**The influence of short-term events on the hydrographic and biological structure of the
southwestern Ross Sea**

by

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Abstract

The Ross Sea continental shelf supports very high productivity and phytoplankton biomass. Conventional methods, including ship-based sampling, instrumented moorings, satellite imagery, and modelling, have illustrated the typical patterns of seasonal progression of the phytoplankton blooms. While variability in the Ross Sea is substantial over relatively large scales, our understanding of smaller scales of variability (on the order of a few hours or several kilometers) is limited. Utilizing data from an autonomous glider, we examined the mechanisms driving both the transitions between stages of the phytoplankton bloom and the short-term perturbations in chlorophyll concentrations. Three phases within the bloom were defined based on chlorophyll changes: an accumulation phase, a dissipation period, and a post-dissipation phase. Short-term perturbations in chlorophyll were repeatedly observed and correlated with wind speed. These chlorophyll perturbations were strongly influenced by the degree of temporal coupling between wind events and the depth of mixing, which varied among phases. Delays of 12-24 h between wind events and chlorophyll changes were observed during the accumulation phase, but shortened to 2-12 h in the dissipation phase. Furthermore, while physical factors contributed to the observed short-term reductions in biomass and the appearance of chlorophyll at depth throughout the study, we hypothesize that turbulence during the period of maximum biomass induced aggregate formation and led to rapid vertical flux. These results suggest that the small-scale, short-term physical perturbations may induce substantial vertical redistribution of biogenic material, which in turn can have significant biogeochemical impacts.

Keywords: mesoscale; Ross Sea, Antarctica; winds; chlorophyll; mixed layers

1. Introduction

The Ross Sea continental shelf is one of the most studied regions in the Southern Ocean, and has been characterized as being the most productive area in the Southern Ocean, contributing 28% of the total productivity (Arrigo et al., 1998b). Studies have indicated the importance of this intense productivity and the substantial phytoplankton standing stocks in sustaining the food web (Smith et al., 2014). However, the substantial phytoplankton biomass appears to be relatively spatially or temporally uncoupled from the distributions of middle and upper trophic level species (Arrigo et al., 2000; Pinkerton and Bradford-Grieve, 2014; Ainley et al., 2015). Furthermore, despite the substantial spatial variability observed (e.g., Arrigo and McClain, 1994; Arrigo et al., 1999; Smith et al., 2013), the dominant scale of variability appears to be temporal (Smith et al., 2000). For example, temporal variations in chlorophyll ranges were up to $25 \mu\text{g L}^{-1}$ over two months, whereas variations in space at any single date tend to be far less extreme. Assessments of the variability within a restricted location in the upper ocean, however, are rare. Smith et al. (2011a) collected data using *in situ* fluorometers at two locations and three depths, and demonstrated the degree of variability in chlorophyll during austral summer. Satellite climatologies also have indicated the scales of temporal variations at single locations, although the satellite data is rarely available on a continuous basis due to frequent cloud cover.

In situ sampling has revealed a consistent temporal pattern in phytoplankton composition in the Ross Sea. In general the austral spring is dominated by the haptophyte *Phaeocystis antarctica* (Smith and Gordon, 1997; Arrigo et al., 1999). This species reaches very high biomass levels (up to $25 \mu\text{g chlorophyll L}^{-1}$; Smith et al., 2003), but its biomass rapidly declines in late December and early January (Smith et al., 2000, 2011b, 2014), after which biomass of all phytoplankton groups generally remains low. Occasionally, a diverse assemblage of diatoms forms large blooms in late summer (Peloquin and Smith, 2007), the magnitude of which can be

equal to those found in spring (Smith et al., 2011a). Although the magnitude of phytoplankton biomass varies among years, the temporal patterns have been corroborated by both satellite and in situ observations (Arrigo and McClain, 1994; Arrigo and van Dijken, 2003, 2004; Arrigo et al., 2008; Smith and Comiso, 2008).

The major reasons for this observed seasonal bloom progression are believed to be both physical and biological. Beginning with bloom initiation in late October, the dominant controlling mechanisms are related to irradiance availability: decreasing ice coverage, seasonally increasing irradiance, lengthening photoperiods, and shoaling mixed layer depths (Smith and Gordon, 1997; Smith et al., 2000). Irradiance penetration through sea ice and snow cover is very low during spring and fall, with an attenuation of ~97% (Arrigo, 2014). Following rapid sea-ice retreat and moderate water column stabilization, phytoplankton productivity and biomass quickly increase, leading to large accumulations of particulate organic matter in the surface layer (Smith and Gordon, 1997; Arrigo et al., 1998a; Smith et al., 2000). A biomass maximum typically occurs in mid- to late December, and further growth is apparently limited by availability of the micronutrient iron (Sedwick et al., 2000, 2011). Overall, the temporal patterns in phytoplankton biomass result from changes in environmental conditions dictated by irradiance, vertical water column structure, and the degree of iron limitation (DiTullio and Smith, 1996; Arrigo et al., 1999; Peloquin and Smith, 2007; Sedwick et al., 2011).

The termination of the biomass maxima is largely thought to be the result of increased vertical flux of aggregated phytoplankton material, passive phytoplankton sinking (with rates enhanced due to nutrient stress), and losses resulting from grazing (Dunbar et al., 1998; Smith and Asper, 2001; Smith et al., 2000, 2011b). Smith et al. (2000, 2011b) demonstrated that two independent environmental factors created a temporal disconnect of ~30 days between the maxima of productivity and later biomass maxima in early January. Initially, limited iron

availability leads to a reduction in phytoplankton production, with grazing and particle aggregation/sinking losses eventually becoming significant, thus leading to a reduction of biomass. Decreases in chlorophyll concentrations of 90% can occur over a period of a few days (Smith et al., 2011a). However, the temporal progression of these seasonal changes in chlorophyll are disrupted by “events” in which chlorophyll decreases over very short periods (hours) and recovers rapidly thereafter. Smith et al. (2011a) hypothesized that these were due to storms that altered surface stratification and diluted surface layers with low biomass water from depth, but provided no data on the physical forcing of such events. They also speculated that some events were the result of advection of low biomass waters, but again had no data to support this speculation.

This study investigated the temporal dynamics of a restricted region in the southern Ross Sea using a glider, an autonomous underwater vehicle that is propelled by buoyancy, which undulates from the surface to depth, and relays the data collected to a remote base station via satellite link (Eriksen et al., 2001). Gliders have recently been deployed in the Ross Sea, a region that has heavy ice cover and frequent storms, both of which restrict oceanographic sampling (Kaufman et al., 2014; Ainley et al., 2015; Queste et al., 2015). Gliders have slow speeds of $< 0.25 \text{ m s}^{-1}$ and typically cover horizontal distances of $\sim 0.5\text{-}5 \text{ km dive}^{-1}$ or $\sim 15\text{-}30 \text{ km d}^{-1}$ (depending on the dive depth; Rudnick et al., 2004); as such, they are typically limited to resolving features with scales of a few days and a few km. Given that the spatial coverage of our glider was limited to an area of ca. $25 \times 50 \text{ km}$, the temporal variability over scales of months was assessed to elucidate the mechanisms that generate both the seasonal patterns as well as the short-term or “event scale” variations that commonly occur. We hypothesized that short-term variations in physical forcing had significant impacts on phytoplankton vertical distribution and potentially had important biogeochemical impacts on the entire southern Ross Sea.

2. Materials and Methods

A glider (iRobot Seaglider™ Model 1KA) was deployed from the fast ice near Cape Crozier, Ross Island (77.438°S, 169.746°E) on November 22, 2012 at 09:25 UTC. It initially headed to the northeast for the first 50 dives before returning close to the point of deployment. For the remainder of the study (dives 81-571), the glider followed a ‘radiator’ pattern (with three main latitudinal transects and several perpendicular, longitudinal transects; 25 km E/W x 50 km N/S), sampling multiple times along certain transects (Figure 1) and diving to approximately 800 m. The glider completed 571 dives over 78 d and was retrieved (76.771°S, 167.729°E) by the *RVIB Palmer* on February 8, 2013 at 00:44 UTC. The glider’s sensors included a SeaBird CT Sail, an Aanderaa Oxygen Optode 4330F, and Wet Labs ECO Triplet Puck. Data were collected approximately every 5 seconds; the Wet Labs ECO Puck was disabled below 250 m to conserve battery power. Resolution of all variables was on average 0.70 m or 5.7 s in the vertical and 1.3 km or 1.7 h in the horizontal (Jones, 2015). All data are available at <http://www.bco-dmo.org/dataset/568868>.

Due to the absence of ship-based support, glider calibration was limited to one CTD cast during recovery of the glider. The *RVIB Palmer* CTD profiles and water samples were used to calibrate glider temperature, salinity, and dissolved oxygen. Issues with the chlorophyll water samples obtained from the CTD profile led us to utilize a previous chlorophyll calibration for the fluorescence data (Kaufman et al., 2014). Photoquenching was observed in the glider fluorescence signal, though it did not significantly influence the temporal patterns we observed. Hysteresis (the delay between measured and actual values at one depth) occurred through strong gradients in temperature, conductivity, and dissolved oxygen casts. Temperature and conductivity hysteresis were corrected by cross-calibration with the CTD profiles and application

of a temporal offset. These corrections and other pre-analysis calculations were performed on the glider data set using the ‘uea-seaglider-toolbox’ developed by Dr. Bastien Queste (available at <https://bitbucket.org/bastienqueste/uea-seaglider-toolbox/>).

Glider fluorescence and particulate backscattering coefficients ($b_{bp}(\lambda)$, calculated from the glider optical backscattering counts; Stramski et al., 1998, 1999) were processed by regressing data from the final glider dive against *in situ* data ($n=12$). Samples for chlorophyll *a* were analyzed by fluorometry using the acid-addition method on a Turner Designs 10-AU fluorometer (JGOFS, 1996), and particulate matter concentrations were determined by filtering known volumes of seawater through combusted GFF filters, drying at 60°C, and pyrolysis on an elemental analyzer to obtain particulate organic carbon (POC) concentrations (Gardner et al., 2000). Fluorescence counts (FL) were converted into chlorophyll *a* concentrations (Chl) using the regression:

$$\text{Chl} = (\text{FL} + 141) * 0.00225; n = 12; R^2 = 0.94; p < 0.01$$

(Kaufman et al., 2014). To calculate $b_{bp}(\lambda)$, glider optical backscattering counts were converted to total volume scattering, β (117°, 470 nm) by subtracting dark counts and multiplying by a factory-calibrated scale factor, subtracting the volume scattering of seawater, β_w (Morel, 1974) from the total volume scattering to obtain volume scattering of particulates, β_p , and multiplying by $2\pi\chi$, where χ is 1.1 (Boss and Pegau, 2001). A regression based on the calibration cast:

$$\text{POC} = (42,850 * b_{bp}(470 \text{ nm})) + 57.47; n = 11; R^2 = 0.55; p < 0.01$$

was used to convert $b_{bp}(\lambda)$ into POC concentrations.

To prepare the calibrated, core data sets for analyses, individual glider dives were separated into descending and ascending legs, with the deepest point in the dive marking the delineation between the two. With GPS positioning available only when the glider was at the ocean surface, the GPS position obtained just prior to and just following a dive were interpolated to estimate

position data for that dive. A number of physical and biological parameters were derived for each cast. Values for sea surface temperature, salinity, and density were calculated by averaging observations from the top 5 m of the water column. Mixed layer depth (MLD) was computed using a potential density threshold method (Thomson and Fine, 2003; Smith et al., 2013), which defines the MLD as the minimum depth at which a 0.01 kg m^{-3} increase is observed over the density (σ_θ) at 3 m. Integrated chlorophyll concentrations were calculated using trapezoidal integration.

To investigate changes in the physical and biological variables through the seasonal progression of the phytoplankton bloom, temporal periods were identified using average 0-50 m chlorophyll data. Three periods, “accumulation”, “dissipation”, and “post-dissipation”, were defined which corresponded with biomass accumulation (increasing chlorophyll concentrations), biomass maintenance/decrease (decreasing chlorophyll concentrations), and post-decrease (relatively constant and low chlorophyll concentrations), respectively. The accumulation period began when the glider entered into the radiator pattern on December 1 at 00:00 UTC and ended on December 14 at 00:00 UTC, approximately 12 h prior of the maximum average 0-50 m chlorophyll. The accumulation period spanned 13 d and was comprised of 184 casts (casts 152-335). The dissipation period began December 14 at 00:00 UTC (approximately 12 h following the maximum average 0-50 m chlorophyll) and ended on January 7, 2013 at 00:00 UTC, 24 d later. The duration of this period is approximately twice as long as the accumulation period, as the bloom dissipates over a longer time. The end of the dissipation phase was defined by a change in the rate of reduction of chlorophyll. The dissipation period included 366 casts (glider casts 336-701). The post-dissipation period began January 7 at 00:00 UTC and ended 12 d later on January 19 at 00:00 UTC (167 casts; casts 702-868; Table 1). All data were grouped into 10-

m bins for analysis. Selection of the depth scale or temporal period limits did not bias the resultant correlations.

Hourly wind speeds were obtained from the Antarctic Meteorological Research Center (AMRC), University of Wisconsin, Madison (<http://amrc.ssec.wisc.edu/>). The AMRC-Automated Weather Station (AWS) is located east of Ross Island on the Ross Ice Shelf and is named Laurie II (77.451°S, 170.760°E; anemometer height approximately 40 m above sea level). It is located approximately 45 km from the center of the glider deployment (Figure 1).

Bootstrapped 95% confidence intervals of the means ($n = 1000$) were used to statistically compare variable averages between different temporal periods. Significant differences for pairwise comparisons were assessed at the $\alpha = 0.05$ level. To perform two-sample hypothesis tests on variability in the non-normal, autocorrelated glider, mooring, and wind speed time series, bootstrapped data for the temporal periods and variables of interest were fit to an autoregressive integrated moving average (ARIMA) model to obtain a ratio of residual variance. This process was repeated 1000 times to construct a 95% confidence interval for the ratio of residual variances. If the 95% confidence interval did not contain 1, the null hypothesis of equal variance between groups was rejected, and it was concluded that the variances differed significantly. For the event-scale perturbation analyses and moving correlations, Pearson's linear correlation coefficients were calculated for glider and AWS variables. Correlation coefficients with p values ≤ 0.05 were considered statistically significant.

3. Results

Although spatial variability was noted within the small sampling region, the dominant mode of variability was temporal, especially with regard to chlorophyll concentrations. Spatial variability was assessed by comparing hydrographic and biological means and standard

deviations among three east-west transects that were sampled close in time; three independent comparisons were made (Table 2). Similarly, temporal variability was assessed by comparing hydrographic and biological means and standard deviations among the three temporal periods and across the glider deployment (Table 1; Figures 2, 3). Although spatial variability was observed and significant (Table 2), it was less than the seasonal changes observed (Table 1; Figures 2, 3). Specifically, comparisons of chlorophyll standard deviations for north, central, and south east-west transects closely related in time were nearly equal in all three locations, suggesting that spatial variability, while significant, is relatively constant at any one time. For example, during the accumulation phase, the three east-west transect mean chlorophyll levels ranged narrowly from 3.27 – 4.09 $\mu\text{g L}^{-1}$ (minimum 80% of maximum), whereas at the ‘North Horizontal’ transect chlorophyll decreased in time from 4.09 to 0.77 $\mu\text{g L}^{-1}$ (a decrease to 19% of maximum; Table 2). In addition to variability within the surface observations, high variability was observed in the upper 150 m across the glider deployment (Figure 4). Within that ca. 78 d deployment, three defined periods – “accumulation”, “dissipation”, and “post-dissipation” – were based on changes in mean chlorophyll concentrations in the upper 50 m; in addition, marked events within each phase were observed and are discussed below. These mesoscale temporal variations were consistent with the general seasonal patterns in phytoplankton biomass, stratification, and temperature that had been repeatedly observed (Smith et al., 2000, 2014; Kaufman et al., 2014), but included marked short-term variations as part of the seasonal trends. Winds were variable throughout (Figure 5), and surface temperatures increased with time, with a modest decrease in mid- to late January. Mixed layer depths were also relatively constant, but similar in timing to changing temperatures they increased slightly in the post-dissipation phase (Figure 2). Chlorophyll increased to a maximum on December 13 and decreased rapidly to levels approximately 25% of those found earlier (Figure 3). Particulate organic carbon

concentrations also increased with time (Figure 3), and became maximal around December 25. POC also decreased after that, but much more slowly than chlorophyll. As a result, POC:Chl ratios rose with time (Figure 3), as had been noted previously (Kaufman et al., 2014). The mean conditions and standard deviations found during the three periods for each variable are listed in Table 1. Mean vertical profiles of temperature, salinity, chlorophyll and POC also illustrate the temporal variations observed (Figure 6).

Accumulation Phase

The accumulation phase was from December 1 – 14, and average chlorophyll levels increased from 3.8 to 6.2 $\mu\text{g L}^{-1}$ (Figure 3). Within that period, one strong wind event that lasted 48-h (December 4-5) was observed, with the direction of the wind event ranging between 210 and 260° (Figures 5a, 7). Other than this event, relatively quiescent conditions allowed MLDs to decrease and surface temperatures to increase. The stratification during these calm periods facilitated phytoplankton growth and accumulation, as demonstrated by the increased chlorophyll concentrations. A 12-24 h delay between the onset of the wind event and the initiation of a decrease in upper 30 m chlorophyll appeared to result from a delay in transferring the wind energy into changes to the upper water column structure (i.e., a MLD increase). During the accumulation period wind events typically began to affect mixed layer depths once wind speeds had nearly or fully reached the wind event's maximum velocity (Figure 7).

While wind events during the accumulation period were observed to affect chlorophyll after 12-24 h, changes to chlorophyll were observed almost immediately in response to changes to mixed layer depth. The December 4-5 wind event generated a MLD excursion from December 5-6 of 5-10 m to 60-80 m, which lasted for approximately 18 h. The initial responses in chlorophyll were limited to the upper 20 m of the water column and were observed from approximately noon until midnight UTC on December 4, 12 h prior to the onset of the MLD

excursion. The deepening of the MLD started at 0000 UTC on December 5, and chlorophyll decreases were observed in the upper 40 m, and chlorophyll increases between 50 and 80 m (Figure 7).

Dissipation Phase

The dissipation phase began on December 14 (approximately 12 h following the maximum in the mean 0-50 m chlorophyll) and ended on January 7 (24 d). During this period chlorophyll concentrations were variable (ranging from 1.8 to 6.2 $\mu\text{g L}^{-1}$; Figure 3), with several event-scale perturbations observed. A large reduction of chlorophyll from 6.5 to 1.9 $\mu\text{g L}^{-1}$ and a re-establishment to 5.8 $\mu\text{g L}^{-1}$ over 4 d (December 14-18) was noted (Figure 3), but the phase also included several rapid event-scale perturbations (December 18-22) nearly equal in magnitude to the December 14-18 perturbation, but with greater vertical variations (Figure 8). A lack of strong wind events early in the phase (December 14-23) allowed a MLD between 5 and 10 m to develop and chlorophyll remained at approximately 70% of the seasonal maximum (Figures 2, 3). Around December 23, there was a sharp reduction in average (0-50 m) chlorophyll, which coincided with the December 23-25 wind event and resulted in the MLD deepening and chlorophyll concentrations decreasing (Figures 3, 8). Predominantly southwesterly winds have shifted westward to between 215 and 295° (Figure 5b). As in the accumulation period, changes in chlorophyll were observed to coincide with the excursion of the MLD, which, following the onset of the December 23-25 wind event, deepened to approximately 30 m for 12 h. It then further deepened to 40-75 m for 30 h as a succeeding wind event began. However, a reduction in chlorophyll had already begun in the surface 10 m around December 15, *prior* to any MLD excursions. Both the chlorophyll reductions prior to the change in MLD and the MLD increases resulted in the prominent reductions of chlorophyll between 10 and 30 m, and the appearance of chlorophyll between 40 and 100 m approximately 6-12 h before the onset of the MLD excursions

on December 23. The subsequent wind event on December 25-27 and one that followed on December 29-30 continued to disrupt the MLDs, which resulted in a continuation of increased chlorophyll between 30 and 90 m from December 25-31 (Figure 8). Integrated chlorophyll concentrations (0 – 100 m) during the dissipation period did not change as markedly as did those changes observed between 0 and 40 m and 40 and 100 m (Figure 10), suggesting that changes in organic matter were a vertical redistribution and not losses such as might result from zooplankton grazing (and fecal pellet flux) or heterotrophic remineralization of POC.

Post-dissipation Phase

The post-dissipation period had the least variability in chlorophyll concentrations among all periods (Table 1), and there are no large event-scale perturbations in chlorophyll during the phase (Figure 3). Three strong wind events occurred, as well as the beginning of a fourth prolonged, strong wind event that concluded following the end of the post-dissipation period (Figures 2, 9). Wind direction for the stronger wind events remains primarily between 210 and 300°, though slower wind speeds are observed from the northwest and east (Figure 5c). With low levels of chlorophyll (average of 1.12 $\mu\text{g L}^{-1}$; Table 1) in the upper 50 m following the dissipation phase, the four post-dissipation wind events had little marked effect on chlorophyll vertical distribution (Figure 9). Mixed layers, however, ranged from 3.5 to 44 m (Table 1). Chlorophyll was maintained at low levels in the upper 10 m; between 10 and 100 m chlorophyll decreased over the course of the phase (Figure 9).

Correlation Analysis

Overall variability (defined as standard deviations) over the entire season was substantial. Average 0-50 m chlorophyll variability was large and, like SST and SS σ_T , had the greatest variability during the dissipation period. However, variability in both wind speeds and MLDs

generally decreased over the course of the three periods. Finally, bootstrapped 95% confidence intervals of residual variance indicated that there were significant differences in variability between all combinations of SST, SS σ_T , MLD, average 0-50 m chlorophyll, or wind speeds among the three phases (Jones, 2015). As the season progressed and chlorophyll began to dissipate throughout the upper 100 m, trends (such as the shift from increasing to decreasing chlorophyll and redistribution of particulate matter within the upper 100 m) resulted in correlational shifts that were depth-dependent (Figure 11g). While temperature and chlorophyll were strongly, positively correlated during the accumulation period when both were increasing and vertical flux was minimal, a moving correlation suggests a switch to a negative correlation ($r = -0.65$ to -0.41 ; $p < 0.001$) in the upper 30 m around December 15-18 (Figure 12). Temperatures below 40 m during the dissipation remain significantly correlated with chlorophyll ($p < 0.001$), with the correlations strengthening with depth (from $r = 0.28$ from 40-50 m to $r = 0.76$ from 80-90 m; Figure 11h).

Shortly after the accumulation phase December 5 wind event, when wind speeds were low and MLDs had deepened in response to the increased winds, chlorophyll decreases were observed in the upper 30 m and chlorophyll increases were observed between 50 and 90 m. As a result, negative correlations between MLD and chlorophyll in the surface 40 m ($r = -0.43$ to -0.11 ; $p = < 0.001$ to 0.046) and positive correlations between 50 and 80 m ($r = 0.12$ to 0.17 ; $p = 0.002$ to 0.038) were observed (Figure 11d). Wind speeds were positively correlated with chlorophyll in the upper 30 m ($r = 0.21$ to 0.30 ; $p < 0.001$), and were negatively correlated with chlorophyll between 40 and 100 m ($r = -0.51$ to -0.21 ; $p < 0.001$).

During the dissipation period, correlation patterns over the upper 100 m between MLD and chlorophyll did not display a sign change, but the magnitude changed due to increased exchanges

of chlorophyll (Figure 11e). As in the accumulation period, changes to chlorophyll were observed to coincide with the excursions in the MLD. Correlations of MLD with chlorophyll between 10 and 30 m remained negative ($r = -0.26$ to -0.20 ; $p < 0.001$), and between 40-100 m became more strongly positive ($r = 0.15$ to 0.49 ; $p < 0.001$; Figure 11e). The positive correlations during the accumulation period between wind speed and chlorophyll in the upper 30 m and negative correlations from 40 and 100 m reversed during the dissipation period, reflecting the increased temporal alignment (on the order of 2-12 h) between wind events and MLD-chlorophyll dissipations (Figure 11a,b,d,e). As a result, wind speed and chlorophyll display a weak negative correlation from 10-20 m ($r = -0.15$; $p < 0.001$), but positive correlations between 30 and 100 m ($r = 0.14$ to 0.40 ; $p < 0.001$). Moving correlations suggest that the changes in correlation signs occurred on December 9-11, 2012, during the later stages of the accumulation period (Figure 12).

Period-wide trends during the post-dissipation phase (i.e., gradually increasing MLDs and decreasing chlorophyll) likely drive the negative correlations found between MLD and chlorophyll over the upper 100 m ($r = -0.36$ to -0.14 ; $p = < 0.001$ to 0.019 ; Figure 11f). The influence of trends were again highlighted between wind speed and chlorophyll by the shift from positive correlations between 30 and 100 m (weakly negative near the surface) during the dissipation phase to weak, negative correlations ($r = -0.31$ to -0.15 ; $p = < 0.001$ to 0.019) across most of the water column (10-100 m) during the post-dissipation period (Figure 11b,c). A shift in the correlation patterns between temperature and chlorophyll continued during the post-dissipation period, as chlorophyll reductions were observed over much of the upper 100 m and the warming trend continued (Figure 11h,i). With that, negative correlations between temperature and surface chlorophyll deepened from 0-30 m to 10-60 m ($r = -0.57$ to -0.24 ; $p < 0.001$) between the dissipation and post-dissipation periods (Figures 11h,i). Also, the deeper,

positive correlations between temperature and chlorophyll during dissipation ($r = 0.28$ to 0.76 ; $p < 0.001$; 40-100 m) weakened and were limited to 90-100 m ($r = 0.22$; $p < 0.001$). In the upper 10 m, temperature is positively correlated with chlorophyll ($r = 0.27$; $p < 0.001$), indicating the influence of atmospheric forcing near the conclusion of the phytoplankton summer season (Figure 11i). Lastly, moving correlations corroborate the depth dependent nature of these changes, with shifts from positive to negative correlation occurring later as depth increases (Figure 12).

4. Discussion

Biological and hydrographic conditions in the southwestern Ross Sea, as documented by glider observations and weather station data, provided a broad view of the environmental conditions and demonstrate the substantial spatial and temporal variability in the region, as well as the atmosphere-ocean coupling. From our 2012-2013 glider observations, the surface waters were warmer, less dense, and experienced shallow MLDs when compared to other years (e.g., Smith et al., 2011a; Kaufman et al., 2014; Jones, 2015; Queste et al., 2015). Wind data indicated that the 2012-2013 study period had low average wind speeds (Jones, 2015). Previous studies have demonstrated that the Ross Sea continental shelf has experienced decadal trends in increased ice concentrations, decreased open ice duration, and decreased salinities (Stammerjohn et al., 2008; Jacobs and Giulivi, 2010). Therefore, it is possible that a portion of the 2012-2013 seasonal trends may be partially attributable to the broader forcing observed across the entire Ross Sea sector.

It is well understood that environmental conditions in the Ross Sea exhibit high levels of spatial and temporal variability, as do phytoplankton concentrations as determined from ocean color observations and in situ measurements (Arrigo and McClain, 1994; Arrigo et al., 1999;

Smith et al., 2013). In the Ross Sea, this variability may be caused by factors such as ice, vertical or horizontal advection, stratification, and seasonality. Our observations, collected within a small area (25x50 km), were characterized by 1) relatively pronounced mesoscale variability, 2) perturbations in water column variables during and subsequent to wind events, 3) a relatively complete seasonal trend was observed when compared to previous studies using gliders (Kaufman et al., 2014; Queste et al., 2015), and 4) movement of chlorophyll to depth that was related to wind mixing events. The variability observed reflects different environmental factors. For example, differences in the proximity to ice between different study locations may have had an effect on the observed differences in variability. Our glider operated close to pack ice, which especially during the accumulation and dissipation periods may have facilitated the rapid excursions in chlorophyll. Ice can induce a great deal of variability to the vertical stratification and irradiance availability within a water column, as melting ice imparts low salinity water at the surface, changing stratification, which in turn leads to altered phytoplankton growth and accumulation.

The strength of the coupling between wind events and MLD excursions may change due to the shifting wind directions over the course of the accumulation and dissipation periods. For example, wind directions during the accumulation period are typically from the southwest ranging primarily from 210 to 260°. From the accumulation into the dissipation and post-dissipation period, the wind direction shifts westward (the dissipation period ranges from 215 to 295° and the post-dissipation period ranges from 210 to 300°) and wind speeds are reduced. The shift and the lessening are likely due to seasonal changes in regional atmospheric circulation patterns as influenced by the seasonal changes and latitudinal shifts in atmospheric pressure cells and impacts of local topography (e.g., Ross Island). Additionally, the study region is dominated by a west to east current flow along the ice shelf and around Ross Island, that while fairly

consistent during the summer season, interacts with the bathymetry and may contribute to variability (Dinniman et al., 2007). Given that average wind directions do not change dramatically, and that the dominant direction limits winds over open water, it is likely that direction has a secondary role in regulating mixed layer depth at our location.

During the accumulation period short-term disturbances to water column structure driven by wind events had the greatest effect on chlorophyll. Changes resulting from wind mixing typically propagated from the surface to depth and were determined by the speed and duration of a wind event. Wind events changed mixed layer depths as well as the temperature and density vertical profiles in the upper 100 m. While changing mixed layer depths may be indicative of wind induced mixing, it should be noted that wind events may induce complex circulation or mixing not captured by the density threshold measurement of mixed layer depth (Brainerd and Gregg, 1995). For example, wind events on December 4-5 and two wind events from December 23-26 (durations of ~48-96 h) resulted in MLD increases from ~5-10 m to ~25-80 m, where they remained for approximately 36-72 h (slightly shorter than wind events and with a lag; Figures 7, 8). Mixing was also indicated by temperature changes during periods of increased MLDs, where noticeable increases in temperature were observed between 30 and 90 m during the December 23-26 mixing event. Deeper MLDs (60-80 m) increased temperatures between 70 and 90 m (Figures 7, 8). In contrast, relatively calm periods, such as those observed over the first two days of December and at the end of the accumulation period, promoted the shoaling of MLDs to less than 15 m (Figures 5, 7). By limiting deep mixing and maintaining phytoplankton in the well-lit, presumably nutrient-replete surface layer, shallow and stable MLDs generate periods (hours to days) that allow biomass to accumulate and growth to proceed under irradiance-saturated conditions (Smith and Donaldson, 2015).

The delays between the onset of a wind event and the resulting changes to the mixed layer during the accumulation period were on the order of 12-24 h (Figure 7). Therefore, the temporal alignment of accumulated biomass (following calm, shallow MLD periods) and the deepening of MLDs (in response to wind energy inputs) produced significant correlations in the upper 30 m between wind speed or MLD with chlorophyll ($r = 0.21$ to 0.30 and $r = -0.43$ to -0.35 , respectively; $p < 0.001$; Figure 11a,d). However, between 40 and 100 m, these correlations had the reverse relationships, with wind speed and chlorophyll negatively correlated ($r = -0.51$ to -0.21 ; $p < 0.001$) and MLD and chlorophyll positively correlated between 50 and 80 m ($r = 0.12$ to 0.17 ; $p = 0.011$ to 0.002 ; Figure 11a,d).

The contrasting observations of the correlations between wind speed and MLD with chlorophyll, and the change in correlational sign that occurs at the inflection point at 40 m, are both the result of two inter-related processes. First, the temporal uncoupling between a wind event and the resultant mixing response or MLD excursion (ca. 12-24 h) leads to the differences in signs between wind speed and MLD correlations. Importantly, the timing between MLD excursions and chlorophyll responses experience little to no change, thus maintaining the relationship of MLD with chlorophyll between the accumulation and the dissipation periods (Figure 11d,e). Second, the resultant mixing effects of the wind event (and increased MLDs) have depth-dependent influences on chlorophyll. Chlorophyll is reduced in shallow waters and appears in deeper waters as a result of the MLD excursions to depth with little delay. During the accumulation period and the December 5-7 wind event, biomass increased substantially between 40 and 80 m following the wind event (Figure 7), which is unlikely to be fully explained solely by wind induced mixing and MLD excursions due to the relatively shallow effects of wind mixing (MLDs increasing to 25-80 m). The timing and nature of the disappearance of chlorophyll shallower than 40 m and the appearance of chlorophyll between 50 and 90 m

suggests that wind-induced mixing and increased turbulence caused aggregate formation and enhanced vertical flux (Waniek, 2003; Burd and Jackson, 2009). That is, particle-particle interaction was enhanced by turbulence, and resulted in the formation of larger particles (or aggregates) that had significant numbers of chlorophyll-containing cells. These larger particles sank rapidly and were then observed at depth, though aggregates continue to sink and remineralize at deeper depths. For example, chlorophyll disappearances between 0 and 30 m preceded the MLD increases and are aligned temporally with the maximal wind speeds on December 4, 2012. Within 12 h, biomass appeared at 70 to 90 m prior to the influence of the MLD at those depths (Figure 7). Chlorophyll decreases rapidly below this depth (Jones, 2015), likely due to oxidation via heterotrophic growth and grazing. Together, this suggests that vertical biomass displacement may be explained by vertical water movement (the result of wind/MLD events), coupled with aggregate formation and vertical flux (the result of wind-derived turbulence).

The overall abundance of aggregates is a function of production and loss processes that are influenced by several environmental factors, such as particle stickiness and frequency of particle interaction (Burd and Jackson, 2009). Depending on the complex interplay of these factors, aggregate formation and increased particle sizes may develop. Alternatively, wind-driven turbulence may also cause aggregate or particle disruption, where the formation of smaller particles may slow the rates of vertical fluxes and the movement of biomass to depth. Coincident with slower sinking rates, increased remineralization of smaller particles might also occur. However, our data suggest that the net processes during these short intervals resulted in increased aggregate sizes and enhanced vertical flux of chlorophyll to depth.

The majority of biomass reductions in the surface waters and rapid appearances in waters deeper than 40-50 m occurred during the dissipation period. Biomass during this time is

typically dominated by *Phaeocystis antarctica* (Smith and Gordon, 1997; Arrigo et al., 1999; Smith et al., 2013), and it has been suggested that the colonial form of *Phaeocystis* contributes greatly to aggregate formation and subsequent, rapid vertical flux to depth (DiTullio et al., 2000; Smith and Asper, 2001; Smith et al., 2011b). With differences in the carbon to chlorophyll ratios between *Phaeocystis* and diatom species (Kaufman et al., 2014), observations of POC:Chl during the study periods suggest low POC:Chl ratios (*Phaeocystis*-dominated blooms; Kaufman et al., 2014) were present during the accumulation and dissipation periods, while high POC:Chl ratios (diatom-dominated blooms; Kaufman et al., 2014) were present during the post-dissipation period (Figure 3). Therefore, it is likely that aggregate processes on *Phaeocystis*-dominated blooms during the dissipation period likely played a major role in the enhanced vertical flux to depth.

As biomass became maximal and the bloom entered the dissipation period, changes in wind speeds, MLDs, and chlorophyll caused shifts in the correlations among those variables. The most altered correlation (Figure 11b) was between wind speed and chlorophyll, with a negative correlation between 10 and 20 m ($r = -0.15$; $p < 0.001$) and positive correlations between 30 and 100 m ($r = 0.14$ to 0.40 ; $p < 0.001$). This reflects a change in the delay from 12-24 h during the accumulation period to 2-12 h during the dissipation period between the onset of a wind event and the resulting response in the MLD. This is illustrated by the prominent wind events during the dissipation period. Following a calm period from December 14-22 with variable and high chlorophyll levels in the upper 40 m, wind events from December 23-27 quickly reduced average chlorophyll concentrations from 4 - 8 to 1.5 - 3 $\mu\text{g L}^{-1}$ (Figure 8). With the stronger temporal coupling between wind events and a response in MLD, the effect of the increased winds on chlorophyll was rapid. Compared with the accumulation phase, the correlations in the dissipation phase suggest that wind events typically cause chlorophyll to decrease from waters

shallower than 40 m and increase between 40 and 100 m (Figure 8). Integrated concentrations (from 0-100 m) did not change markedly, suggesting a redistribution of pigments rather than a removal (Figure 10b). The redistribution of chlorophyll biomass is observed as an increase in the 0-40 m chlorophyll concentration in Figure 10c following the December 23-25 wind event. Moving correlation analyses suggest that this shift in the signs of the wind speed with chlorophyll correlations occurred just prior to the bloom maxima (December 9-13; Figure 12).

While the correlations between wind speed and chlorophyll changed from the accumulation to the dissipation period due to changes in timing, there are minimal changes for the correlations between MLD and chlorophyll moving into the dissipation period, as changes in chlorophyll remain tightly coupled to MLD. Significant, negative correlations near the surface weakened (10 to 30 m; $r = -0.26$ to -0.20 ; $p < 0.001$), while significant correlations at depth strengthened (40 to 100 m; $r = 0.15$ to 0.49 ; $p < 0.001$; Figures 11d,e). Generally, the underlying mechanism of shoaled MLDs promoting biomass increases and minimal mixing to depth, and increased MLDs carrying phytoplankton biomass to depth and stimulating aggregate formation appears to hold from the accumulation through the dissipation period, leaving the correlation pattern largely intact. During the dissipation period, this pattern is most apparent during the MLD excursions from 5-10 m to 70-75 m that occurred on December 23-26. The decline of chlorophyll between 10 and 40 m on December 22 results in chlorophyll appearances between 50 and 90 m (Figure 8).

As illustrated by a period of low wind speeds and stable, shallow MLDs (December 14-22), the coupling between wind events and MLD excursions with chlorophyll changes may not always be the case when noting the variability in chlorophyll observed between 10 and 50 m. Nevertheless, while little change is noted in MLDs during this low wind period, a short duration (8-12 h) increase in wind speed on December 18 appeared to drive chlorophyll reductions in

shallow waters (10-20 m) and promote its appearance at depth (30-50 m) on December 19 (Figure 8).

By the post-dissipation period (January 7-19), reductions of chlorophyll across the upper 100 m caused the correlations of wind speed and MLD with chlorophyll to weaken (Figure 11c,f). That is, with chlorophyll largely at post-bloom levels, three wind events (January 10, 12-13, and 14-15) and limited MLD excursions have a minimal effect upon absolute chlorophyll vertical distribution (Figure 9). With the end of the growing season approaching (Smith et al., 2011a), trends of increasing wind events, the associated deepening of the MLD, and decreasing chlorophyll appear to be the driving factors in a switch to negative correlations for both wind speed and MLD (Figure 11c,f).

Among several factors that lead to variability in shallow chlorophyll concentrations, these observations suggest strongly that vertical stratification may play one of the greater roles in influencing the variability observed in the upper 100 m of the water column. Observations of seasonal changes to the temporal coupling between wind events and MLD excursions illustrate that alternating calm and perturbation periods rework vertical stratification, so that mid-season wind events are more effective at rapidly deepening mixing and transferring chlorophyll to depth (Smith and Jones, 2015). Other studies have noted substantial chlorophyll concentrations at depth (e.g., DiTullio et al., 2000) and suggested that they indicate substantial vertical fluxes of organic material. Our results support this concept – that is, short-term events induced by atmospheric forcing can result in rapid fluxes of organic matter to depth. Understanding and predicting the magnitude of such perturbations is challenging, but such events may play important roles in the biogeochemical cycles of polar regions.

5. Conclusions

The seasonal progression in surface layer hydrography and biological conditions is influenced by short-term events driven by winds. Wind energy is transferred to the water column, increasing turbulence and in most cases mixed layer depths. These in turn alter vertical chlorophyll distributions throughout the water column by enhancing particle-particle interactions, generating larger aggregates which sink more rapidly to depth. Such events and changes are generally short (less than 2 days) and are interspersed with periods of low winds and strong stratification that promote rapid phytoplankton growth. The coupling of the wind-stratification-chlorophyll linkage depends on the season, and is greatest when biomass is near maximum and least when biomass is increasing or has largely dissipated. These event-scale perturbations impose substantial variability on the seasonal pattern and have the potential for altering vertical distributions of organic matter and its flux to depth.

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References Cited

- Ainley, D.G., Ballard, G., Jones, R.M., Jongsomjit, D., Pierce, S.D., Smith, W.O., Veloz, S., 2015. Trophic cascades in the western Ross Sea, Antarctica: revisited. *Marine Ecology Progress Series* 534, 1–16. doi:10.3354/meps11394
- Arrigo, K.R., 2014. Sea ice ecosystems. *Annual Review of Marine Science* 6, 439–67. doi:10.1146/annurev-marine-010213-135103
- Arrigo, K.R., DiTullio, G.R., Dunbar, R.B., Robinson, D.H., Van Woert, M.L., Worthen, D.L., Lizotte, M.P., 2000. Phytoplankton taxonomic variability in nutrient utilization and primary production in the Ross Sea. *Journal of Geophysical Research* 105, 8827–8846.
- Arrigo, K.R., McClain, C.R., 1994. Spring phytoplankton production in the western Ross Sea. *Science* 266, 261–263. doi:10.1126/science.266.5183.261
- Arrigo, K.R., Robinson, D.H., Worthen, D.L., Dunbar, R.B., DiTullio, G.R., Van Woert, M.L., Lizotte, M.P., 1999. Phytoplankton community structure and the drawdown of nutrients and CO₂ in the Southern Ocean. *Science* 283, 365–367. doi:10.1126/science.283.5400.365
- Arrigo, K.R., van Dijken, G.L., 2003. Phytoplankton dynamics within 37 Antarctic coastal polynya systems. *Journal of Geophysical Research* 108, 3271. doi:10.1029/2002JC001739
- Arrigo, K.R., Van Dijken, G.L., 2004. Annual changes in sea-ice, chlorophyll a, and primary production in the Ross Sea, Antarctica. *Deep-Sea Research II* 51, 117–138. doi:10.1016/j.dsr2.2003.04.003
- Arrigo, K.R., van Dijken, G.L., Bushinsky, S., 2008. Primary production in the Southern Ocean, 1997–2006. *Journal of Geophysical Research* 113, C08004. doi:10.1029/2007JC004551
- Arrigo, K.R., Weiss, A.M., Smith, W.O., 1998a. Physical forcing of phytoplankton dynamics in the southwestern Ross Sea. *Journal of Geophysical Research* 103, 1007–1021.
- Arrigo, K.R., Worthen, D.L., Schnell, A., Lizotte, M.P., 1998b. Primary production in Southern

- 595 Ocean waters. *Journal of Geophysical Research* 103, 15587–15600.
- 596 Brainerd, K.E., Gregg, M.C., 1995. Surface mixed and mixing layer depths. *Deep Sea Research I*
597 42:9, 1521-1543.
- 598 Boss, E.S., Pegau, W.S., 2001. Relationship of light scattering at an angle in the backward
599 direction to the backscattering coefficient. *Applied Optics* 40, 5503–5507.
- 600 Burd, A.B., Jackson, G.A., 2009. Particle aggregation. *Annual Review of Marine Science* 1, 65–
601 90. doi:10.1146/annurev.marine.010908.163904
- 602 Dinniman, M.S., Klinck, J.M., Smith, W.O., 2007. Influence of sea ice cover and icebergs on
603 circulation and water mass formation in a numerical circulation model of the Ross Sea,
604 Antarctica. *Journal of Geophysical Research* 112, C11013. doi:10.1029/2006JC004036
- 605 DiTullio, G.R., Grebmeier, J.M., Arrigo, K.R., Lizotte, M.P., Robinson, D.H., Leventer, A.R.,
606 Barry, J.P., Van Woert, M.L., Dunbar, R.B., 2000. Rapid and early export of *Phaeocystis*
607 *antarctica* blooms in the Ross Sea, Antarctica. *Nature* 404, 595–598. doi:10.1038/35007061
- 608 DiTullio, G.R., Smith, W.O., 1996. Spatial patterns in phytoplankton biomass and pigment
609 distributions in the Ross Sea. *Journal of Geophysical Research* 101, 18467–18477.
610 doi:10.1029/96JC00034
- 611 Dunbar, R.B., Leventer, A.R., Mucciarone, D.A., 1998. Water column sediment fluxes in the
612 Ross Sea, Antarctica: Atmospheric and sea ice forcing. *Journal of Geophysical Research*
613 103, 30741–30759. doi:10.1029/1998JC900001
- 614 Eriksen, C.C., Osse, T.J., Light, R.D., Wen, T., Lehman, T.W., Sabin, P.L., Ballard, J.W.,
615 Chiodi, A.M., 2001. Seaglider: a long-range autonomous underwater vehicle for
616 oceanographic research. *IEEE Journal of Oceanic Engineering* 26, 424–436.
617 doi:10.1109/48.972073
- 618 Gardner, W.D., Richardson, M.J., Smith, W.O., 2000. Seasonal patterns of water column

- 619 particulate organic carbon and fluxes in the Ross Sea, Antarctica. *Deep-Sea Research II* 47,
620 3423–3449.
- 621 Jacobs, S.S., Giulivi, C.F., 2010. Large multidecadal salinity trends near the Pacific–Antarctic
622 continental margin. *Journal of Climate* 23, 4508–4524. doi:10.1175/2010JCLI3284.1
- 623 JGOFS, 1996. Protocols for the Joint Global Ocean Flux Study (JGOFS) Core Measurements,
624 IOC SCOR Rpt. 19. Bergen, Norway.
- 625 Jones, R.M., 2015. The Influence of Short-Term Events on the Hydrographic and Biological
626 Structure of the Southwestern Ross Sea. The College of William and Mary.
- 627 Kaufman, D.E., Friedrichs, M.A.M., Smith, W.O., Queste, B.Y., Heywood, K.J., 2014.
628 Biogeochemical variability in the southern Ross Sea as observed by a glider deployment.
629 *Deep-Sea Research I* 92, 93–106.
- 630 Morel, A., 1974. Optical properties of pure water and pure sea water, in: Jerlov, N.G., Nielsen,
631 E.S. (Eds.), *Optical Aspects of Oceanography*. Elsevier, New York, pp. 1–24.
- 632 Peloquin, J.A., Smith, W.O., 2007. Phytoplankton blooms in the Ross Sea, Antarctica:
633 Interannual variability in magnitude, temporal patterns, and composition. *Journal of*
634 *Geophysical Research* 112, C08013. doi:10.1029/2006JC003816
- 635 Pinkerton, M.H., Bradford-Grieve, J.M., 2014. Characterizing foodweb structure to identify
636 potential ecosystem effects of fishing in the Ross Sea, Antarctica. *ICES Journal of Marine*
637 *Science* 71, 1542–1553.
- 638 Queste, B.Y., Heywood, K.J., Smith, W.O., Kaufman, D.E., Jickells, T.D., Dinniman, M.S.,
639 2015. Observations of dissolved oxygen dynamics during a phytoplankton bloom in the
640 Ross Sea polynya. *Antarctic Science* 27, 362–372. doi:10.1017/S0954102014000881
- 641 Rudnick, D.L., Davis, R.E., Eriksen, C.C., Fratantoni, D.M., Perry, M.J., 2004. Underwater
642 gliders for ocean research. *Marine Technology Society Journal* 38, 73–84.

- doi:10.4031/002533204787522703
- Sedwick, P.N., DiTullio, G.R., Mackey, D.J., 2000. Iron and manganese in the Ross Sea, Antarctica: Seasonal iron limitation in Antarctic shelf waters. *Journal of Geophysical Research* 105, 11321–11336. doi:10.1029/2000JC000256
- Sedwick, P.N., Marsay, C.M., Sohst, B.M., Aguilar-Islas, A.M., Lohan, M.C., Long, M.C., Arrigo, K.R., Dunbar, R.B., Saito, M.A., Smith, W.O., DiTullio, G.R., 2011. Early season depletion of dissolved iron in the Ross Sea polynya: Implications for iron dynamics on the Antarctic continental shelf. *Journal of Geophysical Research* 116, 1–19. doi:10.1029/2010JC006553
- Smith, W.O., Ainley, D.G., Arrigo, K.R., Dinniman, M.S., 2014. The oceanography and ecology of the Ross Sea. *Annual Review of Marine Science* 6, 469–487. doi:10.1146/annurev-marine-010213-135114
- Smith, W.O., Asper, V.L., 2001. The influence of phytoplankton assemblage composition on biogeochemical characteristics and cycles in the southern Ross Sea, Antarctica. *Deep-Sea Research I* 48, 137–161. doi:10.1016/S0967-0637(00)00045-5
- Smith, W.O., Asper, V.L., Tozzi, S., Liu, X., Stammerjohn, S.E., 2011a. Surface layer variability in the Ross Sea, Antarctica as assessed by in situ fluorescence measurements. *Progress in Oceanography* 88, 28–45. doi:10.1016/j.pocean.2010.08.002
- Smith, W.O., Comiso, J.C., 2008. Influence of sea ice on primary production in the Southern Ocean: A satellite perspective. *Journal of Geophysical Research* 113, C05S93. doi:10.1029/2007JC004251
- Smith, W.O., Dennett, M.R., Mathot, S., Caron, D.A., 2003. The temporal dynamics of the flagellated and colonial stages of *Phaeocystis antarctica* in the Ross Sea. *Deep-Sea Research II* 50, 605–617. doi:10.1016/S0967-0645(02)00586-6

- 667 Smith, W.O., Donaldson, K., 2015. Photosynthesis–irradiance responses in the Ross Sea,
668 Antarctica: a meta-analysis. *Biogeosciences* 12, 3567–3577. doi:10.5194/bg-12-3567-2015
- 669 Smith, W.O., Gordon, L.I., 1997. Hyperproductivity of the Ross Sea (Antarctica) polynya during
670 austral spring. *Geophysical Research Letters* 24, 233–236. doi:10.1029/96GL03926
- 671 Smith, W.O., Jones, R.M., 2015. Vertical mixing, critical depths, and phytoplankton growth in
672 the Ross Sea. *ICES Journal of Marine Science* 72, 1952–1960. doi:10.1093/icesjms/fsu234
- 673 Smith, W.O., Marra, J., Hiscock, M.R., Barber, R.T., 2000. The seasonal cycle of phytoplankton
674 biomass and primary productivity in the Ross Sea, Antarctica. *Deep-Sea Research II* 47,
675 3119–3140.
- 676 Smith, W.O., Shields, A.R., Dreyer, J.C., Peloquin, J.A., Asper, V.L., 2011b. Interannual
677 variability in vertical export in the Ross Sea: Magnitude, composition, and environmental
678 correlates. *Deep-Sea Research I* 58, 147–159. doi:10.1016/j.dsr.2010.11.007
- 679 Smith, W.O., Tozzi, S., Long, M.C., Sedwick, P.N., Peloquin, J.A., Dunbar, R.B., Hutchins,
680 D.A., Kolber, Z., DiTullio, G.R., 2013. Spatial and temporal variations in variable
681 fluorescence in the Ross Sea (Antarctica): Oceanographic correlates and bloom dynamics.
682 *Deep-Sea Research I* 79, 141–155. doi:10.1016/j.dsr.2013.05.002
- 683 Stammerjohn, S.E., Martinson, D.G., Smith, R.C., Yuan, X., Rind, D., 2008. Trends in Antarctic
684 annual sea ice retreat and advance and their relation to El Niño–Southern Oscillation and
685 Southern Annular Mode variability. *Journal of Geophysical Research* 113, C03S90.
686 doi:10.1029/2007JC004269
- 687 Stramski, D., Reynolds, R.A., Kahru, M., Mitchell, B.G., 1999. Estimation of particulate organic
688 carbon in the ocean from satellite remote sensing. *Science* 285, 239–242.
689 doi:10.1126/science.285.5425.239
- 690 Stramski, D., Reynolds, R.A., Mitchell, B.G., 1998. Relationships between the backscattering

- 691 coefficient, beam attenuation coefficient and particulate organic matter concentrations in the
692 Ross Sea, in: Ackleson, S., Cambell, J. (Eds.), Proceedings of Ocean Optics. Kona, Hawaii,
693 pp. 1–10.
- 694 Thomson, R.E., Fine, I. V., 2003. Estimating mixed layer depth from oceanic profile data.
695 Journal of Atmospheric and Oceanic Technology 20, 319–329.
- 696 Waniek, J.J., 2003. The role of physical forcing in initiation of spring blooms in the northeast
697 Atlantic. Journal of Marine Systems 39, 57–82. doi:10.1016/S0924-7963(02)00248-8
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Table 1. Mean and standard deviations of conditions measured during the three phases within the seasonal progression of hydrographic and biological variables. Numbers in brackets are the ranges encountered. SST = sea surface temperature; SSS = sea surface salinity; SS σ_T = sea surface density.

Variable/Phase	Accumulation	Dissipation	Post-Dissipation
Wind Speed [m s^{-1}]	4.1 ± 3.3 [0.0 to 16.6]	3.8 ± 3.0 [0.0 to 13.8]	3.2 ± 2.6 [0.0 to 12.1]
SST [$^{\circ}\text{C}$]	-1.14 ± 0.22 [-1.74 to -0.22]	-0.04 ± 0.58 [-1.31 to 1.43]	0.70 ± 0.43 [-0.54 to 2.00]
SSS	34.409 ± 0.104 [33.690 to 34.632]	34.131 ± 0.245 [33.310 to 34.460]	34.222 ± 0.101 [33.824 to 34.400]
SS σ_T [kg m^{-3}]	27.687 ± 0.086 [27.109 to 27.842]	27.408 ± 0.212 [26.526 to 27.727]	27.442 ± 0.097 [27.030 to 27.612]
Mixed layer depth [m]	21.7 ± 24.9 [3.1 to 166.4]	16.3 ± 18.1 [3.0 to 91.8]	13.4 ± 7.7 [3.5 to 44.3]
Chlorophyll [$\mu\text{g L}^{-1}$]*	4.79 ± 1.14 [2.55 to 7.08]	3.34 ± 1.16 [1.28 to 7.50]	1.12 ± 0.33 [0.58 to 2.12]

*: Average for upper 50 m

Table 2. Mean and standard deviations of hydrographic and biological variables measured during three transects of the glider that were sampled closely in time. SST = sea surface temperature; SSS = sea surface salinity; SS σ_T = sea surface density.

Variable/Phase	Transect 1	Transect 2	Transect 3
Location	North Horizontal	North Horizontal	North Horizontal
Date and duration of sampling	19-22 Dec 2012 (63:13)	17-19 Jan 2013 (59:53)	25-28 Jan 2013 (59:20)
SST [°C]	0.20 ± 0.65	0.98 ± 0.59	-0.26 ± 0.21
SSS	33.861 ± 0.265	34.162 ± 0.163	34.220 ± 0.050
SS σ_T [kg m ⁻³]	27.159 ± 0.232	27.376 ± 0.167	27.490 ± 0.049
Mixed layer depth [m]	5.6 ± 2.3	17.9 ± 12.4	39.7 ± 25.9
Chlorophyll [$\mu\text{g L}^{-1}$]*	4.09 ± 1.03	0.83 ± 0.11	0.77 ± 0.08
Location	Central Horizontal	Central Horizontal	Central Horizontal
Date and duration of sampling	22-23 Dec 2012 (29:19)	2-3 Jan 2013 (23:34)	22-24 Jan 2013 (34:23)
SST [°C]	0.09 ± 0.41	0.71 ± 0.25	-0.12 ± 0.10
SSS	33.979 ± 0.368	34.252 ± 0.103	34.199 ± 0.050
SS σ_T [kg m ⁻³]	27.288 ± 0.322	27.473 ± 0.094	27.473 ± 0.038
Mixed layer depth [m]	8.1 ± 8.6	12.5 ± 7.0	32.0 ± 17.7
Chlorophyll [$\mu\text{g L}^{-1}$]*	3.29 ± 1.13	2.55 ± 0.33	0.81 ± 0.10
Location	South Horizontal	South Horizontal	South Horizontal
Date and duration of sampling	24-26 Dec 2012 (44:17)	12-14 Jan 2013 (52:35)	21-22 Jan 2013 (28:00)
SST [°C]	-0.73 ± 0.17	0.34 ± 0.25	-0.14 ± 0.12
SSS	34.356 ± 0.033	34.216 ± 0.076	34.179 ± 0.018
SS σ_T [kg m ⁻³]	27.623 ± 0.024	27.461 ± 0.046	27.452 ± 0.018
Mixed layer depth [m]	53.5 ± 25.1	14.1 ± 7.8	51.7 ± 35.5
Chlorophyll [$\mu\text{g L}^{-1}$]*	3.27 ± 0.40	0.97 ± 0.17	0.82 ± 0.14

*: Average for upper 50 m

712 **Figure Legends**

- 713 Figure 1. Map depicting the study location and glider track in the southern Ross Sea. Red
714 square in insert indicates the study area.
- 715 Figure 2. Temporal variations in a) wind speeds, b) sea surface temperatures (SST), c)
716 mixed layer depths (MLD) over the entire study period. Dashed lines indicate the
717 delineation of accumulation, dissipation, and post-dissipation periods.
- 718 Figure 3. Temporal variations in a) average 0-50 m particulate organic carbon (POC)
719 concentrations, b) average 0-50 m chlorophyll (Chl) concentrations, and c) the
720 ratio of POC to chlorophyll over the entire study period. Dashed lines delineate
721 the accumulation, dissipation, and post-dissipation periods.
- 722 Figure 4. Upper 150 m section plots for the entire glider deployment for a) temperature (C),
723 b) density (kg m^{-3}), c) chlorophyll concentration ($\mu\text{g L}^{-1}$), and d) particulate
724 organic carbon concentration ($\mu\text{g L}^{-1}$).
- 725 Figure 5. A “direction-intensity histogram” of wind speed and direction for the
726 accumulation, dissipation, and post-dissipation periods.
- 727 Figure 6. Mean vertical profiles of during the accumulation (solid lines), dissipation (long
728 dashed lines), and post-dissipation (short dashed lines) phases for a) temperature,
729 b) salinity, c) σ_T , d) chlorophyll, and e) particulate organic carbon.
- 730 Figure 7. Temporal changes within the accumulation period in a) wind speeds and mixed
731 layer depths (MLD), and average chlorophyll concentrations (Chl; solid line) and
732 temperatures (T; dashed line) from b) 0-10 m, c) 10-20 m, d) 20-30 m, e) 30-40
733 m, f) 50-60 m, g) 70-80 m, and h) 90-100 m.

- 734 Figure 8. Temporal changes within the dissipation period in a) wind speeds and mixed layer
735 depths (MLD), and average chlorophyll concentrations (Chl; solid line) and
736 temperatures (T; dashed line) from b) 0-10 m, c) 10-20 m, d) 20-30 m, e) 30-40
737 m, f) 50-60 m, g) 70-80 m, and h) 90-100 m.
- 738 Figure 9. Temporal changes within the post-dissipation period in a) wind speeds and mixed
739 layer depths (MLD), and average chlorophyll (Chl; solid line) concentrations and
740 temperatures (T; dashed line) from b) 0-10 m, c) 10-20 m, d) 20-30 m, e) 30-40
741 m, f) 50-60 m, g) 70-80 m, and h) 90-100 m.
- 742 Figure 10. Depth dependent changes to chlorophyll in response to wind events. Changes to
743 a) wind speed, b) 0-100 m average chlorophyll, and c) 0-40 m chlorophyll (solid
744 line) and 40-100 m chlorophyll (dashed line).
- 745 Figure 11. Pearson product-moment correlation coefficients between wind speed and average
746 10 m binned chlorophyll during the a) accumulation, b) dissipation, and c) post-
747 dissipation periods; between mixed layer depth and 10 m binned chlorophyll
748 during the d) accumulation, e) dissipation, and f) post-dissipation periods; and
749 between 10 m binned temperature and chlorophyll during the g) accumulation, h)
750 dissipation, and i) post-dissipation periods. Asterisks indicate significance at the p
751 < 0.05 level.
- 752 Figure 12. Moving correlations utilizing a moving window of 7 days between a) 0-10 m, b)
753 10-20 m, c) 20-30 m, d) 30-40 m, e) 50-60 m, f) 70-80 m, and g) 90-100 m
754 average 10 m binned chlorophyll with wind speed (blue lines), mixed layer depth
755 (MLD; red lines), and the corresponding 10 m average temperature (Temp; black
756 lines). Data are aligned on the x-axis at the window center.
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Fig. 5

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792 Fig. 12