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Unusual subpolar North Atlantic phytoplankton bloom in 2010: Volcanic fertilization or North Atlantic Oscillation?

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[1] In summer and autumn 2010, a highly anomalous phytoplankton bloom, with chlorophyll concentration more than double that of previous years, was observed in the Irminger Basin, southwest of Iceland. Two unusual events occurred during 2010 which had the potential to promote the unusual bloom. First, in spring 2010, the Eyjafjallajökull volcano in Iceland erupted, depositing large quantities of tephra into the subpolar North Atlantic. Second, during the winter of 2009/2010 the North Atlantic Oscillation (NAO) became extremely negative, developing into the second strongest negative NAO on record. Hydrographic conditions were highly anomalous in the region, with an influx of freshwater spreading through the basin, and unusual nutrient and mixed layer depth conditions. Here we use a combination of satellite, modeled and in situ data to investigate whether the input of iron from the volcanic eruption or change in hydrographic conditions due to the extreme negative NAO were responsible for the anomalous phytoplankton bloom. We conclude that changes in physical forcing driven by the NAO, and not the volcanic eruption, stimulated the unusual bloom.

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1. Introduction

[2] In April/May 2010, the Eyjafjallajökull volcano in Iceland erupted, disrupting air travel throughout Europe and producing an estimated 270 million cubic meters of airborne tephra, of which roughly half fell in Iceland, with much of the remainder deposited in the surface waters of the North Atlantic [*Gudmundsson et al*., 2012; *Karlsdóttir et al*., 2012]. Because volcanic ash contains iron, an important micronutrient for phytoplankton growth, eruptions are postulated to have potentially large effects on oceanic primary production (PP) [*Duggen et al*., 2010]. Marine sediment records contain evidence of volcanic ash deposition events throughout Earth's history, and results from field investigations suggest that these events may have been responsible for large-scale climatic changes [*Robock*, 2000]. The eruption of Mt. Pinatubo in 1991, for example, coincided with a drawdown of CO_2 in the Northern Hemisphere, hypothesized to be driven by increased PP stimulated by natural iron fertilization following ash deposition [*Sarmiento*, 1993]. Satellite-derived ocean color data have recently been used to investigate localized increases in PP following volcanic eruptions in Mikakejima, Japan [*Uematsu et al*., 2004], Montserrat [*Duggen et al*., 2007], Kasatochi, Alaska [*Langmann et al*., 2010], and Anatahan, Mariana Islands [*Lin et al*., 2011].

[3] Another unusual event with the potential to alter phytoplankton bloom dynamics in the subpolar North Atlantic also occurred in 2010. The NAO index has been mostly in a positive state since the early 1970s, with occasional negative years (supporting information Figure S1). However, the winter of 2009/2010 had exceptional NAO conditions, being the second strongest negative NAO on record. Negative NAO conditions are typified by reduced westerly winds and a southward displacement of the North Atlantic storm track. The altered heat fluxes result in a tripolar structure in sea surface temperature, being warmer in subpolar and subtropical regions and cooler at midlatitudes [e.g., *Visbeck et al*., 2003]. A rapid wind-driven response in surface currents follows interannual variations in the NAO. In negative NAO periods, the North Atlantic Current shifts eastward and heat and freshwater transports into the northern North Atlantic increase with the expansion of the subtropical gyre [*Bersch*, 2002].

[4] The role of the NAO in driving interannual variability in phytoplankton populations has been explored in both the Continuous Plankton Recorder data [*Reid et al*., 1998;

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Edwards et al., 2001] and in biogeochemical modeling studies [*Henson et al.*, 2009; *Patara et al.*, 2011]. The sustained positive NAO period in the 1980s and 1990s resulted in reduced phytoplankton populations in both the eastern and western North Atlantic [*Edwards et al.*, 2001]. Correlations at the decadal scale between the NAO and phytoplankton abundance are fairly weak [*Barton et al.*, 2003]; however, interannual variability in mixing driven by NAO-induced changes in heat and freshwater fluxes are well correlated with phytoplankton bloom timing and magnitude [*Henson et al.*, 2009]. Negative NAO periods typically exhibit a decrease in phytoplankton abundance in the subtropics and an earlier, enhanced bloom in subpolar regions [*Patara et al.*, 2011].

[5] We observed a widespread, anomalously strong phytoplankton bloom in satellite ocean color data throughout the Irminger Basin in the subpolar North Atlantic during spring and summer 2010. The Irminger Basin lies between Greenland and Iceland and is a site of occasional deep convection [*Bacon et al.*, 2003]. Phytoplankton dynamics in the basin are typical of the classic North Atlantic spring bloom (in the sense of *Sverdrup* [1953]), with deep mixed layers in winter ensuring that phytoplankton growth is light limited [*Henson et al.*, 2006a]. The spring bloom is initiated by a shoaling mixed layer associated with a period of net heat flux into the ocean [*Waniek and Holliday*, 2006; *Henson et al.*, 2006a]. Typically, diatoms dominate the initial stages of the bloom, becoming outcompeted by smaller flagellates as silica is depleted [*Henson et al.*, 2006b; *Heath et al.*, 2008]. However, unlike most regions of the subpolar North Atlantic, nitrate is not completely depleted during the growing season, with residual concentrations of $\sim$2–4 μmol l^{-1} at the end of summer [*Henson et al.*, 2003]. This observation has led to the suggestion that phytoplankton growth is iron limited in summer in the Irminger Basin [*Sanders et al.*, 2005]. To directly test this hypothesis, two cruises to the region in spring and summer 2010 were planned as part of the Irminger Basin Iron Study (IBIS) program. The eruption of the Eyjafjallajökull volcano occurred coincidentally just prior to the first cruise allowing the investigation of the influence of volcanic ash deposition on the biogeochemistry of the region [*Achterberg et al.*, 2013].

[6] In addition to the two cruises in 2010, the Irminger Basin was also visited in spring and summer 2002, and so here we are able to compare the unusual conditions in 2010 to the more typical conditions observed in 2002. We use a combination of satellite data, in situ data and model output to investigate both unusual forcing events—the volcanic eruption and the extreme negative NAO—as possible mechanisms to explain the anomalous bloom.

2. Data and Methods

[7] MODIS Aqua Level-3 satellite-derived chlorophyll concentration (chl) data (R2012.0) at 8 day, 9 km resolution for July 2002–December 2010 were obtained from NASA (http://oceancolor.gsfc.nasa.gov/). Satellite-derived primary production PP estimates were made using the standard Vertically Generalized Productivity Model [*Behrenfeld and Falkowski*, 1997]. Volcanic ash carried in the atmosphere can interfere with satellite ocean color retrievals, resulting in a false increase in chl [*Claustre et al.*, 2002]. To investigate whether airborne ash may have interfered with the satellite chl retrieval, the time series of MODIS Level-3 aerosol optical thickness (AOT) data in the Irminger Basin was examined. This showed only two 8 day periods during which AOT was greater than the range found in previous years (1–8 May and 18–25 June; supporting information Figure S2). The first period occurs during the active eruption phase of the volcano, but the elevated AOT does not coincide with an anomalous increase in chlorophyll (Figure 1b). The second period of unusually high AOT occurs more than 1 month after the eruption ended, and again does not correspond with anomalously high chl. MODIS Level-3 processing by NASA has a high level of quality control, with 27 flags including those for failed atmospheric correction and failed chlorophyll retrieval. In this case, we conclude that the NASA processing successfully excluded the majority of airborne ash-affected pixels from the Level-3 chl product. In situ chl measurements from 2002 and 2010 in the central Irminger Basin (Figure 1c) agree well with the satellite values, giving further confidence that volcanic ash is not unduly impacting the satellite-derived chl data.

[8] Monthly wind data for 1998–2010 were obtained from the NCEP/NCAR Reanalysis project [*Kalnay et al.*, 1996]. Monthly optimally interpolated fields of salinity and temperature for 1998–2010 were obtained from the Hadley Centre (EN3 data set derived from a variety of data sources, including Argo floats and the World Ocean Database). Full details of the quality control and data processing can be found in *Ingleby and Huddleston* [2007]. Density was calculated from temperature and salinity fields using the International Equation of State for Seawater (IES80). Mixed layer depth was estimated from individual Argo float profiles collected in the central Irminger Basin from 2001 to 2011 using the *Holte and Talley* [2009] density algorithm.

[9] Two cruises to the Irminger Basin took place aboard the *RRS Discovery* during (26 April–9 May 2010; D350) and after (4 July–11 August 2010; D354) the Eyjafjallajökull eruption as part of the IBIS program [*Achterberg et al.*, 2013]. Here we compare nutrient and chl concentrations measured on the 2010 cruises to those measured during the UK Marine Productivity program to the same region in 2002 [*Holliday et al.*, 2006]. The 2002 cruises took place during similar periods as in 2010 (18 April–27 May 2002; D262 and 25 July–28 August 2002; D264). On all cruises, macronutrients were determined using a Skalar San Plus autoanalyser [*Kirkwood*, 1996] and samples for chl were filtered through 25 mm Whatman GF/F filters, extracted in 90% acetone and determined fluorometrically. Surface chl samples were taken from the ship's continuous thermosalinograph outflow from an intake of 5 m depth. Nutrient data are reported as integrated euphotic zone values, where euphotic depth was estimated from satellite chl using the method of *Morel* [1988]. The temperature and salinity characteristics of the Irminger Basin in 2010 are compared to data collected south of Greenland in 2008 (21 August–25 September 2008; D332) [*Bacon*, 2010]. A map showing all the cruise stations used in this study is presented in supporting information Figure S3. The "central Irminger Basin" is defined here as the region bounded by the 1500 m depth contour and north of 58°N, east of 42°W.

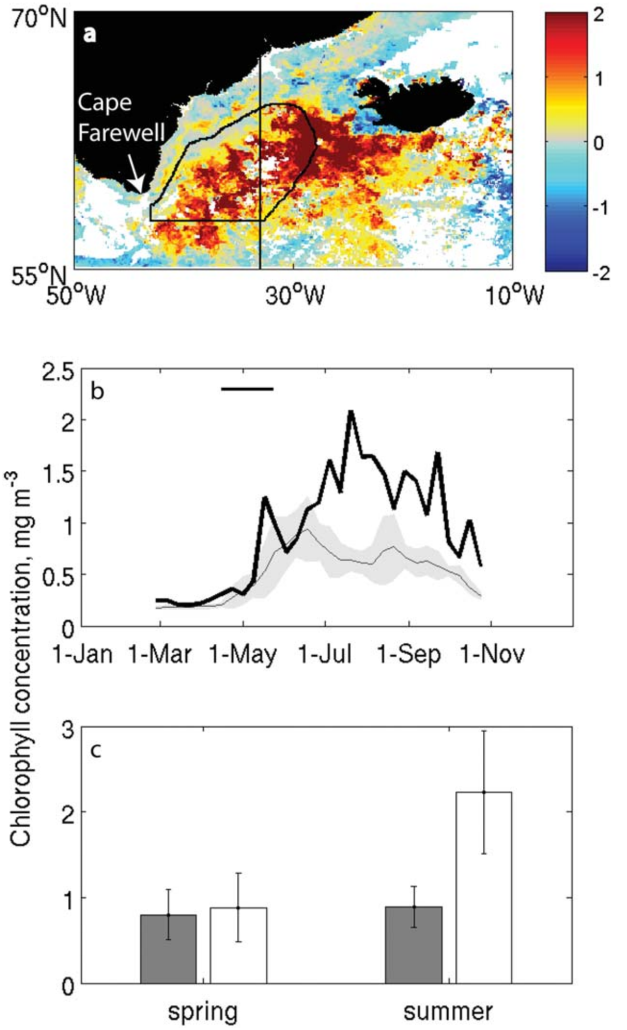

Figure 1. Anomalous chlorophyll conditions in the Irminger Basin in 2010. (a) Chlorophyll anomaly (mg m^{-3}) for 12–19 July 2010, relative to the mean of the same period in 2003–2009 and 2011. The region denoted in this study as the "central Irminger Basin" is outlined in black. The position of the transect plotted in Figure 3 is marked with a vertical black line. (b) Satellite-derived time series of chlorophyll concentration in the central Irminger Basin in 2010 (thick black line), the mean (thin gray line), and standard deviation (gray shading) of years 2003–2009 and 2011. Black horizontal line marks the period of the volcanic eruption. (c) In situ measured chlorophyll concentration (collected from the underway system at ~5 m depth) in the central Irminger Basin collected on cruises in spring and summer 2002 (gray bars) and 2010 (white bars). Error bars denote the spatial variability of data within the basin (standard deviation).

Results reported for the central Irminger Basin represent a spatial average of all measurements made in this region.

3. Results

3.1. Chlorophyll Concentration

[10] Satellite-derived chlorophyll images reveal the extent of the anomalous bloom in the Irminger Basin in summer 2010. In early to mid-July, the central Irminger Basin experienced unusually high chl concentrations, as revealed in the anomaly map in Figure 1a. The chl concentration in individual pixels was as much as 5 mg m^{-3} greater than the mean of previous years. Examination of the time series of satellite chl data averaged over the central Irminger Basin (Figure 1b), shows that the unusual chl signal in 2010 was a function of the anomalous timing of the bloom peak, in addition to the extended period of high chl concentration, which continued throughout the autumn. The exceptionally high chl concentration in summer 2010 is confirmed by comparison of in situ chl samples with the conditions observed in 2002 (Figure 1c). Although spring chl was similar in both years (0.8–0.9 mg m^{-3}), chl was significantly higher in summer 2010 (2.2 mg m^{-3}) than in 2002 (0.9 mg m^{-3}). In 2003–2009 and 2011, the evolution of chl through the year follows the typical North Atlantic bloom pattern with a rapid increase in chl in late April/early May, followed shortly after by maximum chl values of ~1 mg m^{-3} in late May/early June. A reduction in chlorophyll during summer is followed by a smaller autumn bloom in late August. In 2010, the bloom progresses in an altogether different manner. The bloom starts as normal in early May, but instead of rapidly reaching a peak before dying back in summer, the chl concentration continues to rise, finally reaching its peak in mid-July, with elevated concentration persisting through autumn 2010. Chlorophyll concentrations then returned to normal levels in 2011. Clearly, the normal seasonal cycle of phytoplankton growth was perturbed in the Irminger Basin in 2010, but was this due to changes in physical forcing or a volcanic fertilization effect?

3.2. Hydrographic Conditions

[11] The NAO in winter 2009/2010 was in an extreme negative phase (supporting information Figure S1). During negative phases of the NAO, surface pressure increases in the Icelandic low and decreases in the Azores high, resulting in weaker westerly winds and a southward shift of the storm track [*Visbeck et al*., 2003]. In the case of 2010, the wind during December 2009 to June 2010 is either northward (in January and March) or eastward (in February and May), that is, in the opposite direction entirely, compared to mean conditions. Wind speed is also exceptionally high in May 2010 (Figure 2).

[12] The hydrographic conditions in the Irminger Basin in 2010 were also anomalous. Transects of density anomaly along 33°W (Figure 3) show a fresh anomaly intruding into the Irminger Basin from ~April onward, inundating the entire basin by August and persisting until December. The anomaly extends over only the top 50–80 m of the water column, that is, it likely originates from anomalous surface, or near-surface, conditions. Theta-S plots from 2010 provide insight into the likely source of the anomalous water mass in the central Irminger Basin (Figure 4). The theta-S characteristics of the upper ocean in the central Irminger Basin in 2010 are very similar to data collected along a transect south of Cape Farewell in early September 2008 [*Bacon*, 2010]. Although the 2010 surface waters are warmer than in 2008, they occupy a theta-S space that is absent in the 2008 data; that is, a mixing region between the typical central Irminger Basin surface layer (salinity of

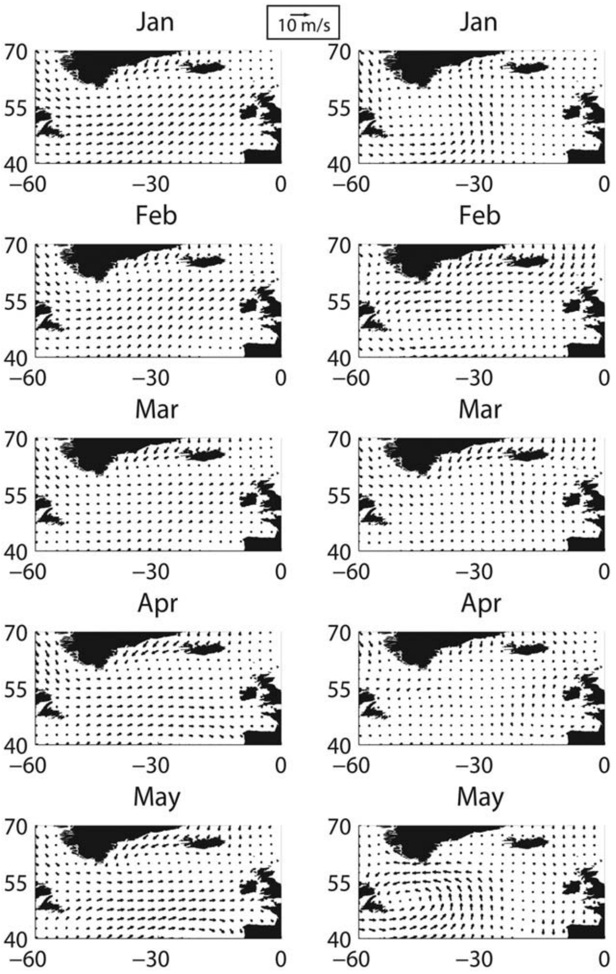

Figure 2. (left) Mean (1998–2009) wind speed over the North Atlantic in selected months (at 10 m above sea level). (right) Wind speed in selected months of 2010. Note that the wind direction in February 2010 is reversed with respect to the mean and that northward wind speeds in May 2010 are anomalously strong.

~34.87–35.05, as in 2008) and the central gyre fresh pool (salinity of ~34.65–34.75). A plot of isohalines in the region also demonstrates the anomalous extent of freshwater in the Irminger Basin in 2010 (Figure 5). In March, the location of the isohaline (salinity of 35.02) in 2010, relative to previous years, is unusually far east at 52–54°N. By June 2010, the isohaline is further north and east than is typical at 58–60°N, indicating a significant change in surface water properties, and by September 2010 the isohaline is north and east of its usual position, indicating anomalously fresh water has inundated a large portion of the Irminger Basin. The unusual hydrographic conditions are also reflected in the mixed layer depth, which remained shallower than normal throughout the autumn (Figure 6).

[13] The freshwater intrusion appears to have also introduced anomalous macronutrient conditions, particularly in the southern Irminger Basin. Comparison of nutrient concentrations during the 2010 cruises to similarly timed spring and summer cruises in 2002 suggest that nitrate concentration in spring was similar in the 2 years, but the summer nitrate was lower in summer 2010 than 2002 (Figure 7). The silica concentration was slightly lower in spring 2010 than in 2002, but similar during the summer cruise (Figure 7). The ratio of N : Si consumed between the spring and summer cruises was 1 : 1 in 2002, the expected value for a diatom dominated bloom [*Brzezinski*, 1985], but was 3 : 1 in 2010.

3.3. Volcanic Dust Deposition

[14] In addition to the unusual hydrographic conditions observed in 2010, the Eyjafjallajökull volcano in Iceland was actively erupting between 15 April and 23 May. The ash ejected from the volcano was approximately 7.6% iron oxide by weight [*Achterberg et al.*, 2013] and so could potentially have added substantial quantities of iron to the surface ocean. Estimates of the iron deposition over the region were made by *Achterberg et al.* [2013] using an atmospheric Lagrangian particle dispersion model for the spatial distribution of ash deposition [*Stohl et al.*, 2011] and estimates of ash sinking rates and modeled iron dissolution. A map of iron deposition from the volcanic eruption (Figure 8a) shows that a maximum of ~0.08 nmol l^{-1} of iron were likely deposited in the very northern part of the Irminger Basin, and ~0.005 nmol l^{-1} in the central basin, over the course of the eruption. The time series of iron deposition in this region (Figure 8b) shows that the vast majority was deposited on just 1 day: 14 May 2010.

[15] Chlorophyll concentrations and hydrographic conditions returned to normal in 2011, and the NAO index, although still negative, was within the range of previous winter conditions (supporting information Figure S1). This implies that the anomalous conditions of 2010 resulted from a one-off perturbation to the system, rather than a prolonged regime shift, and that winter mixing in 2010/11 "reset" the system for the following year's spring bloom.

4. Discussion

[16] Satellite-derived chl from the Irminger Basin typically depicts a short, intense bloom in spring, followed by a summer minimum [*Henson et al.*, 2006a]. However, 2010 exhibited highly anomalous conditions, with an extended period of high chl persisting from early May through October, with a peak in mid-July. Total annually integrated primary production in the Irminger Basin, estimated from satellite-derived data was ~50% greater in 2010 than the mean of 2003–2009 and 2011, implying potentially significant impacts on CO_2 drawdown and carbon export.

[17] The atmospheric conditions in winter 2009/2010 reflect the "Warm Arctic-Cold Continents" climate pattern, which has occurred only three times in the last 160 years [*Overland et al.*, 2010]. The particularly cold winter in northern Europe and the eastern USA in 2009/2010 was in contrast to exceptionally warm Arctic temperatures, e.g., in Baffin Bay, northeastern Canada, where winter air temperature was ~8°C warmer than usual [*Overland et al.*, 2010]. These conditions resulted in increased ice melt over much of the Arctic [*Stroeve et al.*, 2011]. As indicated by the unusual progression of isohalines (Figure 5) and the theta-S characteristics of waters in the southern central Irminger Basin in 2010 (Figure 4), this upper ocean freshwater signal is likely to have originated from south of Cape Farewell

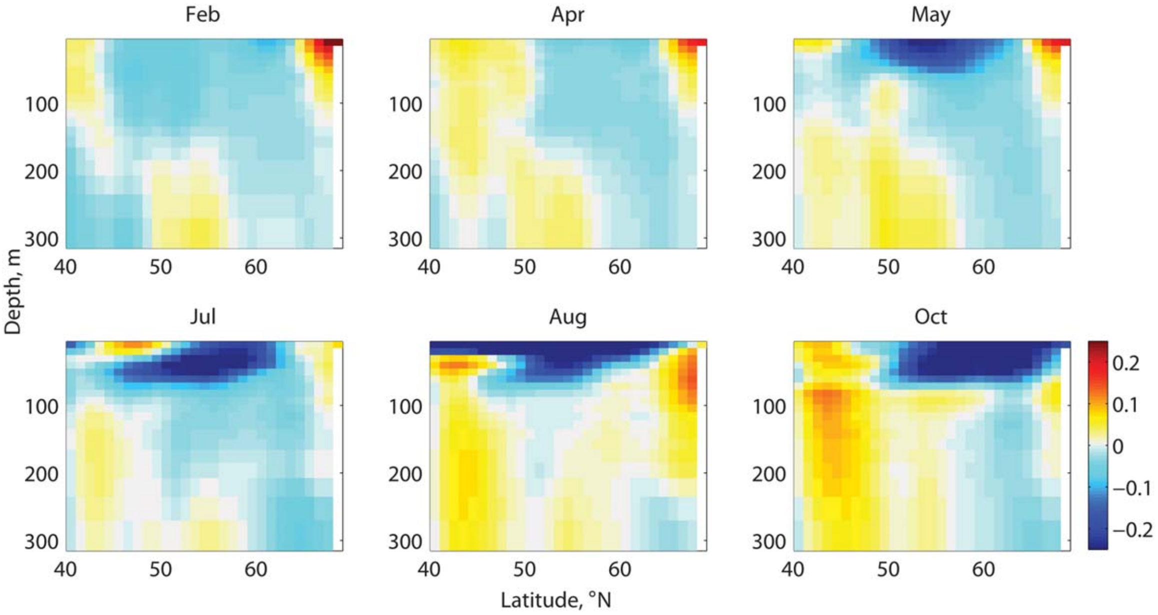

Figure 3. Density anomaly (kg m^{-3}) from the Hadley Centre EN3 data set for selected months in 2010, relative to the mean of 1998–2009, along a transect at 33°W. Note the presence of an anomalously light, fresh water mass entering the Irminger Basin during May, capping the entire region by August and persisting until October.

where a pool of relatively fresh water typically resides [*Reverdin et al*., 2002; *Holliday et al*., 2007]. The region of the freshwater pool is strongly influenced by the Labrador Current that flows south-eastward, transporting very fresh water from Davis Strait to Newfoundland. Data from Lagrangian drifters have demonstrated that the Labrador and North Atlantic currents intensified under the negative NAO conditions of 1996, although the recirculation in the Irminger Basin weakened [*Flatau et al*., 2003]. In addition, an increased northward transport of heat and freshwater

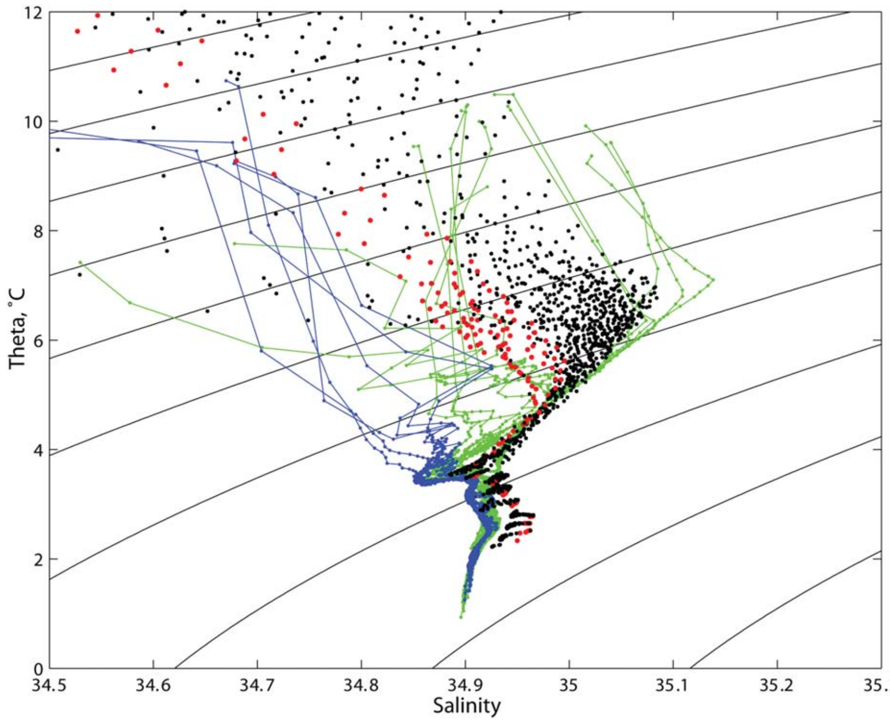

Figure 4. Theta-S plot showing data from early September 2008 along a transect south of Cape Farewell (blue lines) and in the southern Irminger Basin along 59°N (green lines). Also plotted are data from August 2010 in the central (black dots) and southern (55–56°N) Irminger Basin (red dots).

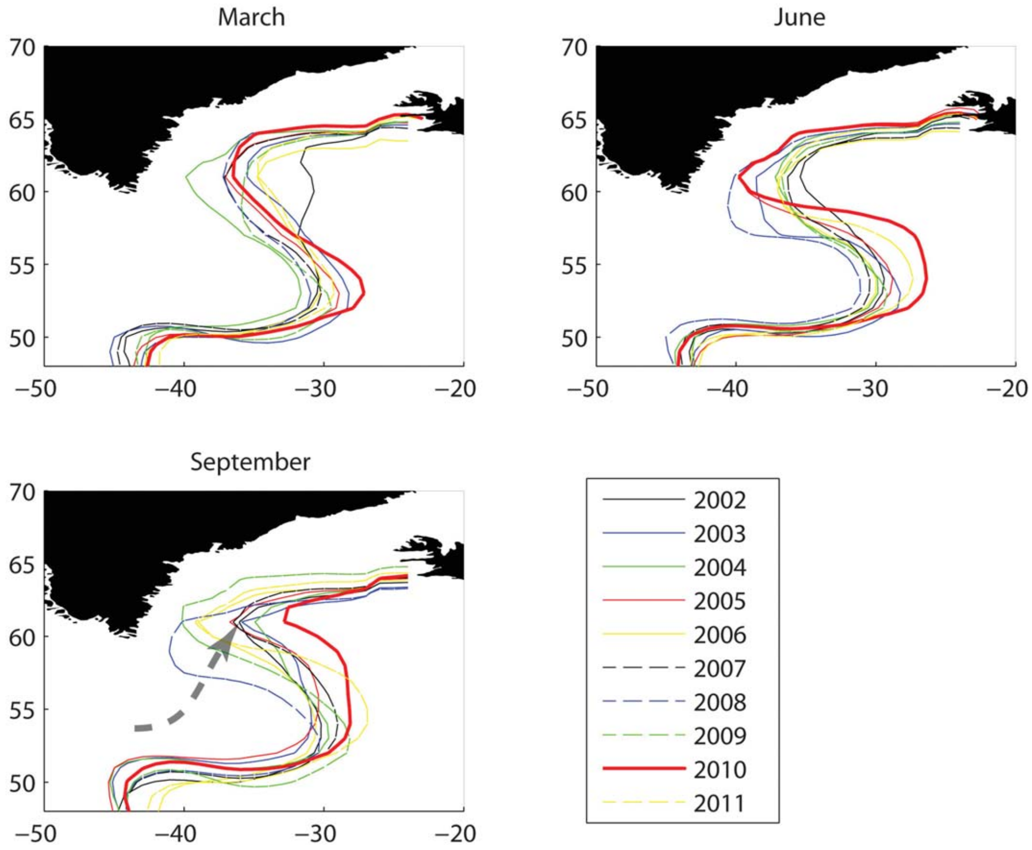

Figure 5. Position of isohalines (salinity = 35.02) in March, June, and September for years 2002–2011. The position of the isohaline in 2010 is marked with a thick red line. Note that in 2010 the appearance of the freshwater anomaly perturbs the isohaline to the northeast, suggesting the water mass originated to the southwest. Dashed gray arrow in last plot is a sketch of the hypothesized path of the freshwater intrusion.

into the Irminger Basin along the western side of the mid-Atlantic Ridge was observed [*Bersch*, 2002]. In 2010, the extreme negative NAO may have resulted in intensification of the Labrador Current and hence freshwater influx to the central subpolar gyre. Anomalous wind conditions in 2010 may have then driven the fresh surface waters northeastward, overcapping the central Irminger Basin.

[18] The freshwater intrusion may also have altered nutrient conditions in the region, as Labrador Current water has an excess of both phosphate and silicate relative to nitrate [*Harrison and Li*, 2007] and transports Arctic species of plankton south to the Newfoundland shelf [*Head and Pepin*, 2010]. The unusual nutrient drawdown ratios shown in Figure 7 arise partly from the lower euphotic zone silica concentrations in spring 2010, relative to nitrate and phosphate, which seems counter to a Labrador Current origin for the freshwater anomaly. Historical macronutrient data extracted from the World Ocean Database for 1980–2004 suggests that silica concentrations in spring 2010 were indeed lower than the mean, suggesting that the diatom bloom had already begun prior to the spring cruise. Silica concentrations in the central Irminger Basin were, in general, around average in summer 2010 (supporting information Figure S4), except along the southern transect (Figure 7f). The anomalously high silica concentrations in the southern Irminger Basin may be a remnant signature of the enhanced nutrient concentrations introduced to the basin with the freshwater anomaly. The anomalous nutrient supply conditions are also reflected in the unusual spring to summer N:Si drawdown ratio, which was 3:1 in 2010

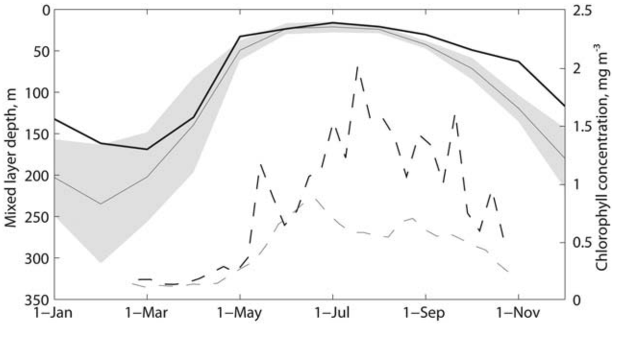

Figure 6. Monthly mixed layer depth in the central Irminger Basin in 2010 (thick black line), the mean (thin gray line), and standard deviation (gray shading) of years 2001–2009 and 2011. Note that the mixed layer in 2010 is shallower than the mean in winter and throughout autumn. Also plotted are chlorophyll concentration in 2010 (black dashed line) and the mean of 2003–2009 and 2011 (gray dashed line) taken from Figure 1b.

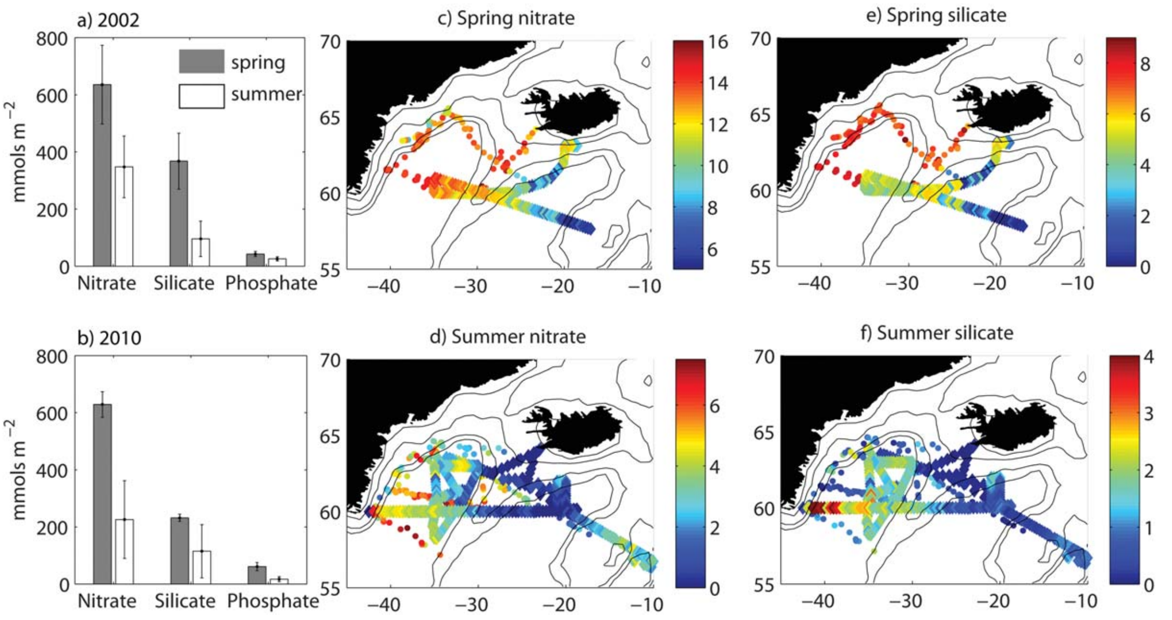

Figure 7. Comparison between integrated euphotic depth macronutrients in the central Irminger Basin in (a) 2002 and (b) 2010 from cruises in spring (gray bars) and summer (white bars). Error bars denote the spatial variability of data within the basin (standard deviation). (c) Nitrate and (e) silicate concentrations (mmol m^{-3}) collected underway at ~5 m depth on cruises in spring 2002 (small dots) and 2010 (diamonds), and similarly in summer for (d) nitrate and (f) silicate.

rather than the typical 1:1, suggesting that either the seasonal consumption by phytoplankton of silica relative to nitrate was reduced in 2010 relative to 2002, or that an additional input of silica occurred between May and August, perhaps associated with the freshwater anomaly. This latter hypothesis is consistent with our observation that the spring cruise occurred before the freshwater had fully inundated the basin (Figures 3 and 5), but by the time of the summer cruise, the freshwater anomaly had arrived at ~60°N coinciding with higher than normal silicate concentrations in the southern central Irminger Basin (Figures 5 and 7).

[19] A change in the N:Si drawdown ratio may also imply a change in phytoplankton community structure in 2010, from the typical diatom dominated population expected in spring in the Irminger Basin [*Sanders et al.*, 2005] to a phytoplankton population that consumed nitrate in greater quantities than silica. This observation would be consistent with the canonical North Atlantic seasonal succession, where diatoms are succeeded by other phytoplankton functional types as silica becomes depleted. However, silica concentrations were greater than 2 mmol m^{-3} in most of the Irminger Basin during the summer cruise, i.e., above the level where diatoms may be outcompeted by other species [*Brown et al.*, 2003]. Nevertheless, the N:Si drawdown ratio was not 1:1 as expected for diatoms, suggesting that the summer phytoplankton constituted a mixed community. A full taxonomic analysis of the phytoplankton community structure for the 2010 cruises is ongoing, but initial observations confirm the presence of an unusually large proportion of *Coccolithus pelagicus* (A. Poulton, personal communication, 2012), a species abundant in the Labrador Sea [*Ziveri et al.*, 2004]. *C. pelagicus* may thus have been advected into the Irminger Basin with the freshwater anomaly and contributed to the unusual nutrient drawdown ratios in 2010.

[20] Either the influence of the warmer than usual air temperatures in the region, and/or the freshwater intrusion resulted in an unusually shallow mixed layer depth in

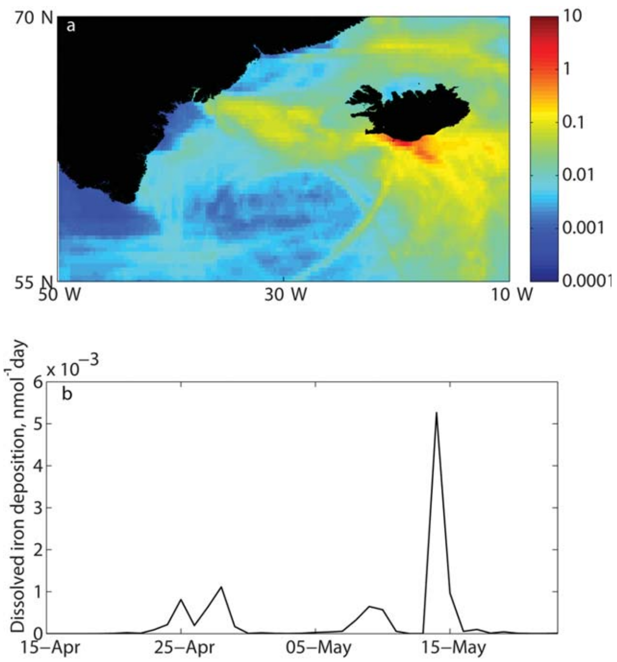

Figure 8. (a) Modeled dissolved iron deposition (nmol l^{-1}) from the eruption of the Eyjafjallajökull volcano (15 April—23 May 2010). For details of the model and estimation of iron deposition see *Stohl et al.* [2011] and *Achterberg et al.* [2013]. (b) Daily modeled iron deposition rates (nmol l^{-1} day^{-1}) in the central Irminger Basin during the volcanic eruption.

autumn 2010. This is likely to have extended the phytoplankton growing season (as seen in the satellite-derived chl data; Figure 1b), which is typically light limited in late autumn and winter [*Henson et al*., 2006a]. Even if the light conditions were suitable for phytoplankton growth for an extended period of time in 2010, this is unlikely to be the sole reason for the anomalous bloom, as sufficient nutrients would also be required. Although an additional source of nutrients may have been introduced to the region by the unusual hydrographic conditions, it seems unlikely, if not impossible, that this alone could have fueled the exceptional increase in PP (satellite-derived estimates suggest annually integrated PP was ~50% greater in 2010 than the mean). Any additional nitrate introduced to the region would likely need to be supplemented by increased rates of nutrient recycling to sustain these high rates of PP.

[21] In situ measurements of ammonium concentration within the mixed layer revealed a 500 fold increase between the spring and summer cruises, confirming that intense recycling was occurring during the summer bloom (E. Achterberg, unpublished data, 2012). We have no other ammonium data from this region, and so cannot establish whether the concentrations were unusually high compared to other more typical years. However, atypically for a high latitude, open ocean region, very high ammonium concentrations (up to 2 μmol l^{-1}) were observed immediately below the mixed layer in summer. The freshwater cap present in the Irminger Basin in 2010 may have allowed retention of the ammonium within the euphotic zone, potentially providing a significant quantity of nitrogen to support the high productivity observed throughout the summer. Our results suggest that the relatively warm spring/summer of 2010, combined with the shallow mixed layer, may have favored increased rates of nitrogen recycling relative to other years.

[22] The anomalous bloom may also have been influenced by reduced grazing pressure. The abundance and behavior of mesozooplankton in the central Irminger Basin in summer 2010 was comparable to other years, based on measurements of individual body size, gut clearance rates and weight-specific ingestion (S. Giering, unpublished data, 2012). However, zooplankton were only grazing ~0.5–5% of PP per day, i.e., top-down control of the bloom was very weak. Zooplankton seem to have been unable to increase their grazing rate to respond to the anomalous bloom in 2010, so that, in contrast to normal conditions, PP remained high throughout summer and into autumn. Analysis of a decadal time series of copepods from the Continuous Plankton Recorder survey suggested that the abundance and population dynamics of *Calanus finmarchicus* are insensitive to climate variability in this region [*Heath et al*., 2008]. Our results suggest that the mesozooplankton population was incapable of altering its behavior in response to variability in conditions, so that grazing rates did not increase despite anomalously high phytoplankton abundance. Although the bloom may not have been under top-down control, maintaining the high phytoplankton biomass through summer and autumn would still have required sufficient additional nutrients.

[23] An alternative explanation for the anomalous bloom evolution in 2010 lies with the eruption of the Eyjafjallajökull volcano. Ash deposited onto the sea surface during an eruption will leach iron and other metals into the upper ocean. However, for any additional deposited iron to have a fertilization effect, phytoplankton growth must be iron limited. Some evidence for iron limitation in this region comes from iron addition experiments conducted in summer 2007 in the Iceland Basin and in summer 2010 in the Irminger Basin, which both demonstrated an increase in chl compared to controls [*Nielsdottir et al*., 2009; *Ryan-Keogh et al*., 2013]. Indirect evidence suggesting potential iron limitation in the Irminger Basin also came from the Marine Productivity cruises in 2002. Typically, one would expect that after silica becomes depleted, diatoms would become outcompeted by other phytoplankton functional types, which would consume the remaining nitrate. However, *Sanders et al*. [2005] established that, although silica is depleted in surface waters in summer, nonlimiting concentrations of nitrate remain at the end of the growing season, suggesting that iron limitation of phytoplankton growth may inhibit complete nitrate drawdown in the Iceland, and possibly Irminger, Basins.

[24] Results from an atmospheric transport model, combined with estimates of iron content and dissolution rates (see *Achterberg et al*. [2013] for details), suggest that ash from the eruption may have contributed an additional ~0.005 nmol l^{-1} of dissolved iron to the central Irminger Basin, the majority of it in one day, 14 May (Figure 8). Elevated chl concentration (~0.3 mg m^{-3} higher than the mean) is evident in the Irminger Basin in anomaly maps for the 9–16 May period (supporting information Figure S5), but has receded by the following week, when chl is lower than the mean.

[25] The relatively transient nature of any stimulation of phytoplankton growth by ash-borne iron was confirmed by shipboard experiments conducted with ash from the Eyjafjallajökull eruption. Bioassays conducted during the summer 2010 cruise found that addition of 9 mg l^{-1} of ash (≈4.6–16 nM DFe) to Iceland Basin populations induced an increase in phytoplankton biomass, although the response was weaker than to addition of 2 nM $FeCl_3$, i.e., much of the dissolved iron released from the ash in seawater was not bioavailable [*Achterberg et al*., 2013]. The ash also sank rapidly, with settling experiments suggesting that the particles fell through the upper mixed layer (~30 m deep in spring) in ~90 min, during which time all the salts leached from the particles [*Achterberg et al*., 2013]. This is even faster than the previously reported residence time of atmospheric iron inputs in the surface ocean of 2–7 weeks [*Sarthou et al*., 2003]. These results are consistent with the observation during the summer 2010 (i.e., posteruption) cruise of a lack of elevated surface iron concentrations in the region offshore of Eyjafjallajökull [*Achterberg et al*., 2013]. Taken together these observations imply that any fertilization effect due to ash deposition was short-lived.

5. Conclusions

[26] A highly anomalous chlorophyll bloom occurred in the central Irminger Basin in 2010, persisting throughout the summer and autumn. We present here several observations of unusual conditions in 2010 that may have influenced the development of the anomalous phytoplankton bloom:

[27] 1. The North Atlantic Oscillation was in an extreme negative phase in winter 2009/2010; the second most

negative on record since winter 1968/1969 (supporting information Figure S1).

[28] 2. Wind direction was reversed in spring 2010 compared to normal conditions, coming from the south or east (Figure 2).

[29] 3. A surface freshwater anomaly was observed extending throughout the central Irminger Basin in summer 2010 to a depth of ~30 m (Figure 3).

[30] 4. The freshwater anomaly had characteristics similar to that of the central gyre fresh pool, situated south of Cape Farewell (Figure 4), and migrated northeastward into the Irminger Basin during summer 2010 (Figure 5).

[31] 5. The mixed layer shoaled earlier than usual and stayed relatively shallow through autumn (Figure 6).

[32] 6. The central Irminger Basin phytoplankton community composition had an unusually large proportion of *Coccolithicus pelagicus*, usually abundant in the Labrador Sea.

[33] 7. Nutrient utilization was anomalous, with N:Si drawdown ratios of 3:1, rather than the usual 1:1 (Figure 7).

[34] 8. Zooplankton grazing rates did not increase in tandem with PP during the summer.

[35] 9. The Eyjafjallajökull volcano erupted in April/May 2010 depositing ash in the northern Irminger Basin, which may have added a total of ~0.005–0.08 nmol l^{-1} of iron to the region (Figure 8).

[36] Although we can provide no definitive answer to what caused the anomalous bloom, the balance of evidence points to the unusual hydrographic conditions of 2010, with the volcanic eruption playing a relatively minor role.

[37] Ongoing analysis of the cruise data collected as part of the IBIS program may reveal some additional clues to the causes, and importantly, consequences of the anomalous bloom. A key question is whether the bloom resulted in an increase in organic carbon export, and therefore may have had a significant impact on the biogeochemistry of the region. Introduction of iron to the surface ocean via deposition of volcanic ash has been postulated as a driver of reductions in atmospheric CO_2 [*Watson*, 1997], due to the stimulation of additional carbon export. In the case of the Eyjafjallajökull eruption and the Irminger Basin, the volcanic eruption seems to have had a fairly minor effect. A volcanic eruption would need to produce substantially more ash, over a longer period, with direct deposition over a high nutrient-low chlorophyll area in order to affect atmospheric CO_2 concentrations.

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References

Achterberg, E. P., et al. (2013), Natural iron fertilisation by the Eyjafjallajökull volcanic eruption, *Geophys. Res. Lett.*, *40*, 921–926, doi:10.1002/grl.50221.

Bacon, S. (2010), RSS Discovery Cruise 332, 21 Aug–25 Sep 2008, Arctic Gateway (WOCE AR7), *Natl. Oceanogr. Cent. Southampton Cruise Rep. 53*, National Oceanography Centre Southampton, Southampton, U. K.

Bacon, S., W. J. Gould, and Y. L. Jia (2003), Open-ocean convection in the Irminger Sea, *Geophys. Res. Lett.*, *30*(5), 1246, doi:10.1029/2002gl016271.

Barton, A. D., C. H. Greene, B. C. Monger, and A. J. Pershing (2003), The continuous plankton recorder survey and the North Atlantic oscillation: Interannual- to multidecadal-scale patterns of phytoplankton variability in the North Atlantic Ocean, *Prog. Oceanogr.*, *58*, 337–358.

Behrenfeld, M. J., and P. G. Falkowski (1997), Photosynthetic rates derived from satellite-based chlorophyll concentration, *Limnol. Oceanogr.*, *42*(1), 1–20.

Bersch, M. (2002), North Atlantic Oscillation-induced changes of the upper layer circulation in the northern North Atlantic Ocean, *J. Geophys. Res.*, *107*(C10), 3156, doi:10.1029/2001JC000901.

Brown, L., R. Sanders, G. Savidge, and C. H. Lucas (2003), The uptake of silica during the spring bloom in the Northeast Atlantic Ocean, *Limnol. Oceanogr.*, *48*(5), 1831–1845.

Brzezinski, M. A. (1985), The Si-C-N ratio of marine diatoms—Interspecific variability and the effect of some environmental variables, *J. Phycol.*, *21*(3), 347–357.

Claustre, H., A. Morel, S. B. Hooker, M. Babin, D. Antoine, K. Oubelkheir, A. Bricaud, K. Leblanc, B. Queguiner, and S. Maritorena (2002), Is desert dust making oligotrophic waters greener?, *Geophys. Res. Lett.*, *29*(10), 1469, doi:10.1029/2001gl014056.

Duggen, S., P. Croot, U. Schacht, and L. Hoffmann (2007), Subduction zone volcanic ash can fertilize the surface ocean and stimulate phytoplankton growth: Evidence from biogeochemical experiments and satellite data, *Geophys. Res. Lett.*, *34*, L01612, doi:10.1029/2006gl027522.

Duggen, S., N. Olgun, P. Croot, L. Hoffmann, H. Dietze, P. Delmelle, and C. Teshner (2010), The role of airborne volcanic ash for the surface ocean biogeochemical iron-cycle: A review, *Biogeosciences*, *7*, 827–844, doi:10.5194/bg-7–827-2010.

Edwards, M., P. Reid, and B. Planque (2001), Long-term and regional variability of phytoplankton biomass in the Northeast Atlantic (1960–1995), *ICES J. Mar. Sci.*, *58*(1), 39–49, doi:10.1006/jmsc.2000.0987.

Flatau, M. K., L. Talley, and P. P. Niiler (2003), The North Atlantic Oscillation, surface current velocities, and SST changes in the subpolar North Atlantic, *J. Clim.*, *16*(14), 2355–2369, doi:10.1175/2787.1.

Gudmundsson, M. T., et al. (2012), Ash generation and distribution from the April–May 2010 eruption of Eyjafjallajökull, Iceland, *Nat. Sci. Rep.*, *2*(572), doi:10.1038/srep00572.

Harrison, W. G., and W. K. W. Li (2007), Phytoplankton growth and regulation in the Labrador Sea: Light and nutrient limitation, *J. Northwest Atl. Fish. Sci.*, *39*, 71–82.

Head, E. J. H., and P. Pepin (2010), Spatial and inter-decadal variability in plankton abundance and composition in the Northwest Atlantic (1958–2006), *J. Plankton Res.*, *32*(12), 1633–1648, doi:10.1093/plankt/fbq090.

Heath, M. R., et al. (2008), Spatial demography of Calanus finmarchicus in the Irminger Sea, *Prog. Oceanogr.*, *76*(1), 39–88, doi:10.1016/j.pocean.2007.10.001.

Henson, S. A., R. Sanders, J. T. Allen, I. S. Robinson, and L. Brown (2003), Seasonal constraints on the estimation of new production from space using temperature-nitrate relationships, *Geophys. Res. Lett.*, *30*(17), 1912, doi:10.1029/2003gl017982.

Henson, S. A., I. Robinson, J. T. Allen, and J. J. Waniek (2006a), Effect of meteorological conditions on interannual variability in timing and magnitude of the spring bloom in the Irminger Basin, North Atlantic, *Deep Sea Res., Part I*, *53*(10), 1601–1615, doi:10.1016/j.dsr.2006.07.009.

Henson, S. A., R. Sanders, C. Holeton, and J. T. Allen (2006b), Timing of nutrient depletion, diatom dominance and a lower-boundary estimate of export production for Irminger Basin, North Atlantic, *Mar. Ecol. Prog. Ser.*, *313*, 73–84, doi:10.3354/meps313073.

Henson, S. A., J. P. Dunne, and J. L. Sarmiento (2009), Decadal variability in North Atlantic phytoplankton blooms, *J. Geophys. Res.*, *114*, C04013, doi:10.1029/2008jc005139.

Holliday, N. P., J. J. Waniek, R. Davidson, D. Wilson, L. Brown, R. Sanders, R. T. Pollard, and J. T. Allen (2006), Large-scale physical controls on phytoplankton growth in the Irminger Sea Part I: Hydrographic zones, mixing and stratification, *J. Mar. Syst.*, *59*, 201–218.

Holliday, N. P., A. Meyer, S. Bacon, S. G. Alderson, and B. de Cuevas (2007), Retroflection of part of the east Greenland current at Cape Farewell, *Geophys. Res. Lett.*, *34*, L07609, doi:10.1029/2006gl029085.

Holte, J., and L. Talley (2009), A new algorithm for finding mixed layer depths with applications to Argo data and Subantarctic Mode Water formation, *J. Atmos. Oceanic Technol.*, *26*(9), 1920–1939, doi:10.1175/2009jtecho543.1.

Ingleby, B., and M. Huddleston (2007), Quality control of ocean temperature and salinity profiles—Historical and real-time data, *J. Mar. Syst.*, *65*, 158–175, doi:10.1016/j.jmarsys.2005.11.019.

Kalnay, E., et al. (1996), The NCEP/NCAR 40-year reanalysis project, *Bull. Am. Meteorol. Soc.*, *77*(3), 437–471.

Karlsdóttir, S., Á. G. Gylfason, Á. Höskuldsson, B. Brandsdóttir, E. Ilyinskaya, M. T. Gudmundsson, and Þ. Högnadóttir (2012), The 2010 Eyjafjallajökull eruption, Iceland, ICAO Rep., edited by B. Þorkelsson, Icelandic Meteorological Office, Reykjavik, Iceland.

Kirkwood, D. S. (1996), Nutrients: Practical notes on their determination in seawater, *ICES Tech. in Mar. Environ. Sci. Rep. 17*, Int. Counc. for the Explor. of the Seas, Copenhagen.

Langmann, B., K. Zaksek, M. Hort, and S. Duggen (2010), Volcanic ash as fertiliser for the surface ocean, *Atmos. Chem. Phys.*, *10*(8), 3891–3899.

Lin, I.-I., et al. (2011), Fertilization potential of volcanic dust in the low-nutrient low-chlorophyll western North Pacific subtropical gyre: Satellite evidence and laboratory study, *Global Biogeochem. Cycles*, *25*, GB1006, doi:10.1029/2009gb003758.

Morel, A. (1988), Optical modeling of the upper ocean in relation to its biogenous matter content (Case-I waters), *J. Geophys. Res.*, *93*(C9), 10,749–10,768, doi:10.1029/JC093iC09p10749.

Nielsdottir, M. C., C. M. Moore, R. Sanders, D. J. Hinz, and E. P. Achterberg (2009), Iron limitation of the postbloom phytoplankton communities in the Iceland Basin, *Global Biogeochem. Cycles*, *23*, GB3001, doi:10.1029/2008gb003410.

Overland, J., M. Wang, and J. Walsh (2010), Atmosphere, in *Arctic Report Card 2010*, NOAA.

Patara, L., M. Visbeck, S. Masina, G. Krahmann, and M. Vichi (2011), Marine biogeochemical responses to the North Atlantic Oscillation in a coupled climate model, *J. Geophys. Res.*, *116*, C07023, doi:10.1029/2010jc006785.

Reid, P. C., B. Planque, and M. Edwards (1998), Is observed variability in the long-term results of the Continuous Plankton Recorder survey a response to climate change?, *Fish. Oceanogr.*, *7*(3–4), 282–288, doi:10.1046/j.1365–2419.1998.00073.x.

Reverdin, G., F. Durand, J. Mortensen, F. Schott, H. Valdimarsson, and W. Zenk (2002), Recent changes in the surface salinity of the North Atlantic subpolar gyre, *J. Geophys. Res.*, *107*(C12), 8010, doi:10.1029/2001JC001010.

Robock, A. (2000), Volcanic eruptions and climate, *Rev. Geophys.*, *38*(2), 191–219, doi:10.1029/1998rg000054.

Ryan-Keogh, T. J., A. I. Macey, M. C. Nielsdóttir, M. I. Lucas, M. C. Stinchcombe, E. P. Achterberg, T. S. Bibby, and C. M. Moore (2013), Spatial and temporal development of phytoplankton iron stress in relation to bloom dynamics in the high latitude North Atlantic Ocean, *Limnol. Oceanogr.*, *58*, 533–545.

Sanders, R., L. Brown, S. Henson, and M. Lucas (2005), New production in the Irminger Basin during 2002, *J. Mar. Syst.*, *55*(3–4), 291–310, doi:10.1016/j.jmarsys.2004.09.002.

Sarmiento, J. L. (1993), Carbon-cycle—Atmospheric CO2 stalled, *Nature*, *365*(6448), 697–698, doi:10.1038/365697a0.

Sarthou, G., et al. (2003), Atmospheric iron deposition and sea-surface dissolved iron concentrations in the eastern Atlantic Ocean, *Deep Sea Res., Part I*, *50*(10–11), 1339–1352, doi:10.1016/s0967-0637(03)00126-2.

Stohl, A., et al. (2011), Determination of time- and height-resolved volcanic ash emissions and their use for quantitative ash dispersion modeling: The 2010 Eyjafjallajokull eruption, *Atmos. Chem. Phys.*, *11*(9), 4333–4351, doi:10.5194/acp-11–4333-2011.

Stroeve, J. C., J. Maslanik, M. C. Serreze, I. Rigor, W. Meier, and C. Fowler (2011), Sea ice response to an extreme negative phase of the Arctic Oscillation during winter 2009/2010, *Geophys. Res. Lett.*, *38*, L02502, doi:10.1029/2010gl045662.

Sverdrup, H. U. (1953), On conditions for the vernal blooming of phytoplankton, *J. Cons. Cons. Int. Explor. Mer*, *18*, 287–295.

Uematsu, M., M. Toratani, M. Kajino, Y. Narita, Y. Senga, and T. Kimoto (2004), Enhancement of primary productivity in the western North Pacific caused by the eruption of the Miyake-jima Volcano, *Geophys. Res. Lett.*, *31*, L06106, doi:10.1029/2003gl018790.

Visbeck, M., E. P. Chassignet, R. G. Curry, T. L. Delworth, R. R. Dickson, and G. Krahmann (2003), The ocean's response to North Atlantic Oscillation variability, in *The North Atlantic Oscillation: Climatic Significance and Environmental Impact, Geophys. Monogr. Ser.*, vol. *134*, edited by J. W. Hurrell, pp. 113–146, AGU, Washington, D. C.

Waniek, J. J., and N. P. Holliday (2006), Large-scale physical controls on phytoplankton growth in the Irminger Sea, Part II: Model study of the physical and meteorological preconditioning, *J. Mar. Syst.*, *59*(3–4), 219–237, doi:10.1016/j.jmarsys.2005.10.005.

Watson, A. J. (1997), Volcanic iron, CO2, ocean productivity and climate, *Nature*, *385*(6617), 587–588, doi:10.1038/385587b0.

Ziveri, P., K. H. Baumann, B. Böckel, J. Bollmann, and J. R. Young (2004), Present day coccolithophore-biogeography in the Atlantic Ocean, in *Coccolithophores. From Molecular Processes to Global Impact*, edited by H. R. Thierstein and J. R. Young, pp. 403–428, Springer, Berlin.