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Impacts of sea ice on the marine iron cycle and phytoplankton productivity

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Abstract

Iron is a key nutrient for phytoplankton growth in the surface ocean. At high latitudes, the iron cycle is closely related to sea ice. In recent decades, Arctic sea ice cover has been declining rapidly and Antarctic sea ice has exhibited large regional trends. A significant reduction of sea ice in both hemispheres is projected in future climate scenarios. To study impacts of sea ice on the iron cycle, iron sequestration in ice is incorporated to the Biogeochemical Elemental Cycling (BEC) model. Sea ice acts as a reservoir of iron during winter and releases iron to the surface ocean in spring and summer. Simulated iron concentrations in sea ice generally agree with observations, in regions where iron concentrations are lower. The maximum iron concentrations simulated in the Arctic sea ice and the Antarctic sea ice are 192 nM and 134 nM, respectively. These values are much lower than observed, which is likely due to missing biological processes in sea ice. The largest iron source to sea ice is suspended sediments, contributing fluxes of iron of 2.2×10^8 mol Fe month^{-1} to the Arctic and 4.1×10^6 mol Fe month^{-1} to the Southern Ocean during summer. As a result of the iron flux from ice, iron concentrations increase significantly in the Arctic. Iron released from melting ice increases phytoplankton production in spring and summer and shifts phytoplankton community composition in the Southern Ocean. Simulation results for the period of 1998 to 2007 indicate that a reduction of sea ice in the Southern Ocean will have a negative influence on phytoplankton production. Iron transport by sea ice appears to be an important process bringing iron to the central Arctic. Impacts of iron fluxes from ice to ocean on marine ecosystems are negligible in the current Arctic Ocean, as iron is not typically the growth-limiting nutrient. However, it may become a more important factor in the future, particularly in the central Arctic, as iron concentrations will decrease with declining sea ice cover and transport.

1 Introduction

Sea ice covers a large portion of the high latitude oceans and is vulnerable to climate change. Arctic sea ice has experienced a large reduction in summer ice extent (Serreze et al., 2007; Kwok et al., 2009), thickness (Rothrock et al., 1999), and multi-year ice cover (Maslanik et al., 2007; Nghiem et al., 2007). Overall Antarctic sea ice remains relatively unchanged. While there is a mild increase in ice cover (Kurtz and Markus, 2012), there are strong interannual variations in ice extent (Oza et al., 2011) and large regional trends (Parkinson and Cavalieri, 2012). Climate projections suggest that sea ice will decline rapidly in future decades in both hemispheres (Holland et al., 2006; Stroeve et al., 2012). The onset of sea ice formation and melting will also change (Boe et al., 2009).

Sea ice has strong impacts on ocean biogeochemical cycles and marine ecosystems (Ducklow et al., 2012; Sedwick and DiTullio, 1997). Phytoplankton blooms are often observed in the marginal ice zone, where there is recent melting of sea ice (Smith and Nelson, 1985; Taylor et al., 2013; Fitch and Moore, 2007). Heavy sea ice cover blocks light and limits phytoplankton growth rates. The seasonal ice-ocean freshwater exchange and brine rejection during ice formation can modify the mixed layer depth and drive vertical mixing, which affects irradiance and nutrients experienced by phytoplankton (Taylor et al., 2013; Stabeno et al., 2010). Sea ice removes nutrients from seawater during ice formation, catches dust deposited on it, and releases nutrients and dust during ice melting. Therefore, changes in sea ice will alter the nutrient cycling and the timing that nutrients are available to phytoplankton.

Iron is an essential nutrient for phytoplankton growth. It is required for important biological processes, such as photosynthesis and producing enzymes. The demand for iron varies in different phytoplankton group, therefore iron concentrations in seawater also influences phytoplankton competition and marine ecosystem structure (Wang and Moore, 2011). It has been broadly recognized that the Southern Ocean is the largest High Nutrient Low Chlorophyll (HNLC) region, where the primary productivity is often limited by low iron availability (Wang and Moore, 2011; Arrigo et al., 2003; Boyd, 2002; Boyd et al., 2001).

Potential iron sources for high latitude surface oceans include diffusion from sediments (Moore and Braucher, 2008), sediment resuspension (Planquette et al., 2013), dust deposition (Jickells et al., 2005), snow and sea ice melting (Lannuzel et al., 2008; Aguilar-Islas et al., 2008; Tovar-Sanchez et al., 2010), river input (Klunder et al., 2012), glacial runoff (Bhatia et al., 2013) and iceberg melting (Lin et al., 2011; Shaw et al., 2011; Smith et al., 2007). In polar regions, atmospheric deposition is obstructed by ice cover during the cold season, when dust accumulates in the sea ice (Maenhaut et al., 1996; Tovar-Sanchez et al., 2010; Sedwick and DiTullio, 1997). Sedimentary materials are important iron sources in shallow oceans (Moore and Braucher, 2008; Tagliabue et al., 2012). Suspended sediments can be incorporated into sea ice during ice formation and released during melting (Measures, 1999; Thomas and Dieckmann, 2002). Meanwhile, sediment in sea ice can be transported away from shallow regions by ice movement (Nurnberg et al., 1994; Grotti et al., 2005). Sea ice formation may also remove dissolved iron in surface waters by the process of entrainment in ice itself (Lancelot et al., 2009; Lannuzel et al., 2011). Previous field studies reported that iron concentrations in sea ice can be one- or two- orders of magnitude higher than iron concentrations in the underlying seawater (van der Merwe et al., 2011a; Aguilar-Islas et al., 2008; Lannuzel et al., 2007, 2008). Therefore, sea ice may be a significant iron source at high latitudes, play an important role in controlling the location and timing of iron fluxes to seawater, and regulate phytoplankton production through its impacts on the iron cycle.

There have been some research efforts about the role of sea ice in nutrient cycling (Lannuzel et al., 2011; van der Merwe et al., 2011b; Aguilar-Islas et al., 2008; Measures, 1999; Lancelot et al., 2009). Observational data on iron concentrations in sea ice and the surrounding seawater are scarce due to harsh conditions and difficulty in making measurements. Questions about impacts of sea ice on the iron cycle remain. Given the rapid change in Arctic sea ice and the potential for large reductions in both

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hemispheres in the future, there is an urgent need to explore the role of sea ice in ocean biogeochemical cycles.

Modeling can provide a valuable addition to field measurements. A previous modeling study conducted by Lancelot et al. (2009) suggests that sea ice-ocean iron exchanges alter the geographic pattern of iron distribution in the Southern Ocean. Simulated annual mean iron uptake in sea ice was on the same order of magnitude as iron sediment fluxes in the western Ross Sea (Lancelot et al., 2009). This study is a first attempt to incorporate iron in the sea ice component of the Community Earth System Model (CESM). Different sources of iron in sea ice are incorporated. This work allows us to assess impacts of sea ice on the iron cycle at high latitudes in both hemispheres.

2 Methods

2.1 The ocean and sea ice components of the CESM

This work is based on the marine ecosystem and biogeochemistry module of the Community Earth System Model (CESM), which is known as the Biogeochemical Elemental Cycling (BEC) model. In this study, we use active ocean and sea ice components, and data atmosphere and land components that prescribe observationally-based forcing information. The BEC module runs within the ocean physics component of the CESM1.0, the Parallel Ocean Program version 2 (POP2, Smith et al., 2010). It has a nominal horizontal resolution of 1° and 60 vertical levels. The thickness of each vertical level is 10 m in the upper 150 m and increase with depth below 150 m. Details of the model as incorporated in the CESM1.0 are described by Danabasoglu et al. (2012). The wind speed–mixing relation in the model was adjusted to better match the observed mixed layer depths in the Southern Ocean (de Boyer Montégut et al., 2004).

The sea-ice component in the CESM is the Community Ice CodE version 4 (CICE4) (Hunke and Lipscomb, 2010). CICE4 operates on the ocean horizontal grid. The CICE4 includes elastic–viscous–plastic dynamics, energy-conserving thermodynamics, and

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a subgrid-scale ice thickness distribution (ITD). The ITD includes five categories within each grid cell, which have different thickness, different surface properties, and different melt and growth rates as computed by the thermodynamics. This version of the CICE model includes a more sophisticated shortwave radiative transfer algorithm and allows for both passive and active tracers (Holland et al., 2012). An additional tracer, iron, is incorporated in the CICE4 as a passive tracer. Iron in sea ice comes from different sources and can be transported by ice motion.

2.2 The biogeochemical model

The BEC model used here includes four phytoplankton functional groups, one zooplankton group and biogeochemical cycling of multiple growth limiting nutrients (nitrate, ammonium, phosphate, iron and silicate) (Moore et al., 2004). The four phytoplankton groups are diatoms, diazotrophs, small phytoplankton, and coccolithophores. In the BEC model, the diatom group represents larger phytoplankton, which are less efficient at nutrient uptake and export carbon more efficiently than other phytoplankton groups.

The light-, nutrient-, and temperature-dependencies of phytoplankton growth rate are modeled multiplicatively. Phytoplankton growth rates decrease under nutrient stress according to Michaelis-Menten nutrients uptake kinetics. Phytoplankton photoadaptation is described by varying chlorophyll to nitrogen ratios based on the model of Geider et al. (1998). Phytoplankton groups also have variable Fe/C ratios, which vary as a function of ambient iron concentrations. Dissolved iron sources to the ocean in the BEC model include atmospheric dust deposition and sedimentary diffusion. There is one "dissolved" iron pool that is assumed to be bioavailable in the BEC model. Iron is removed from the dissolved pool by biological uptake and particle scavenging. A fraction of the scavenged iron is assumed lost to the sediments (Moore and Braucher, 2008). Ecosystem parameters were chosen based on field and laboratory data and described in detail by Moore et al. (2002, 2004). The CESM 1.0 BEC model is evaluated against observational datasets for the 1990s by Moore et al. (2013). This work

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also addresses the impacts of 21st century climate change on marine biogeochemistry and ecosystem dynamics.

2.3 Modifications related to the iron cycle

In the previous CESM-BEC model, sea ice acted as a physical barrier for air–sea gas exchange, but had no impact on atmospheric dust deposition. Dust deposition over ice covered model grid cells was released into seawater directly. There was no iron transfer between the ocean and sea ice either.

Both dissolved iron and particulate iron were observed in sea ice. Iron concentrations in sea ice can be up to two-orders higher than the underlying seawater (van der Merwe et al., 2011a; Aguilar-Islas et al., 2008; Lannuzel et al., 2007, 2008; Measures, 1999). However, processes of accumulation, transport and removal of iron in sea ice had not been studied extensively and mechanisms governing these processes were unclear. We made several assumptions and simplifications to simulate iron sequestration in sea ice and explored impacts of sea ice on iron fluxes and marine ecosystems.

First, we assumed that the bioavailable iron frozen in sea ice from suspended sediments was proportional to iron concentrations in seawater, since iron concentrations in shallow regions were mainly controlled by sedimentary input (Moore and Braucher, 2008). The ratio of sediment areas in bottom cells (percSed) were calculated using the high-resolution ocean bathymetry from the ETOPO2 version 2.0, similar to the approach used in estimating sedimentary iron source (Moore and Braucher, 2008). The amount of sedimentary iron flux to ice was proportional to the sediment area. Thus, the sedimentary iron in sea ice ($Fe_{\text{sed-ice}}$) was computed as $Fe_{\text{sed-ice}} = f_{\text{sed}} \cdot dFe_{\text{water}} \cdot \text{percSed}$, where dFe_{water} represented the iron concentration in seawater, and f_{sed} was a factor needed to better match observations. Different f_{sed} values had been tested to obtain a better match to observations (not shown). Sedimentary iron was only incorporated into sea ice where sediments were shallower than 50 m. Different values of the sediment depth had been tested. A limit of 50 m was used to avoid

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excessive sedimentary iron in the Arctic. However, sediments in ice may be underestimated in winter deep convection regions (to be discussed later).

Second, we assumed that the amount of iron removed during ice formation was proportional to dissolved iron in the surrounding seawater. Instead of pumping iron from seawater to obtain a fixed concentration of iron in the ice (Lancelot et al., 2009), a constant percentage of iron was removed from seawater and trapped in sea ice. We had tested different removal fractions ranging from 50 % to 100 % and compared simulated iron concentrations in sea ice with observations. The differences in correlations between simulation and observation were small. A 60 % removal fraction, a moderate value, was chosen for a better match to observations.

The role of sea ice on the flux of dust iron into oceans was considered in the modified CESM-BEC. When dust was deposited over sea ice covered regions, it was stored in the ice and passively follows ice mechanical movements. Iron in sea ice, regardless of its source, was transported with the ice. This allows ice to transport sedimentary iron from shallow regions to deep, iron-depleted regions. We assumed there was no chemical removal or biological uptake of iron in sea ice. We assumed there was no diffusion of iron between ice and seawater, and iron was only released to seawater during melting. Iron remained bioavailable when it was released into seawater.

2.4 Model experiments

The initial distributions of nutrients, inorganic carbon and alkalinity were based on the World Ocean Atlas database (Garcia et al., 2006) and the GLODAP database (Key et al., 2004). Dissolved iron initialization was based on simulations from Moore and Braucher (2008). The atmospheric dust deposition data used in this study are from Mahowald et al. (2005), assuming a dust iron content of 3.5 % and a constant 2 % solubility. The sedimentary iron input data is estimated by Moore and Braucher (2008), based on a high resolution bathymetry and estimated Fe flux constrained by the carbon export.

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The atmospheric forcing information was based on the National Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) meteorological reanalysis climatology data (Kalnay et al., 1996) with a number of modifications (Large and Yeager, 2009). The dataset covered the period of 1948 to 2007. Simulations were forced by repeating reanalysis data for 210 yr as a spin-up, and then different modifications are incorporated. Simulations ran from this state for 30 yr, which was long enough so that drifts in upper ocean fluxes declined to negligible levels and the state of marine ecosystems approached steady state. The control simulation (CTRL) was conducted using the old CESM-BEC without any modification.

Four 30 yr simulations were conducted following the initial spin-up of 210 yr (summarized in Table 1). All modifications described above were incorporated in the FULL simulation. Impacts of sea ice on the iron cycle and phytoplankton production were evaluated by comparing the FULL simulation with the CTRL simulation. The relative importance of each iron source to sea ice was discussed by comparing different simulations. The differences between the NOdust and the FEdust indicated the contribution of dust to iron accumulation in sea ice. The contribution of seawater iron can be estimated by comparing the FEsw and FEdust. Comparisons between the FULL and the FEsw showed the importance of sediments in ice iron accumulation. Interactions between different iron sources were not considered in this study.

The iron cycle and marine ecosystems are sensitive to the physical climate system. While the CESM generally does a good job of simulating sea ice and ocean physics, there are some biases in the model at high latitudes (Danabasoglu et al., 2012; Gent et al., 2011). Physical biases were significantly reduced in our simulations, due to the use of the NCEP/NCAR reanalysis climatology data as forcing (Danabasoglu et al., 2014). We mainly focus on comparisons between different experiments with the same physical environment to discuss the relative importance of iron sources to sea ice and impacts of iron delivered by sea ice on marine ecosystems.

Simulated iron distributions in sea ice and ocean waters were compared with observations in Sect. 3.1. Impacts of sea ice on the iron cycle and marine ecosystems were

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evaluated by comparing the FULL simulation and the CTRL simulation in Sect. 3.2. We also compared the relative importance of different iron sources to sea ice (Sect. 3.3), followed by an examination of time-varying impacts of iron sequestration in sea ice from 1998 to 2007 (Sect. 3.4).

3 Results and discussion

3.1 Iron distribution in sea ice and water

Simulated iron concentrations in surface seawater and sea ice in the phytoplankton blooming month are shown in Fig. 1. Iron concentrations in high latitude seawater are less documented than other ocean regions. There have only been a few studies reporting iron concentrations in sea ice. Here we review observed iron concentrations in ice and seawater in both the Southern Ocean and the Arctic region. We also compare simulated results with observations.

In the Southern Ocean seawater, observed surface dissolved iron (dFe) concentrations are typically in the range of 0.1–0.5 nM, with lower values in sub-Antarctic regions and higher values in shelf and shallow regions (Tagliabue et al., 2012; Moore and Braucher, 2008). The model does a good job of capturing the general pattern of iron distributions and the magnitudes reported in field studies, with highest iron concentrations in surface waters around the Antarctic Peninsula (Fig. 1).

Observed iron concentrations in Antarctic pack ice vary considerably: 0.2–81 nM in the East Antarctic (Lannuzel et al., 2007; van der Merwe et al., 2009; van der Merwe et al., 2011a), 0.7–36.8 nM in the Weddell Sea (Lannuzel et al., 2008), and 1.1–30.2 nM in the Bellingshausen Sea (Lannuzel et al., 2010). Samples of sea ice in the East Antarctic and the Ross Sea have lower iron concentrations: 0.14–4.3 nM (van der Merwe et al., 2009) and 1.1–6.0 nM (Grotti et al., 2005), respectively. Iron in sea ice is replenished in winter and then decreases as summer progresses (Lannuzel et al., 2010).

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In our simulations, dFe concentrations in Antarctic sea ice are generally lower than 2 nM, except in sea ice near the Antarctic Peninsula. The lower stock is likely due to missing iron inputs to sea ice in the Southern Ocean. Sedimentary iron is only incorporated in sea ice where the water column is shallower than 50 m in our simulation. However, ocean bathymetry is often deeper than 50 m in the ocean model grid. This parameterization may cause an underestimate of entrainment of suspended sediments. In regions, such as the Ross Sea, where there is deep convection during winter, sediments may be brought up by the deep mixing. It is also possible that certain iron sources are missing in our simulation, such as continuous biological transfer of iron by ice algae (Lannuzel et al., 2010). Simulated iron concentrations in sea ice above shallow water near the west Antarctic Peninsula are up to 134 nM during late winter months. This is higher than reported dFe concentration in Antarctic sea ice. However, simulated dFe concentrations in sea ice in the Bellingshausen Sea in spring and summer months are lower than 26 nM, which agree with previous observations (Lannuzel et al., 2010). Simulated dFe concentrations are higher in sea ice near the continental shelf and decline offshore. There are elevated dFe concentrations along the margin of sea ice, especially on the east side of the Antarctic Peninsula. It is likely due to the flow of sea ice, which transports iron-rich sea ice formed near the Antarctic Peninsula downstream. Iron concentrations are higher during winter and decrease rapidly in November and December, which agree with the seasonal variation shown in observations.

Compared to a previous model study of iron sources in the Southern Ocean (Lancelot et al., 2009), simulated iron concentrations in Antarctic sea ice are generally lower, except in the west Antarctic Peninsula. This is mainly due to the different parameterization of transferring ocean dFe to sea ice used in this study. Lancelot et al. (2009) set a maximum iron concentration of 16.5 nM in sea ice and moved iron from seawater to ice to ensure this value. The parameterization caused extremely low dFe concentrations in seawater and relatively high iron stocks in sea ice. In this study, a fixed fraction of iron is removed from seawater to ice during ice formation, ensuring a reasonable

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amount of dFe remains in the water. There are no strong observational constraints on the fraction of dFe incorporated into the sea ice during ice formation.

In the Arctic region, previous field studies reported high dissolved iron concentrations in the Laptev Sea, the Fram Strait, Bering Strait, and Chukchi Sea: > 10 nM (Klunder et al., 2012), 10 nM (Tovar-Sanchez et al., 2010), 5–10 nM (Nishimura et al., 2012), and 8.1–16.5 nM (Nishimura et al., 2012), respectively. Surface dFe concentrations in the central Arctic are in the range of 0.5 nM - 2 nM (Klunder et al., 2012). Iron concentrations in the Canada Basin and the Bering Sea are 1–3.2 nM (Nakayama et al., 2011) and 0.8–3.14 nM (Aguilar-Islas et al., 2008), respectively. Surface dFe concentrations in the Siberian shelf seas are relatively low during August and September (< 1 nM) (Klunder et al., 2012). Compared to the CTRL simulation, simulated surface iron distributions in the FULL experiment are improved in the central Arctic, the Bering Strait, the Chukchi Sea, and the Canada Basin, but overestimate iron concentrations in the offshore areas of the Siberian shelf seas. A previous study suggested that the depletion of iron in the Barents Sea and Kara Sea is caused by biological uptake (Klunder et al., 2012), however, simulated biomass in this region is biased low. Simulated iron concentrations along the Fram Strait are similar in the CTRL and FULL experiment, and biased low compared to observations. This is likely due to positive biases in biological production and the large consumption of iron by phytoplankton in this region. The general pattern of iron distributions in our experiment agrees well with observations, except for this Fram Strait region.

Observations of iron concentrations in the Arctic sea ice are scarce. Aguilar-Islas et al. (2008) reported that dFe concentrations in sea ice in the Bering Sea varied from 2.92 to 376 nM, with a geometric mean of 22.9 nM. Simulated dFe stocks fall in this range of observations. The highest simulated dFe concentrations in April, May and June are 119 nM, 159 nM and 192 nM in the Bering Strait and Chukchi Sea. Observed mean iron concentrations in multiyear sea ice along the Greenland current and Fram Strait are 863.9 nM in the surface layer and 531.7nM in the deep layer (Tovar-Sanchez et al., 2010). The minimum values of dFe observed in this study were 238.9 nM in

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surface ice and 219.9 nM in deep ice, which are much higher than our simulations (< 2 nM dFe). Multiyear ice may accumulate more iron through interactions between brine channels and sea ice biota, processes not included in the model.

3.2 Impacts of sea ice on the iron cycle and marine ecosystems

Differences of net iron fluxes to the surface ocean between the CTRL and the FULL simulations (FULL – CTRL) are shown in Fig. 2a. There is a relative negative influx of iron to the surface ocean in the FULL experiment in ice forming areas. There are two reasons for the differences. First, sea ice in the modified model blocks the dust deposition, which causes a decrease in iron supply to seawater. Second, a fraction of the iron is incorporated into the sea ice during ice growth, which reduces iron concentrations in the seawater. When ice melts, both dust and iron that is in the ice are released back to the seawater. Suspended sediments trapped in ice are also released to the seawater as "new" iron available for phytoplankton growth, during spring-summer when there is more light and macronutrients available in the mixed layer. Compared to the CTRL, the iron input during the melting season shifts northward in the Southern Ocean, and to the Bering Sea, the Fram Strait, and the west Arctic.

Impacts of sea ice on the iron concentrations in seawater and marine ecosystems highly depend on the limiting factors of phytoplankton growth. The increased iron input can stimulate phytoplankton growth, when iron is the most limiting factor for growth. However, iron concentrations in seawater may not change significantly due to rapid biological uptake. When phytoplankton growth is not limited by iron, iron concentrations in seawater will increase with the increased iron input.

In the Arctic region, primary production is mainly limited by light and macronutrient concentrations, except for some areas in the Bering Sea and the Sea of Okhotsk. Thus ambient iron concentrations increase with larger iron inputs. Ice melting in spring causes large increases of iron concentrations in the surface Arctic waters (Fig. 2b). The largest increase appears in the east Arctic, where there is wide shallow continental shelf. The impact of ice on surface iron distribution is generally confined in the seasonal ice zone. Only a small amount of iron from ice is laterally transported to the Barents Sea and the south Bering Sea.

Figure 2a shows clear changes of the total iron input to seawater in the Southern Ocean. Simulated iron fluxes from sea ice to ocean are less than 1 µmol Fe m^{-2} month^{-1} in the Southern Ocean, except in the Bellingshausen Sea. The flux is lowest in the Australian sector of the Southern Ocean. This pattern is quite different from results in the previous study of Lancelot et al. (2009), due to the different parameterizations of iron incorporation into ice. In the previous study, the iron flux is highest in the Weddell Sea and East Antarctic and quite low in the Amundsen-Bellingshausen Sea (< 0.25 µmol Fe m^{-2} month^{-1}, Lancelot et al., 2009). In both the study of Lancelot et al. (2009) and this study, iron concentrations in seawater show no remarkable changes after incorporating the iron cycle to sea ice (Fig. 2b), because iron is often the most limiting factor for phytoplankton growth in the Southern Ocean (Moore et al., 2004; Wang and Moore, 2011; Boyd, 2002; Arrigo et al., 2003). Thus, iron from melting ice is rapidly consumed by phytoplankton.

Due to iron sequestration in sea ice, iron is "preserved" in winter and released to seawater in spring and summer when there is sufficient light for phytoplankton growth. As a result, biological production increases in spring and summer in both hemispheres. Differences in export production during the growing season are shown in Fig. 3. In the Arctic region, export production is mostly unchanged, except in the Bering Sea and the Sea of Okhotsk. Aguilar-Islas et al. (2008) found strong evidence for this ice-iron linkage leading to phytoplankton blooms in the Bering Sea. Export production in spring and summer is generally enhanced by iron supply from ice in the Southern Ocean. This agrees with previous field studies, which suggest that iron released by sea ice melting fuels summer blooms in the Southern Ocean (Sedwick and DiTullio, 1997; Tagliabue and Arrigo, 2006; Gerringa et al., 2012). Overall, the effect of iron released by sea ice on biological production is moderate. Export production in regions south of 60° S increase less than 4 % in the growing season.

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Sea ice shifts the timing and locations of iron supply to surface water. Different phytoplankton groups show different responses to melting ice (Fig. 4). Small phytoplankton biomass increases rapidly along margins of sea ice in spring. Elevated production of diatoms only appears near the west Antarctic Peninsula in November, where iron concentrations in sea ice are highest. During December, increases in small phytoplankton biomass are mostly between 60–65° S, while diatom biomass increases at higher latitudes. This is because diatoms are more abundant in near shore waters and require higher iron supply for growth (Wang and Moore, 2011).

The simulated small phytoplankton and diatoms groups have different responses to the iron supply from sea ice in the Northern Hemisphere (not shown). There is an increase of diatom biomass in the Bering Sea and the Sea of Okhotsk in spring, while small phytoplankton biomass remains similar, with a slight decrease in the south Bering Sea in June. Therefore, iron supply from sea ice has an impact on the phytoplankton community structure. Changes in the phytoplankton composition may contribute to observed ecosystem shifts at high latitudes (Cooper et al., 2012; Grebmeier, 2012). The effect of sea ice on phytoplankton may grow larger with a continuous decline of sea ice in the future. Arrigo et al. (2008) suggests that the Arctic phytoplankton production may increase to 3 times the current level. Such a large increase in phytoplankton biomass would consume a large amount of iron and could cause iron limitation for phytoplankton growth. In this case, sea ice may become a more important source for bioavailable iron in the Arctic. Iron from ice may also affect denitrification at high latitudes (Arrigo et al., 2008).

3.3 Iron sources to sea ice

Three sources of iron are considered for iron accumulation in sea ice: atmospheric deposition, seawater iron, and suspended sediments in shallow regions. Dust iron and sedimentary iron are “new” to the system. Iron originally from seawater is temporarily removed and released in spring and summer. Our focus is the role of sea ice as an iron source for phytoplankton production. Therefore, we estimate the importance of

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different iron sources by comparing the iron flux from sea ice to the ocean surface during summer (Fig. 5).

Mineral dust and sedimentary iron contribute roughly equally to dissolved iron in the global ocean. However, at high latitudes atmospheric deposition is relatively low (Moore and Braucher, 2008). In the Northern Hemisphere, dust deposited on sea ice during winter contributes to 8.5×10^6 mol month^{-1} iron flux to surface seawater in summer. The flux is highest in the northwest Sea of Okhotsk (Fig. 5a). Dust from Asia is accumulated in ice over the winter, and released to the surface waters in spring. This delayed release of dust iron causes the enhanced production discussed in the previous section. The role of dust as an iron source to sea ice is weaker in the high latitude Southern Ocean. There is an averaged flux of 2.2×10^6 mol dust Fe month^{-1} from sea ice meltwater (below 45° S). Distributions of the flux are relatively uniform, but set the northern boundary of increased productivity. The total amount of dust iron in sea ice meltwater is less than 1 % of global dust deposited on the oceans.

Iron in meltwater, which originally comes from winter seawater, contributes to a flux of 1.7×10^7 mol Fe month^{-1} during the Northern Hemisphere summer. The distribution of the iron flux follows simulated iron concentrations, higher in the east Arctic and lower in the west (Fig. 5b). The flux is double the flux contributed by dust. Mean iron flux around Antarctic contributed by seawater is 1.6×10^6 mol Fe month^{-1} during summer. Though dust deposition over the seasonal ice zone is low, seawater has the smallest total contribution for iron concentrations in sea ice in the Southern Ocean. This is largely attributed to the iron depletion caused by biological uptake. Iron fluxes contributed by seawater in our study are lower compared to results of Lancelot et al. (2009), in which iron in surface water is largely transferred to sea ice during winter leading to very low concentrations in seawater. Unlike the influence of dust deposition, influences of seawater and sediments have a distinct geographical pattern. Iron from seawater mainly contributes near the Antarctic continent where ambient iron concentrations are elevated.

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Processes of incorporating sediments in sea ice and transporting sediments offshore are considered for the first time in this modeling study. Figure 5c shows iron fluxes attributed to sediments in ice. Sediments frozen in ice lead to a large iron flux into the ocean in our model. Total sedimentary iron carried by sea ice is equivalent to $\sim 8\,\%$ of monthly averaged iron fluxes from sediments in the global ocean, a flux of $2.2 \times 10^{8}\,\mathrm{mol\,Fe\,month^{-1}}$ in Arctic and $4.1 \times 10^{6}\,\mathrm{mol\,Fe\,month^{-1}}$ near Antarctic in summer. Distributions of the iron flux show that iron has been transported from the Siberian seas toward the central Arctic, which supports the hypothesis of iron being transported by "dirty" sea ice (Measures, 1999; Klunder et al., 2012).

3.4 Sensitivity to the varying iron supply from sea ice

Climate models have projected a large reduction in sea ice. It is necessary to understand the influence of declining sea ice on the iron cycle and marine ecosystems to estimate the future primary production and the ocean carbon cycle at high latitudes. There has been a significant decrease of the Arctic sea ice and large variations in Antarctic sea ice in the past decade. Simulations of this period provide insight on the sensitivity of iron distributions and phytoplankton production to changes of sea ice. We compare differences between the FULL simulation and the CTRL simulation from year to year, focusing on the role of iron released by sea ice. Changes of diatoms biomass and small phytoplankton biomass are attributed to iron from ice and its interactions with the physical environment at that time. Interannually varying effects on diatoms and phytoplankton in December from 1998 to 2007 are shown in Fig. 6.

Comparing simulations of the Antarctic Peninsula in 2002 and 2004, the mixed layer depths in the Weddell Sea and the Bellingshausen Sea are similar (Fig. 6). Ice extents around the Antarctic Peninsula are similar, but ice volume in 2004 is lower than 2002. More iron is removed from seawater and incorporated into sea ice in winter. As a result, there is more iron released in the summer of 2002. Consequently, a large increase of phytoplankton production is simulated in 2002. Differences in the region north of the Amery Ice Shelf between 2000 and 2001 also suggest the important role of iron from sea ice in phytoplankton growth. Iron concentrations are similar in this region in 2000 and 2001. The larger ice extent in 2000 releases more iron to seawater and cause a clear increase in phytoplankton biomass. Thicker and more extensive ice in the Amundsen Sea in 2001 releases more iron to the surface ocean. This is the reason why there is higher biological productivity in this region during the spring in 2001 than the following years. Changes in marine ecosystems attributed to iron from sea ice vary interannually. For example, time series of ice volume anomalies and differences of iron concentration anomalies and phytoplankton biomass anomalies in the Amundsen Sea region are shown in Fig. 7. As expected, iron concentrations anomalies are influenced by ice anomalies. Positive phytoplankton biomass anomalies are often associated with higher than normal ice volume, vise versa. Comparisons of these differences suggest that reduction in sea ice in the future will have a negative effect on biological production attributed to the influence of ice on the iron cycle.

Sea ice has a strong influence on depths of the mixed layer. Ice melting increases the stratification of surface waters, thus iron from meltwater is often released to areas with a shallow mixed layer. Light is often sufficient for phytoplankton growth in these areas and iron becomes the main limiting factor controlling production. This is why small phytoplankton biomass often shows an immediate increase along margins of ice due to enhanced iron fluxes from sea ice (Fig. 6). However, iron fluxes from ice may not be the dominant control of phytoplankton production. For example, iron fluxes from ice are similar in the Ross Sea in 2006 and 2007. However, the increase of diatoms biomass is larger in 2006. The remaining iron in surface water is higher in 2007, which suggests phytoplankton growth is limited by lower light due to the thicker ice in 2007.

Iron sequestration in sea ice affects the competition between different phytoplankton functional groups. Changes of marine ecosystems in the Atlantic sector of the Southern Ocean show a clear example of impacts of varying sea ice on phytoplankton community structure (Fig. 6). Sea ice drifts from the Antarctic continent to the north in this region (Holland and Kwok, 2012). When sea ice forms in iron-rich nearshore waters, it removes iron from seawater causing decreases of iron concentrations in the surface

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ocean. Northward sea ice motion then transports iron away from continental areas. As a result, there is a decrease of diatoms biomass nearshore, especially during heavy ice years, and an increase of small phytoplankton biomass in the north (Fig. 6). Therefore, increases of sea ice in this region will cause phytoplankton community shifts from diatoms to small phytoplankton, and vise versa.

Increases of iron concentrations are generally concurrent with increases of sea ice formation in the Arctic. If the decreasing trend of Arctic sea ice continues, iron concentrations in the Arctic will decline due to lower iron inputs. Impacts of iron sequestration in sea ice on marine ecosystems are more confounding in simulations between 1998 and 2007. Phytoplankton production is usually limited by light intensity and macronutrient (nitrogen and silicon) concentrations in the mixed layer. Though iron is a limiting factor for phytoplankton growth in some areas of the Bering Sea and the Sea of Okhotsk, changes of phytoplankton production are not tightly coupled with iron fluxes from ice. The decoupling between iron fluxes and production is likely to happen, when strong ice melting causes a dilution of other nutrients. This is because concentrations of macronutrients in sea ice are usually lower than seawater (Lin et al., 2011; Tovar-Sanchez et al., 2010). Thus, the biogeochemical impacts of melting ice depend on the surrounding ocean environment. Longer simulations forced with future climate conditions are needed for predicting the possible impact of iron fluxes from sea ice on marine ecosystems.

4 Discussion and summary

Iron sequestration and transport by sea ice are incorporated into the CESM-BEC model to study the role of sea ice in the marine iron cycle and the impacts on biology. We also include suspended sediments as another source of iron to sea ice in shallow regions, which can advance our understanding of the impacts of "dirty" ice on transporting iron and other materials.

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The modified model reasonably captures the general pattern of iron distributions in the high latitude oceans. Compared to the previous BEC model, simulated iron concentrations are in better agreement with observations in central Arctic, the Bering Strait, the Chukchi Sea, and the Canada Basin. Biases in iron concentrations are associated with biases in phytoplankton production. Simulated maximum iron concentrations in sea ice are in the west Antarctic Peninsula (134 nM dFe) and in the Bering Strait and the Chukchi Sea (192 nM dFe). Our simulation cannot reproduce the highest iron concentrations (up to 863.9 nM dFe) observed in the field (Tovar-Sanchez et al., 2010). Previous work suggests that the continuous growth of ice algae at the bottom of sea ice during winter also contributes to the accumulation of iron in sea ice (Lannuzel et al., 2010). As such, the model bias may be improved by incorporating an ice algae component to the model (Deal et al., 2011) or by including other biological iron sources in the sea ice. Our assumption of unified iron distributions in each sea ice cell may also lead to the problem of missing high iron concentrations. The sea ice microstructure and brine dynamics have strong impacts on tracer distributions in ice (Jeffery et al., 2011; Vancoppenolle et al., 2010). Interactions between iron and brine dynamics should also be considered in future studies.

Our results show that sea ice can change the timing and location of iron supply to the ocean. Suspended sediments are an important iron source to sea ice and can significantly affect iron distributions in the Arctic. Iron released from sea ice melting enhances phytoplankton production in the Southern Ocean, the Bering Sea and the Sea of Okhotsk. The magnitude of simulated increases of production may be underestimated, because high iron concentrations in ice are not reproduced in this study. The simulated influence of ice on phytoplankton agrees with conclusions from previous studies (Measures, 1999; Lannuzel et al., 2008; Sedwick and DiTullio, 1997). The ecosystem structure can be changed by iron from sea ice, because different phytoplankton groups respond differently to iron inputs.

Simulations of 1998 to 2007 show that decreases in sea ice have negative impacts on phytoplankton growth and production due to interactions with the iron cycle in the

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Southern Ocean. Iron in sea ice has small impacts on total phytoplankton production. Iron concentrations will likely decrease in the central Arctic as summer sea ice cover continues to decline in the coming decades. Given the apparent importance of sea ice transport in bringing iron and other trace metals to the central Arctic, iron limitation of phytoplankton growth rates is possible in a future, ice-free Arctic Ocean during summer months. Declining sea ice cover also modifies the light regime experienced by phytoplankton and ocean circulation, mixed layer depths, and stratification. Marine ecosystems are affected by the combination of all of these factors (Pabi et al., 2008; Arrigo et al., 2008; Sedwick and DiTullio, 1997; Lannuzel et al., 2008; Lee et al., 2012). The combined physical and biogeochemical effects of sea ice on marine ecosystems need to be considered holistically in the future in the context of the ongoing perturbations of the climate system.

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Table 1. Description of model experiments.

Simulations	Descriptions	Iron in sea ice
CTRL	No iron sequestration in sea ice	No
NOdust	Dust deposited over ice is blocked and lost.	No
FEdust	Dust deposited over ice is stored in ice and released during melting	Yes
FEsw	Dissolved iron is transported from seawater to sea ice during ice formation. Modifications in FEdust is included.	Yes
FULL	Sedimentary iron trapped in sea ice is considered, as well as modifications in FEsw	Yes

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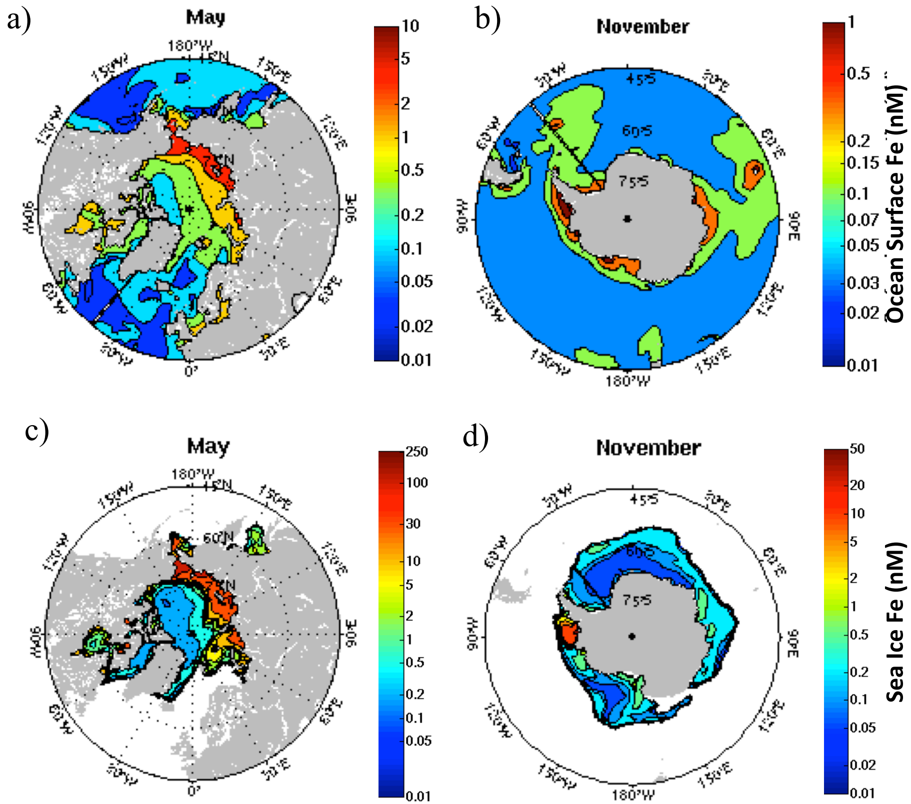

Fig. 1. Simulated iron concentrations in **(a, b)** seawater and **(c, d)** sea ice in **(a, c)** Arctic and **(b, d)** Antarctic. Results are iron concentrations in May in Arctic and November in Antarctic, when phytoplankton biomass starts increasing rapidly. The color scale is logarithmic. **(c)** and **(d)** show data where ice concentration is greater than 15 % in simulation.

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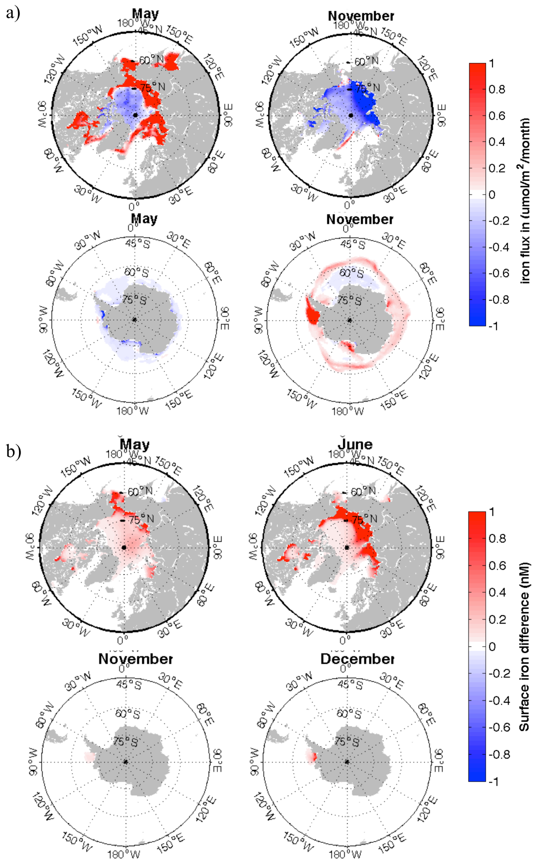

Fig. 2. Differences between the CTRL and the FULL simulations (FULL – CTRL): **(a)** The net iron flux to the surface ocean in May and November. Negative values indicate iron is primarily removed from seawater due to ice formation. Positive values indicate enhanced iron input to the surface ocean due to effects of sea ice. **(b)** Mean surface iron concentrations during the growing season.

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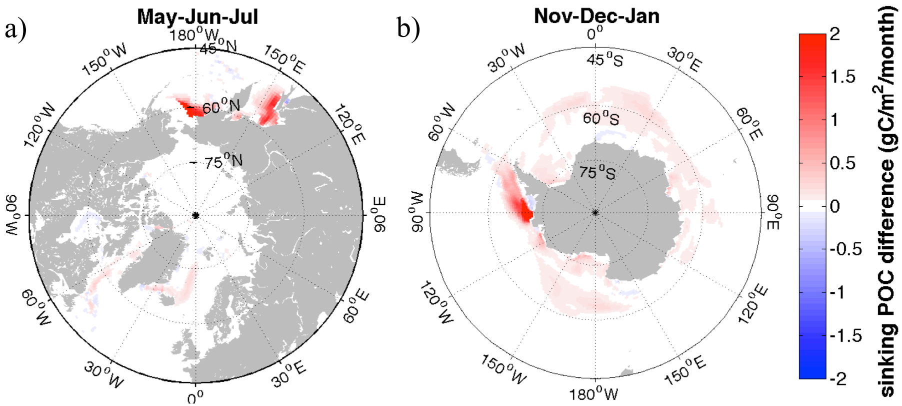

Fig. 3. Impacts of iron supply from sea ice on export production during growing season: **(a)** May, June, and July in the Northern Hemisphere; **(b)** November, December, and January in Southern Hemisphere. The figure shows differences of export production between the FULL and the CTRL simulations.

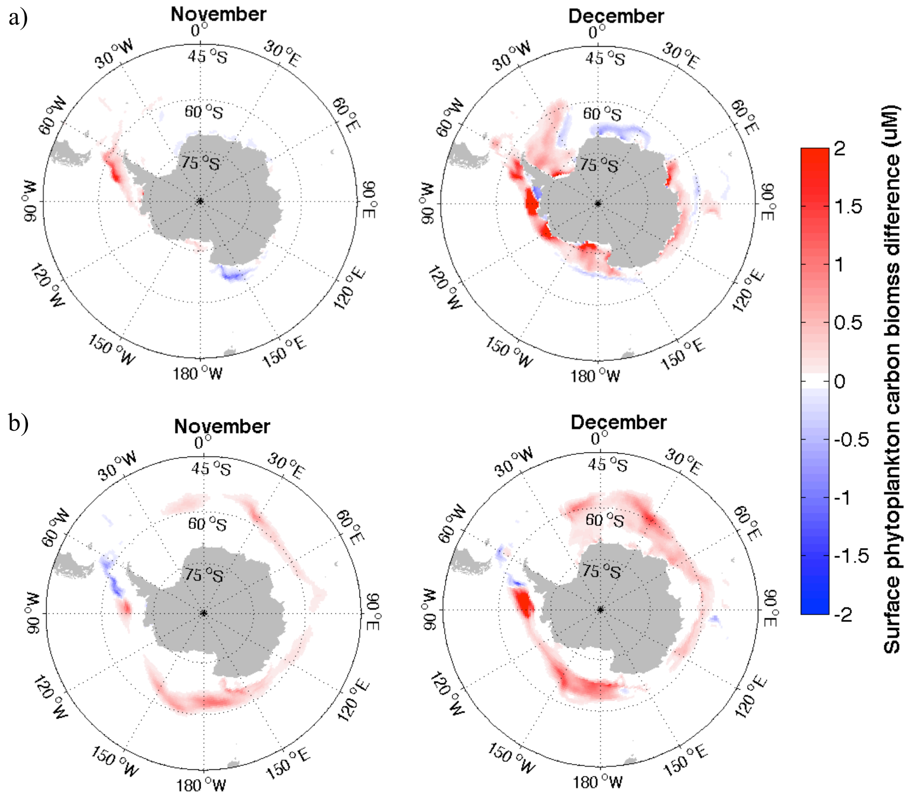

Fig. 4. Impacts of sea ice on phytoplankton biomass during the growing season in the Southern Ocean: **(a)** diatoms; **(b)** small phytoplankton. The figure shows differences of phytoplankton biomass between the FULL and the CTRL simulations.

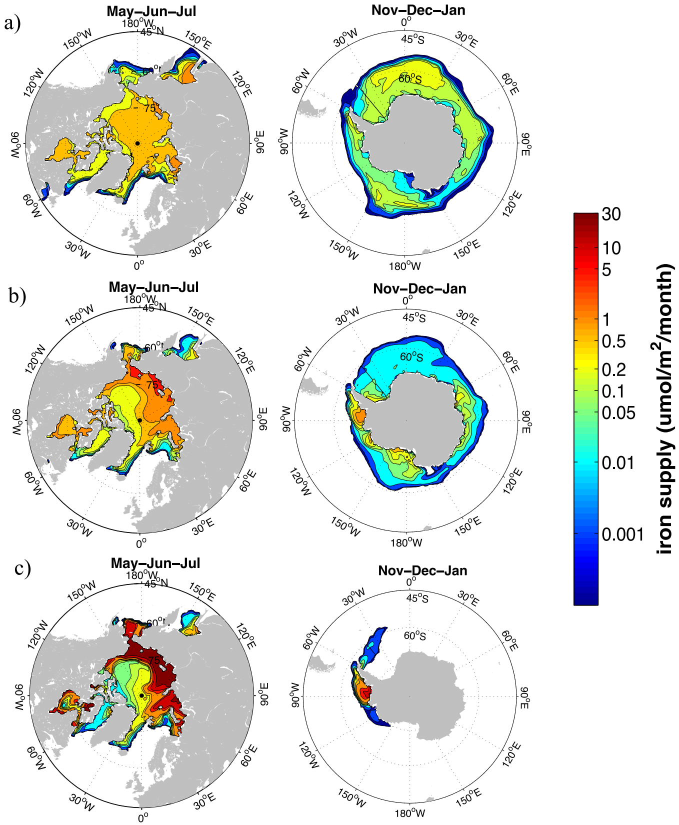

Fig. 5. Iron supply from sea ice contributed by different iron sources in summer: **(a)** dust, **(b)** seawater, and **(c)** sediments. Contributions of different iron sources are differences of the net iron flux into the surface ocean (FEdust – NOdust, FEsw – FEdust, and FULL – FEdust, respectively).

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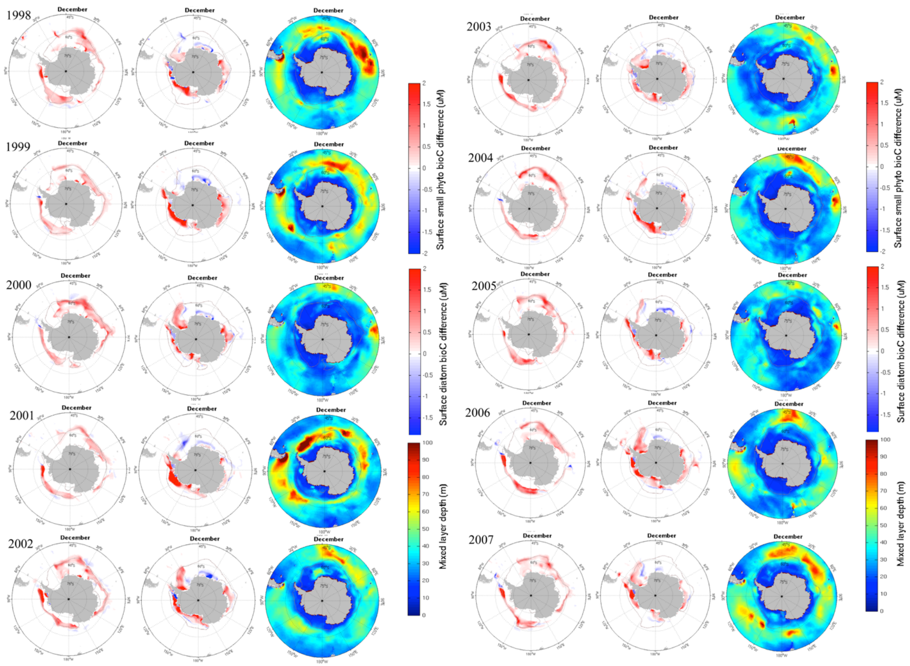

Fig. 6. Differences of biomass in December caused by iron fluxes from ice (from 1998 to 2007) (FULL – CTRL): (left) small phytoplankton and (middle) diatoms. The mixed layer depth in December is shown on the right. The grey line shows 15 % ice concentration.

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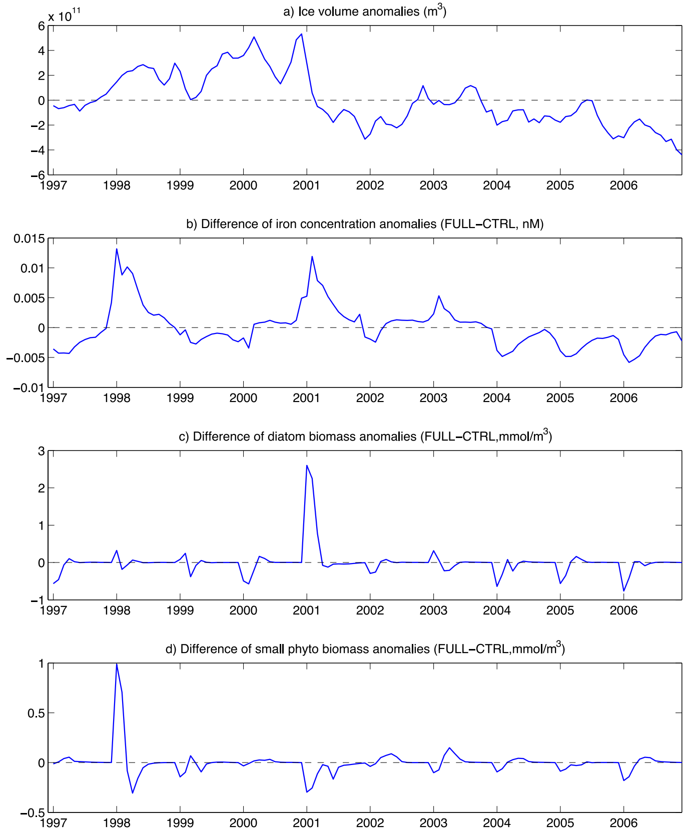

Fig. 7. Time series of the Amundsen Sea region (100–130° W, Antarctic continent to 65° S): **(a)** anomalies of ice volume (i.e., 10 yr mean seasonal cycle removed); **(b)** Differences of iron concentration anomalies (FULL – CTRL); **(c)** Differences of diatom biomass anomalies (FULL – CTRL); **(d)** Differences of small phytoplankton biomass anomalies (FULL – CTRL). Differences between the FULL and the CTRL show effects of iron sequestration in sea ice.

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