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Relaxation of wind stress drives the abrupt onset of biological carbon uptake in the Kerguelen Bloom: a multisensor approach.

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Key Points:

- Phytoplankton onset is detected from in-situ high resolution multisensors data gathered by one mooring and two Biogeochemical Argo floats
- The bloom starts approximately two months after the net heat flux becomes positive, during an abrupt shoaling event of the mixing-layer
- The bloom starts only after a decrease in the depth over which winds actively mix the upper ocean leading to a decrease of the turbulence

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Abstract

We deployed sensors for physical and biogeochemical measurements on one Eulerian mooring and two Lagrangian biogeochemical Argo-floats on the Kerguelen Plateau. High temporal and vertical resolution measurements revealed an abrupt shoaling of both the mixed-layer depth and mixing-layer depth. The sudden stratification was concomitant with the start of significant biological activity detected by chlorophyll-a accumulation, oxygen oversaturation and dissolved inorganic carbon drawdown. The net community production computed in the mixing-layer during the onset period of 9 days was $119 \pm 7 \text{ mmol m}^{-2} \text{ d}^{-1}$. While it is generally admitted that bloom initiation is mostly driven by the onset of positive heat fluxes, our results suggest this is not a sufficient condition. Here we report that the decrease in the depth over which wind mixes the upper layer drives the initiation of the bloom. These results suggest that future atmospheric changes in Southern Ocean could impact the phenology of the blooms.

Plain Language Summary

The region of the Kerguelen Plateau is well known as a naturally iron fertilized region and it supports a productive marine ecosystem. In the present study, we investigate the evolution of the biogeochemical and physical conditions during the 2016 phytoplankton bloom event near the Kerguelen Plateau. We use a unique combination of moored instruments and profiling floats in order to follow the phytoplankton evolution at vertical, spatial and temporal scales and to understand the main physical drivers supporting such an abrupt bloom initiation that occurs only over a 9-days period. The large phytoplankton bloom develops during a major shallowing event bringing in few days the mixed-layer depth from its typical winter value to its typical summer value. This abrupt stratification of the water column is driven by a decline of the wind stress. These results have important implications considering that the wind regimes are predicted to intensify in the future in the Southern Ocean.

1 Introduction

More than 30 years ago the Southern Ocean (SO) has been classified as a High Nutrient Low Chlorophyll region (Minas et al., 1986). During the following decades, convincing arguments have been provided to explain that low nutrient utilization and therefore inefficient biological carbon pump in this ocean resulted primarily from iron limitation of photosynthetic microorganisms (Martin, 1990; Boyd et al., 2000; Blain et al., 2007). However, the increasing number of observations coming both from field studies and autonomous sensors embarked on various in-situ or satellite vectors revealed that the “low chlorophyll” historical view of the SO embedded more complex spatial and temporal patterns. From the examination of seasonal cycles of surface chlorophyll, a patchy regionalization emerged which could not be firmly linked to environmental properties of the usual provinces of the SO (Thomalla et al., 2011). When the mixed-layer depth was taken into account, three different types of SO blooms were defined which coincide with distinct dynamical regions (Sallée et al., 2015). More recently, Ardyna et al., (2017) combined in-situ measurements, satellite observations and model outputs to suggest that environmental controls delineate seven different bioregions with respect to phytoplankton biomass and phenology. All these studies

have made an extensive use of chlorophyll-a concentration (Chla), both at the surface or in the water column, as a proxy of phytoplankton biomass.

The phenology of phytoplankton blooms has received considerable attention because it depends on forcing parameters that might be altered by climate change (e.g. heat flux, wind stress, freshening), and also because it is strongly linked to essential marine ecosystem services (e.g. resources or carbon sink) (Jones et al., 2016). Various conceptual frameworks have been proposed to explain the onset of the spring blooms (Sverdrup, 1953 ; Behrenfeld, 2010 ; Chiswell, 2011 ; Taylor & Ferrari, 2011 ; Brody and Lozier, 2015; Mignot et al., 2018). However other facets of phytoplankton bloom, such as the dynamics of biological CO₂ uptake, have received much less attention.

This is a very challenging task if high resolution or long-term measurements are targeted. In the SO, different strategies have been used. Underway measurements of partial pressure of CO₂ (pCO₂) in the surface waters (Bakker et al., 2015) largely contributed to the assessment of Southern Ocean CO₂ sink and its variability (Landschützer et al., 2015). Lagrangian surface drifters (CARIOCA buoy) equipped with pCO₂, fluorescence or O₂ sensors revealed the high spatial and temporal variability of the carbon fluxes in response to physical and biological changes (Boutin et al. 2008, Resplandy et al. 2014). The net community production (NCP) that can be used cautiously as a proxy of the carbon export, was also derived from the CARIOCA measurements (Merlivat et al., 2015). Provided appropriate alkalinity (Alk) parametrization, Biogeochemical-Argo profiling floats (hereafter BGC-Argo) equipped with pH and O₂ sensors have recently been proposed to estimate pCO₂ in various provinces on the SO over full seasonal cycles (Gray et al., 2018). A large spatial coverage of NCP in the SO was reported using underway (O₂/Ar) measurements (Cassar et al., 2007). Eulerian approaches to the estimate of pCO₂ in the SO have also been pioneered by instrumented moorings deployed in the Sub Antarctic zone of Tasmania from which temporal description and quantification of different carbon fluxes were successfully derived (Weeding & Trull, 2014 ; Shadwick et al., 2015).

In our study conducted in the naturally iron fertilized region of Kerguelen (Blain et al., 2007), south of the polar front, we take advantage of the in-situ high resolution multisensor dataset gathered by one mooring and two BGC-Argo floats to look carefully at biological signatures during the early stage of the bloom from a multiple parameter perspective. In addition to changes in Chla, temporal variations in dissolved oxygen (O₂) and dissolved inorganic carbon (DIC) provide new insights on the short-term variability of carbon stocks and fluxes. We assess bloom initiation dynamics in relation to heat flux and wind stress changes.

2 Data Sources and Processing

During the SOCLIM (Southern Ocean and Climate - Field Studies with Innovative Tools) cruise (DOI:10.17600/16003300) an instrumented mooring and two profiling BGC-Argo floats were deployed in spring 2016 on the central Kerguelen plateau. The atmospheric context at the time of measurements is documented using atmospheric reanalysis. The climatological and synoptic context of our area of interest is described using a combination of satellite and

in-situ observations. Full details of all materials and methods are provided in the Acknowledgment and Supporting Information. Here we provide a brief outline.

2.1 SOCLIM Mooring

One anchored mooring was deployed from October 18, 2016 to April 06, 2017 in the central part of the Kerguelen Plateau, at 50°37'135 S and 072°06'179 E (Figure. 1), where the waters are naturally enriched in iron leading to enhanced phytoplankton bloom development (Blain et al., 2007, 2008; Jouandet et al., 2008). This mooring was equipped with a package of sensors located at 42 m depth. A Sea-Bird SBE-16 sensor measured hourly conductivity and temperature. An Andraea optode, with post cruise calibration by the manufacturer, provided hourly O₂ measurements. A Carioca sensor, calibrated using DIC and Alk profiles at the mooring site in October 2016 and January 2017 (OISO cruise), according to the protocols described in Merlivat et al. (2017), provided hourly measurements of pCO₂.

The mooring line between 42-300 m was equipped with 10 Sea-Bird SBE-37 (conductivity, temperature, and pressure) and 22 Sea-Bird SBE-56 sensors (temperature) were also deployed between the SBE-37 loggers. Altogether these sensors provided profiles between 42-300 m with a temporal resolution of 30 minutes and a vertical resolution of 30 m and 5 m for salinity and temperature, respectively.

2.2 BGC-Argo Floats

Two BGC-Argo profiling (WMO 6902737 and WMO 6902736) were also deployed on October 18, 2016, near the mooring location (Figure. 1). Both floats sampled the water column between 300 m up to the surface with a vertical resolution of 1 m, once a day during 18 days (October 18, 2016 to November 05, 2016). Afterwards the temporal resolution was relaxed to one profile every two days during 114 days (November 05, 2016 to February 27, 2017) and one profile every four days from February 27, 2017 to the end of batteries (April 08, 2018 for WMO 6902736 and June 12, 2018 for WMO 6902737). For the purpose of this study, we only display data over a 2-months' time period between October 18, 2016 and December 20, 2016.

Both BGC-Argo floats were equipped with a SBE41CP Seabird CTD, a WET Labs ECO sensor including a Chl_a fluorometer (excitation at 470 nm and emission at 695 nm), an OC4 radiometric sensor that measures the photosynthetically available radiation (PAR). The measurements were quality-controlled according to internationally agreed procedures (Schmechtig & Thierry, 2016; Roesler et al., 2017). The instantaneous profile of PAR_i was first converted into a daily average profile as described in Mignot et al. (2014). Subsequently, the daily average profile was further averaged over the mixed and mixing-layer (resp. PAR_{MLD} and PAR_{MixLD}). Additionally, the euphotic zone depth, Zeu (m), was computed as the depth where PAR reaches 1 % of its surface-value.

2.3 Atmospheric parameters

Net air-sea heat flux, 10 m wind speed, and wind stress are estimated at the mooring location for the period 2016-2017 from 4 different products (Figure. 2c and d): the Japanese

55-yr reanalysis, NCEP1 reanalysis, ERA5 reanalysis and a product derived from satellite observations OAFflux + CERES.

The air-sea CO₂ flux is derived using the 10 m wind speed at the mooring location from the Japanese 55-yr reanalysis for the period 2016-2017 and atmospheric pCO₂ recorded at Crozet Island.

2.4 Estimation of the Mixed-layer and Mixing-layer Depths

The properties and vertical extent of the mixed-layer are central metrics for understanding phytoplankton dynamics (Sverdrup, 1953). The mixed-layer depth (MLD) refers to the depth of vertically homogeneous profile of temperature, salinity and density. MLD are computed from both the BGC-Argo floats and the mooring profiles. For the mooring, the temperature profiles have a higher vertical resolution (5 m) than the density profile (30 m). Consequently, MLD are estimated from a temperature-based criterion (temperature difference threshold from the surface temperature of $\Delta T = 0.2$ °C) (de Boyer Montégut et al., 2004; Sallée et al., 2006; Holte & Talley, 2009; Pellichero et al., 2017). One important limitation is we do not have access to the surface temperature from the mooring dataset. We therefore assume the surface temperature to be equal to the shallowest available temperature at 42 m, and apply the temperature threshold from there. The MLD from the profiling floats computed both with and without assuming well-mixed temperature within the upper 42 m are very close, however, a few particular events of shallow mixed-layer are missed (Supporting Information S1). In addition, using a temperature threshold rather than a density threshold gives similar results on MLD derived from profiling floats data, which provides us confidence in our MLD detection methods (Supporting Information S1).

While the MLD is an important parameter, contemporary studies have also highlighted the significance of the “mixing-layer” to study the phenology of the phytoplankton bloom (Taylor and Ferrari, 2011; Brody and Lozier, 2015). The mixing-layer can be defined as the depth in which turbulence is fully and actively driven by surface forcing (Stevens et al., 2011). We estimated the mixing-layer depth as the Ozmidov length from the mooring time series and JRA-55 wind stress. Wind stress from different reanalysis are also introduced in Supporting Information S2 as well as all details of calculation of the mixing-layer depth.

2.5 Carbon Fluxes and Net Community Production

The period of interest highlighted in this study, namely the “onset period”, which corresponds to a rapid modification of the biogeochemical parameters, is defined from the multisensors analysis of the DIC, oxygen, and Chla that attests the start of a strong biological activity between October 27 to November 05, 2016. This bloom is indicated by the red shading on all the figures.

DIC concentration was calculated using CARIOCA pCO₂, temperature and alkalinity derived from salinity with a regional relationship as in Merlivat et al., (2015). Based on measurements of Chla and density profiles in the upper layer of the North Atlantic Ocean, Lacour et al. (2019) have shown that at the beginning of the spring the productive layer is shallower than the mixed-layer. This result is consistent with our BGC-Argo floats observations (Figure. 3d and Supporting Information S3) showing higher values of the Chla above the base of the mixing-layer which is shallower than the mixed-layer. Hence, in the following, we consider that the NCP occurs in the mixing-layer only, and that phytoplankton

cells are homogeneously distributed by the mixing in that layer so that the biological carbon uptake is also vertically homogeneous. The NCP, over a few days' time interval, is derived assuming that (1) DIC at 42 m depth is within the mixing-layer, (2) horizontal advection and (3) vertical mixing are both negligible. These hypotheses are supported during the onset period between October 27 and November 05, 2016, by several evidences. First, the 42 m sensors are within the mixing-layer estimates (Figure. 3a), except for a short period around November 03, 2016 (indicated as dotted line on Figure. 3a). However, taking this short period into account or not does not affect our NCP estimates. Second, the Chla recorded by the mooring and by the Lagrangian floats qualitatively vary similarly (Supporting Information S4) and finally the mooring temperature regularly increases as expected in a near surface layer isolated from the subsurface layers (Figure. 3c). We neglect the eddy diffusion term because the DIC gradient at the basis of the mixing-layer is expected to be much smaller than the one at the basis of the mixed-layer. Therefore, NCP is estimated from the slope in time of DIC integrated over the mixing-layer depth (DIC_{int}) corrected from the air-sea flux contribution:

$$NCP = \left(\frac{\Delta DIC_{int}}{\Delta t} \right)_{bio} = \left(\frac{\Delta DIC_{int}}{\Delta t} \right)_{meas} - k \times s \times (pCO_{2atm} - pCO_{2sw}) \quad (1)$$

The first term at the right-hand side of the Equation. 1, "meas", corresponds to the measured temporal change of DIC_{int} derived from pCO_2 observations, the second term corresponds to DIC change due to air-sea CO_2 flux. (further details in Supporting Information S5 and in other studies (Mehrbach et al., 1973; Dickson & Millero, 1987; Park et al., 2008; Wanninkhof, 2014)).

3 Results and Discussion

3.1 Temporal variability of Physical and Biological Parameters at the Mooring Site

The climatological seasonal cycle of the MLD around the mooring site (Figure. 2a) varies from 180 m in winter (J-A-S) to 80 m in summer (J-F-M). The mooring 2016-2017 MLD is very close to the climatological values except during spring (O-N) where the mooring records faster and stronger restratification as well as deeper MLD than usual in late winter. The MLD rapidly shallowed from about 250 m at the end of October 2016, to ~70-80 m, 10 days later. During the same period the BGC-Argo WMO 6902737 float recorded a strong increase of Chla (0.5 to 2.5 $mg\ m^{-3}$; Figure. 2b). Similar temporal changes of Chla were also observed with the mooring and with the second BGC-Argo float (Supporting Information S4). A second bloom was also detected later in the season, also consistent with the climatology (see Supporting Information S6 for more details on the Chla climatology). In this paper we only focus on the first bloom as during the second bloom the BGC-floats were already 100 km away from the mooring and this later bloom occurred at a date when additional vertical DIC fluxes (i.e. entrainment) took place at the base of the MLD making Equation. 1 less reliable.

Heat fluxes and wind speed are two major drivers of the turbulence in the MLD (Kraus & Turner, 1967; Dong et al., 2008; Sallée et al., 2010; Belcher et al., 2012). The 2016-2017 seasonal variability of the net heat fluxes is overall consistent across 4 different reanalysis products (Figure. 2c), despite a few significant differences between them (e.g period where the heat fluxes switch from negative to positive). The heat flux minimized in winter with a mean value of $-50\ W\ m^{-2}$ (JRA-55), and maximized at the end of spring / early summer with a mean value around $160\ W\ m^{-2}$. When heat flux becomes null or slightly

positive (in 2016, early August for NCEP and ERA5, mid-August for OAFlux and early September for JRA-55) it acts as a stabilizer transferring heat from the atmosphere into the ocean and then stratifying the water column.

Similarly, the wind stress estimated from the same 4 reanalysis products (Figure. 2d) is consistent for the 2016-2017 season with a climatological cycle that maximizes in winter with values ranging from 0.28 N m^{-2} (OAFlux) to 0.4 N m^{-2} (JRA-55) and gradually decreases over the season until reaching its minimum value at the end of spring / early summer.

3.2 Atmospheric drivers of rapid change in surface layer

While the 2016-2017 environmental seasonal cycle at the mooring site, as described by MLD and Chla appears consistent with the climatological mean seasonal cycle, the 2016-2017 is marked by abrupt changes at the end of October, characterized by rapid shallowing of the MLD, and marked increase of Chla between October 27 and November 05, 2016 (Figure. 2a, b). Zooming on the MLD time-series, it appears however that the rapid increase in Chla leads the marked shallowing of the MLD by a few days (Figure. 3a). Some studies have argued that Chla increase would precede shallowing of the MLD, because active mixing in the MLD would be reduced days before the stratification increase reducing the MLD (e.g. Taylor and Ferrari, 2011; Brody and Lozier, 2014). Many studies have used air-sea heat flux as a proxy to define when the active mixing in the MLD is stopped at the end of winter (e.g. Taylor and Ferrari, 2011), assuming active mixing stops when air-sea heat flux switches from negative to positive values (which warm the ocean surface, thus stratifying the water column).

In the present study, the 4 heat flux products are already positive several weeks before MLD and Chla rapid increase in late October (Figure. 2c). Therefore, while positive heat fluxes may be a necessary condition of reduced mixing activating Chla increase, this is not sufficient to initiate biomass accumulation. Interestingly, winds are at their annual maximum strengths in August-September, gradually reducing until late October, suggesting that even if the heat fluxes are stratifying, the wind could be strong enough to maintain enough turbulence into the surface and keep the mixed-layer at depth, hence preventing bloom onset. This is in line with Brody and Lozier (2015) analyses suggesting that decreases in the depth of active mixing are a result of the transition from buoyancy-driven to wind-driven mixing and control the timing of the spring bloom. Here we focus on reduced turbulence rather than change of advection (lateral or vertical) that could potentially impact the stratification, because our observations show that the mixed-layer is actually destratified (i.e. deep MLD), when the Chla increase starts. If that Chla increase would be supported by change of large-scale advection changing the stratification (e.g. change in wind-stress curl), we would observe a fingerprint in the density field, i.e. observe a change of MLD. Similarly, at short temporal scales and small spatial scales as studied here, submesoscale dynamics within the mixed-layer have been reported to play a role in restratifying the mixed-layer (Lévy et al. 2018; Siegelman et al., 2020). However, if submesoscale processes would be the main cause of the stratification event, it would have a fingerprint on the density field, hence, would be associated with a shallowing of the MLD which is not observed on October 27. Submesoscale might be active and act to restratify after that period when turbulence relaxes (even though some analyzes suggest that it might be a weak effect in our region; Rosso et al., 2016). Instead, we

hypothesize here that the density field remains unchanged, but the local turbulence reduces, allowing a Chla increase (e.g. Taylor and Ferrari, 2011; Brody and Lozier, 2015).

This hypothesis is confirmed through the analysis of temporal variation of the Ozmidov length (Figure. 3a, Supporting Information S2). This scale gives a measure of the typical depth over which mixing occur under stable stratification and at a given turbulence rate, here estimated to come from wind stress (Denman & Gargett, 1983; Riley & Lelong, 2000; Brody & Lozier, 2015). Using this scale to quantify the balance between both wind stress effect and buoyancy forcing effect over the ocean surface, it appears that despite restratifying forces from air-sea heat fluxes, the wind actively mixes the surface layer up to 130 m before October 27, and rapidly switches to mix a much shallower layer of ~50 m, after October 27 (Figure. 3a). Our results therefore suggest that a shoaling of the mixing-layer (i.e. decrease of the mixing) precedes a shoaling of the MLD, and corresponds to the onset of the bloom. These observations suggest that variations in the wind stress first modulate the variations of the mixing-layer and second of the MLD, consistent with wind-induced mechanical mixing (Dong et al., 2008; Mahadevan et al., 2010, 2012; Gille et al., 2014; Carranza & Gille, 2015; Carranza et al., 2018). We now investigate how this change of surface mixing translate into change of biological parameters and carbon uptake, beyond the simple fingerprint on Chla concentration.

3.3 Multi-parameters detection of the rapid bloom onset

A large decrease of DIC (Figure. 3b), a rapid increase in the O₂ oversaturation as well as an increase of the temperature (Figure. 3c) is observed at the mooring site at 42 m depth from Oct 27 to Nov 05. These gathered parameters unambiguously attest the start of a strong biological activity on October 27, (i.e. the onset date of the phytoplanktonic bloom). The estimate of NCP integrated over the mixing-layer is equal to $119 \pm 7 \text{ mmol m}^{-2} \text{ d}^{-1}$ and compares well with the average daily NCP of $99 \pm 37 \text{ mmol m}^{-2} \text{ d}^{-1}$ computed by Jouandet et al. (2008) at the same site using a seasonal DIC budget approach. It is not surprising that the present estimate is above this seasonal mean value because, at this site, the daily NCP is higher at the beginning of the season than later when high heterotrophic bacterial activity takes place (Obernosterer et al., 2008).

During this period of intense biological activity, the phytoplankton bloom develops as indicated by a large and rapid increase of the Chla (Figure. 3d). Over both the entire mixing-layer and MLD the averaged Chla also increases (Supporting Information S7) but with a higher magnitude in the mixing-layer. NCP derived by assuming that DIC is vertically homogeneous over the mixing-layer ($119 \pm 7 \text{ mmol m}^{-2} \text{ d}^{-1}$), greatly differs from the estimate obtained by assuming that DIC is vertically homogeneous over the whole mixed-layer ($212 \pm 7 \text{ mmol m}^{-2} \text{ d}^{-1}$, see Supporting Information S5). This result is in line with an overestimation of the Chla stock calculated by multiplying the surface Chla by the depth of the MLD rather than by the depth of the mixing-layer in which the biology actively develops (Lacour et al. 2019).

The changes in the various parameters (DIC, O₂, Chla) pinpoint to a rise of the biological activity concomitant with the rapid shallowing of the mixing-layer (Figure. 3) suggesting a direct link between both processes. The shallowing of the mixing-layer is associated with an increase of the daily average PAR within both the mixed and mixing-layers (Figure. 3d) from less than $1.3 \text{ mol quanta.m}^{-2}\text{d}^{-1}$ on the 26 of October to peak at 5.2 mol

quanta $\text{m}^{-2}\text{d}^{-1}$ two days later. These values are in close agreement to regional average values reported for the productive zone eastward of the Kerguelen plateau (Blain et al., 2013). Concomitantly the euphotic zone depth, Z_{eu} , decreases from about 60 m just before the bloom on October 27, to a minimum of ~ 30 m on November 02 (not shown). These observations suggest that the seasonal rise of the surface PAR combined with the rapid shallowing of the mixing-layer, increased $\text{PAR}_{\text{MIXLD}}$ above the winter value and triggered the abrupt increase of the biological activity. We also note that $\text{PAR}_{\text{MIXLD}}$ decreased when the mixing-layer stopped shallowing and then increased again when the mixing-layer shallowed (around November 10, 2016) and part of the surface biomass was exported below the mixing-layer around mid-November. This temporal pattern is consistent with previous observations (Blain et al., 2013) suggesting that self-shading imposed an upper limit to the light available during the development of the bloom and had a strong influence on the maximum amount of biomass that can accumulate in the mixing-layer.

4 Conclusion

A multiplatform investigation acquiring multidisciplinary data at the appropriate temporal resolution is used to refine a mechanistic understanding of the environmental setup required for the phytoplankton spring bloom onset on the Kerguelen plateau. Our primary conclusion is that, while the reversal of heat fluxes from destratifying to stratifying force may be a necessary condition for spring bloom onset, it is not a sufficient one when wind is properly strong to actively mix the surface layer deep enough. The bloom described in the present study starts nearly six weeks after the heat flux reversal. The present study emphasizes the critical role of wind stress which can continue to maintain phytoplankton biomass at its winter level. It is only when wind stress diminishes to a given threshold that turbulence becomes sufficiently alleviated and the mixing-layer reduced for the bloom to start, thanks to a sufficient light availability within the reduced mixing-layer. While such a mechanism was already suggested at play for bloom initiation in the North Atlantic by Brody and Lozier (2015), it is the first time that it is documented in the SO, together with the associated DIC decrease. In such a configuration, our data highlight the abrupt DIC decrease as a mirror of biomass increase. We further suggest that such DIC changes might represent alternate metrics for bloom onset characterization in terms of timing and magnitude.

More generally, our results confirm wind regime is an important driver of the phenology of the bloom of the central Kerguelen Plateau as previously reported in the SO (Gille et al., 2014; Carranza & Gille, 2015; Carranza et al., 2018). These observations have significant implications for the understanding of the variability of blooms. At the interannual time scale, variability in wind stress could likely be responsible for the observed variability in the bloom timing. Furthermore, the wind regimes are predicted to change in the future as a result of climate change, with potential complex counter-balanced effect of increased stability of the water-column due to surface intensified warming, and increased wind stress (Bracegirdle et al., 2013). The Kerguelen bloom sustains a unique ecosystem which includes important marine resources.

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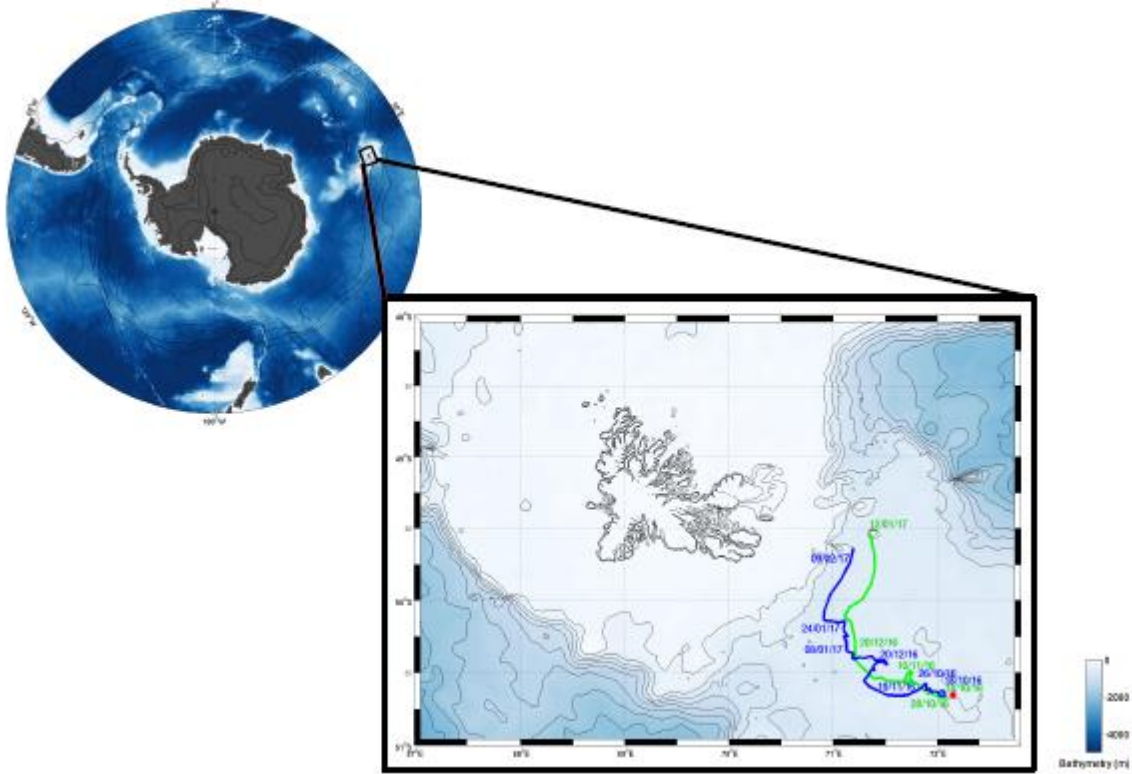


Figure 1. Map of the Southern Ocean, showing major topography and a zoom on the studied area near Kerguelen. The mooring is located near Kerguelen (red point); bathymetry contours spaced every 500 m are in black. The two BGC-Argo floats trajectories within 100 km from the mooring are in blue (WMO 6902736) and green (WMO 6902737) from October 18, 2016 to February 09, 2017 and January 01, 2017 respectively.

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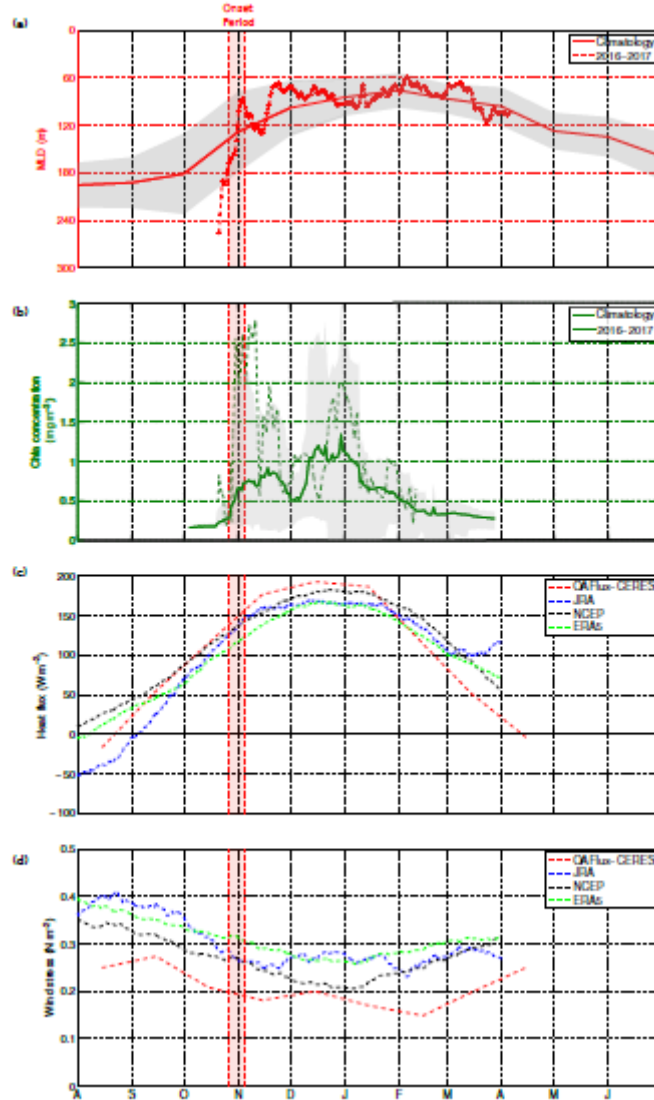


Figure 2. 2016-2017 observations (a,b) and reanalysis (c,d) at the mooring location and comparison to climatology (a,b). (a) MLD (m) from the mooring (dashed line) and climatology (plain line) at the mooring location as derived by Pellichero et al. (2017); standard deviation in grey; (b) Chla averaged over 10 m depth as recorded by WMO 6902737 BGC-float (dashed line) and satellite climatology at the mooring location (plain line; standard deviation in grey); (c) net heat flux for the period 2016-2017 (W m^{-2}) at the mooring site from JRA-55 reanalysis (blue), ERA5 (green), NCEP1 (black) and a product derived from satellite observations OAFux+CERES (red). (d) Wind stress for the period 2016-2017 (N m^{-2}) at the mooring site from JRA-55 reanalysis (blue), ERA5 (green), NCEP (black) and OAFux (red). On each panel, the red shading highlights the period of interest in this study, corresponding to the onset of the bloom period as defined in section 3.3.

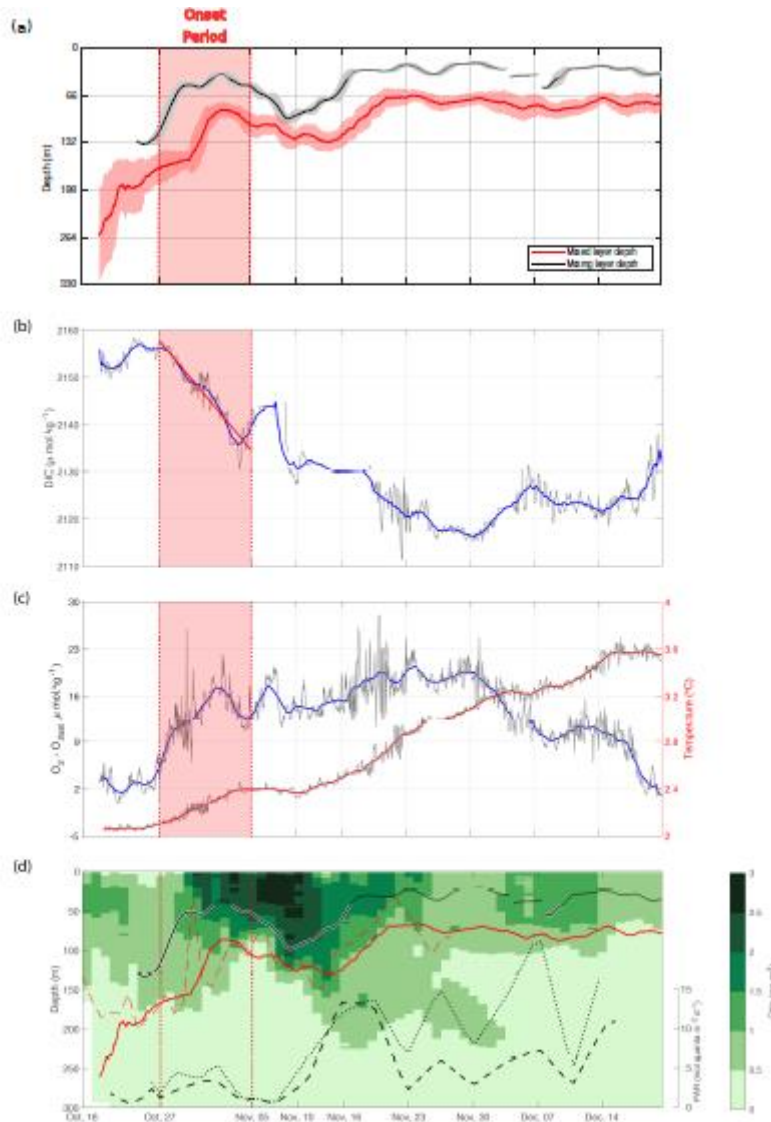


Figure 3. Multiparameter detection of biological activity rise. (a) MLD and mixing-layer (resp. red and black) computed at the mooring site, with 24-hour running standard deviation in shown as shading around the curves. The dotted parts of the black line correspond to period where the mixing-layer is above 42 m. (b) DIC times series (blue) and linear regression from October 27 to November 05, 2016 (red line). (c) Time series of the oxygen oversaturation (blue) and temperature records by the mooring at 42 m (red). (b, c) Data at 1-hour resolution (grey) and averaged with a 2-day running mean (color); (d) Vertical section of Chla (WMO 6902737 BGC-Argo float); MLD and mixing-layer computed at the mooring site are shown as plain red and black lines respectively, and MLD from the float time-series is shown as red dashed line for comparison; PAR_{MLD} and $\text{PAR}_{\text{MIXLD}}$ are shown as black dashed and dotted lines, respectively. On all panels, the onset period of the bloom (October 27, 2016 to November 05, 2016) is shown as the two vertical red dashed lines as well as red shading for panel a-c.