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BGD

8, 11489–11527, 2011

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Abstract

Due to its importance as a limiting nutrient for phytoplankton growth in large regions of the world's oceans, ocean water column observations of concentration of the trace-metal iron (Fe) have increased markedly over recent decades. Here we compile > 13 000 global measurements of dissolved Fe (dFe) and make this available to the community. We then conduct a synthesis study focussed on the Southern Ocean, where dFe plays a fundamental role in governing the carbon cycle, using four regions, six basins and five depth intervals as a framework. Our analysis reveals the importance of biological activity and dFe inputs in governing the inter-region and inter-basin differences in surface dFe, respectively. In deep waters, the major controls of inter-region and inter-basin dFe variability are ligand distributions and deep dFe inputs or water mass characteristics, respectively. We find that even in regions where many dFe measurements exist, the processes governing the seasonal evolution of dFe remain enigmatic, suggesting that, aside from broad sub-Antarctic–Antarctic trends, biological activity might not be the major driver of dFe variability. Nevertheless, missing measurements during key seasonal transitions make it difficult to better quantify and understand surface water replenishment processes and the seasonal Fe cycle. Statistical differences exist in the measured dFe between measurements taken over the period 1989–2002 and 2003–2008, which may reflect progress in clean sampling and analysis techniques. Finally, we detail the degree of seasonal coverage by region, basin and depth. By synthesising prior measurements we suggest a role for different processes and highlight key gaps in understanding, which we hope can help structure future research efforts in the Southern Ocean.

1 Introduction

Since the advent of trace metal clean techniques in the late 1970s/early 1980s (e.g., Bruland et al., 1979), the role of iron (Fe) as a key micronutrient that regulates phytoplankton growth rates, primary production and the biological carbon pump in the

BGD

8, 11489–11527, 2011

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



so-called “High Nutrient Low Chlorophyll” regions of the world’s Oceans is well established (e.g., De Baar et al., 2005; Boyd et al., 2007). Of these regions, most attention has focussed on the Southern Ocean since it is typified by the largest residual stocks of unused macronutrients and plays an important role in governing the global air-sea CO₂ balance due to the extensive production of dense deep waters (Caldeira and Duffy, 2000). Indeed, the “Iron Hypothesis” of John Martin (Martin, 1990) posits that the glacial interglacial cycles of atmospheric CO₂ recorded in ice cores could have been controlled by changes in the supply of Fe to the Southern Ocean from aeolian deposition. Phytoplankton production in the modern Southern Ocean is clearly controlled to some degree by Fe, with the highest rates of productivity found alongside known Fe sources such as islands, the Antarctic continental shelf and frontal regions (e.g., Arrigo et al., 2008).

The cycling of Fe in seawater is complicated, as compared to other macronutrients, and its distribution is controlled by a variety of chemical, physical and biological processes. At the surface, Fe is consumed due to biological uptake, but also precipitation and scavenging by particles. Processes such as photochemistry, redox chemistry and dissolution are important drivers of its chemical speciation between soluble (usually < 0.02 μm), colloidal (0.02–0.2 μm) and particle (> 0.2 μm) fractions (e.g., Bowie and Lohan, 2009). A key component governing the cycling and distributions of dissolved Fe (dFe, < 0.2 μm) in the ocean is the role of Fe binding organic ligands. Such ligands complex Fe in the soluble and colloidal fractions (e.g., Wu et al., 2001; Boye et al., 2010) and reduce losses due to precipitation/scavenging, thereby increasing the residence time of dFe in the ocean. These processes result in dFe having a vertical profile that is typical of a nutrient and a scavenged element (Boyd and Ellwood, 2010). Predominant Fe sources to the Southern Ocean are associated with dust deposition close to continents (e.g., Gaiero et al., 2003; Tagliabue et al., 2009), shallow continental margins (e.g., Blain et al., 2007; Bowie et al., 2009; Pollard et al., 2009; Tagliabue et al., 2009), sea ice melting (e.g., Lannuzel et al., 2008; Van der Merwe et al., 2011) and, in the deep ocean, hydrothermal vent systems (e.g., Tagliabue et al., 2010).

Southern Ocean Iron Distributions

A. Tagliabue et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

I◀

▶I

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Due to its acknowledged importance as a limiting nutrient, measurements of dFe in the ocean, and in the Southern Ocean in particular, have increased markedly over recent years. Advancements in measurement techniques and inter-laboratory comparisons and evaluation projects (e.g., Bowie et al., 2004; Johnson et al., 2007) have been of major importance. The first compilation of dFe data (354 observations at 30 stations) was performed by Johnson et al. (1997) and then expanded by Parekh et al. (2005), which aided in understanding the importance of scavenging of dFe in ocean models. More recently, Moore and Braucher (2008) published an expanded database of 6540 observations to aid in the evaluation of their global ocean model. They found that surface dFe concentrations were bi-modal and decoupled from major inputs, highlighted the importance of the removal of dFe at low concentrations in subsurface waters and noted the importance of non-aeolian Fe sources (Moore and Braucher, 2008). In recent years, the number of dFe observations has increased still further, though international programs such as GEOTRACES (www.geotraces.org), especially in regions where sampling was difficult such as the Southern Ocean and the deep ocean. These newer observations in the deep Southern Ocean were recently used to demonstrate the importance of hydrothermal Fe sources in governing the ocean's deep water dFe inventory (Tagliabue et al., 2010). Thus, accessible databases of dFe data compilations are useful for evaluating ocean models, examining the importance of different Fe cycle processes and demonstrating the importance of different Fe sources to the ocean.

In this paper, we present a new compilation of > 13 000 measurements of dFe in the global ocean for use by the community. We focus on a synthesis and statistical summary of the > 3000 measurements now available for the Southern Ocean by examining the regional distribution of observations at different depths in this important ocean region. In doing so, we can examine where observations are distributed, what we can learn from them and where future observational efforts are needed. Moreover, we use our database to examine the role of different processes in governing the variability in dFe between different ocean basins and regions, at different ocean depths, as well as investigating seasonal trends in well sampled locations.

2 Methodology

Building on the dFe dataset collected by Moore and Braucher (2008), which contained 6540 individual observations, we compiled an additional 6585 dFe observations (mostly from recent campaigns) to arrive a total of 13 125 global observations with collection dates that span 30 yr from 1978 to 2008. The observations used filter pore sizes ranging from 0.2–0.45 μm as an operational cut-off for dFe and are often means from duplicate and triplicate samples. Obviously a wide range of sampling, processing and analytical techniques have been employed over this 30 yr period (see review of Achterberg et al., 2001). The data was mostly collected by literature review and manual notation of data and submission of data from investigators.

Our focus here is to synthesise the measurements taken in the Southern Ocean in greater depth, since dFe plays a critical role in governing the ocean carbon cycle in this region. To this end, we applied a latitudinal cut off at 40° S, which approximately corresponds to the Sub-Tropical Front (STF) and results in 3332 “Southern Ocean” observations. To further regionalise this variable ocean, we sub-divided these 3332 observations using a variety of different criteria. We firstly separated the “Shelf” region, since this is often typified by high rates of dFe input (bottom depth < 2000 m) from the “Offshelf” region. The “Offshelf” waters were further subdivided into “Antarctic” and “sub-Antarctic” using the northern branch of the Polar Front (nPF) as a meridional cut-off. The nPF has a highly variable position as a function of longitude, so we derived the nPF using maps of absolute dynamic topography, which is the sum of satellite altimetry anomaly data and a mean dynamic topography (Rio and Hernandez, 2004). The nPF is defined using a constant isoline of sea surface height, using a technique first described by Sokolov and Rintoul (2007) (for more details see Swart and Speich, 2010). By analysing the timeseries of nPF positions between 1998–2008 we determined that the temporal variability in the nPF at each point in longitude, relative to the mean position used here, to be only $0.72^\circ \pm 0.35$. The longitude and latitude of each dFe observation in “Offshelf” waters was therefore examined as to whether it

BGD

8, 11489–11527, 2011

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



was north (defined as “sub-Antarctic”) or south (defined as “Antarctic”) of the nPF at that particular longitude. The major ocean basins were defined as the Atlantic (ATL, 65° W to 20° E), the Indian (IND, 20° E to 180°) and the Pacific (PAC, 180° to 65° W) all within the “Offshore” region. In total we used 4 regions (“Shelf”, “Offshelf”, “Antarctic” and “sub-Antarctic”) and 6 basins (ATL-Antarctic, ATL-subAntarctic, IND-Antarctic, IND-subAntarctic, PAC-Antarctic, PAC-subAntarctic). Within each of these regions and basins, we further subdivided the data into 5 depth ranges (0–100 m, 100–500 m, 500–1000 m, 1000–2000 m and 2000–6000 m). The distribution of the data, as well as the regional break down is shown in Fig. 1. In our analysis, we were interested in examining the variability in dFe (mean, standard deviation etc), the number of observations and for which months of the year observations are available.

3 Results

3.1 dFe distributions between regions and basins

Mean dFe concentrations versus depth (the mid-point of the particular depth range) for both ocean regions and geographic ocean basins are compiled to examine the variability in dFe profiles within regions and basins (Fig. 2a,b). Throughout, the distribution of mean dFe concentration generally behaves like other nutrients/scavenged elements, with low surface water concentrations (0.1–0.5 nM, aside from the Shelf region), due to biological uptake that persists below the mixed layer due to scavenging, and increased concentrations at depth (> 0.4 nM) due to remineralization (Fig. 2a,b, Tables 1 and 2). Nevertheless, there is a great deal of variability between our 4 regions and 6 basins.

Turning first to the different ocean regions (Fig. 2a), the Shelf region has a high mean dFe concentration throughout the water column, with surface water enrichment of 0.61 ± 1.14 nM ($n = 382$), an intermediate water minima of 0.60 ± 0.35 nM ($n = 31$) and deep water values of 0.53 ± 0.17 nM ($n = 20$). Note that data points below 2 km were removed by our shelf adjustment/definition. At the other extreme, the “Offshelf”

BGD

8, 11489–11527, 2011

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



data shows a much clearer nutrient/scavenged element profile, with a surface water minima of 0.31 ± 0.45 ($n = 999$) that increases with depth to 0.54 ± 0.26 nM ($n = 301$), which is slightly less than the reported concentration of Fe-binding ligands in Southern Ocean deep waters (e.g., ~ 0.6 – 1 nM, Boye et al., 2010; Ibisamni et al., 2011; Thuroczy et al., 2011). From within this “Offshelf” dataset, the “sub-Antarctic” observations are lower at the surface (0.23 ± 0.27 nM, $n = 426$, suggestive of either biological uptake or lesser inputs), but much higher in deep waters (0.64 ± 0.31 nM, $n = 71$) and therefore have a much steeper gradient between surface and deep values. On the other hand, the “Antarctic” region has higher surface concentrations (0.38 ± 0.55 nM, $n = 573$), alongside lower deep water values (0.51 ± 0.24 nM, $n = 230$) and thus a much flatter profile. The variability in surface values might reflect differences in the degree of productivity, Fe inputs, inter-annual variability or how much of the seasonal cycle has been measured (see Sect. 3.2), whereas deep water values may reflect regional variability in deep water ligand concentrations or different deep water Fe sources (e.g., hydrothermal vents, Tagliabue et al., 2010). It is noteworthy that at intermediate depths (for both the 500–1000 and 1000–2000 m depth ranges, Fig. 2a, Table 1) there is no noticeable difference in the mean dFe concentrations for the Antarctic and sub-Antarctic regions.

The dFe measurements show a great deal of intra- and inter-basin variability at all depths across our 6 Southern Ocean basins (Fig. 2b, Table 2). Surface concentrations in the PAC-Antarctic and PAC-subAntarctic are the lowest we found and very similar for these two regions (0.15 ± 0.08 nM, $n = 141$ and 0.14 ± 0.10 nM, $n = 45$, respectively). Outside of the Pacific, the regional variability noted in Fig. 2a is retained. Both ATL-subAntarctic and IND-subAntarctic are lower than ATL-Antarctic and IND-Antarctic regions, but within each region, the ATL basin is consistently higher than the IND at surface (0.30 ± 0.55 nM, $n = 58$ and 0.47 ± 0.69 nM, $n = 226$ for ATL-subAntarctic and Antarctic, respectively; 0.23 ± 0.20 , $n = 323$ and 0.43 ± 0.51 , $n = 206$ for IND-subAntarctic and Antarctic, respectively, Fig. 2b, Table 2). Again, this could reflect differences in biological activity, surface Fe inputs, or the degree to which the seasonal

5 cycle has been captured in the observations. Conversely, at the deeper depths, the IND basin is always greater than the ATL basin for both Antarctic and sub-Antarctic regions (Fig. 2b, Table 2). ATL basin intermediate water (500–2000 m) has greatly elevated dFe concentrations ($\sim 0.4\text{--}0.5$ nM) for both Antarctic and sub-Antarctic regions, relative
10 to IND and PAC basins. Overall, this leads to much steeper dFe profiles in the IND basin because of both lower surface concentrations and higher deep concentrations (relative to the corresponding region of the ATL basin). In the PAC basin, dFe concentrations increase almost linearly up to a value of 0.48 ± 0.12 nM ($n = 2$) for subAntarctic and 0.36 ± 0.05 nM ($n = 3$) for Antarctic in the 1000–2000 m depth bin (unfortunately
15 there is no PAC data deeper than 2000 m). At intermediate depth (500–2000 m) the PAC basin shows similar dFe concentrations to those measured in the IND basin for both the Antarctic and sub-Antarctic regions. There is a surface enrichment in IND-Antarctic that is robust in the 0–100 and 100–500 m depth (Fig. 2b) that may be related to the influence of the Antarctic continental shelf sources on dFe at stations with bottom
20 depths > 2000 m (e.g., Sedwick et al., 2008). It is also noteworthy that despite a large number of samples, the ATL-Antarctic shows a very flat profile with little difference in dFe from surface (0.47 ± 0.69 nM, $n = 226$, 0–100 m depth) to deep (0.49 ± 0.21 nM, $n = 177$, 2000–6000 m depth).

25 Since we are not considering replicates of identical samples, the standard deviation of the mean dFe calculated for a particular region or basin is actually better viewed as a metric of the degree of variability in measured dFe therein. For the different ocean regions (Fig. 2c), the standard deviation typically decreases with depth and is greatest in Shelf surface waters (> 1 nM). This suggests a greater degree of variability in surface dFe, probably driven by seasonal and interannual trends, and more stable dFe concentrations in deeper waters. Between the sub-Antarctic and Antarctic regions, the differences in standard deviation mirror those seen for the dFe concentration (compare Fig. 2c with a). Examining the ocean basins, there is more variability at the surface, relative to intermediate waters, but variability increases again between 2000–6000 m (Fig. 2d). Within this pattern, the IND-subAntarctic shows a consistently lower standard

Southern Ocean Iron Distributions

A. Tagliabue et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

deviation than the ATL basin and the IND-Antarctic and both regions of the PAC basin (which had the lowest dFe concentrations) have very low standard deviations. A high standard deviation (seen as the degree of variability) in a particular region/basin or depth range could result from seasonal or inter-annual variability in dFe or incomplete data coverage where “extreme” observations (e.g., those close to hydrothermal Fe sources, Klunder et al., 2011) have a disproportionate weight.

Figure 3(a,b) presents a synthesis of the statistical variability in dFe for the different ocean basins and regions and as such combines the information present in the different panels of Fig. 2. They highlight regions where dFe is high, but also highly variable (e.g., the Shelf between 0–100 m, Fig. 3a). On the other hand, the basin by basin breakdown shows basins where dFe is low, but varies little (such as the PAC-subAntarctic, 0–100 m) or those where dFe can be high, but with large variability (e.g., IND-subAntarctic, 2000–6000 m).

3.2 Frequency of sampling for dFe

3.2.1 Number of observations

Tables 1 and 2 summarises the sampling of dFe and presents the number of unique months and number of dFe observations with respect to depth for the 4 regions and 6 ocean basins, respectively. Throughout, all parameters (number of months and observations) decrease with increasing depth regardless of the particular region or basin considered.

In more detail, the Shelf region has the lowest number of observations (Table 1), with a maximum of 382 in the upper 100 m. That said, this is still a substantial number of measurements given the relatively small size of this region and results from a long history of repeated field campaigns to particular shelf regions such as the Ross Sea and the Kerguelen plateau. Considering the Offshelf waters, there are up to ~ 1000 observations at the surface and, in general, there are more observations in the Antarctic region with 573, relative to the sub-Antarctic region with 426 (apart from

BGD

8, 11489–11527, 2011

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



the 1000–2000 m depth interval). In the deep ocean (2000–6000 m), there are around 300 measurements in total for Offshelf region, but almost all are concentrated in the Antarctic region (230).

Unsurprisingly, the different ocean basins also display a decreasing number of observations with respect to depth within a specific sector (Table 2). As seen previously, the Antarctic region of a particular basin is generally more sampled than the corresponding sub-Antarctic region. An exception to this is the IND basin, where its sub-Antarctic region is better sampled than its Antarctic counterpart. Within the sub-Antarctic region upper waters (0–500 m), the PAC basin has less observations (85) than the ATL basin (116), which has less observations than the IND basin (603). For the Antarctic upper waters (0–500 m), the PAC basin is still the lowest (225, but a factor ~ 3 greater than the PAC-subAntarctic), but now the ATL basin has more observations (492, a factor > 4 greater than the ATL-subAntarctic) than the IND basin (353, which is a factor ~ 2 less than the IND-subAntarctic). At depths greater than 500 m, the relatively high degree of sampling in the IND-subAntarctic decreases and the ATL-Antarctic becomes the most sampled basin (Table 2). Unfortunately, there are no dFe observations deeper than 2000 m in the both the Antarctic and sub-Antarctic regions of the PAC basin.

3.2.2 Seasonal measurements

In terms of the seasonal coverage, dFe observations are generally concentrated in the period October to April at best (the Austral spring/summer), although there is one set of winter time observations in July (Tables 1 and 2). On the Shelf, observations are available between October to April coverage down to 1000 m, deeper than which observations are only available from February to April. There is a similar degree of coverage in Offshelf waters, apart from the sub-Antarctic region, which has the only winter dFe observations (in July, Ellwood, 2008). Unlike the Shelf, October sampling is absent below 500 m in all Offshelf waters, but July data is present to depths of 2000 m. At depths > 2000 m, there is only coverage from December to April in the Offshelf waters of the Antarctic region, whereas January is absent from the sub-Antarctic region. In

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



the IND-subAntarctic, there is complete coverage from October to April down to 500 m, below which certain months disappear with only December and April present deeper than 2000 m. For the IND-Antarctic, March, as well as the winter data in July, is missing from the October to April coverage at the surface and there is less seasonal coverage than the IND-subAntarctic in intermediate waters. That said, the IND-Antarctic does have data for January in addition to that for December and April present in the IND-subAntarctic basin in the 2000–6000 m depth range. There is lower seasonal coverage throughout the water column of the ATL-subAntarctic, only 5 months (October, December, February, March and April) have been sampled in the upper 100 m and only February, March and April remain below 500 m depth. Conversely, there is more seasonal coverage in the ATL-Antarctic, with 6 months sampled in the upper 500 m (December is missing from the October–April period) and below 500 m only February, March and April has been sampled. Despite the lowest number of total observations (45), the PAC-subAntarctic data still manages to cover 5 months between October–April at the surface (October and February are missing), although the number of months sampled decreases dramatically with depth (in parallel to the decreasing total number of observations). A similar pattern is found in the PAC-Antarctic, with also 5 months sampled at the surface (February and March are missing on this occasion) and a sharp decrease in the number of months sampled in subsurface waters. Overall, despite the great effort made over past decades, knowledge of the dFe distribution (even in surface waters) is completely lacking for the months of May, June, August and September (i.e., the autumn–winter and winter–spring transitions), which may be crucial in understanding the seasonal replenishment and depletion of this important limiting nutrient.

3.3 Case studies

3.3.1 SR3

Some relatively constrained areas of the Southern Ocean have seen extended efforts of sampling over many years. The two best examples are the SR3 transect south of

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Tasmania (between 1994–2008) and the South-Western Ross Sea (between 1990–2006) on the Antarctic continental shelf. For surface waters (0–100 m) there are a total of 294 and 240 observations in our defined “SR3” and “Ross” sectors, respectively (see Fig. 4 legend). Because of the differences between the Antarctic and sub-Antarctic regions, we further split the “SR3” sector using the mean position of the PF there (see Fig. 1), leaving 189 and 105 total observations for the sub-Antarctic-SR3 and Antarctic-SR3, respectively. The seasonal cycle of surface waters dFe (on a monthly basis) from these three regions (SR3-subAnt, SR3-Ant and ROSS) was then extracted and is presented in Fig. 4. We note that the dFe data from Lai et al. (2008) had to be excluded from the SR3 analysis (but retained for our larger scale synthesis) since their dataset was strongly significantly different from other observations from January, February and March (Wilcoxon test, $p < 0.0001$) and thus biased the monthly dFe during the Austral summer to too high a value.

Climatological satellite derived weekly chlorophyll-*a* for the identical locations in the SR3-subAntarctic and Antarctic sectors for which we have dFe measurements (Fig. 5a,b) show that Chl-*a* increases gradually in the SR3-subAntarctic from a (non-zero) winter minima to maximum values between January and March with a high degree of variability (both spatial and interannual, Fig. 5a), while SR3-Antarctic Chl-*a* concentrations are lower but with a slightly larger amplitude from winter values and reach a seasonal maxima by late November (Fig. 5b). One might expect high Chl-*a* values to correspond to lower dFe values (due to biological uptake), but for the SR3-subAntarctic, the Chl-*a* maxima in January–February is actually associated with the highest dFe levels (Figs. 4a and 5a), although the Chl-*a* decline that follows is mirrored in the dFe concentrations. It is plausible that the variability in dFe for January–March period (Fig. 4a) is driven by similar inter-annual variability in Chl-*a* over this period (Fig. 5a) or that both are connected to variability in dFe inputs. For the SR3-Antarctic, the November maxima in Chl-*a* corresponds to a minima in dFe of ~ 0.14 nM (Figs. 4b and 5b, Sedwick et al., 2008), but dFe then increases through the December–February period before declining again by April (Fig. 4b), while Chl-*a* levels fall over the same

BGD

8, 11489–11527, 2011

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



period (Fig. 5b). Despite the mean seasonal trend, Chl-*a* values are highly variable across the November–April period (Fig. 5b) and this could preclude the presence of a distinct seasonal trend in dFe. Nevertheless, the general positive covariation of Chl-*a* values and dFe concentration (the putative limiting nutrient) over the growing season presented in Figs. 4a,b and 5a,b is somewhat surprising and might indicate that the dominant driver of dFe variability is not phytoplankton biomass as alluded to by Chl-*a*, but rather exogenous inputs and/or ocean circulation.

An important aspect of the seasonal cycle from the SR3 sub-Antarctic sector is the presence of the only Southern Ocean measurements of dFe in the austral winter of ~0.1 nM in the upper 100 m (Ellwood et al., 2008). Within the paradigm of a winter “reset” of dFe levels by vertical mixing, these values are initially surprisingly low. July concentrations are only slightly higher than values of ~0.05 nM in April (Bowie et al., unpublished data, similar to the SR3-Antarctic in April) and November concentrations are only ~0.14 nM (Sedwick et al., 2008). Thus there is a gradual trend of only a slight increase in dFe from 0.05 nM in April, to 0.1 nM in July and 0.14 nM in November that is followed by much higher values in January and February (to ~0.35 nM, Fig. 4a) that tracks the Chl-*a* trend (Fig. 5a). Taken at face value, this suggests that the “reset” of dFe concentrations in the sub-Antarctic region of the SR3 sector might actually be a springtime phenomenon (i.e., occurring after November), possibly driven by atmospheric Fe deposition, vertical supply or advection of sub-Tropical waters (via the east Australian Current extension, which is particularly important for samples in the northern part of our SR3 sub-Antarctic sector) into the sub-Antarctic region (Boyd et al., 2004; Ellwood et al., 2008; Sedwick et al., 2008, Bowie et al., 2009). Biological activity in January and February then depletes these values to ~0.23 nM in March (Sedwick et al., 2008) and then to their April minimum levels. However, it is important to note that there are no measurements between July and November, a period over which Chl-*a* levels show an increasing trend (Fig. 5a), which suggests some dFe re-supply is not altogether unlikely. Indeed, seasonal mixed layer depth climatologies suggest that the deepest mixed layers in this region are actually in the August–September period (de

BGD

8, 11489–11527, 2011

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Boyer-Montegut et al., 2004). Thus it appears plausible that there is some increase in dFe between July and November due to vertical mixing that is then further augmented by other dFe sources (either dust, sub-Tropical water or further vertical mixing) in January and February. For example, Sedwick et al. (2008) have noted the influence of continental airmasses deep into the SR3-subAntarctic sector in January, while samples in the northern part of this sector will have dFe inputs from sub-Tropical water (Bowie et al., 2009). The absence of winter observations from the SR3-Antarctic sector means that it is not possible to assess whether (or not) the winter dFe value is higher than that measured in the SR3-subAntarctic in July or how, when and if surface water dFe stocks are replenished in order to fuel biological production in the subsequent growing season. Only new dFe observations from the autumn-winter-spring transition period can help resolve these questions. Using the vertical profiles from Fig. 2b, our synthesis would suggest that the maximum possible “winter reset” to dFe concentrations in this region would be to around 0.3 nM (i.e., the maximum values in the 100–500 and 500–1000 m depth bins, Fig. 2b, Table 2).

3.3.2 The Ross Sea

The high degree of variability in the magnitude and timing of the seasonal cycle of phytoplankton biomass in the Ross Sea is mirrored by a large degree of variability in dFe where numerous observations (over the period 1990–2006) have been made. Median dFe concentrations are low (< 0.25 nM) throughout the November to January period, but are associated with high variability for, in particular, the months of November and December (Fig. 4c). The Central Ross Sea polynya is normally associated with a bloom of *Phaeocystis antarctica* that displays a peak in December, but shows variability associated with the timing and extent of open water over the period (Arrigo and Van Dijken, 2004). The dFe observations span the period 1990–2008 and thus encompass years typified by large anomalies in the timing and magnitude of the phytoplankton bloom that are primarily driven by variable sea-ice dynamics due to El Niño (1997/8, Arrigo and Van Dijken, 2004) and the discharge of large ice-bergs (2000/1

BGD

8, 11489–11527, 2011

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



and 2002, Arrigo et al. (2002), Arrigo and Van Dijken, 2003), as well as the location of sampling (e.g., bottom depth, vicinity to melting sea ice, Sedwick et al., 2011). Thus, the variability in measurements of dFe in November and December (which both contain observations spanning 1994–2006) include years of early and intense blooms (probably associated with low dFe), as well as years with delayed and low biomass blooms (more likely to result in higher dFe levels), thus driving a high degree of variability in dFe (Fig. 4c). Moreover, studies that sampled near known Fe sources (sea ice, shallow bathymetry), as compared to those undertaken in open, probably Fe depleted (Sedwick et al., 2011), polynya waters also contributes to dFe variability. Accordingly, the high dFe values for February (Grotti et al., 2001) are probably more indicative of sampling close to fast ice rich in continental Fe, rather than a seasonal trend. Nevertheless, dFe values can remain low throughout the period October–January (Fig. 4c), which is indicative of a rapid utilisation of the winter reservoir (Sedwick et al., 2011). Reconciling low dFe values in November, with a biomass peak around December led Sedwick et al. (2011) to speculate about additional Fe sources during this period. Nevertheless, a constraint on the seasonal maxima in dFe, which is critical in calculating seasonal depletion, remains lacking at the either the start or end of the growing season. Regional model results suggest that convective overturn in winter will cause a winter maxima in dFe (Tagliabue and Arrigo, 2006, unlike the SR3-subAntarctic, Sect. 3.3.1) but there are no observations later than February (i.e., after the major peak in productivity). Given the large body of dFe observations already collected in the Ross Sea (already 240 between 0–100 m) and the information gained regarding the importance of a variety of processes in connecting dFe cycling to phytoplankton productivity, we would hope that future studies (especially those aimed at constraining the seasonal maxima in dFe) will continue in order to better understand the seasonal cycle of dFe in this important natural laboratory for Southern Ocean systems.

BGD

8, 11489–11527, 2011

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



4 Discussion

4.1 dFe distribution and processes

Overall, surface dFe is higher in the Antarctic region than the sub-Antarctic region, while within each of these regions the ATL basin is characterised by higher dFe than the IND basin. We speculate that these differences relate to differing levels of biological activity and the degree of Fe inputs. The Southern Ocean primary productivity (PP) estimates of Arrigo et al. (2008) show greater rates of PP in the sub-Antarctic region ($> 100 \text{ gC m}^{-2} \text{ a}^{-1}$), relative to the Antarctic region ($< 100 \text{ gC m}^{-2} \text{ a}^{-1}$), suggesting that higher surface dFe in the Antarctic region (Fig. 1a, Table 1) might result from lower rates of biological activity (most likely due to sea ice cover/reduced open water duration and low light levels for half of the year), with the opposite true for the sub-Antarctic. It is also plausible that the upwelling of upper circumpolar deep water (UCDW) that is enriched with deep-water dFe in the Antarctic region is depleted during its Ekman transport northwards to the sub-Antarctic region (Hoppema et al., 2003), thus also contributing to the inter-region surface dFe differences and lowering the Antarctic region's PP. Turning to geographic regions, Arrigo et al. (2008) report mean annual PP of $70.2 \text{ gC m}^{-2} \text{ a}^{-1}$ for their "Weddell Sea" geographic sector (which closely corresponds to the area of our ATL basin), while the combined mean PP of the "South Indian Ocean" and "Southwest Pacific Ocean" sectors (which encompass most of our IND basin) was $\sim 46 \text{ gC m}^{-2} \text{ a}^{-1}$ (Arrigo et al., 2008). Thus, it appears that PP is most likely higher in the ATL, relative to the IND basin, and high ATL basin PP is associated with higher dFe concentrations in surface waters (Fig. 2b, Table 2). So while the low dFe of the sub-Antarctic region is associated with high PP, the opposite is true when inter-basin trends are compared (low dFe in the IND basin corresponds to lower PP), which suggests that PP rates alone cannot explain inter-basin differences in dFe. Thus, when combined with our evidence from the SR3 region (Sect. 3.3.1), it appears that biological activity is not able to explain the seasonal or inter-basin dFe trends since either PP or Chl-*a* covary positively with dFe.

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Inter-basin trends in surface dFe from our synthesis might more reflect differences in the degree of Fe inputs. Since PP differences between the ATL and IND basins are opposite to what one might anticipate explaining the dFe trends, we must examine whether other processes might be at work. The obvious remaining candidate is that the ATL basin receives higher rates of Fe input than the IND basin, so that despite higher rates of PP, the ATL basin is still typified by higher surface dFe concentrations. The major sources of Fe to Southern Ocean surface waters are dust deposition and supply from shallow continental margins, as well as seasonal melting of sea-ice (e.g. Lannuzel et al., 2008; Tagliabue et al., 2009; Van der Merwe et al., 2011). For example, the ATL basin is close to Patagonian dust sources of Fe (e.g., Gaiero et al., 2003), as well as the large Patagonian plateau and associated rivers and glaciers, the continental margins of the numerous islands present in the ATL basin, and the Antarctic peninsula (e.g., Boyd and Ellwood, 2010; Klunder et al., 2011). Additionally, higher rates of upper ocean eddy kinetic energy (eddy genesis and meandering, EKE) in the south Atlantic are caused by current convergence/divergence predominantly at the Brazil-Malvinas Confluence (Maamaatuaiahutapu et al., 1998; Peterson and Stramma, 1990), the Agulhas Current Retroflexion and over the Atlantic mid-ocean ridge (Swart and Speich, 2010). Higher EKE provides a plausible mechanism whereby deeper waters with higher Fe concentrations can be supplied to the euphotic zone through upwelling processes (Archer and Johnson, 2000; Levy et al., 2001; Klein and Lapeyre, 2009). On the other hand, the IND-basin's dust sources are at more northerly latitudes and only the Kerguelen plateau and the Tasmanian shelf are potential margin sources (e.g., Bowie et al., 2009; Boyd and Ellwood, 2010), which may mean that this basin receives less exogenous input of dFe. Therefore, we speculate that while the inter-region differences in dFe can be explained by higher PP rates in the sub-Antarctic region, the inter-basin differences are due to the greater Fe inputs, both exogenous and vertical, experienced in the ATL basin.

Southern Ocean Iron Distributions

A. Tagliabue et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[I◀](#)[▶I](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

Opposite to the surface trends, the deep sub-Antarctic region has higher dFe than the Antarctic region and within each region, the IND basin is now greater than the ATL. These differences could result from variability in the concentrations of Fe binding ligands in the deep ocean, or differing degrees of Fe inputs from deep sources such as hydrothermal vents. While there are of course not as many measurements of Fe ligands as for dFe, Thuroczy et al. (2011) report measurements from both the sub-Antarctic and Antarctic regions of the ATL basin. At depths of ~ 4 km, ligand concentrations were ~ 1 nM in the sub-Antarctic ATL, while the Antarctic ATL concentrations were only ~ 0.7 nM (Thuroczy et al., 2011), which matches well with our higher deep dFe in the ATL-subAntarctic as compared to the ATL-Antarctic (Fig. 2b, Table 2). Interestingly, Thuroczy et al. (2011) found ligand concentrations > 1 nM in the intermediate waters of their ATL sub-Antarctic stations, which corresponds well with the increased intermediate depth dFe (Fig. 2b, Table 2). If these inter-region differences in ligand concentrations were consistent across the Southern Ocean, then it appears that variability in ligands might dictate deep dFe trends between the sub-Antarctic and Antarctic regions (Fig. 1a, Table 1). Comparing the deep water values of Thuroczy et al. (2011) from the ATL (~ 1 nM) and those measured by Ibsanmi et al. (2011) in the IND basin (0.58–0.83 nM), within the sub-Antarctic region suggests that the increasing dFe (> 2000 m depth) trend between the ATL-subAntarctic and the IND-subAntarctic is not reflected by a similar increasing trend in ligand concentrations. This may mean that inter-basin differences between the ATL and IND reflect different Fe inputs. The only plausible inputs at these depths are associated with hydrothermal activity, and indeed the hydrothermal Fe fluxes proposed by Tagliabue et al. (2010) do show greater inputs in the IND basin, relative to the ATL basin, due to the faster spreading rates of hydrothermal systems in the IND basin, which results in greater dFe input. Overall, this suggests that the greater deep dFe in the sub-Antarctic region is related to greater concentrations of Fe binding ligands, while dFe concentrations are increased in the deep IND basin due to more hydrothermal activity. If remineralisation of organic matter is a ligand source (Ibsanmi et al., 2011), then greater ligand concentrations in

Southern Ocean Iron Distributions

A. Tagliabue et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

the sub-Antarctic region are consistent with the region's higher PP. In the PAC basin, where hydrothermal inputs are proposed to be greatest, we lack the deep dFe data to investigate the potential importance of this Fe source. A plausible additional explanation for the different vertical profiles is that the upwelling of Fe-enriched UCDW (Hoppema et al., 2003) in the ATL basin causes a flatter profile than the IND.

4.2 Sampling frequency and seasonality

In the past decades, a large number of dFe measurements have been collected that provide seasonal variability data in the Southern Ocean (e.g., Martin et al., 1990; de Baar et al., 1995, 1999; Löscher et al., 1997; Lannuzel et al., 2011; Fitzwater et al., 2000; Sedwick et al., 1997, 1999, 2000, 2008, 2011; Sohrin et al., 2000; Measures and Vink, 2001; Bowie et al., 2004, 2009; Coale et al., 2005; Chever et al., 2010; Klunder et al., 2011). However, despite these efforts, there are some months in different ocean regions and basins in which no dFe measurements have yet been made. Our synthesis (Tables 1 and 2) shows that there is general coverage from October to April, but, apart from one set of observations in the IND-subAntarctic from July, no measurements outside of this period. By basin, the PAC stands out as having the lowest number of observations in total and no measurements whatsoever below 2000 m. That said, it has similar seasonal coverage to the ATL basin in the upper water column (0–1000 m). It is also notable that the ATL basin has no measurements in November and January (for the sub-Antarctic) or December (for the Antarctic), despite this being in the austral summer period.

While acknowledging the difficulty of sampling for Fe outside of the Austral spring-summer-autumn period, the lack of observations from the winter-spring and autumn-winter “transition” periods, as well as the winter in general can hinder attempts to understand the Southern Ocean Fe cycle. For example, given the seasonal cycle in dFe, missing months can bias the mean dFe we calculate by basin and region, which may have implications for the processes thought to be governing dFe distributions. In addition, dFe measurements are now being used to evaluate the performance of

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



complex ocean biogeochemical models (e.g., Moore and Braucher, 2008; Tagliabue et al., 2010), but if no observations are available outside of the October–April period then the model is not assessed at these times. In practise, this means that the models are only compared when dFe levels are relatively low in the growing season and the maximum dFe concentrations (which likely occur outside of this period) cannot be constrained. This is important, because it means that the “winter stock” of dFe in a given model, which more or less dictates the total net primary productivity achievable, cannot be compared to observations. Finally, our case studies from the SR3 and Ross Sea sectors of the Southern Ocean show that the lack of measurements from the winter-spring and autumn-winter “transition” periods result in a poorly constrained seasonal cycle. Most importantly, the only winter measurements from the Southern Ocean appear to contradict the paradigm of a winter reset in dFe concentrations (at least for the IND-subAntarctic) and highlights the need of obtaining dFe measurements in this difficult period of the year for sampling.

As months of the year, or specific basins without dFe observations leave us with difficulties in understanding the seasonality of dFe in the Southern Ocean we propose that our synthesis can help target future dFe observations. We would argue that a major priority, regardless of location, would be to obtain any measurements outside of the well sampled October–April period. Targeting the seasonal transitions, when the system is in “flux” appear to be important in understanding how stocks of dFe are replenished for subsequent growing seasons. This may be best achieved through mooring programs such as the Southern Ocean Time Series (SOTS), if such mission can return clean samples collected on a weekly-to-biweekly basis to shore based laboratories for analysis (Trull et al., 2010). Poorly sampled regions, such as the PAC basin in general and the ATL for some months, would also be important to better characterise inter-basin variability in the sources and cycling of dFe. For example, if hydrothermal sources are indeed important, then the largest signal of this should be in the deep PAC basin (Tagliabue et al., 2010), where observations are, up to now, absent. Although significant dFe enrichment was observed in Southeast Pacific Deep Slope Water masses

BGD

8, 11489–11527, 2011

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



near Drake Passage that have transited from the east Pacific rise and supporting high dFe inputs in the deep PAC basin that, in that study, were transported to and measured in the ATL basin (Klunder et al., 2011).

4.3 Comparing recent dFe measurements with early determinations

5 Over the 1989–2008 period for which we have synthesised Southern Ocean dFe measurements there has been a steady convergence of techniques and ever improving precision. For example, limits of detection of early Fe techniques were of the order of 0.1 nM (e.g., Achterberg et al., 2001; Bruland and Rue, 2001), whereas more recently methodologies permit dFe concentrations < 0.05 nM to be precisely determined
10 (Bowie et al., 2006; Klunder et al., 2011). In addition, intercalibration projects such as IRONAGES (Bowie et al., 2006) and SAFE (Johnson et al., 2007) resulted in a number of improvements in protocols and SAFE, in particular, produced a surface and deep water “reference” sample for which the concentrations had been precisely determined, which has proved invaluable for scientists to “validate” new methods. Importantly, the
15 advent of clean sampling rosettes (e.g., Measures et al., 2008) for trace metals that are capable of deep-water deployments has also played an important role in the collection of contamination-free samples. New intercalibration efforts under the GEOTRACES program are continuing to improve our ability to produce high quality dFe data in the open-ocean. Because of these issues, we were interested in examining whether there
20 has been a statistical difference in the properties of the dFe measurements between earlier observations that had higher detection limits and no reference samples, against more recently collected dFe data. To that end, we split our Southern Ocean dFe dataset into two subsets, one containing measurements from 1989–2002 (a total of 1458 observations) and the other from 2003–2008 (a total of 1874 observations), which were
25 then subdivided into the Shelf, Offshelf, Antarctic and sub-Antarctic regions, as well as across the 5 depth ranges as previously described (see Fig. 1, further subdivision by basin would have risked having too little data for each timeframe).

BGD

8, 11489–11527, 2011

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Southern Ocean Iron Distributions

A. Tagliabue et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

Table 3 presents the results of a Wilcoxon/Mann Whitney test between the 1989–2002 and the 2003–2008 dFe data by region and depth range. In the upper 500 m, 1989–2002 dFe data was significantly lower than that measured between 2003–2008 for all regions except the sub-Antarctic. In the shelf region, 1989–2002 dFe was much higher between 0–100 m than that derived from 2003–2008 data (means of 0.898 and 0.292 nM, respectively, Table 3). For Offshelf waters and the Antarctic region, 0–100 m dFe concentrations were on the order of 0.1 nM lower in 2003–2008 observations, relative to 1989–2002, with similar offsets in the 100–500 m depth interval (Table 3). On the other hand, the deepest samples (depths > 1000 m) showed no significant differences between 1989–2002 and 2003–2008, while the Offshelf and sub-Antarctic regions showed significantly higher dFe from 2003–2008, relative to 1989–2002, in intermediate waters (500–1000 m, mean differences of ~ 0.06–0.07 nM, Table 3).

While in intermediate and deep waters, the low number of observations in the 1989–2002 dataset may hinder the detection of significant difference, it appears that upper water differences are robust since ample observations are present in both subsets. Thus, we conclude that determinations of dFe concentrations from the upper 500 m of the Southern Ocean taken between 2003–2008 are indeed significantly lower than those from 1989–2002 for the Shelf, Offshelf and Antarctic regions. Differences between 1989–2002 and 2003–2008 in Offshelf waters are not related to differential sampling of the seasonal cycle between these time periods, as similar months are sampled in all regions except for a small number of July observations from 2003–2008 (which are too few to have significant weight). In the Antarctic region, 1989–2002 data contains samples from January and December, which are absent from the 2003–2008 data, but since these months should be typified by lower dFe levels they appear unlikely to explain the significantly higher dFe from 1989–2002. The sub-Antarctic zone had no significant differences between 1989–2002 and 2003–2008 despite 185 and 241 observations, respectively, and while 1989–2002 data lacks December and February observations, the samples from 2002–2008 have no October and March observations. It is therefore plausible that 1989–2002 observations show higher dFe concentrations

than those from 2003–2008 in the Shelf, Offshelf and Antarctic zones due to methodological differences between these periods, probably driven by the large reductions in the limit of detection, as well as better sampling systems, for dFe. Finally, it is also worth drawing attention to the fact that the 2003–2008 dataset (especially in the deep ocean) always has many more dFe observations than the 1989–2002 dataset, despite covering a much smaller time period. This is striking evidence of the progress being made in collecting dFe samples from the Southern Ocean as part of the International Polar Year and ongoing GEOTRACES efforts.

5 Conclusions

We have compiled 13 125 dFe observations covering the global ocean and encompassing the period 1978–2008 and make this available to the community for future studies (please contact one of the corresponding authors). Moreover, we have conducted a more detailed synthesis and analysis of the 3332 observations taken the Southern Ocean where Fe plays a fundamental role in regulating primary productivity and the carbon cycle. This analysis was performed within 4 different ocean regions and 6 different ocean basins across 5 depth intervals and highlights inter-basin and inter-region variability in the profiles of dFe and their variability. At the surface, differences in Fe inputs (both exogenous and vertical) or biological activity might explain the variability we find between the sub-Antarctic and Antarctic regions or the ATL and IND sector, while at depth we suggest a role for heterogeneity in Fe binding ligands and deep Fe sources such as hydrothermal vents. The profile of dFe might be linked to upwelling of UCDW south of the Polar Front. In surface waters, dFe data collected in recent years (2003–2008) is significantly different from that collected previously, which suggests refinements to methods result in statistically different dFe values and we also note the substantial progress made in Southern Ocean dFe sampling in recent years (more than half of all Southern Ocean observations are from the past 5 yr) as part of the GEOTRACES programme. However, despite these efforts the seasonal cycle of dFe

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



and its relationship to biological productivity remains enigmatic, even in some of the most well sampled regions of the Southern Ocean. In fact, we find that neither the SR3 seasonal nor inter-basin trends (between the ATL, IND and PAC) in dFe are driven by biological activity, which suggests that, apart from broad Antarctic–sub-Antarctic trends, other processes might be the dominant driver of dFe variability. However, there are no dFe measurements available for the Southern Ocean from May, June, August and September and we highlight in particular the nature of the autumn-winter-spring transition in dFe as a key gap in our knowledge. We hope that identifying these issues can help in the planning of future voyages to the region.

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BGD

8, 11489–11527, 2011

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Arrigo, K. R., van Dijken, G. L., Ainley, D. G., Fahnestock, M. A., and Markus, T.: Ecological impact of a large Antarctic Iceberg, *Geophys. Res. Lett.*, 29, 1104, doi:2001GL014160, 2002.

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Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Bowie, A. R., Archtenberg, E. P., Croot, P. L., de Baar, H. J. W., Laan, P., Sarthou, G., and Worsfold, P. J.: A community-wide intercomparison exercise for the determination of dissolved iron in seawater, *Mar. Chem.*, 98(1), 81–99, 2006.

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Southern Ocean Iron DistributionsA. Tagliabue et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

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Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



han, M., Mendez, J., Milne, A., Obata, H., Ossiander, L., Plant, J., Sarthou, G., Sedwick, P., Smith, G. J., Sohst, B., Tanner, S., van den Berg, C. M. G., and Wu, J.: Developing standards for dissolved iron in seawater, *EOS*, 88(11), 131–132, 2007.

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Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Sohrin, Y., Iwamoto, S., Matsui, M., Obata, H., Nakayama, E., Suzuki, K., Handa, N., and Ishii, M.: The distribution of Fe in the Australian sector of the Southern Ocean, *Deep-Sea Res. Pt. 1*, 47, 55–84, 2000.

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Table 1. A summary of the dFe sampling frequency, number of months sampled, number of observations (*N*) and mean concentration (\pm the standard deviation) for each depth range and Southern Ocean region.

Ocean regions	J	A	S	O	N	D	J	F	M	A	M	J	# of months	<i>N</i>	Mean dFe \pm std deviation (nM)	
0–100 m																
Shelf				█									7	382	0.61 \pm 1.14	
Offsh	█			█									8	999	0.31 \pm 0.45	
ANT				█									7	573	0.38 \pm 0.55	
SANT	█			█									8	426	0.23 \pm 0.27	
100–500 m																
Shelf				█									7	280	0.45 \pm 0.43	
Offsh	█			█									8	875	0.33 \pm 0.33	
ANT				█									7	497	0.40 \pm 0.40	
SANT	█			█									8	378	0.25 \pm 0.20	
500–1000 m																
Shelf				█									7	31	0.60 \pm 0.35	
Offsh	█				█									7	241	0.36 \pm 0.18
ANT					█									6	139	0.36 \pm 0.19
SANT	█				█									7	102	0.35 \pm 0.16
1000–2000 m																
Shelf											█		3	20	0.53 \pm 0.17	
Offsh	█										█		7	203	0.47 \pm 0.27	
ANT											█		6	150	0.46 \pm 0.30	
SANT	█										█		7	53	0.49 \pm 0.17	
2000–6000 m																
Shelf															ND	
Offsh							█						5	301	0.54 \pm 0.26	
ANT							█						5	230	0.51 \pm 0.24	
SANT							█		█		█		4	71	0.64 \pm 0.31	

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract Introduction

Conclusions References

Tables Figures

◀ ▶

◀ ▶

Back Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Table 2. A summary of the dFe sampling frequency, number of months sampled, number of observations (*N*) and mean concentration (\pm the standard deviation) for each depth range and Southern Ocean basin.

Ocean regions	J	A	S	O	N	D	J	F	M	A	M	J	# of months	<i>N</i>	Mean dFe \pm std deviation (nM)
0–100 m															
PAC SANT													5	45	0.14 \pm 0.10
PAC ANT													5	141	0.15 \pm 0.08
ATL SANT													5	58	0.30 \pm 0.55
ATL ANT													6	226	0.47 \pm 0.69
IND SANT													8	323	0.23 \pm 0.20
IND ANT													6	206	0.43 \pm 0.51
100–500 m															
PAC SANT													5	40	0.22 \pm 0.07
PAC ANT													4	84	0.21 \pm 0.12
ATL SANT													5	58	0.30 \pm 0.28
ATL ANT													6	266	0.49 \pm 0.48
IND SANT													8	280	0.24 \pm 0.19
IND ANT													5	147	0.32 \pm 0.24
500–1000 m															
PAC SANT													3	10	0.28 \pm 0.08
PAC ANT													2	16	0.32 \pm 0.09
ATL SANT													3	24	0.50 \pm 0.20
ATL ANT													3	76	0.42 \pm 0.22
IND SANT													6	68	0.30 \pm 0.11
IND ANT													4	47	0.28 \pm 0.14
1000–2000 m															
PAC SANT													1	2	0.48 \pm 0.12
PAC ANT													1	3	0.36 \pm 0.05
ATL SANT													3	39	0.52 \pm 0.17
ATL ANT													3	110	0.48 \pm 0.33
IND SANT													3	12	0.39 \pm 0.11
IND ANT													3	37	0.41 \pm 0.20
2000–6000 m															
PAC SANT															ND
PAC ANT															ND
ATL SANT													3	55	0.63 \pm 0.33
ATL ANT													3	177	0.49 \pm 0.21
IND SANT													2	16	0.70 \pm 0.25
IND ANT													3	53	0.57 \pm 0.29

Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract Introduction

Conclusions References

Tables Figures

◀ ▶

◀ ▶

Back Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Southern Ocean Iron Distributions

A. Tagliabue et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Table 3. Differences in the distribution of dFe data collected between 1989–2002 and 2003–2008 is tested using the Wilcoxon/Mann–Whitney test for non-parametric data by region and depth. The statistical significance or p value associated with the difference in location parameters (where a positive value indicates greater dFe in the 1989–2002 date of collection subset) is shown below the table. To aid clarity, negative differences in the location parameter are in italics, and p values ≤ 0.01 (i.e., highly significant) are in bold. “ND” indicates where there was not enough data to perform the statistical test. Also reported is the mean dFe concentration (with the number of observations in parentheses) for each region/depth.

Region	Depth range (m) Depth range	Difference in location (nM)	Mean dFe (nM, 1989–2002, n)	Mean dFe (nM, 2003–2008, n)
Shelf	0–100	0.171 ^{****}	0.898 (198)	0.292 (184)
Offshelf		0.048 ^{****}	0.354 (555)	0.27 (444)
SUB-ANTARCTIC		~ 0 ^{NS}	0.244 (185)	0.221 (241)
ANTARCTIC		0.080 ^{****}	0.409 (370)	0.317 (203)
Shelf	100–500	0.090 ^{****}	0.541 (144)	0.350 (136)
Offshelf		0.060 ^{****}	0.400 (401)	0.278 (474)
SUB-ANTARCTIC		0.040 ^{**}	0.308 (132)	0.217 (246)
ANTARCTIC		0.050 ^{***}	0.444 (269)	0.345 (228)
Shelf	500–1000	0.256 [*]	0.820 (9)	0.503 (22)
Offshelf		<i>-0.053</i> ^{***}	0.301 (62)	0.376 (179)
SUB-ANTARCTIC		-0.040 ^{NS}	0.296 (20)	0.358 (82)
ANTARCTIC		<i>-0.062</i> ^{***}	0.303 (42)	0.39 (97)
Shelf	1000–2000	ND	ND (0)	0.532 (20)
Offshelf		-0.010 ^{NS}	0.453 (42)	0.469 (161)
SUB-ANTARCTIC		-0.068 ^{NS}	0.419 (6)	0.499 (47)
ANTARCTIC		0.012 ^{NS}	0.458 (36)	0.458 (114)
Offshelf	2000–6000	0.060 [*]	0.607 (47)	0.528 (254)
SUB-ANTARCTIC		0.270 ^{**}	0.870 (6)	0.622 (65)
ANTARCTIC		0.0500 ^{NS}	0.568 (41)	0.495 (189)

NS = $p > 0.1$ (or deemed “not significant”)

* = $0.1 \geq p > 0.05$, ** = $0.05 \geq p > 0.01$, *** = $0.01 \geq p < 0.001$, **** = $p < 0.001$.

Southern Ocean Iron Distributions

A. Tagliabue et al.

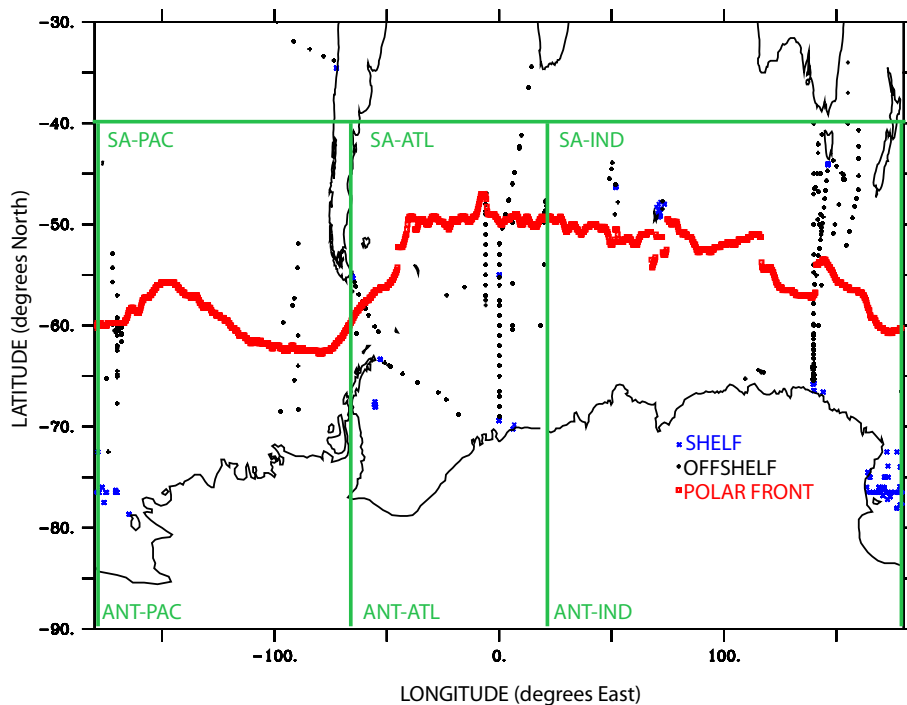


Fig. 1. Distribution of the data and the regional breakdown into different ocean regions and basins.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Southern Ocean Iron Distributions

A. Tagliabue et al.

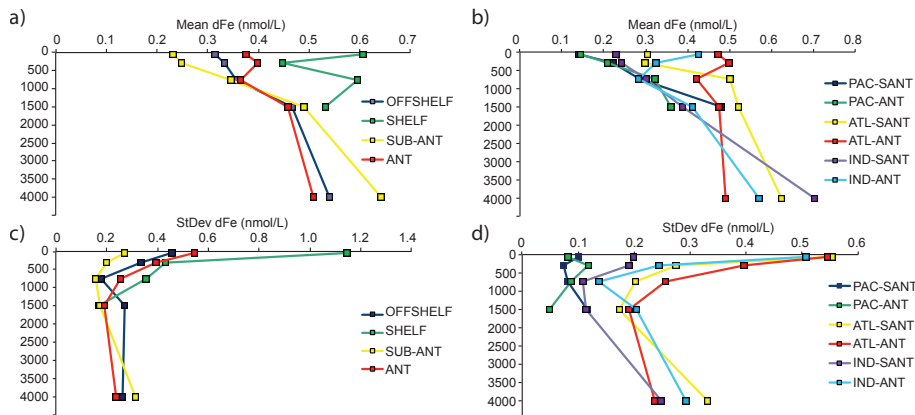


Fig. 2. Mean dFe by (a) region and (b) basin as a function of depth. Standard deviation of dFe measurements by (c) region and (d) basin as a function of depth. The points are plotted at the midpoint of the particular depth range.

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper

Title Page

Abstract Introduction

Conclusions References

Tables Figures

◀ ▶

◀ ▶

Back Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Southern Ocean Iron Distributions

A. Tagliabue et al.

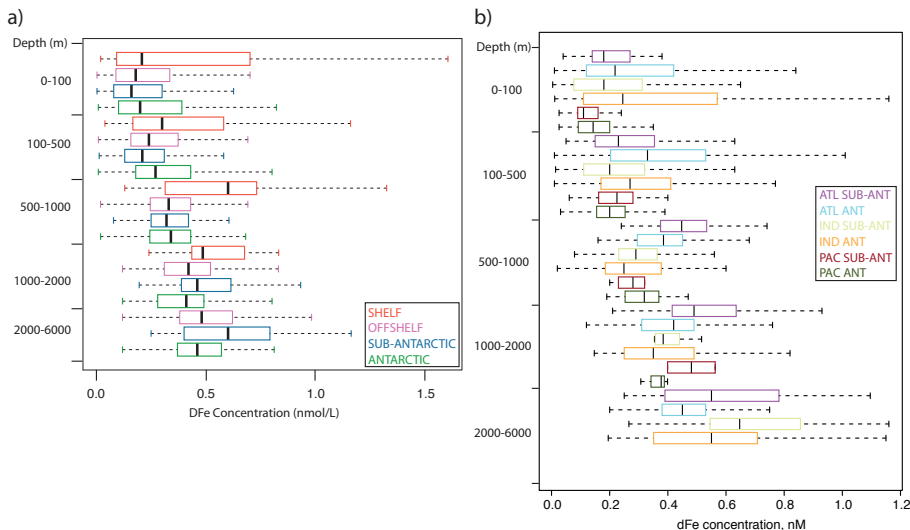


Fig. 3. Box and Whisker plots of dFe by **(a)** region and **(b)** basin. The size of the box represents the 1st to 3rd quartiles, with the vertical bar corresponding to the median and the whiskers representing 1.5 times the inter-quartile range.

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Southern Ocean Iron Distributions

A. Tagliabue et al.

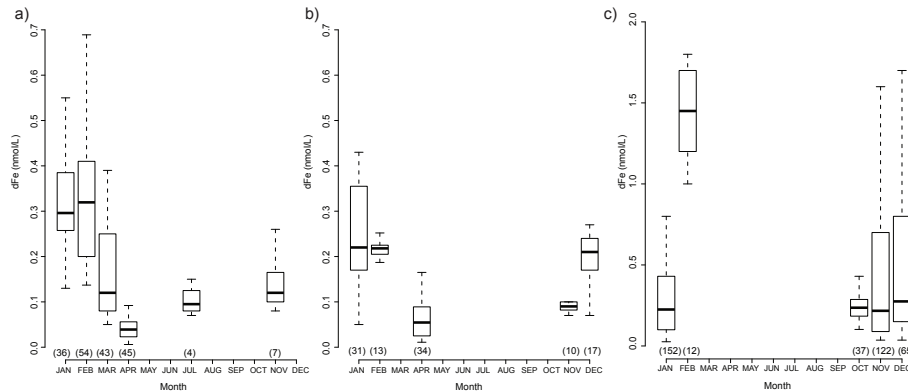


Fig. 4. Box and Whisker plots of the seasonal cycle of dFe in the upper 100 m for **(a)** the sub-Antarctic SR3 (189 total observations from 1995–2008), **(b)** the Antarctic SR3 (105 total observations from 1994–2008) and **(c)** the Ross Sea (240 total observations from 1990–2008). The SR3 region is defined as that encompassing 135° E to 155° E and the Ross Sea region is between 155° W to 180° W. The sub-Antarctic and Antarctic zones are separated using the mean position of the nPF as illustrated in Fig. 1. The size of the box represents the 1st to 3rd quartiles, with the vertical bar corresponding to the median and the whiskers representing 1.5 times the inter-quartile range.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



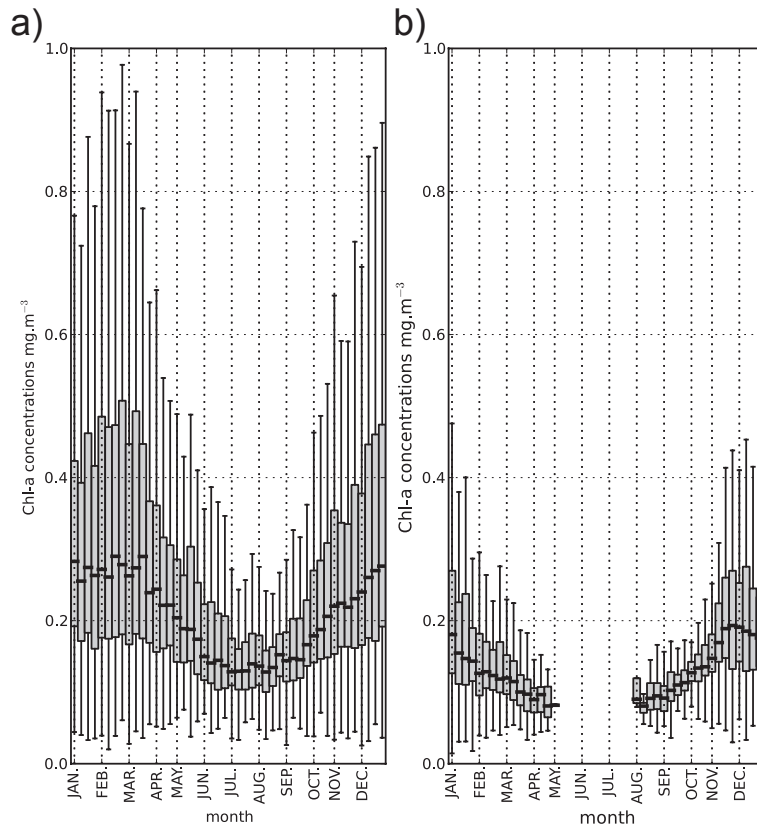


Fig. 5. The mean seasonal cycle in Chl-*a* (mg m^{-3}) within our SR3 sector from a climatology of Globcolor ocean color data over the 1998–2010 period for **(a)** the sub-Antarctic and **(b)** Antarctic. The size of the box represents the 1st to 3rd quartiles, with the vertical bar corresponding to the median and the whiskers representing 1.5 times the inter-quartile range.