



# Unexpected quasi-independence of coloured dissolved organic matter absorption from chlorophyll-*a* concentration in the Southern Ocean

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Received: 5 November 2025 – Discussion started: 24 November 2025

Revised: 20 April 2026 – Accepted: 26 April 2026 – Published: 8 May 2026

**Abstract.** The absorption coefficient of coloured dissolved organic matter (CDOM),  $a_y$ , plays a critical role in driving ocean optical properties and thereby light attenuation and light-dependent biogeochemical cycles. In the Southern Ocean (SO), however,  $a_y$  remains poorly documented because of the scarcity of in situ measurements and the absence of suitable bio-optical models. To address this gap, we derived  $a_y$  in surface waters from the diffuse attenuation coefficient ( $K_d$ ) derived from radiometric measurements performed by Biogeochemical-Argo floats. Sensitivity analyses using Monte Carlo simulations indicated that the uncertainty of our estimates is mainly driven by the uncertainty in  $K_d$ , and is overall  $\sim 18\%$  for  $a_y$  at 380 and 412 nm. Our derived  $a_y$  vs. chlorophyll-*a* concentration (Chl) relationships for low-latitude waters are consistent with previously published relationships. They, however, diverge in the SO, with a larger relative contribution of  $a_y$  to the absorption budget for clear waters ( $\text{Chl} < \sim 0.2 \text{ mg m}^{-3}$ ) and the opposite for greener waters, leading to a weaker dependence of  $a_y$  on Chl. Lower-than-expected CDOM absorption mostly happens during the austral summer, suggesting significant photobleaching or lower biologically-mediated production. The relative contributions of CDOM and phytoplankton to the absorption budget are also found to diverge from what bio-optical models predict, with implication for interpretation of satellite ocean colour observations in the SO.

## 1 Introduction

Coloured dissolved organic matter (CDOM) in oceanic waters is the fraction of the dissolved organic matter (DOM) pool that absorbs light in the ultraviolet and visible region of the electromagnetic spectrum. The corresponding absorption coefficient is hereafter denoted  $a_y$  ( $\text{m}^{-1}$ ), with the subscript referring to the *yellow substance* denomination used in previous studies (Morel and Gentili, 2009). CDOM absorption reduces light penetration within the water column, thereby influencing phytoplankton dynamics, nutrient cycling, primary productivity, and the overall biological carbon pump (Nelson and Siegel, 2002; Siegel et al., 2002; Nelson and Siegel, 2013; Mannino et al., 2014). Therefore, it plays a significant role in regulating biogeochemical and photochemical processes within the global carbon cycle (Gruber et al., 2009, 2019; Hauck et al., 2023; Boyd et al., 2024). Accordingly, it is important to quantify the CDOM distribution for better understanding of the biogeochemical processes underlying its variability.

In addition, CDOM absorption in the blue part of the spectrum is superimposed to phytoplankton absorption, which means that accurately quantifying  $a_y$  is also of paramount importance to a proper estimate of the phytoplankton chlorophyll-*a* concentration (Chl,  $\text{mg m}^{-3}$ ) from satellite ocean colour measurements, which combine re-

flectance measurements in several spectral bands including in the blue (generally around 440 nm). A higher (lower) CDOM contribution to absorption in the blue than assumed in semi-analytical ocean colour algorithms will lead to (under)overestimating Chl. Several studies have indeed pointed to significant biases when comparing satellite-derived Chl with field measurements in the SO (e.g., Johnson et al., 2013; Chen et al., 2021) while others did not identify such an issue (e.g., Haëntjens et al., 2017). Atmospheric correction issues have been suggested as a possible reason for these degraded performances in the SO, although only for studies focusing on coastal areas (Salyuk et al., 2025). Therefore, no consensus exists about the reasons for the poor performance of satellite Chl algorithms (e.g., Morel and Maritorena, 2001; Hu et al., 2012) in the SO. Since these algorithms have been developed primarily from low-latitude bio-optical data sets, the question arose as to whether the SO bio-optical properties significantly differ from what they are in low-latitude oceans, making the application of current satellite ocean colour algorithm problematic.

Studies have indeed shown that bio-optical properties of the SO are statistically different from low-latitude waters, both for phytoplankton (Robinson et al., 2021) and non-algal particles (NAP; Li et al., 2024), with impact on the ocean reflectance (Dierssen and Smith, 2000). The role of CDOM absorption as another source of misinterpretation of the satellite ocean colour signal in terms of Chl has not, however, been thoroughly investigated. Therefore, the main objective of this study is to assess whether the relationship between Chl and  $a_y$  in the SO differs from other oceanic regions and, if it does, to discuss possible reasons.

Whether CDOM concentration, hence the amplitude of  $a_y$ , is high or low in the surface layers of the oceans depend on the balance between CDOM production and losses. In the open oceans, production is essentially local from biological activity, and losses can occur either through photobleaching, biological degradation, dilution through vertical mixing with CDOM-poor waters or enrichment if mixing occurs with CDOM-rich deep waters (Siegel et al., 2005; Nelson and Siegel, 2013; Fichot et al., 2023; Yamamoto et al., 2024). Surface circulation can either lead to increases or decreases of  $a_y$  depending on which water masses are advected. These processes occur in all oceans, yet some peculiarities of the SO might lead to a different balance between CDOM production and losses.

The SO is characterized by strong vertical mixing in winter, low photobleaching in the low-irradiance winter yet strong photobleaching in summer when irradiance can be as high as it is in the equatorial belt (Campbell and Aarup, 1989). Phytoplankton populations are different to what they are in low-latitude environments (e.g., Wright et al., 2010). Sea ice melting is another potential source of CDOM (Ortega-Retuerta et al., 2010b) affecting waters in the seasonal ice zone. It is therefore legitimate to expect that

this rather peculiar combination of characteristics and processes might lead to changes in the  $a_y$  vs. Chl relationship.

While the number of CDOM absorption measurements are increasing in global databases, as for any dynamic variable, in situ observations will always under-sample the ocean. This is even more true in the SO where logistical difficulty and the harsh environment mean that we have extremely limited in situ studies of CDOM. Therefore, addressing our question using ship-based  $a_y$  and Chl measurements was not possible. The deployment of autonomous profiling Biogeochemical-Argo (hereafter BGC-Argo) floats in the SO by, e.g., the Southern Ocean Carbon and Climate observations and Modelling (SOCCOM; Sarmiento et al., 2023) or the Remotely-sensed Biogeochemical Cycles in the Ocean (RemOcean; Claustre et al., 2020) programs has dramatically improved the availability of in situ data, and made this study feasible. Here we used floats equipped with radiometers, allowing a semi-analytical derivation of  $a_y$  from the diffuse attenuation coefficient of downward irradiance,  $K_d$  ( $\text{m}^{-1}$ ). Our method uncertainties were quantified through sensitivity analyses and Monte Carlo simulations. We then compared  $a_y$ -related bio-optical properties and relationships between the SO and low-latitude waters to explore potential mechanisms underlying their differing distributions.

## 2 Data and methods

### 2.1 Data selection from BGC-Argo floats

We used data from a total of 60 BGC-Argo floats deployed in the SO (south of 40° S in this study) between 29 November 2013 and 2 May 2025, and 211 floats deployed in low-latitude regions (from 40° S to 60° N) from 22 October 2012 to 26 December 2024. These floats are equipped with Seabird CTD sensors for temperature and salinity, Seabird/Satlantic OCR-500 multispectral radiometers collecting downward plane irradiance ( $E_d(z, \lambda)$ ,  $\mu\text{W cm}^{-2} \text{nm}^{-1}$ ) at 380, 412 and 490 nm, and Seabird/WET Labs ECO-series sensors providing the total optical backscattering coefficient at 700 nm ( $b_b$  (700),  $\text{m}^{-1}$ ) and chlorophyll fluorescence. Overall, these floats had collected 10 579 (SO) and 38 615 (low latitudes) profiles during the period indicated.

For each of the floats, we first eliminated profiles collected in shallow waters (depth < 200 m) based on the global relief ETOPO1 data base (NOAA, 2009), as well as profiles for which the sun elevation was < 15° at the end of the up-cast. Then, for chlorophyll, backscattering and radiometry, we only kept profiles flagged “A” (100 % of good data) or “B” (at least 75 % of good data), as per the nomenclature of the Argo data management team (Argo data management, 2025). For the profiles passing this first screening, only data points with a quality flag set to either 1 (good), 2 (probably good), 5 (value changed) or 8 (interpolated value) were kept.

The total of data points flagged either 1 or 2 was from 80 % to 98 % of the entire data set depending on the parameter.

The locations of profiles that passed these quality controls (roughly one third of the total) are displayed in Fig. 1, and the number of profiles eliminated after each step of quality control are summarized in Table S1 in the Supplement. The temporal coverage of the selected profiles across years and months is displayed in Fig. S1 in the Supplement. The distribution of the sun zenith angles is depicted in Fig. S2a, while Fig. S2b shows the irradiance just above the surface for  $\lambda = 490$  nm, limited to cases within 20 % of the theoretical clear-sky value calculated following Gregg and Carder (1990).

## 2.2 From radiometric measurements to $K_d$

The overall workflow we used to then process the BGC-Argo data is displayed in Fig. S3. We did not correct the radiometry data for dark deep values, which have been shown to be negligible (Organelli et al., 2016). We checked these values and indeed they were always lower than  $10^{-3}$   $\text{mW cm}^{-2} \text{nm}^{-1}$ , with a distribution centred on  $10^{-4}$   $\text{mW cm}^{-2} \text{nm}^{-1}$ .

Then a 4th order polynomial was fitted to the data to clean the  $E_d(z, \lambda)$  profiles from changes due to possible changes in the above-water downward irradiance caused by clouds and from near surface fluctuations generated by waves. This fit was only performed if more than 20 valid data points were available, otherwise the profile was eliminated. This fitting procedure is similar to what Organelli et al. (2016) did, although we did not find it necessary to repeat the 4th order polynomial in order to get smooth profiles.

The  $K_d$  were then calculated in three different ways from the fitted  $E_d$  profile, to allow a sensitivity study of the  $K_d$  value. The first one ( $K_d(0-20\text{ m})$ ) was calculated from  $E_d(z=0^-)$  and  $E_d(z=20\text{ m})$ . This approach mimics the methodology used in most of the field data sets by Morel (1988) and later revised by Morel and Maritorena (2001) (hereafter MM01), and it is taken here as the reference for the low-latitude environments. At that time, profiling radiometers were not yet available; instead, radiometers were deployed using winches and stabilized at successive depths where measurements were collected. A depth of about 20 m was typically chosen, as irradiance fluctuations were sufficiently dampened to ensure reliable  $E_d$  measurements. The second  $K_d$  calculation ( $K_d(Z_{pd})$ ) was similar but used  $E_d$  at the first optical depth ( $Z_{pd}$ ) instead of at 20 m. This depth was calculated for each wavelength and corresponds to the point where  $E_d$  is reduced to  $1/e$  of its below-surface value. At this stage, we added another quality control by eliminating profiles when  $Z_{pd}$  deviated by more than a factor of 2.5 (either greater or lower) from the value predicted from Chl using MM01. The third calculation took the mode of the distribution of local  $K_d$  values, computed at each measurement depth within successive 5 m intervals from just below the sur-

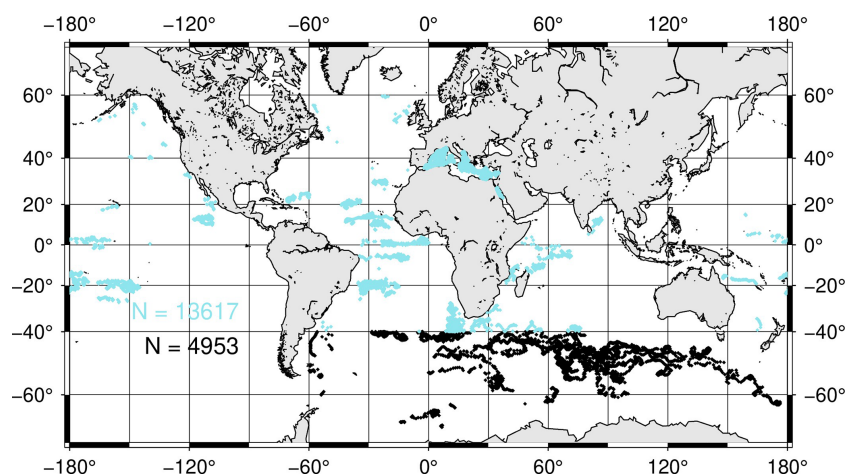
face down to the first optical depth.  $K_d(0-20\text{ m})$  is the one used in subsequent analyses.

## 2.3 Chlorophyll, backscattering and mixed-layer depth from BGC-Argo floats

The Chl values delivered by the BGC-Argo program are derived from chlorophyll fluorescence profiles corrected for possible non-photochemical quenching (Xing et al., 2018; Schmechtig et al., 2023), then scaled to Chl using manufacturers calibration parameters and further divided by a factor of 2 following recommendation by Roesler et al. (2017). A similar correction of the fluorescence-to-Chl ratio was recommended for SO phytoplankton by Schallenberg et al. (2022) with, however, a factor of 3.79 instead of 2, which we have used here. Each Chl and total backscattering profiles were adjusted by shifting the whole profile so that the average value between 200 and 400 dbar equals the mode of the distribution of deep values calculated over the same depth range from all profiles of all floats. This adjustment was performed to account for the potential bias between different measurement technologies and for possible instrument drift. These deep values were  $5.5 \times 10^{-3}$   $\text{mg Chl m}^{-3}$  and  $3 \times 10^{-4}$   $\text{m}^{-1}$  for the backscattering measurements.

After this procedure, we found 2070 values of surface Chl lower than  $0.02$   $\text{mg m}^{-3}$  (15 % of the data). This is unrealistic, as the minimum concentrations ever measured in the upper layers of the ocean are about  $0.02$   $\text{mg m}^{-3}$ , e.g., in the southeast Pacific gyre (Morel et al., 2007b). The use of a single factor of 2 for the fluorescence to Chl conversion is likely responsible for such underestimations, which is consistent with the high variability actually reported for this factor by Roesler et al. (2017). It can also be partly due to the impact of CDOM absorption at depth on the chlorophyll fluorescence efficiency, although this effect was mostly observed for coastal CDOM-rich waters (McKee et al., 2007). Instead of artificially truncating the data set at Chl values  $< 0.02$   $\text{mg m}^{-3}$ , we re-adjusted the deep values to an average of  $0.02$   $\text{mg m}^{-3}$ . This admittedly subjective adjustment allowed avoiding unrealistic low surface Chl values while keeping consistency in the deep adjustment.

Similarly to what was done for the radiometry profiles, a 4th order polynomial was fitted to the inherently noisy Chl and backscattering profiles using data from the top 50 m only. Finally, average surface Chl,  $b_{bp}(700)$ , temperature ( $T$ , °C) and salinity ( $S$ , psu) were calculated over the first optical depth for  $\lambda = 380$  nm determined from the radiometry profiles. The average  $T$  and  $S$  were subsequently used to calculate the seawater backscattering coefficient ( $b_{bw}$ ,  $\text{m}^{-1}$ ) according to Zhang and Hu (2009) and Zhang et al. (2009), which is subtracted from the total backscattering coefficient to get the particulate backscattering coefficient,  $b_{bp}$ . The resulting distributions for Chl and  $b_{bp}$  are illustrated in Fig. S2c, d. The contribution of seawater to the diffuse attenuation coefficient for downward irradiance,  $K_w$ , is ap-



**Figure 1.** Surface locations of the BGC-Argo float profiles used in this study for the SO (black) and elsewhere (blue), after various screenings have been applied to the full data set (see methods).

proximated as  $a_w + b_{bw}$ , where  $a_w$  is the absorption of seawater and its value can be found in Lee et al. (2015). This  $K_w$  value is used to derive the contribution of all non-water components to  $K_d$  as in Morel and Maritorena (2001), as  $K_{bio} = K_d - K_w$ .

The temperature and salinity profiles were used to calculate the depth of the mixed layer (MLD) based on a density criterion, by which MLD is the depth where the density is different by  $0.03 \text{ kg m}^{-3}$  from its average value in the top 10 m (de Boyer Montégut et al., 2004). Density calculations were performed using the swSigmaT R function that uses the UNESCO formulae (IOC, SCOR, and IAPSO, 2010).

## 2.4 Ship-based measurements

The particulate and CDOM absorptions,  $a_p$  ( $\text{m}^{-1}$ ) and  $a_y$  ( $\text{m}^{-1}$ ), form the total non-water absorption. Therefore, to determine  $a_y$  we need as realistic as possible estimates of  $a_p$ . For the low-latitude oceans, we used the  $a_p$  vs. Chl relationships from Bricaud et al. (1998). For the SO, we used ship-based field data acquired during two Southern Ocean research voyages: the Antarctic Circumpolar Expedition (ACE) aboard the RV *Akademik Tryoshnikov* during the Austral Summer from 20 December 2016 to 19 March 2017 (Robinson et al., 2021), and the Southern Ocean Large Areal Carbon Export (SOLACE) research voyage aboard the RV Investigator (voyage IN2020\_V08) from 5 December 2020 to 16 January 2021.

Water samples were collected during the ACE and SOLACE either 3-hourly from the underway seawater supply (sampling depth  $\sim 5$  m) or from the shallowest depth of the CTD (conductivity, temperature, and depth) rosette casts. Phytoplankton pigment concentrations were determined using high performance liquid chromatography (HPLC, see details in Ras et al., 2008 and references therein). Total Chl was defined as the sum of mono- and divinyl chlorophyll *a* con-

centration, chlorophyllide *a*, and the allomeric and epimeric forms of chlorophyll *a* (Hooker and Zibordi, 2005; Reynolds et al., 2016). Particulate absorption ( $a_p$ ) measurements were made on the same filters analysed for pigments. A full description of the measurement protocols and the data are available in Antoine et al. (2021) and Robinson et al. (2021). The resulting  $a_p$  vs. Chl relationships are displayed in Fig. S4.

Measurements of  $a_y$  are unfortunately seldom carried out at sea, leaving us with few options for validating the  $a_y$  estimates. We did not have any such data for the SO. For the low-latitude areas, we used three data sets of field  $a_y$  measurements. The first one is from the Bouée pour l'acquisition d'une Series Optique à Long terme (BOUSSOLE) in the Mediterranean Sea (Antoine et al., 2006). Measurements were carried out at this site from 2011 to 2015, and the initial years of data have been presented by Organelli et al. (2014). The second data set is from the BIOgeochemistry and Optics South Pacific Experiment (BIOSOPE) that occurred in 2004 in the Southeast Pacific Ocean (Claustre et al., 2008), with the  $a_y$  data analysed by Bricaud et al. (2010). The third data set (18 data points out of the SO) was extracted from the NASA NOMAD data base (Werdell and Bailey, 2005). The Mediterranean Sea is known to display higher-than-average CDOM absorption per Chl, while the Southeast Pacific Ocean exhibits the opposite pattern (Morel et al., 2007b). Therefore, the BOUSSOLE data set is expected to match the upper part of the distribution of the  $a_y$  values derived here when plotted as a function of Chl, while the BIOSOPE data would rather match the lower part of that distribution.

## 2.5 $a_y$ inversion model

The  $K_d(\lambda)$  can be expressed as a function of IOPs as follows (Gordon, 1989):

$$K_d(\lambda) = 1.0395 \frac{a(\lambda) + b_b(\lambda)}{\mu_d(\lambda)} \quad (1)$$

where  $\mu_d$  is the average cosine of  $E_d(0^-, \lambda)$ , and  $a(\lambda)$  and  $b_b(\lambda)$  are the total absorption and backscattering coefficients. This equation is based on radiative transfer calculations without inelastic scattering. The absorption and backscattering coefficients can be expanded as follows:

$$a(\lambda) = a_w(\lambda) + a_p(\lambda) + a_y(\lambda), \text{ and} \\ b_b(\lambda) = b_{bw}(\lambda) + b_{bp}(\lambda) \quad (2)$$

The contribution of CDOM to scattering is neglected in this study (Dall'Olmo et al., 2009). When substituting Eqs. (2)–(3) into (1),  $a_y(\lambda)$  can be solved as:

$$a_y(\lambda) = \frac{K_d(\lambda) \mu_d(\lambda)}{1.0395} - a_w(\lambda) - a_p(\lambda) - b_{bw}(\lambda) - b_{bp}(\lambda) \quad (3)$$

where  $a_w(\lambda)$  is assumed constant (values from Lee et al., 2015) and  $b_{bw}(\lambda)$  is calculated using measured temperature and salinity by BGC-Argo floats following Zhang and Hu (2009) and Zhang et al. (2009). Assuming that non-algal particles covary with Chl, the total particulate absorption can be described as a function of Chl based on in situ relationships. For the SO, to account for the high contribution of NAP in oligotrophic waters (Li et al., 2024), a background constant was added to the power-law regression between  $a_p(\lambda)$  and Chl:

$$a_p(\lambda) = \text{const}(\lambda) + \chi(\lambda) \text{Chl}^{e(\lambda)} \quad (4)$$

where the exponent  $e(\lambda)$  and the factor  $\chi(\lambda)$  are derived from concurrent measurements of Chl and  $a_p(\lambda)$  in the SO (see Fig. S4) or from Bricaud et al. (1998) for the low-latitude waters. Note that the tabulated data from Bricaud et al. (1998) do not include wavelengths < 400 nm, however, so we estimated values at 380 nm by extrapolating from their Fig. 4.

$b_{bp}(\lambda)$  is converted from  $b_{bp}(700)$  following

$$b_{bp}(\lambda) = b_{bp}(700) \left( \frac{700}{\lambda} \right)^\eta \quad (5)$$

where  $\eta$  equals to 1.08 for the SO, which is the mean value based on data collected during the ACE and SOLACE cruises (Li et al., 2024). While for the low-latitude waters, a value of 1.03 is adopted to be consistent with the value used in the GSM01 model developed by Maritorena et al. (2002) for non-polar waters. Chl and  $K_d(\lambda)$  are obtained from the floats' measurements (see above). The average cosine,  $\mu_d$ , which is a function of Chl,  $\lambda$  and sun zenith angle ( $\theta_s$ , equals to 90 minus sun elevation) under clear or overcast sky conditions, was derived using the lookup tables (LUT) developed

by Morel et al. (2002) and Morel and Gentili (2004). To determine whether a profile is collected under clear or overcast sky conditions, the spectral solar irradiance model of Gregg and Carder (1990) was implemented to generate the downward irradiance at 490 nm just below the ocean surface. If the absolute difference between the calculated and measured  $E_d(0^-, 490)$  is within 20 %, then the sky is assumed clear, otherwise it was classified as overcast.

## 2.6 Sensitivity studies

### 2.6.1 Individual parameters

The many steps of quality control performed on the  $E_d$  profiles might not fully eliminate bad data from unsupervised BGC-Argo measurements. Their impact on deriving  $K_d$  must be assessed, as it is the first source uncertainty when deriving  $a_y$  using Eq. (4). The three  $K_d$  estimates presented above were derived for this purpose.

The average cosine of the downward irradiance,  $\mu_d$ , is a second source of uncertainty when using Eq. (4). The  $\mu_d$  is taken from the LUTs that have been generated through a bio-optical model, which cannot be always appropriate for any bio-optical conditions (e.g., Morel et al., 2007a). The sensitivity study was conducted by either using the clear vs. cloudy sky test (Fig. S2), in which case  $\mu_d$  was taken from the corresponding LUT (referred to as  $\mu_d$  actual), or by using only the  $\mu_d$  for clear sky or only the  $\mu_d$  for overcast conditions. In doing this, we assumed that the difference in  $\mu_d$  between the clear-sky ( $\mu_d$  between 0.68 and 0.92) and the overcast conditions ( $\mu_d = 0.8$ ) is of the same order of magnitude than the difference caused by variability in bio-optical properties. The third significant source of uncertainty comes from  $a_p$ . This coefficient was derived from its average relationship to Chl, which cannot account for local departure from these relationships. Three relationships were used to assess the impact on  $a_y$  (Fig. S4): our SO relationship with (referred to SO dataset (Eq. 5) and without (SO dataset) a constant background value, and the one from Bricaud et al. (1998).

No individual sensitivity study was performed on  $b_{bw}$  and  $b_{bp}$  because of their small contribution in Eq. (4) and the rather well-constrained values for  $b_{bw}$ . The  $a_w$  value only represents a large contribution to the total absorption in clear waters at 490 nm. Therefore, uncertainties on its value were not assessed individually here.

### 2.6.2 Monte Carlo approach

The sensitivity studies to individual parameters does not provide an overall uncertainty for  $a_y$  as derived through Eq. (4). Therefore, we also conducted a systematic assessment of uncertainty using a Monte Carlo method. This approach involved running Eq. (4) 10 000 times for a given set of inputs, by introducing random uncertainties to each input in each run. For a given parameter, the random uncertainties were

generated by multiplying an average absolute or relative uncertainty (values in Table 1) by a random number within the  $[-0.5, +0.5]$  range. The absolute or relative type B uncertainties are provided in Table 1. The repeated calculations generated a set of 10 000  $a_y$  values for each  $K_d$  value, and the standard deviation of their distribution was used as a measure of uncertainty in  $a_y$ . The advantage of such an approach is that an uncertainty can be derived for each individual  $a_y$  value. This approach does not address potential systematic errors arising from biases in the  $K_d$  values.

### 3 Results

#### 3.1 General $a_y(\lambda)$ distributions

Histograms of retrieved  $a_y(\lambda)$  and corresponding spectral slopes are shown in Fig. 2. The mode values of  $a_y$  in the SO are  $0.0261 \text{ m}^{-1}$  at 380 nm,  $0.0194 \text{ m}^{-1}$  at 412 nm, and  $0.0073 \text{ m}^{-1}$  at 490 nm. For the low-latitude waters, the corresponding values are  $0.0239$ ,  $0.0139$  and  $0.0036 \text{ m}^{-1}$ . Notably, only about 2 % of the  $a_y(380)$  retrievals in the SO are negative, compared with 4 % at 412 nm and 20 % at 490 nm. In the low-latitude waters, the respective percentages are 2 %, 6 % and 29 %. This is expected, as  $a_y(490)$  is significantly smaller than  $a_y(380)$  (due to the exponential decrease with wavelength) and because the method has larger uncertainty at 490 nm. Additionally, the spectral slope of  $a_y(\lambda)$ ,  $S$  ( $\text{nm}^{-1}$ ), was calculated for the 3 possible wavelength pairs, and as the average of the  $a_y$  spectral dependence between 380 and 490 nm and between 412 and 490 nm. The mode value of  $S$  in the SO is  $0.009$  and  $0.015 \text{ nm}^{-1}$  for the low-latitude waters. The latter is close to the value of  $0.014 \text{ nm}^{-1}$  reported by Bricaud et al. (1981).

The latitudinal distributions of the average values of  $a_y(380)$ , the spectral slope of  $a_y$  (average value shown on Fig. 2c, d) and Chl, calculated from all data available in  $2^\circ$  latitude belts, are illustrated in Fig. 3. Generally,  $a_y(380)$  fluctuates between about  $0.01$  and  $0.04 \text{ m}^{-1}$  south of  $30^\circ \text{ S}$ , which is larger than the range observed in low-latitude waters ( $30^\circ \text{ S}$ – $30^\circ \text{ N}$ ), where values around  $0.01 \text{ m}^{-1}$  are quite frequent. This is consistent with the global  $a_y$  distribution that can be derived from Chl by Morel and Gentili (2009) (gold dots; hereafter referred to as MG09), except south of about  $40^\circ \text{ S}$  where the values we derived here are lower; this latitudinal band is also a band of very low continent to ocean ratio. Larger values are observed north of  $30^\circ \text{ N}$  with the increase of Chl towards northern latitudes. The largest spectral slopes are observed in subtropical regions around  $30^\circ \text{ S}$  and  $30^\circ \text{ N}$  and around  $60^\circ \text{ S}$ . The lowest values are in the equatorial region and around  $45^\circ \text{ S}$  and  $45^\circ \text{ N}$ . These distributions vary little seasonally (not shown). Isolated higher values around  $27^\circ \text{ N}$  are from two floats deployed in the northern Red Sea.

#### 3.2 $K_{\text{bio}}(\lambda)$ vs. Chl relationships

The  $K_d(\lambda)$  retrievals underpin the results shown in Figs. 2 and 3. Therefore, we assessed whether these retrievals were consistent with bio-optical relationships previously established for the low-latitude oceans under the form of the  $K_{\text{bio}}$  vs. Chl relationship, where  $K_{\text{bio}}$  is  $K_d - K_w$ , representing the contributions of all non-water components. The relationships for the low-latitude oceans are displayed in Fig. 4a, b, c, along with the MM01 model. The  $\chi$  coefficients and the exponents of the  $K_{\text{bio}}$  vs. Chl relationships are within 15 % of those from MM01 at 380 and 412 nm and differ by about 45 % at 490 nm. The  $r^2$  are accordingly decreasing from 0.5 at 380 to 0.33 at 490 nm. The slopes (exponents) of our relationships are lower than those from MM01. Despite these differences, these results show that the method used here can derive an overall consistent picture of the  $K_{\text{bio}}$  vs. Chl relationship for areas where it is well established. It is therefore supporting its use in the SO, where no such reference exists. Note that we cannot statistically assess the similarity between our relationships and MM01 because the data set that was used to derive the latter is no longer available.

The results of the SO are displayed in Fig. 4d,e,f. Here the Chl range is smaller than in the low-latitude data set, spanning from about  $0.05 \text{ mg m}^{-3}$  (very few points below this value) to  $3 \text{ mg m}^{-3}$ . The  $K_{\text{bio}}(\lambda)$  values do not follow the same decreasing trend as for the low-latitude oceans in the low Chl range ( $< 0.2 \text{ mg m}^{-3}$ ). The MM01 relationships seem to fit our data quite well for  $\text{Chl} > \sim 0.5 \text{ mg m}^{-3}$ . They do not match the data at lower Chl values, and the fit using a function of the form  $K_{\text{bio}}(\lambda) = \chi \text{Chl}^e$  as in MM01 also fails to capture the curvature in this range. A better fit is obtained with a formulation similar to the one used for  $a_p$  (Eq. 5), displayed as the white curves in Fig. 4d, e, f, showing a low dependence of  $K_d$  on Chl below  $\text{Chl} \sim 0.2 \text{ mg Chl m}^{-3}$ . The slopes of the linear fits (on log-transformed data) for the low-latitude waters are statistically different from those of the SO data ( $t$ -test) at 380 and 412 nm but are not at 490 nm, where uncertainties in deriving  $a_y$  are larger.

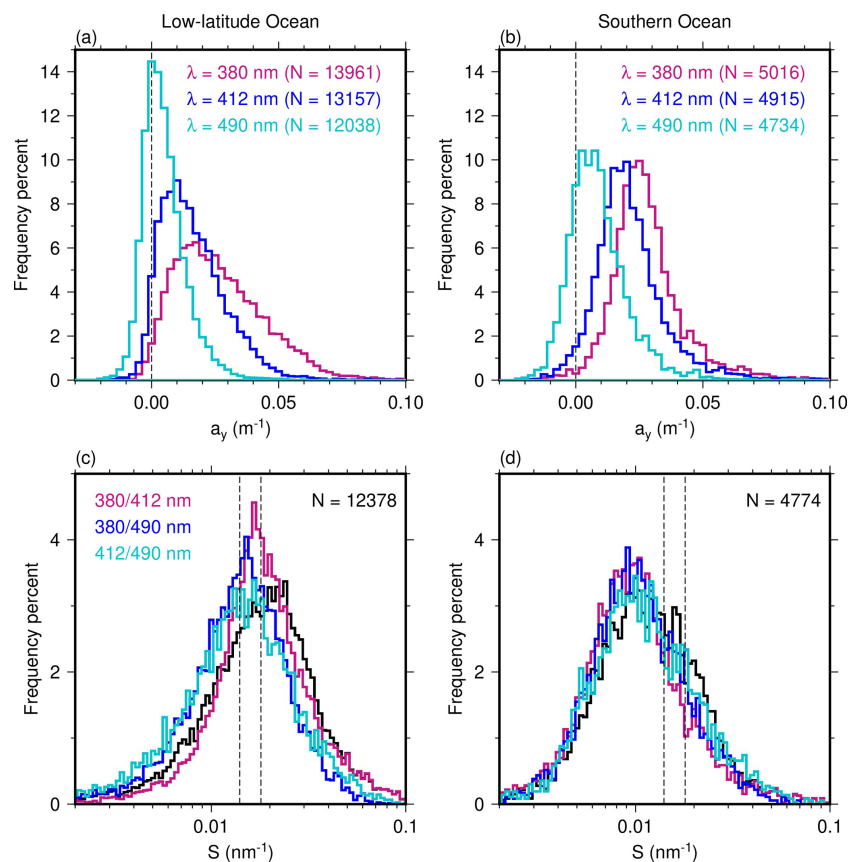
#### 3.3 $a_y(\lambda)$ vs. Chl relationships

Similarly to  $K_{\text{bio}}(\lambda)$ , we analyzed  $a_y$  as a function of Chl (Fig. 5). The relationships we obtained for the low-latitude areas are similar to those proposed by Morel and Gentili (2009), except for  $\lambda = 490 \text{ nm}$ , where the dispersion of the  $a_y$  values is the largest, as expected from the methodology. Therefore, results at this wavelength must be considered with caution. Given that MG09 was originally developed at 400 nm and subsequently extended to other wavelengths using a spectral slope of  $0.018 \text{ nm}^{-1}$ , and our  $a_y$  at 412 nm is the closest match to 400 nm, here we compare it with MG09 at 412 nm to minimize the potential discrepancy that might occur from wavelength conversions involving larger spectral distance. In low-latitude waters, MG09 generally aligns

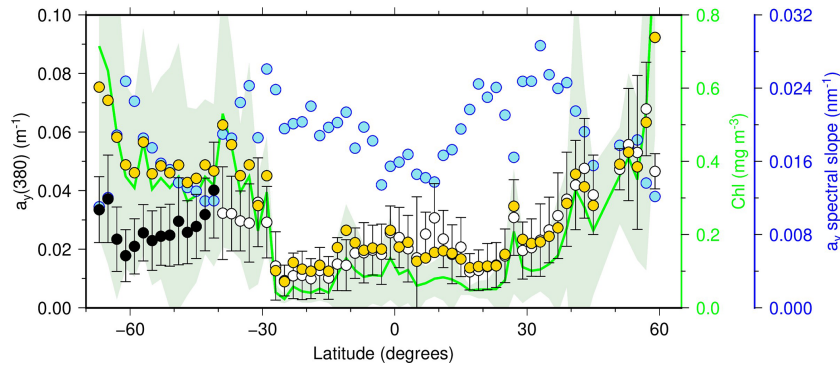
**Table 1.** Nominal individual uncertainties used in the Monte Carlo method.

| Parameter                 |                | Wavelength ( $\lambda$ ) |                        |                        | Comments  |
|---------------------------|----------------|--------------------------|------------------------|------------------------|---|
|                           |                | 380 nm                   | 412 nm                 | 490 nm                 |   |
| $K_d(\lambda)$            | R <sup>1</sup> | ←                        | 30 %                   | →                      | Jamet et al. (2012)   |
| $\mu_d(\lambda)$          | A              | ←                        | 0.1                    | →                      | Twice the standard deviation of $\mu_d$ values calculated for all profiles  |
| Chl                       | R              | ←                        | 35 %                   | →                      | Moore et al. (2009)   |
| $\text{Const}(\lambda)^2$ | A              | 0.00063                  | 0.00056                | 0.00038                |   |
| $\chi(\lambda)^2$         | A              | 0.0018                   | 0.0016                 | 0.0010                 | From the nonlinear regressions in Fig. S4   |
| $e(\lambda)^2$            | A              | 0.075                    | 0.057                  | 0.046                  |   |
| $b_{bp}(\lambda)$         | R              | ←                        | 20 %                   | →                      | Standard deviation in deep values   |
| $a_w(\lambda)$            | A              | 0.0008 m <sup>-1</sup>   | 0.0005 m <sup>-1</sup> | 0.0005 m <sup>-1</sup> | Lee et al. (2015)   |
| $b_{bw}(\lambda)$         | A              | ←                        | 0                      | →                      | Considered negligible (changes are $< 1 \times 10^{-5} \text{ m}^{-1}$ for changes in $T$ and $S$ of 5° or 5 psu, for instance) |

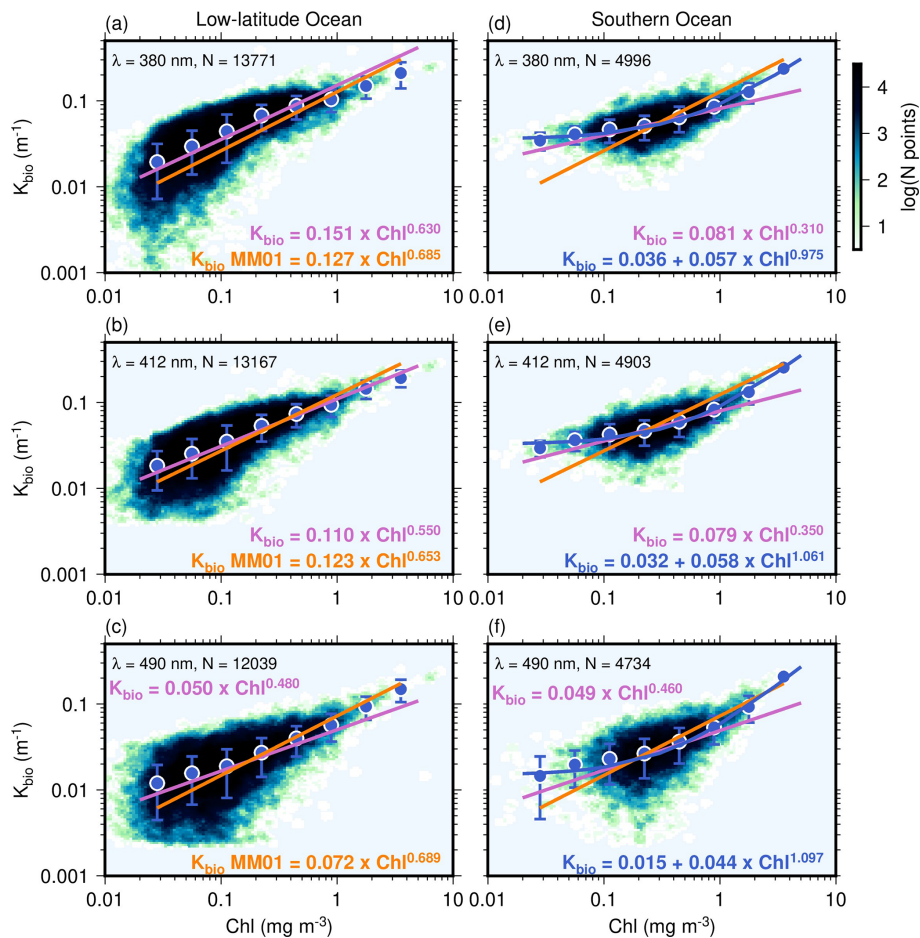
<sup>1</sup> R or A in the second column indicate either a relative or absolute uncertainty. <sup>2</sup> See Eq. (5) for  $a_p$  vs. Chl.



**Figure 2.** Distributions of  $a_y$  as derived from the BGC-Argo data at the three wavelengths indicated and for the low-latitude Ocean (a) and the SO (b). The corresponding spectral slopes are displayed in (c) and (d), both when separately calculated for the three wavelength pairs indicated and when these three estimates are averaged (black line). The dashed lines in (c) and (d) are the  $S$  values proposed by Bricaud et al. (1981) ( $0.014 \text{ nm}^{-1}$ ) and those used by Morel and Gentili (2009) ( $0.018 \text{ nm}^{-1}$ ).



**Figure 3.** Zonal averages and standard deviation of  $a_y(380)$  for  $2^\circ$  latitude bands, calculated across our entire data set (open symbols for latitudes  $> 40^\circ$  S and black symbols for latitudes  $< 40^\circ$  S). The gold symbols are the  $a_y(380)$  estimated from the MG09 relationship for the average Chl values (green curve, with the standard deviation shown as the green shade). The spectral slope of  $a_y$  is also displayed (blue symbols; second scale on the right).



**Figure 4.** Non-water diffuse attenuation coefficient for downward irradiance ( $K_{\text{bio}}$ ) for the three wavelengths indicated in the panels for the low-latitude oceans (left) and the SO (right) data sets. The blue-coloured density plots (scale on the top right) are built from all data obtained from individual float profiles. The large blue dots circled in white and vertical bars are average values and their standard deviation calculated over logarithmically equal Chl intervals. The purple and dark blue solid lines are a linear and a non-linear fits to all data points (log-transformed data; equations provided on each panel). The orange line for both the low-latitude oceans and the SO are for the Morel and Maritorena (2001) model (reported on the left panels as the “ $K_{\text{bio}}$  MM01” equation).

with our predicted  $a_y(412)$  vs. Chl relationship, apart from  $\text{Chl} > 3.0 \text{ mg m}^{-3}$ , where additional data is required for further assessment. This further confirms the validity of our float-based inversion approach. As previously said for  $K_{\text{bio}}$ , we cannot statistically assess the similarity between our relationships and MG09 because we do not have the data set that was used to derive the latter.

The BOUSSOLE data sit on the upper part of the data cloud and the BIOSOPE data rather in the middle of it, with some low values for low Chl, which is consistent with what has already been shown for the Mediterranean Sea and the Southeast Pacific gyre (Morel et al., 2007c). The NOMAD data are also on the high range. This consistency of the derived  $a_y$  with field measurements further validates the approach.

In the SO (Fig. 5d, e, f),  $a_y$  does not vary much across the whole Chl range, with slopes of the  $a_y$  vs. Chl relationships much lower than those of the low-latitude data set and the MG09 model (equations reported on each panel of Fig. 5). The regression coefficient of the relationship at 380 nm in low-latitude waters is 0.26, whereas for the SO it is less than 0.1 across all wavelengths. Confidence intervals and a t-test show that all slopes (the  $B$  exponent in the  $A \times \text{Chl}^B$  relationships) are statistically different from zero, showing that the dependence of  $a_y$  on Chl still exist but is weak for the SO.

Reynolds et al. (2001) have reported an  $a_y$  vs. Chl relationship for the Ross Sea and Antarctic Polar Front Zone, expressed as  $a_y(400) = 0.046 \text{ Chl}^{0.298}$  ( $r^2 = 0.55$ ,  $N = 55$ ). When extrapolated to other wavelengths using the spectral slope they got from their data set ( $S = 0.0195 \text{ nm}^{-1}$ ), the slopes of these  $a_y$  vs. Chl relationships sit between those of our relationships and those of MG09 (Fig. 5d, e, f).

### 3.4 Distribution of $a_y$ anomalies.

Figure 5 shows that the  $a_y$  vs. Chl relationship established for low-latitude oceans do not match the SO data. We did not find coherent spatial patterns of the difference between the  $a_y$  derived here in the SO and the values calculated from Chl following MG09.

These differences, hereafter referred to as anomalies (with respect to the model), however display a seasonal pattern (Fig. 6a; black dots), with small differences during austral winter (June–September) and large negative anomalies in summer. When only clear waters are considered ( $\text{Chl} < 0.2 \text{ mg m}^{-3}$ ; blue dots) the anomalies are small and do not exhibit the same seasonal pattern. A seasonal change is also observed in the  $a_y$  spectral slope (orange dots on Fig. 6), with higher values in summer that are close to the average values often considered for the low-latitude oceans ( $0.014 \text{ nm}^{-1}$ ; Bricaud et al., 1981), and lower average values in winter, down to about  $0.009 \text{ nm}^{-1}$ . These anomalies are plotted as a function of the MLD in Fig. 6b, showing the largest negative values for MLDs between about 50 and

100 m. These MLD values are typical of summer months, as shown in the insert of Fig. 6b (December to March/April).

### 3.5 Relative contributions of the absorption and scattering terms and uncertainties of $a_y$ estimates

The relative contributions of the five terms into which  $[(K_d \mu_d)/1.0395]$  is split (Eq. 4) were derived for the entire data set (Fig. 7a, b) and also calculated from bio-optical models (Fig. 7c). The larger the contribution of  $a_y$ , the lower the sensitivity of its derivation through Eq. (4) will be to the values of the other four terms. The first observation is that  $a_y$  shows the largest relative contribution at 380 nm, often around 50 % for both the low-latitude and SO waters. As expected from the spectral dependence of  $a_y$ , the contribution is smaller for longer wavelengths, with percentages ranging from about 30 % to 50 % at 412 nm and from about 20 % to 30 % at 490 nm (see Fig. S5). The large relative contributions for the shortest wavelengths creates favorable conditions to operate Eq. (4).

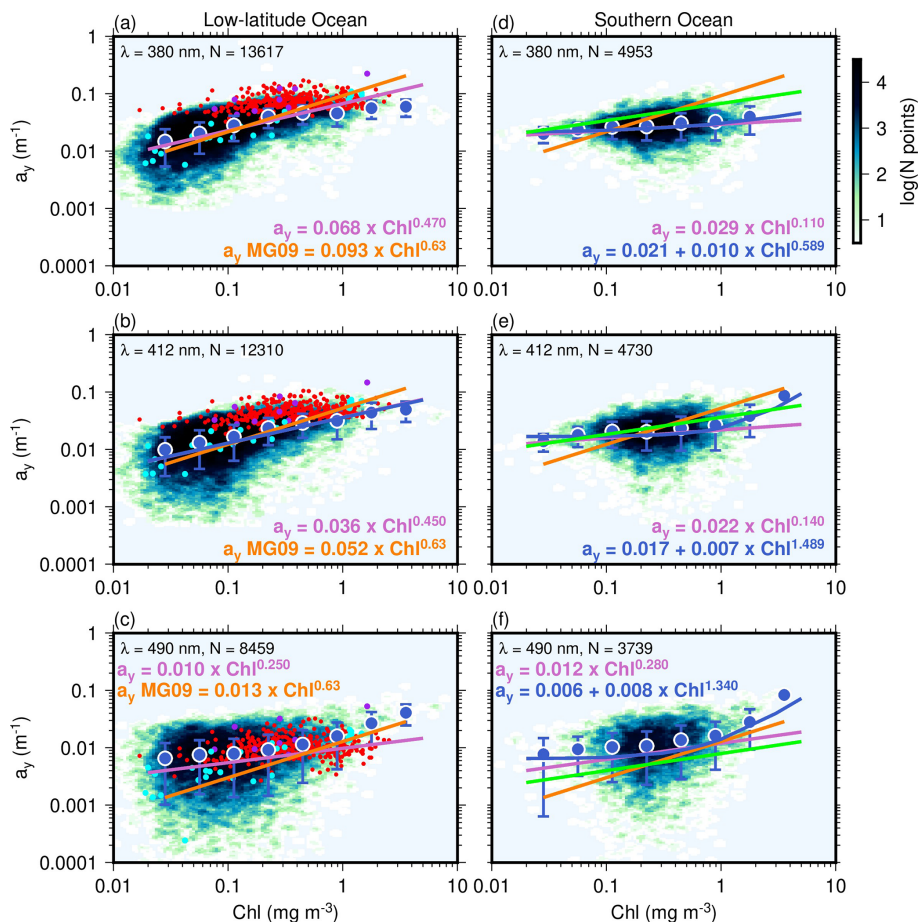
Figure 7 also shows that the relative importance of absorption by particulate matter for both the SO and the low-latitude oceans remains relatively constant around 20%–25 % for  $\text{Chl} < \sim 0.2 \text{ mg m}^{-3}$ , and then increases beyond this concentration to reach about 50 %. This is constrained here by the use of the Bricaud et al. (1998) parameterization and our Eq. (5).

The relative contribution of  $a_y(380)$  for the low-latitude oceans increases from about 30 % for the lowest Chl to  $\sim 60$  % for  $\text{Chl} \sim 0.25 \text{ mg m}^{-3}$ , similarly to what bio-optical models predict (dashed line on Fig. 7b). However, beyond  $\text{Chl} \sim 0.5 \text{ mg m}^{-3}$  the model and the observations evolve in opposite ways, the latter showing the relative contribution of CDOM decreasing to 40 %. For the SO waters, this contribution is steadily around 55 % for  $\text{Chl} < \sim 0.2 \text{ mg m}^{-3}$ , which is larger than for the low-latitude waters, and then regularly decreases down to 30 % when Chl is  $\sim 2 \text{ mg m}^{-3}$ . These changes for SO waters do not match what the bio-optical models predict over the entire range of Chl here considered.

There are several sources of uncertainty when deriving  $a_y$  from  $K_d$  using Eq. (4) without having concomitant measurements of the various parameters of the equation such as  $a_p$ . These uncertainties were assessed as described in section 2.6. At 380 nm in the SO, there is little sensitivity of the overall distribution of the derived  $a_y$  to different approaches to obtain  $K_d$  (Fig. 8a),  $\mu_d$  (Fig. 8b) and  $a_p$  (Fig. 8c).

Associated statistics are given in Table 2. As expected, uncertainties in  $K_d$  contribute the most to differences in the retrieved  $a_y$ , followed by  $a_p$  and  $\mu_d$ . Results are similar at 412 and 490 nm and for the low-latitude waters, except for  $a_p$  at 412 nm in low-latitude oceans. They show increasing sensitivity to the three parameters with increasing wavelength.

The results of the Monte Carlo analysis applied to the SO data set are displayed in Fig. 9, as the distribution of the coefficient of variation (CV, defined as  $100 \times$  standard devia-



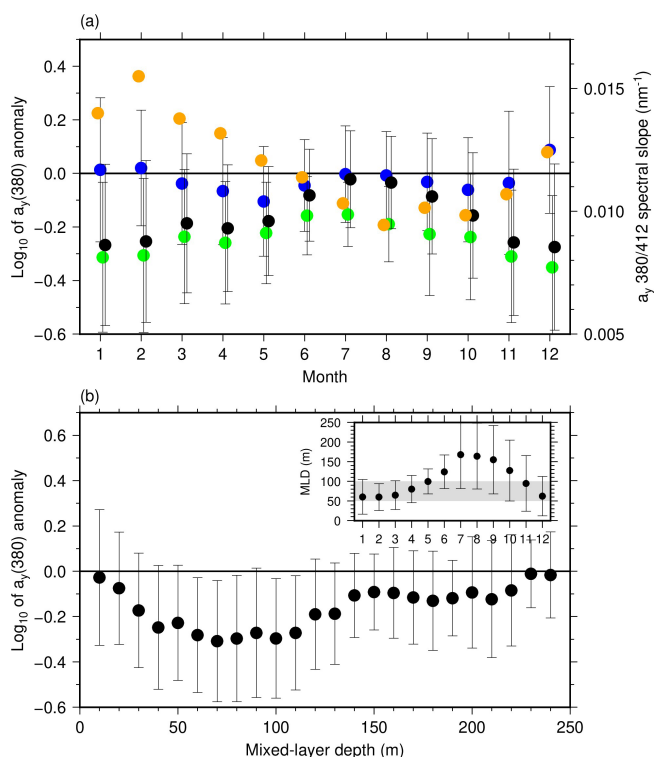
**Figure 5.** CDOM absorption ( $a_y$ ) for the three wavelengths indicated and for the low-latitude waters (left) and the SO (right) data sets. The blue-coloured density plots (scale on the top right) are built from all data obtained from individual float profiles. The large blue dots circled in white and vertical bars are average values and their standard deviation calculated over logarithmically equal Chl intervals. The purple and dark blue curves are a linear and non-linear fits to all data points (log-transformed data), the orange lines are from the Morel and Gentili (2009) model, whose equations are also reported as “ $a_y$  MG09”. The green lines are from Reynolds et al. (2001). In panels (a), (b) and (c), the coloured dots are in situ measurements of  $a_y$  from the BOUSSOLE site in the Mediterranean Sea (red dots), the BIOSOPE research voyage in the Southeast Pacific gyre (turquoise), and the NOMAD data set (purple) that covers various oceans.

tion divided by the mean) of  $a_y(\lambda)$  values obtained for each individual estimate of  $K_d(\lambda)$ . Each CV results from 10 000 runs of Eq. (4) using randomly picked errors on the individual terms of the equation (see methods). The modes of the histograms show that an uncertainty around 18 % can be generally expected for  $\lambda = 380$  and 412 nm, and 25 % for  $\lambda = 490$  nm. Cumulative curves (not shown) indicate that 85 % of uncertainties are lower than 20 % for  $\lambda = 380$  nm, 60 % at 412 nm and 20 % at 490 nm, reemphasizing that the band at 490 nm is far less adapted to deriving  $a_y$  from Eq. (4) than the two other bands.

## 4 Discussion

### 4.1 Uncertainties of $a_y$ estimates

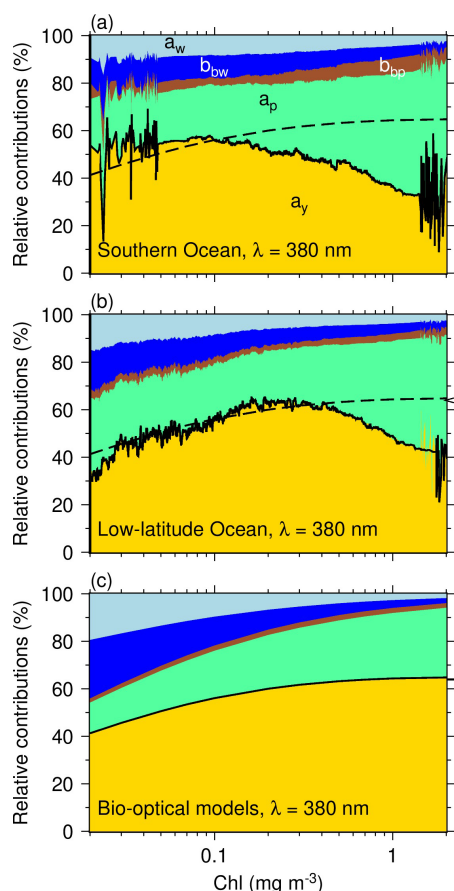
Since the estimation of  $a_y$  is based on a series of empirical and semi-analytical bio-optical relationships, it is subject to several sources of uncertainty. The individual sensitivity analyses (Fig. 8, Table 2) and the Monte Carlo analysis (Fig. 9) have shown that uncertainties on  $a_y$  are non negligible. Nevertheless, meaningful  $a_y$  vs. Chl relationships could be derived thanks to the large size and dynamic range of the float data set, and to the normal distribution of errors. Our uncertainty assessment did not consider possible systematic large biases in the initial  $K_d$  values. The results in Fig. 4, including the comparison with MM01, did not show evidence that such biases are present.



**Figure 6.** (a) Monthly average values (dots) and standard deviations (vertical bars) of  $a_y$  anomalies as a function of month of year (left scale) for the SO data. The anomalies are expressed as the decimal logarithm of the ratio of observed to modelled  $a_y$ , where the modelled values are from Morel and Gentili (2009). The blue dots are for  $\text{Chl} < 0.2 \text{ mg m}^{-3}$ , the green dots for  $\text{Chl}$  above that threshold, and the black dots for all data. The orange dots are the monthly average values of the spectral slopes calculated between 380 and 412 nm (right scale). (b) The same anomalies as in (a) plotted as a function of the mixed-layer depth (MLD), with the insert showing the seasonal course of the MLD in our data set. The greyed area shows MLD values between 50 and 100 m, corresponding to the largest negative anomalies.

The adjustment we applied to the  $\text{Chl}$  values for the low-latitude areas is another source of uncertainty. When it is not performed, the slope of the  $a_y$  vs.  $\text{Chl}$  changes slightly (e.g., from 0.57 to 0.45 at  $\lambda = 412 \text{ nm}$ ), yet the observation that  $a_y$  in the SO does not vary with  $\text{Chl}$  as strongly as it does in the low-latitude areas still holds.

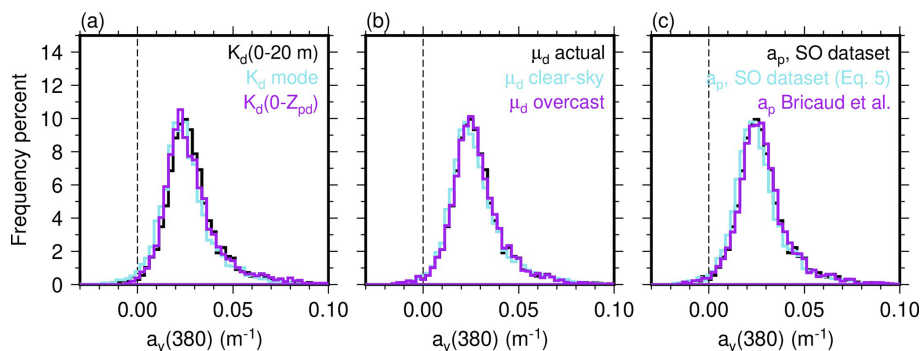
Considering the uncertainty on the Fluorescence-to- $\text{Chl}$  conversion factor, we tested the impact of using the Sauzède et al. (2025) lookup table that provides a global 1-degree resolution map of the fluorescence-to- $\text{Chl}$  ratio instead of using constant factors (3.79 is the SO and 2 elsewhere). In this sensitivity study, the BGC-Argo  $\text{Chl}$  data were re-multiplied by a factor of 2 and divided by the fluorescence-to- $\text{Chl}$  ratio from the lookup table corresponding to the location of the float. The results in terms of global distributions (Figs. 2 and 3) and relationships to  $\text{Chl}$  were not appreciably modified.



**Figure 7.** Relative contributions of  $a_w$  (light blue),  $b_{bw}$  (blue),  $b_{bp}$  (brown),  $a_p$  (green) and  $a_y$  (gold) to  $\frac{K_d(\lambda)\mu_d(\lambda)}{1.0395}$  (Eq. 4) at  $\lambda = 380 \text{ nm}$ , as a function of  $\text{Chl}$ . Panel (a) is for the SO, (b) is for the low-latitude Oceans, and (c) is when using Bricaud et al. (1998) to calculate  $a_p$ , MG09 for  $a_y$ , and MM01 for  $b_{bp}$ . The thick black line delineates the contribution of  $a_y(380)$  to the budget. This modelled relative contribution of  $a_y$  from panel (c) is reproduced in (a) and (b) as a dashed line. The increased noise in that curve for  $\text{Chl} < 0.03 \text{ mg m}^{-3}$  and  $\text{Chl} > \sim 1.5 \text{ mg m}^{-3}$  arises from the low numbers of retrievals in these ranges.

We opted not to use this lookup table, however, because it is based on  $\text{Chl}$  values derived from  $K_d(490)$  using a bio-optical model similar to MM01, which has been shown here not suitable for the SO.

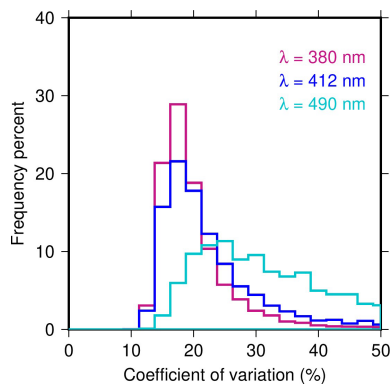
The way we calculated the  $a_y$  spectral slopes is sensitive to errors in individual  $a_y$  values. We nevertheless obtained average values for the low-latitude data set (Figs. 2c and 3) that are close to those generally considered as representative of open-ocean conditions (Bricaud et al., 1981; Morel and Gentili, 2009). The range of values shown in the general distribution of this slope (from about 0.008 to  $0.024 \text{ nm}^{-1}$ ; Fig. 3) is also consistent with previous studies (e.g., Wei et al., 2016; Aurin et al., 2018). Higher values in subtropical areas are expected because of the high impact of photobleaching in stratified waters.



**Figure 8.** Distribution of  $a_y(380)$  resulting from (a) three approaches to obtain  $K_d$  from  $E_d(z, 380)$ , (b) whether the distinction between clear and cloudy sky is applied when calculating  $\mu_d$  or ignored and then  $\mu_d$  being forced to either its clear sky or overcast sky value, and (c) using the three different  $a_p$  vs. Chl relationships displayed in Fig. S4. Data for the SO only.

**Table 2.** Average dispersion (%) of the mean  $a_y$  values with respect to their average for the three instances of each sensitivity study and the three wavelengths. For each parameter ( $K_d$ ,  $\mu_d$  and  $a_y$ ), the dispersion is calculated as the mean absolute difference among average values for this parameter for each of the three sensitivity studies, divided by the average value calculated for the three studies together.

|                     | 380 nm | 412 nm | 490 nm |
|---------------------|--------|--------|--------|
| SO                  |        |        |        |
| $K_d$               | 6.8    | 11.4   | 30.9   |
| $\mu_d$             | 1.5    | 1.1    | 1.1    |
| $a_p$               | 4.4    | 9.6    | 31.7   |
| Low-latitude oceans |        |        |        |
| $K_d$               | 6.4    | 13.5   | 45.0   |
| $\mu_d$             | 1.1    | 1.3    | 3.7    |
| $a_p$               | 7.0    | 0.8    | 21.4   |



**Figure 9.** Distribution of the coefficient of variation of  $a_y(\lambda)$  values obtained by running the Monte Carlo analysis on each of the individual estimates of  $K_d(\lambda)$ .

## 4.2 Comparison with bio-optical models and implication on ocean colour remote sensing

Differences in  $a_y$  vs. Chl relationships between the SO and low-latitude waters may lead to biases when using standard ocean color algorithms to estimate Chl in the SO. Figure 7 shows that the MG09 and Bricaud et al. (1998) bio-optical models predict an overall relative contribution of  $[a_y + a_p]$  to  $K_d(\lambda)\mu_d(\lambda)/1.0395$  (the yellow plus green areas combined) that is close to what we observe for both the SO and the low-latitude waters, except for low Chl values where the contribution from the models ( $\sim 55\%$ ) is lower than from the observations ( $\sim 70\%$ ). These differences do not seem high with respect to the method uncertainties. However, the relative contributions of  $a_y$  and  $a_p$  do not follow the modeled pattern. For low-latitude waters, the divergence starts when  $\text{Chl} > \sim 0.5 \text{ mg m}^{-3}$ , with the relative contribution of  $a_y$  then decreasing significantly towards larger Chl. Below this Chl threshold the predictions and observations are similar. For the SO, the relative contribution of  $a_y$  is slightly larger than what the models predict for  $\text{Chl} < \sim 0.15 \text{ mg m}^{-3}$ . Then the divergence with the model when Chl increases is much larger than what it is for the low-latitude areas.

These observations have implication on the quantification of Chl through satellite algorithms such as the OC4Me (Morel et al., 2007a), which is based on the MM01 bio-optical model, itself consistent with MG09 and Bricaud et al. (1998) used here (Morel, 2009). These bio-optical models underlying OC4Me assume a larger contribution of  $a_y$  than it is shown here for Chl values  $> \sim 0.5 \text{ mg m}^{-3}$ , which could lead to underestimation of Chl in that range when current satellite algorithms are used. If applied to the SO, the algorithm will underestimate even more the large Chl values (the actual relative contribution of  $a_y$  being even smaller), yet it will overestimate low Chl values, in this case because the assumed contribution of  $a_y$  is lower than it actually is. These expected Chl over- or underestimations are actually what several validation studies have shown, as outlined in

the introduction of this study (see also Dierssen and Smith, 2000).

### 4.3 Possible reasons for the different contribution of $a_y$ in the SO as compared to low-latitude waters

Multiple factors can contribute to the discrepancy in  $a_y$  contributions between the SO and low-latitude waters, including differences in source inputs, as well as variations in the corresponding physical and biological processes. Due to the lack of terrestrial input, CDOM in the SO mainly derives from local sources, through in situ biologically-mediated production and consumption in the euphotic zone and redistribution via horizontal and vertical circulation. The redistribution is driven by physical processes such as winter seasonal mixing, subductions, upwelling and storms that can either bring deep CDOM-rich waters to the surface ocean or on the contrary dilute CDOM through mixing of surface waters with CDOM-poor waters (Nelson and Siegel, 2013; Mannino et al., 2014). Among these factors, the deep winter mixing plays a critical controlling role on CDOM dynamics by vertically homogenizing the water column and entraining CDOM-rich deep waters into the surface layer, thereby resetting the upper-ocean CDOM inventory each year. This deep mixing replenishes relatively refractory CDOM at the surface, counteracts cumulative summer photobleaching and microbial alteration, and establishes a consistent winter baseline for CDOM concentration and optical properties. By simultaneously resupplying nutrients that fuel the spring phytoplankton bloom, the winter mixed layer also indirectly regulates subsequent biological CDOM production. As a result, the depth and intensity of winter mixing strongly govern the seasonal amplitude, optical signature, and interannual variability of the CDOM pool in the SO.

The SO is structured by a succession of oceanic fronts that tend to isolate water masses (Park et al., 2019), experiences seasonal sea ice melt that releases organic matter in surface waters (Ortega-Retuerta et al., 2010a; Norman et al., 2011), and is home of pronounced vertical mixing (Olbers and Visbeck, 2005; Hillenbrand and Cortese, 2006). These characteristics create highly heterogeneous environments that influence the sources, transformations, and distribution of CDOM. In addition, the phytoplankton communities in the SO exhibit distinct physiological adaptations to the extreme light-limited conditions, which likely alter their production and release of CDOM compared to those in more illuminated waters (Strzepek et al., 2019). Collectively, these factors introduce substantial variability into  $a_y$  dynamics and apparently weaken its direct coupling with Chl, making it difficult to predict  $a_y$  from Chl in these high-latitude waters.

The local production is related to a wide range of biological processes including viral lysis, bacterial degradation, phytoplankton excretion and zooplankton grazing (Bricaud et al., 1981; Nelson et al., 1998; Nelson and Siegel, 2002; Siegel et al., 2002; Matsuoka et al., 2013; Bonelli et al.,

2021). Loss mechanisms also determine the CDOM balance, including microbial consumption and photooxidation (Siegel et al., 2002; Nelson et al., 2007). Photobleaching is inefficient in winter due to the low incoming irradiation (Fichot et al., 2023). What Fig. 6 shows, however, is that photobleaching is likely significant in summer when the depth of the mixed layer is less than about 100 m and surface irradiance can be as high as it is in the equatorial belt (Campbell and Aarup, 1989). Photobleaching is expected to lead to an increase of the spectral slope of CDOM absorption (e.g., D'Sa and Kim, 2017). This increase is indeed observed in Fig. 6 and must be compensated by an effective decrease of the DOM pool to lead to the observed negative  $a_y$  summer anomaly.

We do not further speculate about other possible causes of the differences we observe between the SO and the low-latitude oceans. Complementary data or model outputs would be needed in complement to what autonomous BGC-Argo floats alone can provide.

### 4.4 Are departures unique to the SO or do they apply to the whole temperate Southern Hemisphere?

Figure 3 highlighted that the difference between the standard bio-optical model and the estimated  $a_y$  depart from one another for latitudes higher than  $30^\circ$  S. As noted above, this latitude also corresponds to where the fractional land (to ocean) contribution decreases rapidly. Whether this reflects the reduced impact of land contribution to the CDOM pool or another feature of temperate SO waters is unknown (note the near absence of BGC-Argo floats equipped with radiometers in the Pacific sector of the SO). However, the departure observed here may point to a much larger region where current bio-optical relationships are distinct at least with respect to CDOM. We also note that the CDOM index derived by Morel and Gentili (2009) does not seem to match our measurements in this region as it suggests higher than average concentration. Overall, the CDOM index often provides high values at high Chl in temperate waters, which fit with our observations in northern regions but not in the southern hemisphere.

### 4.5 Do Southern Ocean waters belong to Case 1 waters?

The concept of Case 1 vs. Case 2 waters (Morel and Prieur, 1977) has been instrumental by providing a global and consistent framework to quantitatively interpret satellite ocean colour observations. The concept is based on the observation that biological matter that drives bio-optical properties and hence ocean colour covaries with phytoplankton in open ocean waters, classified as Case 1 waters. This covariation only emerges, however, when a large dynamic range is considered, e.g., by pooling together data from various trophic levels and oceans. When the dynamic range is small, the correlation generally vanishes. In Case 1 waters, Chl can be used

as a single index of changes in ocean colour, which does not mean that it is the sole responsible for changes. Assuming this general co-variability when deriving empirical chlorophyll algorithms, for instance, does not require separate consideration of how the components of the biological matter individually correlate with Chl (e.g., phytoplankton, detrital matter, CDOM) (Siegel et al., 2005). Variability in these relationships is a large source of uncertainty in the Chl retrieval from satellite ocean colour and has led to questioning whether the concept itself was useful (Mobley et al., 2004). When semi-analytical algorithms are developed, however, phytoplankton, non-algal particles and dissolved substances can vary independently from Chl (e.g., Bricaud et al., 1998; Lee et al., 2002; Maritorena et al., 2002; Siegel et al., 2005; Morel and Gentili, 2009).

The observation from Fig. 7 that the relative contribution of  $[a_y + a_p]$  to  $K_d(\lambda) \mu_d(\lambda) / 1.0395$  predicted by the bio-optical models matches the observations supports the use of Chl to quantify the ocean colour signal following the Case 1 waters paradigm. While the CDOM concentration increases with Chl, this increase is not as strong as in other oceanic regions and the relative contribution starts decreasing at lower Chl ( $\sim 0.15 \text{ mg m}^{-3}$  for the SO and  $0.5 \text{ mg m}^{-3}$  for low-latitude waters). As such while the SO would be expected to be “prototypical” case 1 waters with minimal influence of land and strong influence of biology, other factors – likely physical – have a strong impact on the weak relationship between Chl and CDOM. As a consequence, the relative contribution of CDOM to absorption in the SO, hence to the ocean colour signal, is larger than predicted by the bio-optical models here considered for  $\text{Chl} < \sim 0.1 \text{ mg m}^{-3}$  (55 % instead of 40 % for the lowest Chl) and much lower for Chl above that value (30 % instead of 60 % for  $\text{Chl} = 2 \text{ mg m}^{-3}$ ). This is coherent with the observed underestimation of Chl in that range by current satellite ocean colour algorithms. Improved retrievals of Chl from satellite ocean colour observations over the SO will require revision of how CDOM absorption is parameterized.

**Data availability.** Publicly available datasets were analysed in this study. The BGC-Argo data (<https://biogeochemical-argo.org>, last access: 2 May 2026) were obtained from the Biogeochemical-Argo database accessed through the Coriolis GDAC Ftp site: <ftp://ftp.ifremer.fr/ifremer/argo> (last access: 2 May 2026). These data were collected and made freely available by the International Argo Program and the national programs that contribute to it (<https://argo.ucsd.edu>, last access: 2 May 2026, <https://www.ocean-ops.org>, last access: 2 May 2026). Data collected during cruises are available from Zenodo: <https://doi.org/10.5281/zenodo.3993096> (Antoine et al., 2021), <https://doi.org/10.5281/zenodo.3816726> (Antoine et al., 2020), <https://doi.org/10.5281/zenodo.3660852> (Haumann et al., 2020).

**Supplement.** The supplement related to this article is available online at <https://doi.org/10.5194/bg-23-3073-2026-supplement>.

**Author contributions.** Juan Li: Conceptualization (equal), data curation (equal), formal analysis (equal), investigation (equal), methodology (equal), software (equal), visualization (equal), writing – original draft (lead), writing – review & editing (equal). David Antoine: Conceptualization (equal), data curation (equal), formal analysis (equal), methodology (equal), software (equal), visualization (equal), funding acquisition (lead), resources (lead), supervision (lead), writing – original draft (lead), writing – review & editing (equal). Yannick Huot: Conceptualization (equal), supervision (supporting), writing – review & editing (supporting).

**Competing interests.** The contact author has declared that none of the authors has any competing interests.

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**Financial support.** Juan Li was supported by the Australian Research Council Special Research Initiative, Australian Centre for Excellence in Antarctic Science (ACEAS; project number SR200100008). ACE was funded by Ferring Pharmaceuticals with additional support from the Swiss Polar Institute. Funding from the Australian Research Council Discovery Program (DP160103387) contributed to the exploitation of the ACE data. The SOLACE voyage was supported by a grant of sea time on RV Investigator from the CSIRO Marine National Facility (<https://ror.org/01mae9353>, last access: 2 May 2026).

**Review statement.** This paper was edited by Huixiang Xie and reviewed by Emmanuel Boss and one anonymous referee.

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