Observations of dissolved oxygen dynamics during a phytoplankton bloom in the Ross Sea polynya

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Abstract

The Ross Sea polynya is one of the most productive regions in the Southern Ocean. However, limited access as the polynya opens and high spatial and temporal variability of physical and biological processes limit the use of conventional oceanographic methods to measure broad patterns of primary productivity early in the season. High-resolution observations from two Seagliders provide insights into the timing of a bloom in the southern Ross Sea polynya in December 2010. Changes in chlorophyll $a$ and dissolved oxygen concentrations are used to identify the dynamics of the bloom. Strong winds delayed the onset and its demise was accelerated through wind-induced mixing. Using a ratio of dissolved oxygen to carbon, net primary production is estimated over the duration of the bloom. Estimates suggest a sensitive balance between net community autotrophy and heterotrophy. Both gliders observing spatially distinct regions during the same time period found net community production rates of $-0.670$ and $0.362 \text{ g m}^{-2} \text{ d}^{-1}$ of carbon. The challenge of obtaining accurate estimates highlights the need for increased observational efforts, particularly focusing on sub-surface processes not resolved using surface or remote observations. Without these it will not be possible to determine the net trophic balance of the Ross Sea polynya and quantify the shelf’s importance in carbon export.

1 Introduction

The Ross Sea is considered to be one of the most productive regions in the Antarctic and is associated with significant uptake of $CO_2$ (Arrigo et al., 1998; Smith et al., 2000). The Ross Sea polynya is created by the northward displacement of sea ice by strong winds blowing off the Ross Ice Shelf and intrusions of warmer Modified Circumpolar Deep Water (MCDW) onto the shelf in spring and summer. New ice is formed during winter in the polynya leading to the production of dense brine and High Salinity Shelf Water, which can flow off the shelf as part of the global thermohaline circulation (Orsi and Wiederwohl, 2009).

The timing of phytoplankton blooms in the Ross Sea is strongly dictated by physical factors (ice retreat, solar irradiance and wind mixing). Arrigo et al. (1998) and Arrigo and van Dijken (2004) reported that the opening of the polynya is dependent on the ice cover compactness and thickness, which is regulated by winter air temperatures, rather than by katabatic winds. Once the polynya is open, winds (in conjunction with meltwater input from ice) regulate the onset of the bloom by controlling water column stratification. It has been suggested that strong winds occurring early in the season delay the bloom by reducing water column stability (Arrigo and van Dijken, 2004; Arrigo et al., 1998).

It has been suggested that as the polynya expands, *Phaeocystis antarctica*-dominated blooms occur, followed by increased diatom growth (Arrigo et al., 1999; Smith et al., 2014a). However, more recent observations have found variability in this phenology (Arrigo and van Dijken, 2004; Smith et al., 2006; Peloquin and Smith, 2007). As the *Phaeocystis* and diatom-dominated blooms have little
spatial or temporal overlap, it has been suggested that their occurrence is dependent upon environmental conditions rather than gradual successional dynamics (Arrigo et al., 2000).

Smith et al. (2011a, 2011b) described chlorophyll dynamics in the Ross Sea polynya and found that a combination of physical (advection, wind mixing) and biogeochemical (nutrient limitation, zooplankton ingestion and aggregate sinking) factors regulated fluctuations on short time and small spatial scales (a few days / tens of kilometres). This results in substantial biological patchiness in the Ross Sea (Hales and Takahashi, 2004; Smith et al., 2006). Kaufman et al. (2014) further elucidated the magnitude of these spatial and temporal variations, and found that the scales of the variability were small enough to be missed by most ship-based surveys.

It is unclear what mechanisms limit and terminate the initial phytoplankton bloom (defined by a rapid increase in phytoplankton biomass inferred from chlorophyll a). During spring, irradiance is the main factor limiting phytoplankton growth (Smith et al., 2000, 2012). The termination of the primary bloom and the limiting factor of subsequent blooms is likely to depend on another mechanism. Sedwick and DiTullio (1997) hypothesised that in high light environments, lower iron concentrations could favour diatom growth over P. antarctica as stratification continues to strengthen during austral summer. However, to date there are no data to suggest that the iron uptake capabilities of diatoms differ significantly from that of P. antarctica. Indeed, Sedwick et al. (2011) found that iron concentrations in spring and summer were similar as a result of rapid, early spring removal by Phaeocystis. Smith et al. (2011b) suggest that mixing events may contribute to the termination of blooms, at least in certain locations, by vertically redistributing standing stocks. Smith and Asper (2001) suggested that P. antarctica became iron-limited and sank rapidly from the euphotic zone as aggregates. It is likely that the combination of all of these mechanisms triggers the end of the primary bloom, while changes in iron availability alter phytoplankton composition during the rest of the growing season.

It has been hypothesised that intrusions of MCDW onto the shelf supply iron to the Surface Mixed Layer (SML) through episodic vertical mixing (Dinniman et al., 2011; Sedwick et al., 2011). This in turn would possibly enhance P. antarctica growth (which also is able to grow at low irradiances; Kropuenske et al., 2009). Another source of iron may be ice melt (Sedwick and DiTullio, 1997); however, the majority of sea ice in the central Ross Sea is advected away rather than melting in situ, whereas melting is more common in the western portion. Therefore, a spatial variation of iron input likely occurs and introduces variability in phytoplankton processes within the polynya as well (Arrigo and van Dijken, 2004).

Rates of primary productivity in the Ross Sea exceeding 2 g m$^{-2}$ d$^{-1}$ or 200 g m$^{-2}$ yr$^{-1}$ of carbon with a growing season of approximately 120 days have been reported (Smith and Gordon, 1997; Arrigo et al., 1999, 2000; Arrigo and van Dijken, 2004; Smith et al., 2006). In addition, strong interannual variability in both standing stocks and productivity occurs. Satellite observations (Arrigo and van Dijken, 2004; Peloquin and Smith, 2007), ship-board surveys (Smith et al., 2006, 2011a, 2011b) and modelling studies (Reddy and Arrigo, 2006; Smith and Comiso, 2008) all suggest that interannual variations in ice-melt and polynya opening along with wind mixing impact standing stocks, productivity and phytoplankton composition. A general trend of increasing water column stability, consistent with predictions of climate models of the Southern Ocean (Boyd et al., 2008), may result in increased primary production in the future (Smith et al., 2014b).

Productivity and standing stocks of phytoplankton have been estimated via satellite observations, but satellites are limited by ice and cloud cover and measure only near-surface ocean properties. Ship surveys have described the temporal and spatial variations of productivity and biomass, but such surveys are difficult as the polynya cannot be monitored continuously by ship due to the persistent ice presence and prohibitive costs. Ship-based surveys also struggle to resolve small-scale features, linked to the smaller scale of physical forcing as a function of the smaller Rossby radius at these latitudes O(4-5 km), while obtaining a synoptic view of the wider polynya (Kaufman et al., 2014). As variability in production and standing stocks is very closely tied to physical and biogeochemical
drivers (e.g. solar radiation, wind, currents, nutrient concentrations, plankton composition), it is critical that we resolve the physical, chemical and biological features at these smaller scales if we are to understand the controls on productivity and how these might change in the future.

This study used Autonomous Underwater Vehicles (AUVs) to observe early season bloom dynamics over an extended period (2 months) in the polynya where and when conventional platforms (i.e., ships and moorings) have a limited capability to provide data due to ice presence. In particular, observations presented here were focused on the spatial and temporal variability of phytoplankton blooms and the physical drivers which lead to such heterogeneity. By deploying Seagliders (Eriksen et al., 2001) in the polynya, we intended to understand the relationship between intrusions of MCDW and the distribution of chlorophyll $a$ across the southern Ross Sea. In particular, it was anticipated that the high resolution temporal and spatial data provided by gliders would resolve features that could not be observed using other sampling procedures. The dataset collected was the first to present observations of the initiation of the austral spring bloom throughout the water column across the western Ross polynya at a resolution sufficient to identify the sub-mesoscale variability of processes regulating production within the Ross polynya.

2 Methods

Two Seaglider (Eriksen et al., 2001) autonomous underwater gliders were deployed in the Ross Sea polynya from November 2010 to late January 2011. Each carried sensors for temperature, salinity (Seabird CT sail), Dissolved Oxygen ($DO$; Aanderaa 4330), chlorophyll $a$ fluorescence and optical backscatter at two different wavelengths (470 and 600 nm; Wetlabs Triplet ECOPuck). Seaglider 502 (SG502) was launched into McMurdo Sound on 22 November 2010 and transited beneath a sea-ice bridge into the Ross polynya on 14 December 2010. It performed a repeat zonal transect between 172°E and 180° at ca. 76°40'S and completed a total of 702 dives (Fig. 1). Seaglider 503 (SG503) was launched on 29 November 2010 directly into the Ross polynya, but suffered an instrument failure on the Wetlabs puck after 3 days. This prevented collection of chlorophyll $a$ fluorescence and optical backscatter data, but did not affect collection of temperature, salinity and $DO$ data. It surveyed a meridional bowtie track twice between 76° S and 77° S, crossing SG502’s track, with a total of 923 dives (Fig. 1). Only data when both gliders were in the Ross Sea polynya (between December 14 and January 21) are presented.

Both gliders were launched from the ice edge as no other means of launch were available (Asper et al., 2011). Launch conditions did not allow collection of data for calibration and no moorings were available for cross calibration during the survey. This limitation is one that must be accepted to observe the region this early in the season unless small boat capabilities become available from the nearby bases. All glider data were calibrated against a single CTD cast each from the RVIB Nathaniel B. Palmer during recovery on the January 20 (SG502) and 30 (SG503) of January. Salinity and chlorophyll $a$ sensors on the ship’s CTD rosette were calibrated against in situ samples. Shipboard bottle measurements of chlorophyll $a$ were determined via fluorescence on a Turner Designs Model 10AU fluorometer after filtration and 24 h extraction in 90% acetone. Fluorescence counts from SG502’s final upcast were regressed with bottle measurements of chlorophyll $a$ during recovery ($R^2 = 0.94$), and this regression was applied to all glider fluorescence data to determine estimated chlorophyll $a$ concentrations.

Ship $DO$ data were not calibrated using in situ measurements due to high replicate differences in Winkler titrations. The RVIB Palmer’s CTD sensor package was equipped with a Seabird SBE43 sensor that had recently been calibrated; this provided the best calibration data for the Seaglider optodes. For the densest watermass identified for each glider throughout the survey, a linear regression of $DO$ concentrations against time was performed and the resulting slope used as an estimate of sensor drift. This method assumes that the deepest water mass identified had a constant $DO$ concentrations and hence was minimally affected by biological production or consumption on the
time scales of the study. Figure 2 shows the resulting DO change rates of for waters of different temperature and salinity (T&S) properties after removal of the calculated optode drift.

Erroneous temperature, salinity and DO data points due to sensor noise were removed from the analysis using thresholds for reasonable data on the basis of calibrated data and visual inspection. Temperature, salinity, DO and chlorophyll a data were then binned using a regular 3 hr by 3 db grid. Finally, a vertical linear interpolation was used to fill empty bins. All figures use gridded observations rather than individual data points. A comparison of raw and gridded data (Fig. 3) showed no discernable differences or aliasing of the raw signal.

3 Results and discussion

3.1 Bloom timing and distribution

SG502 travelled east across Ross Bank before returning to the western Ross Sea (Fig. 1). During the eastward transect, SG502 recorded a deepening of a surface warm layer (25 m; Fig. 4a). Below the surface warm layer and above the 27.8 kg m$^{-3}$ isopycnal (dashed line in Fig. 4), an intermediate water mass exhibits temperatures warmer than the near-freezing bottom waters. On the return transect, this same 3-layer pattern remained, but with a much thicker surface warm layer extending to 50 m. At the eastern edge of the bank, where depths approach 250 m, the warm surface layer is not present. Salinity does not follow the same vertical distribution as temperature, with a much more gradual decrease from fresher surface waters (34.3) to the saline deep water (34.7) (Fig. 4b). The warm surface layer creates a strong pycnocline which deepens from 20 to 50 m; this pycnocline is not observed at the eastern edge of Ross Bank. Below this SML, isopycnals follow bathymetric contours in regions shallower than 400 m, but shoal away from the bank when depth exceeds 400 m.

SG503 observed a similar temperature and salinity distribution. A thin warm surface layer appears on 21 December and gradually deepens to 60 m. An intermediate water mass is visible down to the 27.8 kg m$^{-3}$ isopycnal (Fig. 5a). The warm surface layer is observed throughout SG503’s survey, indicating that its absence at the eastern edge of Ross Bank was likely due to increased vertical mixing in the shallow region or strong frontal processes along the eastern edge of the bank. Interestingly, there is little fresh water (< 34.45) visible above the shallower regions of SG503’s transect (Fig. 5b).

As the SML thickness was consistent between both SG502 and SG503, we assume that this SML was present over a much wider area and its formation is governed by large-scale processes, such as increasing solar radiation and changing wind speeds. Wind speeds above the location of each glider (Fig. 6a) suggest that the elevated winds at the start of the mission prevented the formation of a stable SML. Subsequently, wind speeds appear to have remained spatially consistent over both gliders. Another elevated wind speed event (> 10 m s$^{-1}$) began on January 8 but had little visible impact on water column density structure (Figs. 4 & 5). At the same time, SG502 and SG503 observed different isopycnal depths, implying a strong spatial dependence. We interpret this as geostrophic currents following bathymetric contours causing spatial variability in isopycnal depth. These geostrophic flows advect biogeochemical properties and are associated with fronts contributing to a strong physical and chemical heterogeneity in the western Ross Sea polynya.

Optical data were only recorded in the top 250 m for SG502 (Fig. 4d). A large bloom is observed within the Ross polynya while SG502 surveyed the Ross Bank. Chlorophyll a concentrations peaked at 7 mg m$^{-3}$, with the majority of the bloom showing subsurface maxima of 4 mg m$^{-3}$ at approximately 40 m depth and elevated chlorophyll concentrations between 10 and 65 m. Observed chlorophyll a concentrations decreased rapidly after 5 January when SG502 began the westward transect. Fig. 7 shows a 21-day composite of satellite ocean colour data when SG502 observed the bloom over Ross Bank (14 December to 4 January). Despite averaging satellite data over a 3 week period, a large area is devoid of data due to extensive cloud and ice cover. Nonetheless, an extensive
bloom similar to that observed by the Seaglider is revealed (4-7 mg m\(^{-3}\)) extending into the Central Basin. The satellite data indicate that the bloom was not spatially constrained, at least at the surface, to only the Ross bank, but was also present in the Central Basin region.

As this bloom included both the banks and basin (Fig. 7), it is not likely that its onset is regulated by advective processes such as mesoscale intrusions of iron-rich MCDW, as this would cause enhanced localised chlorophyll maxima along the edges of the bank (Kohut et al., 2013). The onset of the bloom in the Ross Sea polynya is traditionally considered to be initiated by increased irradiance, vertical stability and iron release in the water column due to ice melt (Sedwick and DiTullio, 1997; Smith and Gordon, 1997; Smith et al., 2000; Smith and Comiso, 2008). However, we observed a 3-week period between melting of the surface ice and the onset of the bloom (Fig. 6). Fig. 6 compares wind velocity above SG502 and chlorophyll a concentrations in the top 50 m. Our observations suggest a much closer relationship to wind speed than ice melt; a week after wind speeds decrease below 5 m s\(^{-1}\) and the warm SML begins to form, there is a dramatic increase in chlorophyll a concentration. Such rates of increase are consistent with known growth rates in the Ross Sea (e.g., Smith et al., 2000).

Overall, these elements indicate that the factor delaying the onset of the bloom was a lack of vertical stability in the water column. Once the ice receded, sufficient nutrients and light were available for phytoplankton growth, but elevated wind speeds prevented the phytoplankton from growing in a stratified water column and optimal light regime. Once wind speeds decreased, surface warming accelerated stratification of the surface layer, allowing the phytoplankton to grow and accumulate. Hence, ice melt is not the sole control of stratification in the region; strong winds also play a strong role in controlling the onset of the stratification of the water column.

The bloom ended in the first week of January with strong mixing of the water column and a redistribution of chlorophyll a (Fig. 4d) as shown by the reduction in chlorophyll a concentrations and the appearance of the chlorophyll maximum at depth (150 m). This coincides with the second elevated wind speed event (> 10 m s\(^{-1}\); Fig. 6). This agrees with observations by Smith et al. (2011b) that mixing events may cause bloom termination in the Ross Sea polynya. More detailed observations of the mechanisms regulating variability during the demise of this bloom are described by Kaufman et al. (2014).

3.2 Dissolved oxygen distribution

There is net consumption of DO during the survey within most water masses present (Fig. 2). Peak DO consumption occurs both in surface waters and in intermediate waters along the 27.8 kg m\(^{-3}\) isopycnal. Despite overall consumption, there is a net increase in surface dissolved oxygen during the first half of the section (Figs. 4c, 5c & 8). DO was consistent between gliders with a surface layer of oxygen supersaturation down to 50 m for both gliders. This layer was more diffuse and extended deeper (70 m) at the start of the survey. As the warm SML appeared, it exhibited supersaturation of DO (Fig. 8). Peak DO concentrations were observed by both gliders between 21 December and 1 January (Figs. 4c, 5c & 8). Fluctuations in surface saturation were synchronous between both gliders (Fig. 8). Finally, surface saturations decreased again during the final two weeks of the mission as the gliders surveyed the region west of the Ross Bank (west of 175°E). This period reflected a substantial increase in particulate organic carbon that was attributed to a secondary diatom bloom in the same location as the oxygen increase (Kaufman et al., 2014).

The apparent oxygen utilisation (AOU), defined as the difference between the measured dissolved oxygen concentration and saturation concentration, varies in both time and space (Fig. 8). It is used here rather than change in DO concentration to allow for the effects of physical changes, particularly temperature on oxygen saturation. Both SG502 and SG503’s observations show increasing AOU towards the bottom of the water column (Fig. 8). Consumption (i.e. positive AOU) above
90 μmol dm$^{-3}$ is constrained to regions deeper than 400 m according to SG502’s observations (Fig. 8a). It is likely that sinking organic matter from the phytoplankton bloom is remineralised gradually as it sinks and is deposited on the seabed; this remineralisation requires the consumption of DO and often leads to reduction of bottom mixed layer DO (Diaz and Rosenberg, 2008; Queste et al., 2013). Further consumption within the nepheloid layer and during resuspension of settled organic matter enhances AOU near the seabed. In regions deeper than 400 m, the Bottom Mixed Layer (BML) is of a much greater volume (up to 50 m), thereby reducing the impact of oxygen depletion caused by remineralisation of organic matter as there is a larger reservoir of DO. Additionally, the greater depth means more organic matter is remineralised as it sinks as opposed to once it settles on the seabed. This consumes oxygen throughout the water column rather than it concentrating the oxidation near the seabed.

Following these observations of biogenic DO variability, we derived a depth-integrated budget of DO production over time (Fig. 9 & 10). Production is maintained within the top 50 to 70 m of the water column within a SML of gradually changing density. The oxygen data were analysed to estimate production within the survey area in manner similar to Alkire et al. (2012) and Riser and Johnson (2008). Alkire et al. (2012) assessed production during a bloom using floats and gliders. They equated the rate of change of the water-column DO inventory to the sum of oxygen fluxes, including biological production and consumption, horizontal and vertical advection, isopycnal and diapycnal mixing and air-sea exchanges:

$$\frac{\Delta [O_2]}{\Delta t} = F_{\text{biological}} + F_{\text{advection}} + F_{\text{upwelling}} + F_{\text{isopycnal}} + F_{\text{diapycnal}} + F_{\text{ASE}}$$

Alkire et al. (2012) used a Lagrangian experimental design and assumed that advection was negligible. Here we assume that temporal changes of DO in the water column during the gliders’ repeated surveys are far greater than any spatial variability that might be aliased into the time series. For the simple calculations presented here, we further assume that the effects of upwelling and mixing are negligible compared with oxygen utilisation during the bloom. Based on these assumptions, we assume that apparent oxygen utilisation rates measured below the SML are biogenic. Estimation of air-sea exchanges is required for assessing surface mixed layer rates.

A rate of daily DO change was estimated by linearly regressing AOU against time for each depth bin; the slope of the regression is an estimate of $\frac{d[O_2]}{dt}$ (Figs. 9 & 10). A peak in oxygen production was observed at 30 m, decreasing to very low or negative values near the surface and switching to net consumption regime below 50 m. Maximum oxygen production correlates with maximum observed chlorophyll $a$ concentrations (Fig. 9b), suggesting primary production within the deep chlorophyll maximum (DCM) is the cause. Below the DCM, chlorophyll $a$ concentrations decreased gradually and remained relatively high (> 1.5 mg m$^{-3}$) throughout the upper 100 m. This suggests that phytoplankton sank from the euphotic zone and served as a significant source of organic matter to depth. Simultaneously, we observed a sharp decrease in DO below the DCM. Peak apparent consumption occurs at 70 m in both regions. The consumption of DO is strongest just below the DCM and gradually decreases to modest DO consumption rates. This indicates that the majority of organic matter produced within the euphotic zone is rapidly remineralised before it reaches 200 m, a suggestion that is consistent with POC distributions (Kaufman et al., 2014) and with studies performed in other regions around Antarctica (Weston et al. 2013). However, this contrasts with the substantial oxygen removal near the seabed. It is possible that an older water mass moving along the bank could contribute to the high AOU observed near the seabed. Kohut et al. (2013) and Kaufman et al. (2014) both suggest the presence of MCDW along Ross Bank during the 2010/2011 austral summer; this water mass can be identified by its lower oxygen concentration and its unique temperature/salinity features (Orsi and Wiederwohl, 2009). However, this likely has a minimal impact. The elevated AOU is possibly caused by consumption earlier in the season through tidal
resuspension of organic matter deposited on the seabed, allowing for slow but continuous
consumption (DeMaster et al., 1992; Nelson et al., 1996). Further study of the cause of near-bed
oxygen depletion on Ross Bank would be useful.

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\frac{\Delta [O_2]}{\Delta t} \text{ in the water column was } -0.40 \mu \text{mol dm}^{-3} \text{ d}^{-1} \text{ for SG502 and } 0.03 \mu \text{mol dm}^{-3} \text{ d}^{-1} \text{ for SG503. Assuming a water column depth of 400 m, this amounts to } -0.160 \text{ and } 0.012 \text{ mol m}^{-2} \text{ d}^{-1}. \text{ AOU/C ratios are generally thought to be between 0.5 and 2.0, with productive shelf seas closer to the latter value (Thomas, 2002). Using an AOU/C ratio of 2.0, this would suggest production rates of } -0.08 \text{ mol m}^{-2} \text{ d}^{-1} \text{ and } 0.006 \text{ mol m}^{-2} \text{ d}^{-1} \text{ of carbon. In the absence of air-sea exchange, the observed } \frac{\Delta [O_2]}{\Delta t} \text{ would indicate net community production of } -0.96 \text{ g m}^{-2} \text{ d}^{-1} \text{ of carbon for SG502 and } 0.072 \text{ g m}^{-2} \text{ d}^{-1} \text{ of carbon for SG503. As the surface layer is supersaturated (Fig. 8), air-sea gas exchange will lead to the rapid loss of excess DO from the mixed surface layer. Air-sea gas exchange depends on the concentration gradient in surface waters and a gas exchange rate, itself wind speed and temperature dependent. Estimates of sea to air oxygen flux based on ECMWF ERA interim wind speeds, Seaglider measurements of surface temperature, salinity and DO and using an empirical fit to dual tracer data by Johnson (2010) and Nightingale et al. (2000) are ca. 0.025 \text{ mol m}^{-2} \text{ d}^{-1} \text{ (Fig. 11; } \bar{x} = 0.0239 \text{ mol m}^{-2} \text{ d}^{-1}). \text{ By including an air-sea flux of 0.0239 mol m}^{-2} \text{ d}^{-1} \text{ (an additional } 0.29 \text{ g m}^{-2} \text{ d}^{-1} \text{ of carbon), net community production rates estimated from the observations of SG502 and SG503 are } -0.670 \text{ and } 0.362 \text{ g m}^{-2} \text{ d}^{-1}.}

4 Conclusions

It is difficult to reconcile the different estimates of production in this region, as both gliders were observing this region at the same period. The main sources of uncertainty in the estimates are the assumption of horizontal homogeneity and the estimation of air-sea exchanges in a cold region with ice-preservation. Furthermore, a value of 2 for the AOU:C ratio is an upper boundary; this implies that absolute values of carbon production and consumption are likely greater. However, in making these assumptions we come closer to providing a better understanding of the DO dynamics in this region. The contrasting results of SG502 and SG503, the former showing net consumption and the latter net production, highlight that this is a highly variable region, likely exhibiting both net production and net consumption depending on the time and region. The sinking event observed by SG502 at the end of the bloom likely had a disproportionate effect on the calculation of rates. If this sinking event were related to topographically induced flows along the bank, it is possible that the oxygen depletion signature caused by the rapid remineralisation of that organic matter would affect the calculation of rates differently to the data from SG503. Alternatively, the bloom’s demise may occur at different times across the survey region, which was previously observed by Smith et al. (2011a). The satellite imagery showed significant differences in bloom intensity either side of the bank.

It is likely that SG502’s data illustrate a more advanced stage of the bloom’s demise where production has greatly decreased in the SML and respiration has greatly increased as the biomass sinks to depth. SG503’s data represent an earlier stage of the bloom, where production is still occurring at a much greater rate within the SML and the phytoplankton assemblage is maintained within the SML, leading to only limited remineralisation in the sub-euphotic depths of the water column.

This mission served to demonstrate the substantial value of autonomous underwater gliders in remote regions such as the Ross Sea. These instruments provided a wealth of observations, with greater spatial and temporal resolution than any ship-based survey could. Autonomous underwater
gliders could greatly increase our understanding of the region’s biogeochemical cycles and the
influence of small and mesoscale processes on these cycles, and further oceanographic programs
would benefit from the merger of a variety of sampling platforms to resolve the temporal and spatial
variability we observed.

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Figure 1: Seaglider survey locations. Crosses on the solid line show every tenth dive by SG502, circles on the dotted line show every tenth dive by SG503. Dark shading corresponds to the 50% ice cover contour on the day SG502 crossed beneath the ice bridge (December 14th 2010) and entered Ross polynya. Contours correspond to bathymetry in metres (GEBCO, 2010). Contours are located at 200 m intervals through 1000 m. A. McMurdo Sound, B. Central Basin, C. Crary Bank, D. Pennell Bank, E. Ross Bank, F. Ross Island.
Figure 2: Dissolved oxygen change for specific temperature and salinity signatures during the bloom from linear regressions of dissolved oxygen concentrations (µmol dm$^{-3}$ d$^{-1}$) of data after sensor drift correction (a. SG502; b. SG503). Both SG502 and SG503 show very similar patterns with strong consumption in most water masses with the exception of the deep stable water mass which was used to estimate and remove sensor drift by shifting all rates so that the deep stable water mass rate was 0 µmol dm$^{-3}$ d$^{-1}$. 
Figure 3: Short DO section of the Seaglider transect showing data after binning (a) and scattered raw data (b). The binning process maintains all the variability visible in the raw data.
Figure 4: Temperature (a. °C), salinity (b.), DO (c. µmol dm$^{-3}$), AOU (d. µmol dm$^{-3}$) and chlorophyll a (e. mg m$^{-3}$) sections during the bloom observed by SG502 in the Ross Sea polynya. Potential density contours are shown at 0.1 kg m$^{-3}$ intervals from 27.9 kg m$^{-3}$ (solid line) to 27.6 kg m$^{-3}$ (dash-dot line). An estimate of mixed layer depth from Kaufman et al. (2014) is included as a thick solid line near the surface. The bloom study period is identified by the vertical bars.
Figure 5: Temperature (a. °C), salinity (b.), DO (c. µmol dm\(^{-3}\)) and AOU (d. µmol dm\(^{-3}\)) sections during the bloom observed by SG503 in the Ross Sea polynya. Potential density contours are shown at 0.1 kg m\(^{-3}\) intervals from 27.9 kg m\(^{-3}\) (solid line) to 27.6 kg m\(^{-3}\) (dash-dot line). An estimate of mixed layer depth from Kaufman et al. (2014) is included as a thick solid line near the surface. The bloom study period is identified by the vertical bars.
Figure 6: ECMWF ERA-Interm 10 m windspeed over the location of each glider (m s\(^{-1}\)) and mean chlorophyll a concentration (mg m\(^{-3}\)) throughout the top 250 m of SG502's observations shown as a dotted black line (a). Ice cover (%) over the location of each glider as recorded by SSMIS satellite imagery (b). Vertical black lines indicate the start (bloom initiation) and end of the period analysed in this paper. The solid blue line represents data relating to SG502. Dashed red line represents data relating to SG503.
Figure 7: Mean (a) and maximum values (b) of a 30 day composite of MODIS ocean colour surface chlorophyll a concentrations (mg m$^{-3}$) from the 14th of December to the 13th of January. White areas are devoid of data due to the continuous cloud and ice cover during this day period.
Figure 8: Apparent oxygen utilisation (µmol dm$^{-3}$) as observed by SG502 (a) and SG503 (b). 0 and 92 µmol dm$^{-3}$ contours added to highlight supersaturation threshold and peak consumption regions. SG502’s data show that peak apparent oxygen consumption is located on the shallower bank region.
Figure 9: Mean vertical profiles of potential temperature (a, blue, °C), salinity (a, red.), AOU (b, blue, µmol dm\(^{-3}\)) and chlorophyll a (b, red, mg m\(^{-3}\)) during the bloom identified by SG502. Total variability of the profiles (minimum to maximum extent) is displayed as a lighter shade for each variable. Vertical distribution of dissolved oxygen change over time (µmol dm\(^{-3}\) d\(^{-1}\)), as estimated by linear regressions for each depth bin (c). Shading around rates is indicative of one standard deviation for the regression of the rate (c).
Figure 10: Mean vertical profiles of potential temperature (a, blue. °C), salinity (a, red.) and AOU (b, blue. µmol dm$^{-3}$) during the bloom identified by SG503. Total variability of the profiles (minimum to maximum extent) is displayed as a lighter shade for each variable. Vertical distribution of dissolved oxygen change over time (µmol dm$^{-3}$ d$^{-1}$), as estimated by linear regressions for each depth bin (c). Shading around rates is indicative of one standard deviation for the regression of the rate (c).
Figure 11: Sea to air flux of dissolved oxygen (μmol dm$^{-2}$ d$^{-1}$) based on ECMWF ERA interim wind speeds, Seaglider measurements of surface temperature, salinity and DO and using an empirical fit to dual tracer data by Johnson (2010) and Nightingale et al. (2000).