Multi-year robotic observations reveal the seasonality of downward carbon export pathways in the Southern Ocean

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Multi-year robotic observations reveal the seasonality of downward carbon export pathways in the Southern Ocean

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Introductory paragraph

At high latitudes, the export of organic matter from the surface to the ocean interior, the biological carbon pump, has conventionally been attributed to the gravitational sinking of particulate organic carbon (POC). Conspicuous deficits in ocean carbon budgets have recently challenged this long-lived paradigm of a sole pathway. Multiple strands of evidence have demonstrated the importance of additional export pathways, including the particle injection pumps (PIPs). Recent model estimates revealed that PIPs have a comparable downward POC flux to the biological gravitational pump (BGP), but with potentially different seasonal signatures. To date, logistical constraints have prevented concomitant and extensive observations of these pumps, and little is known about the seasonality of their fluxes. Here, using year-round robotic observations and recent advances in optical signal analysis, we concurrently investigated the functioning of two PIPs - the mixed layer and eddy subduction pumps - and the BGP in Southern Ocean waters. By comparing three phytoplankton bloom cycles in contrasting environments, we show how physical forcing and phytoplankton
phenology influence the magnitude and seasonality of these pumps, with implications for carbon sequestration efficiency.

The biological carbon pump (BCP) is considered a major contributor to the Southern Ocean (SO) carbon sink. It removes 3 PgC from the euphotic zone (Ez, upper ~100 m) annually, representing 33% of the global BCP\(^1\). Since the concept of the BCP was originally proposed\(^2\), it has been widely assumed that it was mediated primarily by the sinking of POC, now called the Biological Gravitational Pump (BGP)\(^3\). However, current estimates of biological carbon demand in the mesopelagic (~100-1000 m depth) exceed carbon inputs attributed to sinking material from the euphotic zone by two- to three-fold\(^4\). Such discrepancies highlight the need to reassess the pathways that contribute to downward carbon export.

Recently, other particle injection processes have been invoked as additional pathways that help to balance the carbon budget in the mesopelagic\(^5\). Studies\(^5-7\) have provided evidence that downward export of organic matter also occurs through localized (1-10 km) eddy-driven subduction of POC. This process, known as the Eddy Subduction Pump (ESP), leads to episodic injection of POC-rich waters below the mixed layer. Similarly, the cycle of deep vertical mixing and re-stratification events during the winter to spring transition acts as a physical pump that injects fresh organic matter into the mesopelagic, the so-called mixed layer pump (MLP)\(^8,9\). At basin scale, the ESP and MLP are equivalent to about one quarter of the BGP export\(^3\). However, at the (sub)mesoscale (<100 km) these processes can be equal to, or even greater than the BGP\(^10,11\). This spatial mismatch illustrates the difficulty of carrying out comprehensive intercomparison of these pumps in the field. The mesopelagic migrant pump\(^12\), seasonal lipid pump\(^13\) and large-scale subduction\(^14\) pump are also important contributors to the BCP, although assessing their relative contribution is even more challenging\(^3\) and beyond this study.

Boyd et al. (2019)\(^3\) suggested that the relative importance of both the PIPs and BGP varies seasonally, resulting in substantially different fates for exported carbon with respect to depth of remineralisation, sequestration time scale, and consumption by midwater biota\(^10\). However, concomitant and extensive observations of these pumps are extremely rare, and little is known about the seasonality of their fluxes. Better characterisation of these pumps and their seasonality
will fundamentally push forward our understanding of ocean biological carbon export and sequestration, and help to close regional ocean carbon budgets.

Biogeochemical-Argo (BGC-Argo) floats with multi-year missions and high frequency sampling offer a promising way to jointly investigate the PIPs and the BGP over a broad range of time and space scales. Such platforms have already been successfully used to characterise the MLP at sub-seasonal to seasonal scales\(^8,9\), the ESP at pan-Antarctic scale\(^6\), and the seasonality of the BGP\(^15,16\). In this work we refine and bring together a range of previously developed techniques. Concurrent measurements of chlorophyll \(a\) fluorescence and particulate backscattering (proxies of phytoplankton biomass and POC, respectively), oxygen, temperature and salinity have been used to identify subsurface anomalous features related to the MLP and ESP\(^5,17\). Spikes in particulate backscattering and chlorophyll \(a\) fluorescence have provided insights into the sinking of aggregates through the mesopelagic\(^18\) while short term (order of days) particle accumulation on transmissometer sensors, working as optical sediment traps (OST)\(^19\), allowed characterising the BGP. This novel combination of approaches links a bespoke sensor constellation with recent advances in optical signal analysis to compare the BGP with two PIPs - the MLP and the ESP - through three annual cycles. We focus on a single float, which sampled contrasting biogeochemical provinces across the Pacific sector of the SO, to infer the causal mechanisms that set the magnitude and seasonality of each of these pumps. These mechanisms include bloom timing or seasonality, hereafter called phenology, which influences the concentration, size and composition of the particle assemblage in the upper ocean.

Characteristics of the particle assemblage in contrasting environments

We present here \textit{in situ} observations collected by a BGC-Argo float (WMO 7900791) that travelled >6000 km across the Pacific sector of the SO (Fig. 1a). This float was deployed in May 2016 in sub-polar waters and sampled the water column from 500 m to the surface every 1 to 5 days during three annual cycles. It crossed the Polar Front in September 2016 as evidenced by the abrupt decrease in surface temperature (from 6\(^\circ\)C to 3\(^\circ\)C, Fig. 1c). The float remained in the vicinity of the Polar Front for almost an entire year, until July 2017, and then entered polar waters where near-zero surface temperatures were observed. It was in the vicinity of the sea-ice edge from June to August 2018 (Fig. 1b) in an area characterized by very low salinity surface waters (Fig. 1d) which prevented any deep mixing events during winter. Indeed,
the mixing layer depth (MLD_{bio}, see Methods) remained shallow (<100 m) as opposed to the two previous winters where mixing reached 250-350 m (the black thick line in Fig 1c,d). The float ended its mission due to battery failure in Drake Passage in May 2019. This long trajectory across different oceanic provinces enabled us to characterize the pumps over a broad range of time and space scales in contrasting environments. The downside is the difficulty to study the seasonality of these pumps when the float moves across different water masses. We therefore segmented the trajectory into three bloom cycles during which the contiguous nature of water masses was verified (Fig. S2 and Methods).

These contrasting environments corresponded to clear contrasts in bio-optical proxies for concentration, size and composition of phytoplankton community and associated particles between the three bloom cycles recorded (Fig. 2). The 2016-17 bloom, near the Polar Front, was the most intense as reflected in POC (up to 175 mg C m\(^{-3}\), Fig. 2a) and chlorophyll \(a\) fluorescence (\(F_s\) around 1 mg m\(^{-3}\) chlorophyll, Fig. S7). This bloom was characterised by a very low fluorescence to backscattering ratio, \(F_s / b_{bs}\) (where \(s\) denotes the small fraction of particles, see Methods), a ratio which varies according to changes in phytoplankton community structure, photoacclimation and nutrient status\(^{20-22}\). Here, we attribute low \(F_s / b_{bs}\) values to the growth of coccolithophores, calcifying phytoplankton which form liths – bio-mineral shells – with high refractive index and thus high backscattering signal\(^{23}\). Coccolithophore blooms are recurrent features in the SO, readily detectable from space\(^{24,25}\). Indeed, satellite records confirmed the presence of an intense coccolithophore bloom during this summer in the area where the float was profiling (Fig. S4). Bio-minerals act as ballast by increasing particle specific gravity and sinking speeds, and provide protection from remineralisation\(^{26}\), potentially explaining the massive invasion of POC to 450 m following the coccolithophore bloom (Fig. 2a).

The 2017-18 bloom was less intense with lower phytoplankton biomass. It was characterised by a high \(F_s / b_{bs}\) at its peak (apex), when the mixing layer abruptly shoaled from 300 to <100 m (Fig. 2a, b). Such high \(F_s / b_{bs}\) ratios associated with high-latitude spring blooms, when light is not limiting (Fig. S5), have been attributed to diatom-dominated events\(^{20,21}\), which in this case would have benefited from iron inputs by winter deep mixing and melt waters (Fig. 1c, d).

However, the float-derived iron limitation index\(^{27}\) (Fig. S5 and Methods) indicates that in late December 2017 iron was rapidly depleted by biological consumption, resulting in low phytoplankton biomass relative to the 2016-17 and 2018-19 blooms. Accordingly, the size of particles/aggregates was two-fold lower than for the 2016-17 bloom (Fig. 2c and Methods), but still two-fold higher than the 2018-19 bloom where higher biomass levels were observed.
Diatoms dominating the 2017-18 bloom may explain the relatively high mean particle size. The transition of the bloom to a subsurface chlorophyll maximum (SCM) feature was another indication of nutrient limitation. High $F_l / b_{bl}$ at the SCM (around ~125 m depth, Fig. 2b) was due to photoacclimation as light became limiting in the subsurface. In April 2018, the photoacclimation signal propagated in the entire mixing layer as light levels continued decreasing (Fig. S5) and convective mixing commenced again (Fig. S1b, c).

The 2018-19 bloom was partially sampled as data were missing for nearly two months early in the productive season. The SCM observed from December to February 2018 (Fig. 2b), relatively small particles in the surface layer (Fig. 2c), and a high iron limitation index (Fig. S5) suggest that the bloom was at a late stage when sampled, characteristic of open-water conditions in the seasonal sea-ice zone\(^\text{28}\). The second intense peak of biomass recorded in April 2019 developed in the Drake Passage, near the Antarctic Peninsula, a region known to be influenced by iron-enriched shelf waters\(^\text{29}\). Accordingly, the iron limitation index dropped to low values similar to those observed during the first 2016-17 bloom (Fig. S5).

**Seasonality of multiple carbon export pathways**

The sensor constellation on the BGC-Argo float was used to independently assess the seasonality of the BGP and to concurrently explore the timing of MLP and ESP events. The BGP was characterised by two independent methods, namely the spike index and the OST (see Methods). The spike index quantifies the abundance of large particles in the water column, from both fluorescence and backscattering ($F_l$ and $b_{bl}$ respectively, where $l$ denotes large fraction of particles, Fig. 3b, S10 and Methods). $F_l$ was attributed to live phytoplankton aggregates sinking through the water column and represents a subset of $b_{bl}$ which also includes fecal and detrital matter\(^\text{18}\). These two components were decoupled in time during the 2016-17 bloom. Prior to the maximum in POC (early January, Fig 3a), the dominance of the $F_l$ spike index was attributed to a diatom bloom that generally precede coccolithophore blooms\(^\text{30}\). Indeed, diatom export events have often been described as an initial pulse of fresh aggregates\(^\text{18}\) followed by a second pulse of resting spores, empty frustules and fecal pellets\(^\text{31}\), the latter explaining the significant peak in the $b_{bl}$ spike index observed in mid-December, before the maximum in POC attributed to the coccolithophore bloom. Low $F_l$ and maximum $b_{bl}$ spike indices in mid-January 2017 following the coccolithophore bloom resulted from sinking aggregates partly made of detached coccoliths\(^\text{32}\). A decoupling in $F_l$ and $b_{bl}$ spike indices was also evident during the 2017-18...
diatom bloom. However, the spike index was low relative to the 2016-17 bloom. It is likely that lower surface biomass during the 2017-18 bloom (Fig. 2a) led to lower levels of aggregation\textsuperscript{33}, as evidenced by the lower peak mean particle size (Fig. 2c). Conversely, \( F_l \) and \( b_{bl} \) spike indices were coupled during the 2018-19 bloom (second peak of biomass in April 2019) with the \( b_{bl} \) spike index as the main contributor (Fig. 3b). In all three blooms, \( F_l \) was rapidly attenuated with depth while \( b_{bl} \) remained high as deep as 450 m (Fig. S6) as a result of higher fragmentation and/or consumption rates of fresh and fragile phytoplankton aggregates than of other large sinking particles\textsuperscript{18}.

The OST method provides the opportunity of measuring an \textit{in situ} BGP flux from autonomous floats (see Methods). These measurements are however limited to a single depth, the float ‘parking’ depth, here \( \sim 300 \) m, and are not calibrated in terms of carbon biomass, so they are expressed in \( \text{d}^{-1} \)\textsuperscript{19,34}. This flux was divided into two components, a continuous flux of small slow-sinking particles (slowly accumulating on the transmissometer window) and a pulsed flux of large fast-sinking particles (creating discontinuities in transmissometer records)\textsuperscript{19}. The 2016-17 bloom illustrates well the decoupling between the continuous and pulsed fluxes (Fig. 3c). The latter peaked in mid-January when we observed maximum \( b_{bl} \), while the continuous flux peaked \( \sim 20 \) days later in early February 2017. Assuming both fluxes followed the early January peak in mixing layer POC, the estimated sinking speeds of large (pulsed flux) and small (continuous flux) particles were \( \sim 30 \) m \( \text{d}^{-1} \) and \( \sim 10 \) m \( \text{d}^{-1} \) respectively. These values are in agreement with measured sinking speeds of coccolithophore blooms\textsuperscript{32}. The delay between the continuous and pulsed flux may also reflect disaggregation processes in the mesopelagic which have been reported to account for half of the flux attenuation at high latitudes\textsuperscript{18}. Following the 2017-18 diatom bloom, the low pulsed flux at 300 m was consistent with a high contribution of fresh aggregates (\( F_l \) spike index \( > 80\% \) of the total \( F_l + b_{bl} \) spike index until early December) which were rapidly consumed in the upper mesopelagic, fragmented into smaller particles, and later detected in the continuous flux record (Fig. S12). Note the relatively high continuous flux in late summer and fall when mixing layer POC (Fig. 3a) and particle size (Fig. 2b) were close to the annual minimum. This finding suggests that small low-light low-nutrient adapted phytoplankton contributed \( \sim 50\% \) of the annual continuous flux through slow-sinking of small particles\textsuperscript{35} (Fig. S12).

The ESP contribution to the BCP was quantified as the inventory of POC measured in subsurface anomalous features attributed to localized subduction events\textsuperscript{6} (Fig. S9 and Methods). Such features were detected only during 2016-17 in the vicinity of the Polar Front...
(Fig. 3d), a highly energetic region prone to (sub)mesoscale circulation and ESP events\(^6\). We indeed observed strong spice anomalies (density-compensated changes in temperature and salinity; Fig. S1d, S2 and Methods) over the entire water column revealing different water masses interleaving. The most intense ESP events occurred early in the productive season (September-October 2016), prior to the bloom climax (Fig. S12). Thus, these ESP events actively transported small and freshly-produced organic material as deep as ~400 m (Fig. 2c, 3d and S13). Such labile particles would likely not have reached these depths via the BGP, which illustrates the biogeochemical significance of the ESP. A second peak in ESP events occurred concurrently with the main peak in the F\(_1\) spike index (early December 2016) suggesting that the ESP also transports large fresh aggregates produced in the Ez, in addition to small suspended particles and dissolved compounds\(^9\).

As for the ESP, the MLP was quantified as the inventory of POC measured in MLP remnant layers, delimited by MLD\(_{\text{bio}}\) at the top and MLD\(_{\text{dens}}\) at the bottom (Fig. S11 and Methods). MLD\(_{\text{dens}}\) was defined as the maximum vertical density gradient, and refers to the depth of the seasonal pycnocline\(^8\). All MLP events occurred prior to the bloom apex, each year (Fig. 3d and S12), during the transition between winter deep mixing and spring stratification (~3 months). The strong sub-seasonal variability of the mixing layer depth in 2017 (from 50 to 300 m, Fig. S11), with intermittent stratification and deep mixing events, efficiently transferred organic material out of the mixing layer to depth. Relatively high F\(_s\) content in the remnant layer, as deep as 350 m (Fig. S11), demonstrates the freshness of the exported material, of which a fraction was potentially permanently isolated from the mixing layer\(^8\). The timing of MLP events with respect to bloom timing suggests that large particles produced in the Ez (e.g. in November 2017, Fig. 2c) can also be actively transferred to depth, similar to that observed for the ESP. These results illustrate the possible interplay between a PIP and the BGP, where gravitational sinking could take over downward export following a physical injection event.

Implications for oceanic carbon storage

Multi-year and high-resolution float observations provide a new and more comprehensive view of the various pumps that contribute to the BCP over the annual cycle. The main insight from our study is the seasonal succession of processes contributing to the export of organic carbon to the deep ocean. Export due to the ESP was observed early in the bloom, before the bloom climax, and again at the bloom apex. The MLP was most active during the seasonal transition.
from deep mixing to stratification, and again during sub-seasonal stratification events. Finally, the gravitational pulsed flux was most intense at the bloom apex, while the continuous flux persisted throughout the year, with maximum values delayed in time with respect to the bloom apex – in fall in the case of the 2017-18 bloom. The contribution of all these processes over a complete annual cycle should definitely be accounted for when computing regional mesopelagic carbon budgets.

Another important insight is the identification of the depth strata for the potential interplay between the PIPs and BGP\(^3\). Float-derived optical proxies revealed that PIPs inject large fresh aggregates as deep as 400 m, which then can sink further through gravitational sinking\(^3\). The MLP and ESP have the capacity to transport organic matter at vertical velocities of order of 100-1000 m d\(^{-1}\)\(^{37,38}\), while sinking speeds are order of 10-100 m d\(^{-1}\) for large aggregates\(^18,39\). Thus, physically-mediated processes accelerate the transit through the main remineralisation horizon. The interplay between the PIPs and BGP makes it challenging to assess their relative contribution to carbon sequestration, but the overlap that we find in their timing clearly suggest that their joint contributions can boost the overall efficiency of the BCP.

Large-scale estimates of the PIP contribution to carbon storage require regional high-resolution models able to simulate some of the complex BCP mechanisms we observed\(^11,36\). However, to provide reliable results these models need a larger number of high-resolution, multi-year observations. Process-focused biogeochemical floats with high-frequency sampling are the most cost-effective solution for providing these observations. By combining widespread high-resolution observations with fit-for-purpose models of the full suite of export processes, it should be possible to develop advanced parameterizations that incorporate contemporaneous knowledge of the BCP into Earth System Models with minimal computational cost.

**Methods**

**Float data processing**

The BGC-Argo float used is a Teledyne Webb Research APEX float, equipped with a Sea-Bird SBE41 CTD sensor, WET Labs ECO fluorometer and scattering sensors measuring chlorophyll \(a\) fluorescence and the volume scattering function (at \(\sim 124^\circ\), 470 and 700 nm wavelengths), a Satlantic OC4 radiometer measuring downwelling irradiance at 412, 443, 490 and 550 nm, an Aanderaa Oxygen Optode, and a WET Labs C-Rover transmissometer (660 nm).
The CTD and trajectory data were quality-controlled using the standard Argo protocol. ECO raw signals were converted to chlorophyll $a$ fluorescence (F, in mg Chl m$^{-3}$) and particulate backscattering coefficient ($b_{bp}$ in m$^{-1}$) following the BGC-Argo procedures. Bio-optical data were quality-controlled following the BGC-Argo quality control manual. In addition, F was corrected from the non-photochemical quenching (NPQ) following Xing et al. (2018). Briefly, for each profile where the sun elevation angle was $> 5^\circ$, the maximum F value above the mixed layer depth, defined here as a density difference of 0.01 kg m$^{-3}$. With a reference value at 5 m, was extrapolated toward the surface. As an additional condition, the depth of the extrapolated F value had to be shallower than the 15 μmol photons m$^{-2}$ s$^{-1}$ isolume. As the instantaneous Photosynthetically Available Radiation (iPAR) was not directly measured, we estimated iPAR profiles from the measured downwelling irradiance at 4 wavelengths. At each depth, a spline interpolation was used to compute the irradiance spectra from 400 to 1000 nm (with irradiance value set to zero at 1000 nm). iPAR was then calculated by integrating the interpolated irradiance spectra from 400 to 700 nm. Oxygen data were calibrated using air measurements following Johnson et al. (2015), with a mean gain factor of 1.0557. The $b_{bp}$ at 700 nm was converted into particulate organic carbon (POC) following the relationship in Johnson et al. (2017). Similarly, the $b_{bp}$ at 470 nm was converted into phytoplankton carbon biomass ($C_{phyto}$) following the relationship in Graff et al. (2015). Note that the presence of coccolithophores during the 2016-17 bloom elevates $b_{bp}$ without necessarily elevating POC or $C_{phyto}$, potentially leading to overestimation of these carbon estimates.

The float mission included CTD and bio-optical profiles every 1.5 to 6 days, from 500 m to the surface, with a parking depth of 300 m. The vertical sampling resolution ranged from 3 to 10 m depending on the float ascent speed. All profile data were interpolated at 1 m resolution.

**Quasi-Lagrangian framework**

During its 36-month mission, the float visited different oceanic provinces and crossed strong water mass boundaries, such as in September 2016 when the float crossed the Polar Front. Thus, observed changes in biogeochemical properties are not solely due to temporal changes, making the study of the seasonality of these properties hazardous. We therefore divided the timeseries into three periods in which the contiguous nature of the water masses was verified based on temperature and salinity properties (24-Sep-2016 to 22-May-2017; 10-Oct-217 to 7-Jun-2018; 5-Nov-2018 to 1-May-2019; Fig. S3). The absence of strong water mass contrasts allows us to assume a quasi-Lagrangian framework, where changes in biogeochemical properties can be
interpreted as temporal changes. This approach is commonly used in float studies\(^{48-50}\). For completeness, the figures show the full float timeseries, but only the three quasi-Lagrangian periods are discussed in the text. When calculating rate of change with respect to time (see Methods’ section “Phenology metrics”), the timeseries of the biogeochemical properties of interest were first linearly interpolated on equally spaced 5-day timeseries, and then smoothened with a 30-day (7 points) running average to filter out short-term fluctuations and focus on seasonality.

**Mean particle size**

Mean particle diameter in the upper 50 m, weighted by particle cross-sectional area, was estimated from high-frequency variations in \(b_{bp}\) and \(c_p\) using the variance-to-mean ratio method\(^{51}\), adapted for use on profiling floats following Rembauville et al. (2017)\(^{21}\). This method extracts particle size information from the “spikiness” of optical profiles where particle concentrations are too high to separate individual spikes from a small-particle baseline. Briefly, mean diameter was estimated from \(c_p\) via Eq. (1):

\[
\bar{A}_{cp} = \frac{\text{var}(c_p\text{ detrended}) \cdot V \cdot 1}{Q_c \cdot \alpha(\tau)} \quad (1)
\]

\[
\alpha(\tau) = \begin{cases} 
1 - (3\tau)^{-1}, & \text{if } \tau \geq 1 \\
\tau - \tau^2/3, & \text{if } \tau \leq 1 
\end{cases}
\]

\[
\tau = \left( \frac{t_{res}}{t_{samp}} \right)
\]

\[
\bar{D}_{cp} = 2 \sqrt{\bar{A} \pi^{-1}}
\]

where \(t_{res}\) is the residence time of a particle in the sample volume (0.1 s), \(t_{samp}\) is the duration of a single measurement (1 s), \(V\) is the transmissometer sample volume (12.5 ml), \(Q_c\) is the optical attenuation efficiency of the particles (assumed to be 2 following Bohren and Huffman, 1983\(^{52}\)), \(\text{var}(c_p\text{ detrended})\) is the variance of \(c_p\) after detrending using an 5-point running median, and \(D_{cp}\) is the area-weighted mean particle diameter. The calculation was then repeated for \(b_{bp}\) at both 470 and 700 nm, by replacing \(c_p\) with \(b_{bp}\) in Eq. 1, and using \(V = 0.62\) ml (Briggs et al., 2013), \(t_{samp} = 1\) s, \(t_{res} = 0.06\) s, and \(Q_{bb} = 0.02\)^{51}. In order to reduce “noise” in the individual size estimates, the three estimates were then combined together into 10 d bins, whose medians and 25\(^{th}\) and 75\(^{th}\) percentiles are reported here.

**Spike index**
F and b_{bp} were partitioned into 3 components as in Briggs et al. (2020)\textsuperscript{53}: deep sensor blanks, including a background of small refractory particles (b_{bs} and F_{r}); small, labile backscattering (b_{bs}) and fluorescing (F_{s}) particles; and large, fast-sinking backscattering (b_{bl}) and fluorescing (F_{l}) particles (see supplementary materials in Briggs et al. (2020)\textsuperscript{53} for details on this partitioning). Temporal sections of F_{l} and b_{bl} at 470 and 700 nm are shown in supplementary figure S10. The spike index was computed by depth-integrating F_{l} and b_{bl} from the mixing layer depth to 500 m, and then normalising by their minimum and maximum values. This index makes it possible to quantify the amount of large, fast-sinking particles present in the water column. For the sake of simplicity, only the b_{bl} 470 spike index is shown in figure 3. The b_{bl} 700 spike index shows similar results.

**Optical sediment trap**

The Optical Sediment Trap (OST) method uses the rate of change of particle attenuance (-ln(transmittance), unitless) during the float drifting period at the parking depth of 300 m\textsuperscript{16,19,34}. The accumulation of small slow-sinking particles onto the upward-facing window of the transmissometer drives a smoothly increasing attenuance while the accumulation of large fast-sinking particles produces discontinuities in transmissometer records. Both signals were converted into a continuous flux and a pulsed flux (in units of d\textsuperscript{-1}), respectively, following procedures described in Estapa et al. (2013, 2017)\textsuperscript{19,34} with few modifications to take into account the use of a different platform with different parking behaviour and sampling frequency (every 1h in this study). First, parking phases of duration < 24h were not considered to ensure a sufficient number of data points. The first 3 data points (3h) were removed from the analysis as the float takes time to stabilize at the target depth. Optical spikes, defined as an increase in beam attenuation of 0.002 h\textsuperscript{-1} followed by a decrease within 3h, were also removed and a 5-point median filter was applied. Then, the remaining data points were divided into linear segments interspersed with discontinuities, or ‘jumps’, following the procedure in Estapa et al (2013)\textsuperscript{34}, except that discontinuities were not identified a priori with a threshold criteria but resulted from the subdivision of segments not meeting the fitting criteria. For each parking phase, the continuous flux of slow-sinking particles was computed as the mean slope of all linear segments weighted by their length. The pulsed flux was computed as the sum of all positive discontinuities normalized by the duration of the parking phase.

**Mixed layer pump**
The detection of episodic Mixed Layer Pump (MLP) events follows the methodology described in Lacour et al. (2019). The basis of this method is to distinguish between the mixed layer, the zone of relatively homogeneous water formed by the history of mixing, and the mixing layer, the zone in which mixing is currently active. The mixed and mixing layer depths were computed as the maximum vertical gradient of density (MLD_{dens}) and chlorophyll-a fluorescence (MLD_{bio}), respectively. The underlying concept is that chlorophyll-a fluorescence, as a proxy of phytoplankton biomass, is homogeneous over the whole mixing layer if turbulent mixing overcomes vertical variations in phytoplankton net growth rate. During the winter to spring transition, MLD_{bio} is generally shallower than MLD_{dens} and varies at higher frequency (~1-2 days for MLD_{bio} and ~10 days or more for MLD_{dens}) in response to changes in atmospheric forcing (Fig. S11). On a single float profile, the formation of a remnant layer, delimited by MLD_{bio} at the top and MLD_{dens} at the bottom, can be easily identified and used as a signature of a recent MLP event. To quantify the amount of carbon exported to the mesopelagic by MLP events, each POC profile was depth-integrated over the remnant layer, where organic matter was potentially trapped. To avoid accounting for POC initially present in the mesopelagic, the mean POC value from MLD_{dens} to 500 m was subtracted from the POC profile before the depth integration. Note that in the presence of SCM, typically in summer, MLD_{bio} can be deeper than the mixing layer depth and corresponds more likely to the base of the euphotic zone.

**Eddy subduction pump**

To detect the subduction events contributing to the ESP, we used an updated version of the detection method described in Llort et al. (2018). This method relies on the fact that vertical extent of submesoscale features that drive ESP is of the order of ~10 m and can be smoothed out by averaging float’s vertical variability over larger vertical scales. We applied a 20-bin running average over profiles at 5 m vertical resolution. As a result, we obtained submesoscale-free vertical variability that can be compared to the original profiles to identify anomalous features. We focused only on anomalies found between the bottom of the mixing layer depth and 500 m. As anomalies can be related to other mechanisms than submesoscale vertical circulation, individual profiles were classified as an ESP event if negative anomalies of Apparent Oxygen Utilisation (AOU’) and spice (\( \pi’ \)) coincided in depth. To define the relevant anomalies, we applied detection thresholds to only consider anomalies with AOU’ < -8 \( \mu \text{mol} \text{kg}^{-1} \) and \( \pi’ < -0.1 \). We have also modified the method to better constrain the thickness of the anomalous features, an important metric to estimate the amount of POC exported. In the current
version, we detected the top and bottom depths of both AOU’ and \( \pi' \) as the first and last bin depths where the anomalies were still negative. We then compared the detected depths on AOU and \( \pi' \) to define the deepest (shallowest) of the two as the anomaly bottom (top) depth.

**Iron stress index**

The iron stress index was computed following the method in Ryan-Keogh & Thomalla (2020). The concept of this method is that NPQ variability is linked to iron and light availability and has the potential to provide important diagnostic information on phytoplankton physiology. To remove the effect of in situ light availability on NPQ variability, Ryan-Keogh & Thomalla (2020) proposed to compute \( \alpha_{NPQ} \) the initial slope of the NPQ-PAR curve. Thereby, \( \alpha_{NPQ} \) could be used as a proxy for iron limitation, with higher values being associated with greater iron stress. In our study, NPQ as a function of depth was quantified as the difference between the quenching corrected fluorescence profile and the quenched one, normalized by the latter. For each profile, we plotted our iPAR estimates against NPQ values. We then applied a linear fit to the NPQ-iPAR curve in the region of low iPAR values (between 15 and 75 \( \mu \)mol photons \( m^{-2} s^{-1} \)), where the slope of the linear fit gives \( \alpha_{NPQ} \). Linear fits with \( R^2 < 0.8 \) were rejected (1% of the profiles). We did not fit a Platt-like model as in Ryan-Keogh & Thomalla (2020) because our iPAR values were too low to induce a saturation plateau and because we were interested only in the initial slope of the NPQ-iPAR curve. Note that \( \alpha_{NPQ} \), as a proxy for iron limitation, has to be interpreted with caution as shifts in phytoplankton community composition and changes in the light regime and thus phytoplankton photoacclimation status impact the variability of \( \alpha_{NPQ} \). For that reason, we also computed the median light level within the mixing layer (\( I_{ML} = PAR_{SAT} e^{-0.5 K_d(PAR) MLD_{bio}} \)) which is often used in photoacclimation models. Here, \( PAR_{SAT} \) is the daily mean PAR from MODIS Aqua (4 km) and \( K_d(PAR) \) is the diffuse attenuation coefficient of PAR (in units of \( m^{-1} \)). \( K_d \) was first computed at 490 nm by fitting a fourth-degree polynomial function to the logarithm of the downwelling irradiance \( E_d(490) \) as a function of depth, measured by the float, and then calculating the mean slope over the first 50 m. \( K_d(490) \) was then converted to \( K_d(PAR) \) following equation 9 in Morel et al. (2007). We did not find a clear relationship between \( I_{ML} \) and \( \alpha_{NPQ} \) (Fig. S5). The influence of shifts in phytoplankton community composition on \( \alpha_{NPQ} \) variability cannot be ruled out, but it is worth noting that phytoplankton community also shifts in response to iron availability.

**Phenology metrics**
Phytoplankton phenology can be characterised by three metrics: (i) the bloom onset when bulk phytoplankton biomass starts accumulating (i.e. when the accumulation rate r changes from negative to positive value), (ii) the bloom climax when r reaches its annual maximum value, (iii) the bloom apex when bulk phytoplankton biomass P reaches its annual maximum value. Following Uchida et al. (2019), P was estimated by vertically integrating C_{phyto}, derived from the backscattering, over the whole water column. To avoid including non-phytoplankton particulate matter (e.g. refractory material) in the calculation, C_{phyto} was masked out at depths where $F_s = F - F_l - F_r \leq 0$, where $F_r = 0.03 \text{ mg m}^{-3}$ (Fig. S7). The accumulation rate was then calculated as $r = \frac{1}{P} \frac{\partial P}{\partial t}$. The bloom onset and apex were used as milestones to compare the timing of the 2016-17 and 2017-18 blooms with the seasonality of the carbon export pathways (Fig. S12). The time axis in Figure S12 was rescaled by the onset and the apex of the bloom, so that 0 corresponds to the onset and 1 to the apex.

Acknowledgements and data availability statement

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Author contributions

LL and JL designed the study and conducted the data analysis. LL wrote the manuscript. JL contributed to the data collection and to manuscript review and editing. NB conducted the mean particle diameter analysis and contributed to manuscript review and editing. PS and PWB contributed to the data collection and writing, and helped to design the study.

Competing financial interests

The authors declare no competing financial interests.
Figure 1 Contrasting environmental conditions in the Pacific sector of the Southern Ocean. (a) Float surfacing positions (every 1 to 5 days) during its 36-month mission. Background map is a climatology of Absolute Dynamic Topography (ADT). Light grey line indicates the climatological position of the Polar Front (ADT=-0.48 m) used in Ardyna et al. (2017) and derived from Swart et al. (2010). (b) Minimum distance between float positions and the sea ice edge, defined as the 15% sea ice concentration limit. Vertical sections of (c) temperature and (d) salinity.
and (d) salinity recorded by the float. Light grey lines are isopycnals and the dark line is the mixing layer depth as defined by biological criteria (MLD\textsubscript{bio}, see Methods). Periods with missing data are blank in panels c and d.

**Figure 2** Characteristics of the particle assemblage over three annual cycles evidenced by bio-optical proxies. Vertical sections of (a) particulate organic carbon (POC), and (b) chlorophyll fluorescence to backscattering ratio F\textsubscript{s} / b\textsubscript{bs} 700. Black lines show the mixing layer depth (MLD\textsubscript{bio}). (c) Mean particle size (diameter) in the upper 50 m. The blue line and vertical black lines represent the median and inter-quartile ranges, respectively, over 10-day bins. As milestones, triangles on top of each panel mark the timing of onset (first positive net phytoplankton accumulation), climax (maximum accumulation) and apex (maximum bulk phytoplankton biomass) for each bloom, following the method in Uchida et al. 2019\textsuperscript{65} (see Methods).
Figure 3 Seasonality of multiple carbon export pathways. (a) Stock of particulate organic carbon (POC) in MLDbio. (b) Spike index calculated as depth-integrated $b_{\text{bio}}$ and $F_1$, from MLDbio to 500 m, and normalised by their minimum and maximum values (no units, see Methods). (c) Continuous and pulsed flux measured by the optical sediment trap (OST) at ~300 m. Pulsed flux is defined as discontinuities in the OST record (see Methods). Panels b and c show two complementary approaches to characterise the BGP. (d) Green empty bars show the stock of POC in subsurface layers presenting anomalous POC, Apparent Oxygen Utilization (AOU) and spice, related to ESP events. Brown bars show the stock of POC in the remnant layer, associated with MLP events. In all panels, data has been averaged over 10-day bins. For the ESP and MLP, this averaging avoids double counting POC stocks for events sampled multiple times.
Figure S1 Environmental factors controlling phytoplankton phenology and vertical carbon export. (a) Daily wind speed at 10 m above the ocean surface, along the float trajectory. The black line is a 30-day smooth. (b) Daily net heat flux along the float trajectory, with positive values referring to a flux from the atmosphere to the ocean. Vertical sections of (c) Buoyancy frequency $N^2$ and (d) spice anomaly (see Fig S2) recorded by the float.
Figure S2 Spice anomaly reveals the interleaving of water masses. Examples of spice profile on October 1st 2016 (left), when the float crossed the polar front, and June 7th 2018 (right), when a strong dipole in the spice anomaly was observed at the base of the melt-water layer (see Fig. 1d). Dashed and continuous black lines show a 100-point (or 100 m) and 10-point (or 10 m) moving average of spice profiles, respectively. Red and blue shaded areas indicate positive and negative spice anomaly (difference between 10- and 100-point filtered signals), respectively, as shown on Fig. S1d.
Figure S3 Pseudo-Lagrangian framework. The float temporal section was divided into three periods in which the contiguous nature of the water masses was verified (areas delimited by vertical dashed lines in top panel). The three TS diagrams at the bottom reveal that the float did not cross any strong water mass boundaries during each period of interest. Grey dots in the background show data points of the complete timeseries.
Figure S4 Satellite Particulate Inorganic Carbon (PIC) reveals the presence of a coccolithophore bloom in austral summer 2016-17. Trajectory of the float superimposed on summer (December 2016 to February 2017) composite GlobColour image (25 km) of PIC. Red circles indicate the location of the summer float profiles where we observed a strong decrease in the $F_s / b_{bs}$ ratio.
Figure S5 Iron and light limitation play a key role in shaping phytoplankton phenology in the SO. Top panel, median light level within the mixing layer ($I_{ML}$, blue bars) derived from satellite daily mean PAR (red curve). $I_{ML}$ is an indicator of the light history of phytoplankton cells and has been previously used as a descriptor of photoacclimation. Gaps in the timeseries are due to the winter polar night or the presence of sea ice. Bottom panel, initial slope of the non-photochemical quenching (NPQ) versus instantaneous iPAR curve ($\alpha_{NPQ}$). $\alpha_{NPQ}$ is a proxy for phytoplankton iron limitation, the higher $\alpha_{NPQ}$, the more iron-limited.
Figure S6 Vertical distribution of $F_1$ and $b_{bl}$, summed in 20 m bins over the course of the blooming period (from climax to 30 days after apex for the first two blooms and from climax to the end of the timeseries for the last bloom which shows two distinct biomass peaks few months apart, see Fig. S8) and normalised by their minimum and maximum values. As opposed to $b_{bl}$, $F_1$ was quickly attenuated with depth, although this is less clear for the last 2018-19 bloom.
Figure S7 Bulk phytoplankton biomass derived from optical measurements. Vertical sections of Chlorophyll $a$ fluorescence ($F_s$, top), and phytoplankton carbon ($C_{phyto}$) derived from the backscattering at 470 nm ($b_{bp 470}$, bottom). Both colour bars are on a log scale. Black lines show the mixing layer depth. In the bottom panel, $C_{phyto}$ was masked out where $F_s \leq 0$. The bulk phytoplankton biomass was then calculated by vertically integrating masked $C_{phyto}$ over the whole water column, following Uchida et al. (2019). Note that SCMs are barely discernible in the top figure due to the use of running minimum and maximum filters to compute $F_s$, which partially erases subsurface peaks in fluorescence (see Methods).
Figure S8 Phytoplankton phenology metrics. Top, mixing layer depth (MLD_{bio}). Middle, bulk phytoplankton biomass. Continuous black lines in the top two panels show 30-day moving averages of these variables. Bottom, net growth rate $r$ calculated from smoothed depth-integrated $C_{\text{phyto}}$ (see Methods). Vertical dotted green lines indicate the timing of onset, climax and apex of the three phytoplankton blooms recorded by the float.
Figure S9 Examples of ESP signatures identified from three consecutive AOU, POC and spice profiles in early September 2016. Shaded grey areas highlight features falling within the same water mass (defined by a potential density of 27.31 ± 0.04 kg m$^{-3}$).
Figure S10 Sections of large-particle fluorescence $F_l$ (top), revealing the presence of fresh phytoplankton aggregates, and large-particle backscattering $b_{bl, 470}$ (middle) and $b_{bl, 700}$ (bottom), which additionally include fecal and detrital matter.
Figure S11 Illustration of the MLP during the 2017-18 bloom which shows how rapid and intermittent re-stratifications of the mixing layer isolates fresh particles at depth. The freshness of particles is revealed by the relatively high $F_s / b_{bs}$ ratio (100-200 mg m$^{-2}$) observed in the remnant layer (delimited by MLD$_{bio}$ at the top and MLD$_{dens}$ at the bottom).
Figure S12 Relationship between phytoplankton phenology and carbon export pathways. a) cumulative sum of $\Delta C_{\text{phyto}}$, the temporal difference of bulk phytoplankton biomass. b) Time-integrated continuous (solid lines) and pulsed flux (dashed lines) estimated from optical sediment trap (OST) measurements. c) Cumulative POC stock in eddy-induced subsurface anomalies, and d) MLP remnant layers. POC stocks were first averaged over 10 days, as shown in Fig. 3, to avoid double counting when the float samples the same event multiple times. Each colour represents a different seasonal cycle. The coloured part of the curves indicates the region where we assumed a quasi-Lagrangian framework (See Fig. S3 and Methods). Red dots mark the timing of the bloom climax. The time axis was rescaled by the onset ($t_{\text{onset}}$) and the apex ($t_{\text{apex}}$) of the bloom, so that 0 corresponds to the onset and 1 to
the apex. Note that the last seasonal cycle (2018/2019) is not represented due to missing data early in the productive season.
Figure S13 Mean depths of the detected subsurface anomalies related to ESP events.
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