with a large diazotroph bloom (Foreman et al., 2025).

Bulk iron uptake rates varied ~2.5-fold by month and were, on average, highest in spring ($71 \pm 11 \text{ pM d}^{-1}$, mean and 1 SD) and lowest in summer and fall ($40 \pm 15 \text{ pM d}^{-1}$ in summer and $41 \pm 19 \text{ pM d}^{-1}$ in fall; Fig. 1). Uptake rates generally increased with dFe concentration (Fig. 1b), but also appeared to be driven by other factors. Indeed, uptake rates were similar in May for both 2021 and 2023 (79 ± 36 and $80 \pm 37 \text{ pM d}^{-1}$, respectively) despite a ~2-fold difference in mean mixed layer dFe concentrations (0.50 ± 0.08 and $0.27 \pm 0.02 \text{ nM}$, respectively; Bates & Hawco, 2025). Fe uptake rates were instead correlated with *Prochlorococcus* and picoeukaryote cell counts ($R^2 = 0.47$ and 0.41 , respectively, Fig. 1) and a multiple linear regression model for both populations showed an improved fit with Fe uptake rates ($R^2 = 0.89$, Fig. 1e, further details in Supplement). *Prochlorococcus* and picoeukaryotes together make up the bulk of picophytoplankton biomass and primary production at Station ALOHA (Rii et al., 2016). In contrast, uptake rates did not correlate well with abundances of *Synechococcus* or heterotrophic bacteria ($R^2 = 0.09$ and 0.15 , respectively; Table S1 in Supplement). ^{14}C -based primary production was less variable than Fe uptake rates over these cruises (RSD = 0.13 for primary production compared to 0.40 for Fe uptake). As a result, our inferred Fe:C uptake ratios showed a similar temporal trend to bulk iron uptake rates (Fig. 1f).

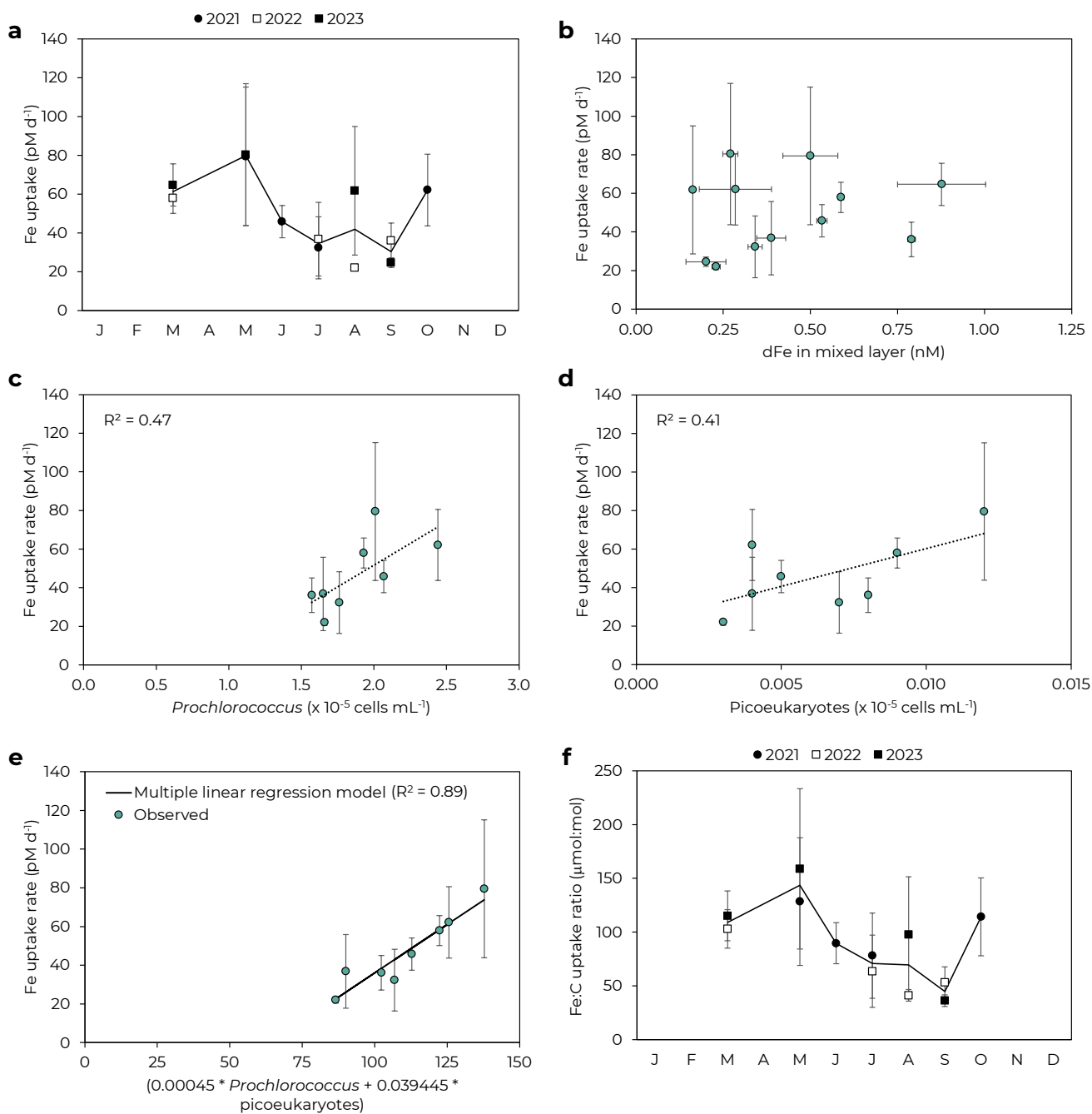


Figure 1: Drivers of Fe uptake. (a) Fe uptake rates at 25 m by month. Error bars represent 1 SD of triplicate incubations. Black line follows mean by month. (b) Fe uptake rate as a function of mixed layer dFe concentration. (c) *Prochlorococcus* and (d) picoeukaryotes at 25 m correlate with Fe uptake rates. Note these panels just include data for 2021-2022 for which flow cytometry data is available. (e) Multiple linear regression of Fe uptake rate as a function of *Prochlorococcus* and picoeukaryote abundance. (f) Fe:C uptake ratio by month.

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Iron uptake rates agreed well with previously reported rates in the subtropical North Pacific. Hawco et al. (2022) found a bulk Fe uptake rate of $48 \pm 13 \text{ pM d}^{-1}$ for June 2019 at Station ALOHA, in agreement with our summer uptake rate of $40 \pm 15 \text{ pM d}^{-1}$. This rate of $\sim 50 \text{ pM d}^{-1}$ was determined to be reasonable based on the kinetic constraints of biological Fe' uptake (Hawco et al., 2022). Similar uptake rates were observed for Fe addition experiments at Station ALOHA ($43 \pm 27 \text{ pM d}^{-1}$ for FeCl_3 , Bundy et al., 2025). The corresponding Fe:C uptake ratios (58 ± 17 and $68 \pm 43 \text{ } \mu\text{mol:mol}$ for Hawco et al. (2022) and Bundy et al. (2025), respectively) also agreed well with our inferred Fe:C ratios. Price et al. (1994) reported oligotrophic Fe uptake rates south of Hawai'i at 15°N , 152.5°W , with Fe additions of 0.06-2 nM used to calculate a maximum uptake rates based on Michaelis-Menten kinetics. Approximate conversions of their reported data to pM Fe d^{-1} units yield an estimated maximum uptake rate of $\sim 76 \text{ pM Fe d}^{-1}$, close to our maximum observed rate of 80 pM d^{-1} in May 2022 and 2023. The maximum Fe:C uptake ratio ($\sim 116 \text{ } \mu\text{mol:mol}$, Price et al., 1994) also fell within 1 SD of our observed Fe:C uptake ratios in May 2022 and 2023 (128 ± 60 and $159 \pm 74 \text{ } \mu\text{mol:mol}$, respectively).

3.2 Estimating biogenic pFe using three different approaches

Quantifying the amount of Fe in biomass is challenging, due in part to issues in quantifying living biomass in the ocean generally. Here, we assess three different approaches to calculating biogenic particulate Fe (pFe_{Bio}). First, following Sofen et al. (2023), we use particulate carbon (PC) and particulate phosphorus (PP), which likely include both living cellular and organic detrital material, although to different extents. For the final approach, we use adenosine-5'-triphosphate (ATP) an approach for quantifying living biomass due to the rapid hydrolysis of ATP in dead biomass (Holm-Hansen & Booth, 1966). The pFe_{Bio} estimated by these approaches is compared to the suspended $\text{pFe}_{\text{Labile}}$, defined using the Berger et al. (2008) labile leach, which has been shown to capture the biogenic pool (Rauschenberg & Twining, 2015). Thus, labile pFe represents an approximate maximum for pFe_{Bio} , if there were no contributions from nonliving labile pFe ($\text{pFe}_{\text{Nonliv}}$, including authigenic pFe).

For the first approach, $\text{pFe}_{\text{Bio_PC}}$ was calculated as:

$$\text{pFe}_{\text{Bio_PC}} = \text{Fe:C} \times \text{PC} , \quad (3)$$

where Fe:C is the molar ratio determined using the bulk iron uptake rates and the HOT primary productivity measurements at 25 m. $\text{pFe}_{\text{Bio_PP}}$ was calculated as:

$$\text{pFe}_{\text{Bio_PP}} = \text{Fe:C} \times \text{PP} \times \text{C:P}_{\text{Phyto}} , \quad (4)$$

where $\text{C:P}_{\text{Phyto}}$ is the ratio of C:P of phytoplankton at Station ALOHA. As the C:P of living biomass at Station ALOHA is not well constrained, we estimated $\text{C:P}_{\text{Phyto}}$ based on the mean of previously reported C:P ratios of *Prochlorococcus*, *Synechococcus*, and picoeukaryotes in the North Pacific Subtropical Gyre (186, 195, and 109 mol:mol, respectively; Lomas et al., 2021), weighted by the relative depth-integrated (0-125 m) carbon biomass of each group at Station ALOHA (mean: 69%, 3%, and 28%, respectively; Rii et al., 2016), resulting in a $\text{C:P}_{\text{Phyto}}$ approximation of 165 mol:mol.



The PC and PP approaches consistently overestimated pFe_{Bio} , which almost always exceeded measurements of pFe_{Labile} (Fig. 2). In the mixed layer, both approaches estimated pFe_{Bio} concentrations to be more than double the pFe_{Labile} concentration (mean: 220% and 256% of pFe_{Labile} , respectively). Both approaches continued to overestimate pFe_{Bio} throughout the 150 m euphotic zone, and, in the mesopelagic, reached concentrations equivalent to $78 \pm 16\%$ (PC) and $77 \pm 23\%$ (PP) of pFe_{Labile} (based on the mean profiles in Fig. 2b). This is likely still an overestimation, as pFe_{Labile} is expected to be dominated by authigenic pFe below the euphotic zone (Sofen et al., 2023).

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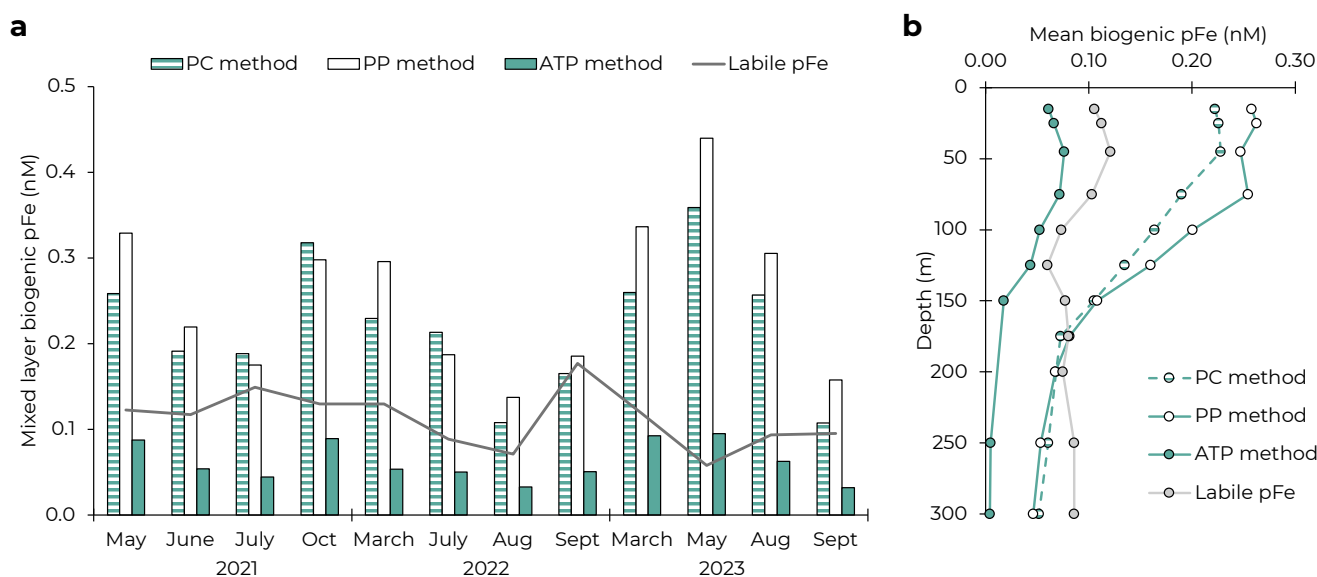


Figure 2: Comparison of three approaches to estimating biogenic particulate Fe using PC, PP, and ATP. (a) Average mixed layer concentrations of biogenic pFe on the 12 cruises. Grey line shows the mixed layer concentration of labile pFe, which should represent a maximum potential biogenic concentration if there were no authigenic contributions. (b) Mean upper ocean profiles of biogenic pFe by the three approaches and labile pFe.

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There are several potential explanations for the overestimation by the PC and PP approaches. First, total PC could include inorganic contributions from calcium carbonate. However, suspended particulate inorganic carbon is very low in oligotrophic gyres (approximately 6% of PC; Gordon, 1971; Hebel & Karl, 2001) and often considered negligible to the total carbon mass balance at Station ALOHA (Gordon, 1971; Karl et al., 2021). The discrepancy may also come from detrital organic matter, including dead cells and fecal pellets, which is estimated to comprise 58-74% of the surface PC pool at Station ALOHA (Henderikx-Freitas et al., 2021; Karl et al., 2022). Unlike organic carbon (as well as N and P), Fe in metalloproteins is not covalently bonded, and so may be released more easily after cell death as cell proteins become denatured. It appears that the Fe:C stoichiometry of this detrital organic carbon is significantly lower than in living cells. With the PP approach, there is also uncertainty in the C:P of living cells and how this ratio varies over time, which could distort the overestimation of pFe_{Bio} . While the stoichiometries of the different phytoplankton groups are expected to be

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flexible (Baer et al., 2017; Geider & La Roche, 2002; Karl et al., 2001) and the phytoplankton community composition varies temporally (Pasulka et al., 2013), we expect these complications to have a relatively small impact on our estimation of $p\text{Fe}_{\text{Bio}}$. Additionally, the calculated $\text{C:P}_{\text{Phyto}}$ used in Eq. 4 omits a major contribution from heterotrophic bacteria to bulk biomass (Karl et al., 2022) and therefore may not represent the true C:P of living biomass. For the mixed layer, we
200 calculated the maximum C:P that would not lead to an overestimation of biogenic pFe ($p\text{Fe}_{\text{Bio}} = p\text{Fe}_{\text{Labile}}$) to be 80 ± 36 mol:mol (average and 1 SD across cruises). This assumes the entire labile pFe pool is biogenic, which would conflict with previous conclusions about the importance of authigenic pFe (Sofen et al., 2023), requiring C:P to decrease further. Importantly, moderate phosphate depletion at Station ALOHA indicates that C:P in cells should exceed the 106:1 Redfield Ratio (Björkman et al., 2000; Karl et al., 2001). Thus, we conclude that the overestimations of $p\text{Fe}_{\text{Bio}}$ by Eq. 3 and 4 are due
205 to nonliving contributions to both the PC and PP inventories.

To avoid the contributions of nonliving organic matter to the PC and PP pools, we used particulate ATP as an estimate of living biomass. An essential molecule for energy generation in all organisms, ATP is turned over quickly via rapid release following cell death and does not easily adsorb to other particulate matter (Holm-Hansen & Booth, 1966). ATP is not commonly measured in the ocean, but has been measured regularly at Station ALOHA since 1989 (Karl et al., 2022),
210 providing a unique method for estimating Fe in living biomass. Previous work using ATP suggests only 26-42% of the surface particulate carbon pool at Station ALOHA is comprised of living biomass (Henderikx-Freitas et al., 2021; Karl et al., 2022). Applying ATP as a biomass indicator restricts the $p\text{Fe}_{\text{Bio}}$ estimate to the Fe quotas in living cells, excluding any biologically-derived Fe in dead cells or biogenic exudates. We calculated $p\text{Fe}_{\text{Bio_ATP}}$ as:

$$p\text{Fe}_{\text{Bio_ATP}} = \text{Fe:C} \times \left(\frac{\text{ATP} \times 250}{12.01} \right), \quad (5)$$

215 where 250 represents the ratio of living C to ATP in g g^{-1} (Karl, 1980). One important limitation for this approach is that the C:ATP ratio must be well constrained. At Station ALOHA, C:ATP has been reported to fall between 250-400 g g^{-1} (Christian & Karl, 1994; Karl, 1980). Recent work based on shipboard experiments and *Prochlorococcus* measurements supports the use of a 250 C:ATP ratio at Station ALOHA (Karl et al., 2022), in agreement with a variety of other field and laboratory studies of marine organisms (Bochdansky et al., 2021; Holm-Hansen, 1970, 1973).

220 The ATP approach produced $p\text{Fe}_{\text{Bio_ATP}}$ values that were consistently less than $p\text{Fe}_{\text{Labile}}$, both in the mixed layer and at depth, thereby satisfying our mass balance constraint on $p\text{Fe}_{\text{Bio}}$. Across the 12 cruises where Fe uptake data were available, $p\text{Fe}_{\text{Bio_ATP}}$ averaged $61 \pm 37\%$ of $p\text{Fe}_{\text{Labile}}$ in the mixed layer (Fig. 2a). Below 150 m, $p\text{Fe}_{\text{Bio_ATP}}$ comprised $<15\%$ of $p\text{Fe}_{\text{Labile}}$ (Fig. 2b).

For all three approaches, there are also limitations in estimating $p\text{Fe}_{\text{Bio}}$ from the calculation and application of the
225 Fe:C uptake ratio. First, there is uncertainty in the Fe:C uptake ratios based on variability in the replicate Fe uptake incubations (error bars in Fig. 1f). The mean replicate RSD from this timeseries was 30% or 27 $\mu\text{mol:mol}$. Additionally, our method of determining Fe:C is based on bulk uptake rates and includes an assumption that the turnover time of the FeDS addition is equivalent to the turnover time of the entire dFe pool (Hawco et al., 2022). The long pre-incubation time allows



for the added Fe to bind with strong Fe ligands routinely observed at Station ALOHA (Bundy et al., 2018; Fitzsimmons et al., 2015; Rue & Bruland, 1995), which makes this approximation more reasonable. In this study, we also added a control incubation to account for production of pFe from colloidal aggregation of Fe oxide precipitation, which yield a median uptake rate of 12.5 pM d⁻¹ (note that this does not account for 'heterogeneous' scavenging or aggregation of dFe onto pre-existing particles). Finally, to estimate pFe_{Bio} throughout the upper water column, the Fe:C uptake ratio at 25 m was assumed to reflect the Fe:C quota of biota regardless of depth, which is a problematic simplification. This ratio likely changes with depth due to changes in dFe concentration, community composition (Twining et al., 2021), and increased Fe quotas under light limitation (Hawco et al., 2022; Hogle et al., 2022; Raven, 1990). Despite these known mechanisms, direct comparisons of phytoplankton Fe quotas between the surface mixed layer and deep chlorophyll maximum (DCM) in the subtropical North Atlantic has not necessarily pointed to meaningful differences in Fe:C (Sofen et al., 2023).

3.3 Estimating nonliving labile pFe

Overall, the ATP approach provides the most reasonable pFe_{Bio} estimates of the three approaches assessed. We therefore used this method to further estimate pFe_{Nonliv} by subtraction, operationally defined by the equation:

$$pFe_{\text{Nonliv}} = pFe_{\text{Labile}} - pFe_{\text{Bio_ATP}} \quad (6)$$

For samples where the pFe_{Bio_ATP} exceeded pFe_{Labile} (n=12, 12% of samples), pFe_{Nonliv} was set to 0. As in Sofen et al. (2023), pFe_{Bio} samples were permitted to be up to 125% of pFe_{Labile} due the heterogeneity of oceanic particles and considering measurement uncertainties. Samples where pFe_{Bio} was greater than 125% of pFe_{Labile} (n=7, 7% of samples) were set to 1.25 times pFe_{Labile}. The majority of these samples were in March 2022 and May 2023, all within the euphotic zone.

There has been considerable interest recently in quantifying the authigenic pFe pool, which is likely generated by different mechanisms than pFe_{Bio} and may be exported with different efficiencies. While the ATP approach offers our most reasonable estimate of the authigenic pFe pool, it is important to consider that using this definition likely includes some particulate Fe outside of authigenic Fe oxyhydroxides precipitated *in situ* from dFe. First, this approach includes Fe in or adsorbed to nonliving organic particles, shown above to be a significant portion of the particulate C and P pools. Additionally, exogenous lithogenic particles from the atmosphere or advected from the Hawaiian Islands also contain a significant fraction of chemically labile Fe that is dissolved by the Berger et al. (2008) leach (~5-20% depending on the source; Shelley et al., 2018). Dust-derived contributions to the pFe_{Labile} pool would be seasonally variable, with peak dust deposition expected in spring (Parrington et al., 1983; Prospero et al., 2003). While the particles might contain amorphous Fe oxyhydroxides that are chemically identical to authigenic pFe, they were not formed *in situ* and so are not strictly authigenic. Regardless, the ATP approach allows for the separation of Fe in living biomass to understand how the behavior of this pool differs from the remaining, nonliving pFe_{Labile} pool.

3.4 Biogenic and nonliving particulate Fe distributions at Station ALOHA

260 The derived profiles of pFe_{Bio} and pFe_{Nonliv} differed substantially, with mean pFe_{Bio} higher in the upper 125 m and pFe_{Nonliv} dominating below 125 m (Fig. 3). The mean pFe_{Labile} profile across the 12 cruises was highest (0.10-0.12 nM) in the upper euphotic zone, lowest in the DCM (100-125 m), and relatively consistent with depth from 150-300 m at 0.07-0.09 nM. Particulate Fe_{Bio} was high throughout the upper euphotic zone (≥ 0.05 nM in the upper 100 m), with no minimum in the DCM. Below 150 m, pFe_{Bio} was < 0.01 nM in the upper mesopelagic, reflecting the steep decline in ATP. The structure of the average pFe_{Bio} profile with depth aligns with upper ocean profiles of *Prochlorococcus* and other phytoplankton at Station ALOHA (Campbell & Vaultot, 1993; van den Engh et al., 2017). In contrast, pFe_{Nonliv} was slightly elevated in the surface mixed layer (0.05 nM on average in the upper 45 m), with a local minimum near the DCM, and increased into the upper mesopelagic. A local minimum in pFe_{Nonliv} at the DCM was observed previously (Sofen et al., 2023), attributed to either the dissolution of authigenic pFe with the dFe minimum, or to increased export efficiency of authigenic pFe in the subsurface.

265 the average pFe_{Bio} profile with depth aligns with upper ocean profiles of *Prochlorococcus* and other phytoplankton at Station ALOHA (Campbell & Vaultot, 1993; van den Engh et al., 2017). In contrast, pFe_{Nonliv} was slightly elevated in the surface mixed layer (0.05 nM on average in the upper 45 m), with a local minimum near the DCM, and increased into the upper mesopelagic. A local minimum in pFe_{Nonliv} at the DCM was observed previously (Sofen et al., 2023), attributed to either the dissolution of authigenic pFe with the dFe minimum, or to increased export efficiency of authigenic pFe in the subsurface.

270 As in the North Atlantic, observations of colloidal Fe at Station ALOHA have also shown a minima around the DCM (Fitzsimmons et al., 2015), as does pFe_{Lith} . From these mean profiles, pFe_{Labile} is majority biogenic in the euphotic zone (59-72% in the upper 125 m), and majority nonliving below the DCM ($>90\%$ at 250-300 m; Fig. 3b). Based on the similarity between pFe_{Nonliv} and pFe_{Lith} , which should be chemically inert, we suggest that the pFe_{Nonliv} profile is primarily driven by particle aggregation and export processes.

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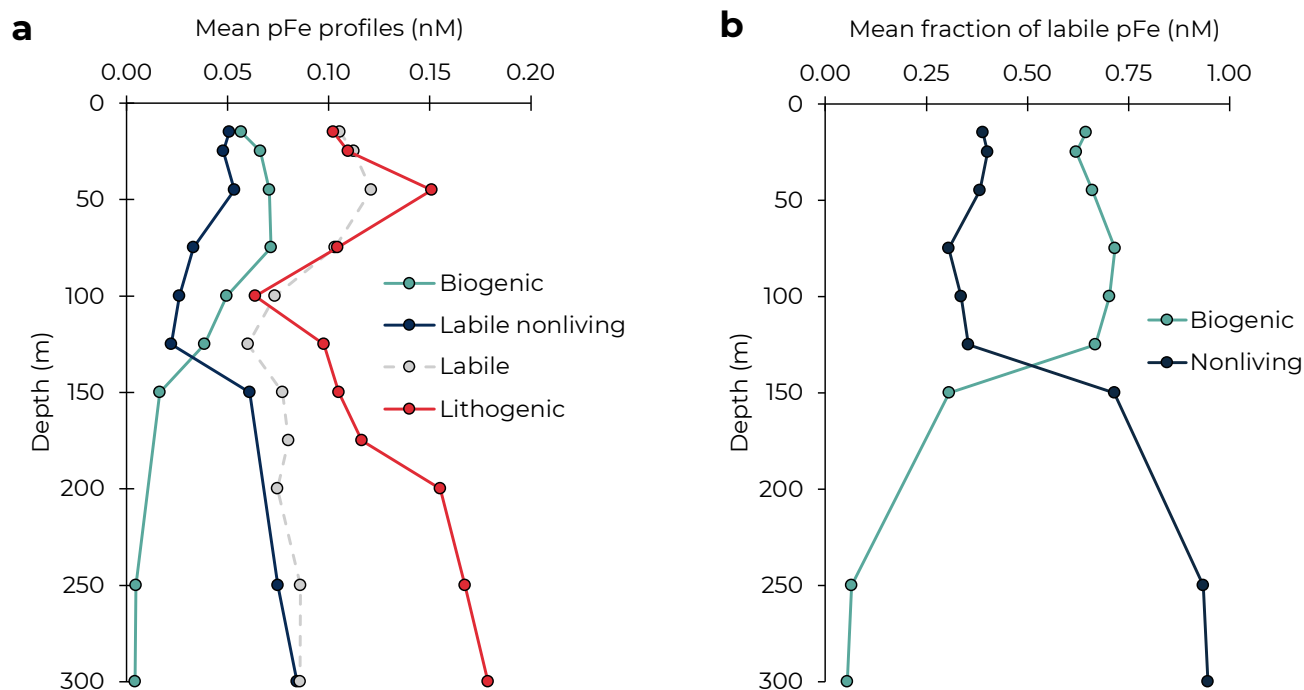
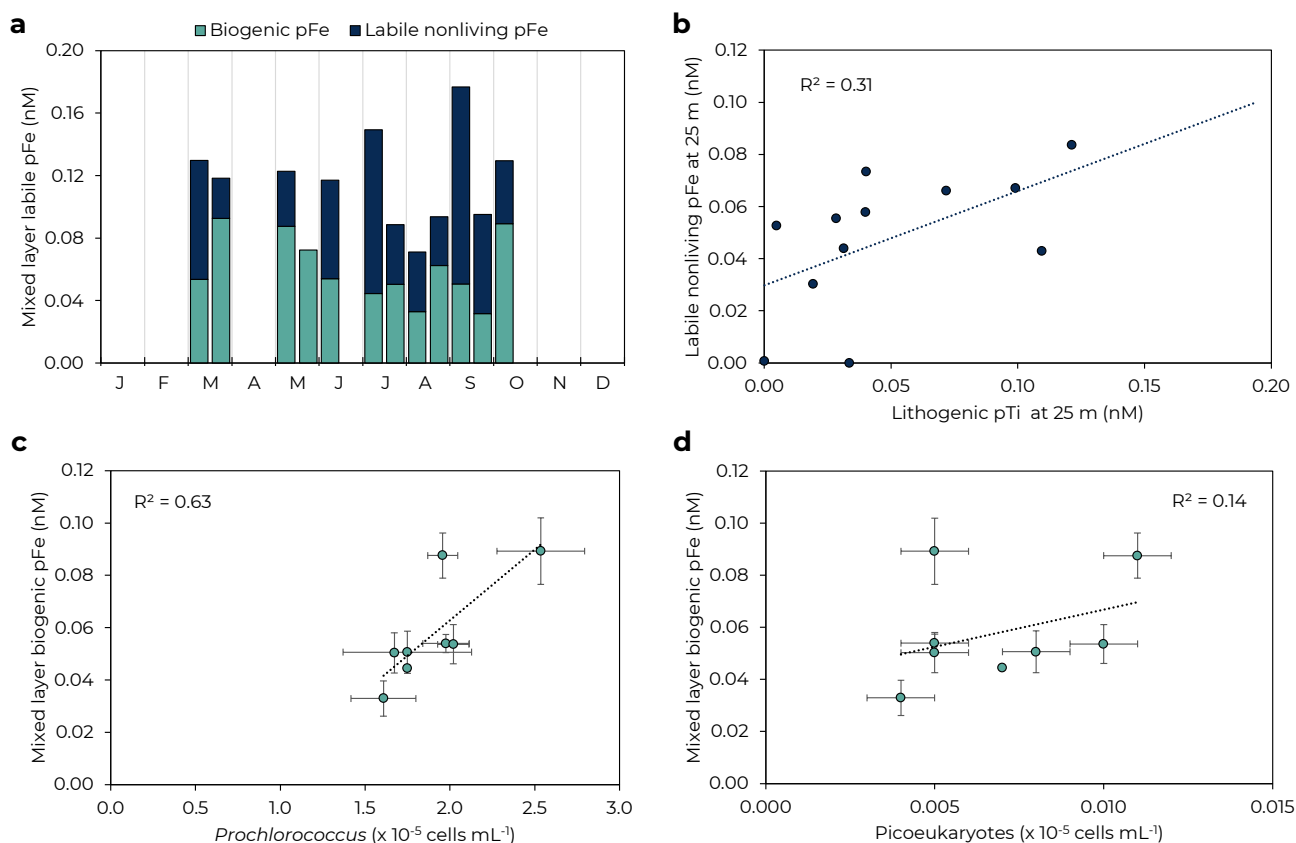




Figure 3: Mean profiles of particulate Fe phases (a). The fraction of labile pFe comprised of biogenic and nonliving pFe, averaged across cruises (b).

280 The seasonal cycle of pFe_{Bio} in the mixed layer was similar to bulk Fe uptake rates, with generally higher concentrations in the spring (ranging 0.06-0.09 nM) and lower concentrations in summer (ranging 0.04-0.06 nM, Fig. 4). pFe_{Bio} also appears to be driven by *Prochlorococcus* and picoeukaryotes abundances ($R^2 = 0.63$ and 0.14, respectively). The correlation between picoeukaryotes and pFe_{Bio} is marred by an outlier in October 2021, when *Prochlorococcus* was 30% greater than the mean abundance and peak pFe_{Bio} occurred.



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Figure 4: Seasonal changes in mixed layer biogenic and nonliving labile particulate Fe. (a) Biogenic (green) and authigenic (navy) pFe by month. (b) Correlation of lithogenic particulate Ti and nonliving labile pFe at 25 m. (c) *Prochlorococcus* and (d) picoeukaryotes in the mixed layer correlate with mixed layer biogenic pFe. Error bars represent 1 SD of samples in the mixed layer. Note panels c-d just include data for 2021-2022; flow cytometry data is not available for 2023.

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Mean pFe_{Nonliv} in the mixed layer overall varied over a larger range than pFe_{Bio} , from 0–0.13 nM, but with no clear seasonal pattern. Mixed layer pFe_{Nonliv} was highest in July 2021 and September 2022 and lowest in March and May 2023. A weak positive correlation with lithogenic particulate Ti at 25 m ($R^2 = 0.31$, Fig. 4b) suggests input of lithogenic material may play a role in hosting authigenic Fe oxyhydroxides, perhaps via scavenging and/or colloidal aggregation, but this correlation

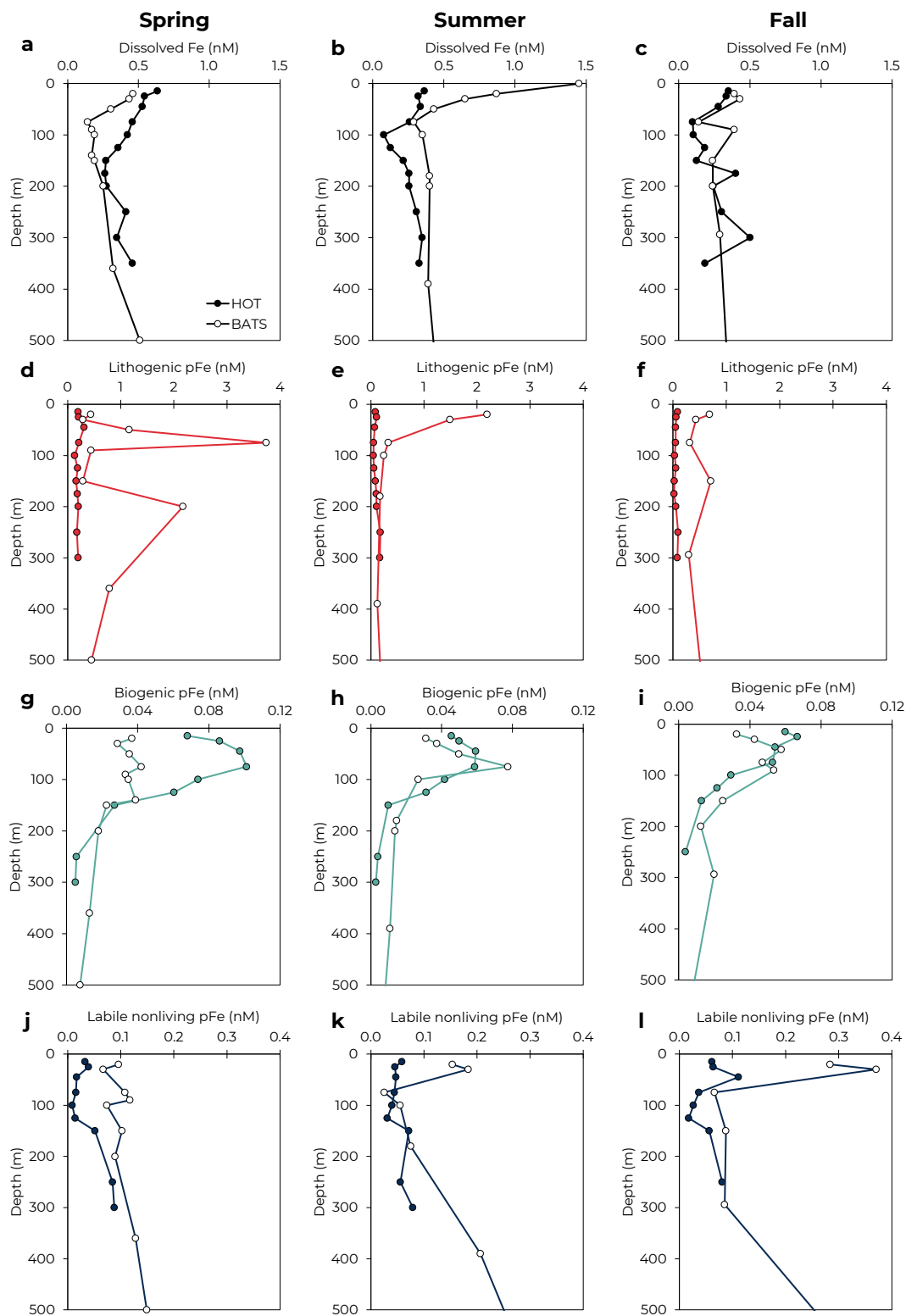


could also be explained by lateral advection of Ti-rich basaltic material containing Fe oxyhydroxides. Nonliving labile pFe
295 was not well correlated with any major phytoplankton groups ($R^2 = 0.04, 0.12, \text{ and } 0.01$ for *Prochlorococcus*,
Synechococcus, and picoeukaryotes, respectively; Table S3 in Supplement), suggesting that variations of pFe_{Nonliv} are not
primarily driven by dead biomass or extracellular Fe precipitates. The concentration of colloidal Fe is thought to be
important in driving production of authigenic pFe minerals (Kunde et al., 2019; Tagliabue et al., 2023). While colloidal Fe
was not measured during our timeseries, the colloidal fraction has been shown to drive dFe variability at Station ALOHA
300 (comprising $64\% \pm 18\%$ of dFe pool; Bergquist et al., 2007; Fitzsimmons et al., 2015; Wu et al., 2001) and so colloidal-
driven production should be traceable by dFe. However, mixed layer dFe and pFe_{Nonliv} were not strongly correlated ($R^2 =$
0.15), suggesting dFe variability (and thus colloidal Fe variability) is not the primary driver of pFe_{Nonliv} variability.

3.5 Comparison to the North Atlantic Subtropical Gyre

As one of the only other open ocean sites with seasonally resolved Fe data, the Bermuda Atlantic Time-series Study
305 (BATS) site in the subtropical North Atlantic serves as a natural comparison to Station ALOHA. While both sites are
oligotrophic with similar annual net community production (Emerson, 2014), the BATS site is expected to receive an order
of magnitude more annual lithogenic dust deposition (Hayes et al., 2017; Kok et al., 2021), indicating the potential for major
differences in Fe cycling. Fortunately, the Fe cycle at BATS has been extensively explored as part of the recent ‘BAIT’
project (Sedwick et al., 2023; Sofen et al., 2023; Tagliabue et al., 2023). A comparison of Fe profiles across seasons
310 highlights the similarities and differences between the sites (Fig. 5). At both sites, dissolved Fe was present at a similar
magnitude, with a similar variability across the seasonal cycle. Mean dFe concentrations in the upper 100 m ranged from
0.28 nM in spring to 0.67 nM in fall at BATS (Sedwick et al., 2023) and from 0.23 nM in fall to 0.64 nM in winter at Station
ALOHA (Bates & Hawco, 2025). As expected, pFe_{Lith} was notably higher at BATS, particularly in August (> 1 nM in the
surface, Fig. 5e) when Saharan dust deposition was highest (Sedwick et al., 2023; Sofen et al., 2023).

315





320 **Figure 5: Comparison of seasonal Fe dynamics between HOT (North Pacific Subtropical Gyre; filled) and BATS (North Atlantic Subtropical Gyre; unfilled). Mean profiles of dissolved Fe (a-c), lithogenic pFe (d-f), biogenic pFe (g-i), and labile nonliving pFe (g-l) in the upper 500 m are shown for spring, summer, and fall. Winter not shown due to little to no sampling coverage for most parameters. BATS data from Sedwick et al. (2023) and Sofen et al. (2023), where pFe_{Auth} is used here labeled as pFe_{Nonliv} .**

Within the labile pFe pool, the contributions of biogenic and nonliving pFe differed substantially between the sites. At BATS, the pFe_{Bio} was consistently lower than pFe_{Nonliv} , with the pFe_{Labile} pool ranging 26-36% biogenic by season above the DCM (Sofen et al., 2023). In contrast, pFe_{Bio} contributions at Station ALOHA were higher, comprising 47-84% of pFe_{Labile} above the DCM by season. Overall, pFe_{Bio} concentrations were similar between the sites, with seasonally averaged
325 concentrations in the upper 100 m ranging 0.03-0.10 nM at HOT and 0.03-0.08 nM at BATS (Fig. 5g-i), in agreement with global estimates for biotic iron pools (Boyd et al., 2015; Sofen et al., 2023). Additionally, BATS and Station ALOHA have similar magnitudes of phytoplankton biomass (Selph et al., 2022), which has a large influence on pFe_{Bio} (Fig. 4), although community composition varies somewhat between the sites (DuRand et al., 2001; Pasulka et al., 2013). Note that the methods of estimating pFe_{Bio} differ between the two sites. Sofen et al. (2023) estimated pFe_{Bio} using Eq. 4 and cellular Fe
330 quotas of eukaryotes determined by synchrotron X-ray fluorescence. This approach includes detrital organic material as part of pFe_{Bio} and assumes eukaryotic Fe quotas are representative of the microbial pool (Sofen et al., 2023). Given the much higher concentration of pFe_{Labile} at BATS, the possible overestimation of pFe_{Bio} , as shown here for Station ALOHA, is less impactful. Much higher pFe_{Nonliv} is evident at BATS, particularly in the surface mixed layer (Fig. 5j-l). Increased pFe_{Nonliv} at BATS could be due, in part, to the increased pFe_{Lith} . Lithogenic particles scavenge colloidal Fe, with scavenging expected to
335 increase with higher Fe (Boyd et al., 2010; Ye & Völker, 2017). Additionally, pFe_{Lith} may also trace input of exogenous labile pFe from dust, which would be counted in the pFe_{Nonliv} pool. Similar to our analysis at Station ALOHA, the higher deposition of lithogenic dust at BATS could drive greater *authigenic* pFe formation, or greater contributions of *exogenous* labile pFe, which does not fully integrate with the dFe pool. This distinction has important implications for the residence time of Fe in the ocean, but future research is needed to separate these possibilities conclusively.

340 4 Conclusions

Understanding the composition of the labile particulate Fe pool in the upper ocean is essential for unraveling the complexities of marine Fe cycling. This study highlights biotic uptake by *Prochlorococcus* and other phytoplankton as drivers of the pFe_{Bio} pool. The pFe_{Nonliv} pool, potentially including contributions from detrital organic matter and dust in addition to authigenic minerals, instead appeared to be correlated with lithogenic material. This relationship suggests that
345 pFe_{Nonliv} may be driven by either scavenging and the precipitation of Fe oxides coatings onto lithogenic material or by direct contributions from dust. Clearly, pFe_{Nonliv} is an important contributor to the pFe_{Labile} pool at Station ALOHA, comprising ~40% in the euphotic zone and >90% below the euphotic zone. As dFe and pFe_{Bio} are transferred into the pFe_{Nonliv} pool, potentially via authigenic mineral formation, more efficient recycling of pFe_{Bio} may be necessary to meet biotic Fe demand. Indeed, the greater portion of pFe_{Bio} at Station ALOHA compared to the North Atlantic may be key to sustaining high rates



350 of Fe recycling previously reported for the North Pacific Subtropical Gyre. However, the actual composition of the $p\text{Fe}_{\text{Nonliv}}$ pool remains elusive. Direct chemical and mineralogical characterization of the operationally-defined $p\text{Fe}_{\text{Nonliv}}$ pool is needed to assess its reactivity, bioavailability, and origin. Continued investigation into the composition and transformations of $p\text{Fe}_{\text{Nonliv}}$ is needed to further our understanding of its role in marine Fe cycling.

Data availability

355 Dissolved, labile particulate, and recalcitrant particulate Fe data is available on BCO-DMO (Hawco & Bates, 2025a, 2025b, 2025c) at <https://doi.org/10.26008/1912/bco-dmo.962986.2>, <https://doi.org/10.26008/1912/bco-dmo.962966.1>, and <https://doi.org/10.26008/1912/bco-dmo.962821.1>. Fe uptake data has been submitted to the Biological and Chemical Oceanography Data Management Office (BCO-DMO). Hawaii Ocean Time-series data were obtained via the Hawaii Ocean Time-series HOT-DOGS application; University of Hawai‘i at Mānoa; National Science Foundation award #2241005 (HOT-DOGS, 2025).

Author contributions

NJH and ESB conceptualized the study. ESB performed the experiments and analyzed the samples. Both authors analyzed the data and wrote the paper.

Competing interests

365 The authors declare that they have no conflict of interest.

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