We respond to the reviewer's comments and provide answers to the remarks made in the annotated file.

Dear Authors ,

I am sorry to be dense but I still find this (revised) manuscript to be significantly lacking

For one, light limitation should be based on **absolute** light flux, not relative. Hence, a mixing layer much deeper than a specific light horizon can support phytoplankton growth as the cells are experiencing a large range of light (your I should be compared to an absolute flux horizon).

It is correct that the calculated value of irradiance is not an absolute light flux. It is calculated as a fraction of the PAR (equation 3) which is estimated to be proportional to the short-wave radiation flux, SW, assuming a constant ratio: PAR=0.473 SW. Hourly values of PAR are calculated from hourly SW measurements on the Azur meteorological buoy. This variability of the data on a short time scale should allow a proper understanding of the mechanisms responsible for the interpretation of the onset of surface phytoplankton growth on March 18, as indicated by the decrease in DIC and the increase in chlorophyll (Fig.5).

Secondly, I understand that you are convinced that no net accumulation occurs before your measurements. However, even your measurements presented in Fig. 2 suggested you missed a biomass accumulation period.

You are right. We have rectified the text: "During periods of a few days, from March 7 to April 5, increases in glider Chla were observed as the result of a series of net accumulation/mixing events. Following the initial growth of surface phytoplankton identified by the DIC decrease on March 18, biomass accumulation was observed a few days later (Fig. 2e) .The same was observed in the Southern Ocean [Pelicherro et al, 2019]."

Rather than argue you capture THE onset, you could describe the spring bloom as starting with a series of net accumulation/mixing events which eventually **lead** to a stratified water column. While not necessarily capturing it all, you captured some of these events convincingly.

It is true that we have only studied a few events during the March-April periods in the years 2016-2019. In all cases, together with a decrease in DIC, we observed the development of a mixing layer in the upper part of the water column. This stratification is the result of the change in atmospheric conditions at the air-sea interface, namely the change in sign of the heat flux and the low wind speed.

Since you are not using 'traditional' metrics of phytoplankton to define the bloom, you should say so in the abstract.

It is now stated lines 27-30 in the abstract.

Satellite imagery is available at daily scale (and can be composited from the 5 satellite

available in your period of measurements). No need to use an 8-day product and then say it is not working.

Our sentence is written from a general point of view regarding the use of satellite data on a global scale.

I am attaching an annotated PDF with more comments that I had as I read the manuscript.

We have answered the comments in the attached .pdf document.

Dear authors, I am often wrong. If you feel my comments are 'off the mark' please contact me directly and I will be more than happy, if convinced, to change my comments. All the best, Emmanuel

1 **Physical mechanisms for biological carbon uptake during the onset of the spring**

2 **phytoplankton bloom in the northwestern Mediterranean Sea (BOUSSOLE site)**

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20 Abstract

21 Several trigger mechanisms have been proposed for the onset of the phytoplankton spring

22 bloom. Among these, that phytoplankton cells begin to bloom when they experience higher

23 average light levels in shallower mixed layers, a result of the surface net heat fluxes becoming

24 positive and wind strength decreasing. We study the impact of these two forcings in the

25 northwestern Mediterranean Sea. We take advantage of hourly measurements of oceanic and

26 atmospheric parameters collected at two neighboring moorings during the months of March

27 and April in the years 2016 to 2019, combined with glider data in 2016. The onset of the

surface phytoplankton growth is concomitant with the start of significant biological activity detected by a sudden decrease in dissolved inorganic concentrations derived from measurements in the upper 10 m of the water column. A rapid reduction in wind stress following high-wind events is observed at the same time. A resulting shallow mixing layer favors carbon uptake by phytoplankton lasting a few days. Simultaneously, the air-sea net heat flux switches from negative to positive linked to changes of the latent air-sea heat flux, which is proportional to the wind speed. This results in an increased thermal stratification of the ocean's surface layers. In 2016, glider data show that the mixing layer is significantly shallower than the mixed layer at the onset of the surface phytoplankton bloom. We conclude that decreases in the mixing and mixed layer depths lead to the onset of the phytoplankton bloom due to the relaxation of wind speed following storms. We estimate net daily community production in the mixing layer over periods of 3 days between 2016 and 2019 as between 38 mmol C m^{-2} and 191 mmol C m^{-2}. These results have important implications as biological processes play a major role in the seasonal evolution of surface pCO2 and thereby the rate of reduction of atmospheric CO_2 by exchange at the air-sea interface.

Introduction

Investigating the physical and biological mechanisms contributing to changes in air-sea carbon dioxide (CO_2) fluxes is essential. It is a critical component of the global carbon cycle and the climate system with the ocean removing about a quarter of the CO_2 emitted into the atmosphere by human activities over the last decade. It depends on the gradient of CO_2 partial pressure, pCO$_2$, between the atmosphere and the surface ocean [Takahashi et al., 2009; Rödenbeck et al., 2013; Merlivat et al., 2015]. pCO$_2$ in the ocean mixed surface layer depends on the relative contribution of dynamic transport, thermodynamics and biogeochemistry. As the biological carbon pump lowers pCO$_2$ and Dissolved Inorganic Carbon (DIC) at the sea surface, the production of plankton biomass during a bloom can be sufficient to reverse the annual cycle of surface-ocean CO_2 uptake. A decrease in surface DIC, after physical processes have been eliminated, indicates the beginning of phytoplankton spring growth. Conceptual frameworks have been proposed to explain when depth-integrated

59 phytoplankton biomass starts accumulating during blooms [Behrenfeld, 2010; Sverdrup,
60 1953]. The timing of the initiation of the surface spring phytoplankton bloom depends in
61 particular on atmospheric forcing. The physical processes of wind stress, heat flux and
62 vertical mixing control the depth of the mixed/mixing layer and thus the availability of light
63 [Siegel et al, 2002, Chiswell, 2011; Taylor and Ferrari, 2011; Brody and Lozier, 2015;
64 Enriquez and Taylor, 2015, Rumyantseva et al, 2019]. Thus, to better understand which
65 fundamental physical processes control the timing of the onset of the surface bloom, in situ
66 measurements are needed at the appropriate hourly to daily timescale. It is important to draw
67 attention to the atmospheric forcing timescale. In biogeochemical ocean modeling studies,
68 typically built on atmospheric models, reanalysis or satellite data, time resolution of at best a
69 few days are required [Mahadevan et al, 2012].
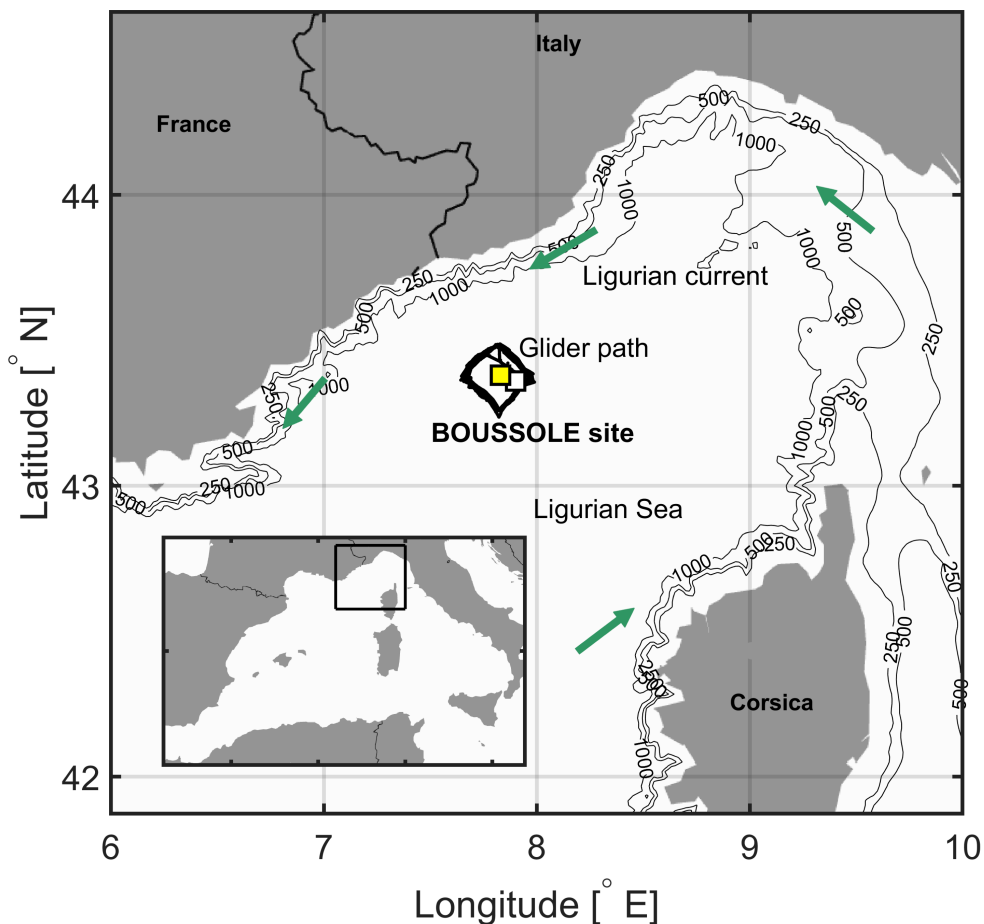
70

71 Figure 1. The area of the northwestern Mediterranean Sea showing the French and Italian
72 Rivieras, the island of Corsica, the main current branches (green arrows) in the Ligurian Sea
73 and the location of the BOUSSOLE buoy (white square) and the AZUR meteorological buoy
74 (yellow square). The black line indicates the glider's path in March-April 2016.

In the northwestern Mediterranean Sea, the BOUSSOLE (Bouée pour l'acquisition de Séries Optiques à Long Terme; Antoine et al., 2006, 2008a,b) buoy is located in the central area of the Liguro-Provençal basin (Fig.1), where the dominant ocean currents are generally weak (<20 cm s^{-1}) [Millot, 1999; Antoine et al., 2008b; Niewiadomska et al., 2008]. The main processes that govern the pCO$_2$ variations at the sea surface on an annual scale, after removal of the temperature effect, are vertical convective mixing and biological production [Begovic and Copin-Montégut, 2002; Copin-Montégut al., 2004; Hood and Merlivat, 2001; Merlivat et al., 2018]. Intense convection resulting from repeated high wind events in winter or early spring when atmospheric temperatures are typically low brings nutrients to the surface layer [Andersen and Prieur, 2000; Antoine et al., 2008b; Marty et al., 2002; Pasqueron de Fommervault et al., 2015].

The objective of this study is to examine the impact of atmospheric forcing on surface phytoplankton bloom initiation in the northwestern Mediterranean Sea. To this aim we use concurrent hourly data of seawater carbon chemistry measured at depths less or equal to 10 m at the BOUSSOLE mooring [Merlivat et al, 2018] and atmospheric flux measurements at the Azur meteorological buoy 3.1 nm (5.8 km) away from BOUSSOLE in March and April 2016 to 2019. Additional measurements of phytoplankton chlorophyll-a (Chla), temperature and density were provided by an underwater glider deployed in March-April 2016. The initiation of biological carbon uptake is revealed by the sudden decrease in DIC measured at the BOUSSOLE buoy.

In the Material and Methods section, we provide the description of data processing, theoretical framework, and calculations used to evaluate mixing regimes in the water column and net community production. The results section presents measured atmospheric and biogeochemical parameters observed in March-April, and NCP estimates from 2016 to 2019. The Discussion section relates our results to previous findings and presents our conclusions regarding the drivers of phytoplankton bloom initiation.

2 Material and methods

2.1- Data from the BOUSSOLE mooring

CARIOCA sensors [Merlivat and Brault, 1995] installed on the BOUSSOLE buoy (43°22'N, 7°54'E; Fig.1) monitored hourly partial pressure, pCO$_2$, at both 3 and 10 m alongside salinity

and temperature (SEABIRD Scientific SBE 37 MicroCAT CTDs). Protocols for calibration of in situ measurements are described in [Merlivat et al, 2018]. CARIOCA measurements of pCO_2 and temperature are combined with total alkalinity (Alk) derived from salinity [Merlivat et al., 2018] to calculate DIC. The dissociation constants of Mehrbach et al., (1973) refitted by Dickson and Millero (1987) are used, as recommended by Alvarez et al. (2014) for the Mediterranean Sea. For the March-April months mean salinity is 38.3, corresponding with mean Alk equal to 2571.4 $\mu mol\ kg^{-1}$ [Copin et al, 2004]. The expected precision of the CARIOCA pCO_2, and DIC is 5 μatm, and 3 $\mu mol\ kg^{-1}$, respectively [Merlivat et al, 2018].

2.2- Glider observations

An iRobot Seaglider model 1KA (SG537) with an ogive fairing operated by the University of East Anglia (UEA) [see Hemming et al., 2017] was deployed during the BOUSSOLE monthly cruise (#169) [Golbol et al 2000] and recovered 1 month later with a ship of opportunity. Diving to 1000 m, 147 dives were completed between March 7 and April 5 2016 repeatedly-sampling a square-shaped path surrounding the BOUSSOLE buoy (Fig. 1) with corners pointing to cardinal directions. Out of 147 dives, a number of them contain data that are within 10 km of the BOUSSOLE mooring, i.e. when it was sampling in the eastern corner of the diamond path. The glider was equipped with a non-pumped SEABIRD Scientific SBE 9 CTD instrument package providing temperature and salinity, and a WET Labs Eco Puck sensor providing relative phytoplankton Chla fluorescence, factory calibrated into Chla concentration units ($mg\ m^{-3}$).

2.3- Azur meteorological buoy

The Azur meteorological buoy has been deployed since 1999 by the French weather forecast Agency (Meteo-France) at 43°23'N, 7°50'E (Fig. 1). It provides hourly measurements of surface wind speed extrapolated to 10 m height, U_{10} ($m\ s^{-1}$), net incoming and emitted long-wave radiation, incoming short-wave radiation, and the sensible and latent heat fluxes. These are needed to estimate the net air-sea heat budget at the sea surface (http://mistrals.sedoo.fr). U_{10} is used to calculate wind stress, τ, as follows: $\tau = \rho_a\ C_d\ U_{10}^2$, where ρ_a is air density (1.2 $kg\ m^{-3}$), and C_d is the drag coefficient (1.4 x 10^{-3}). Photosynthetically Active Radiation (PAR, mol photons $m^{-2}\ d^{-1}$) is related to solar irradiance (i.e. short-wave flux), SW, following PAR = 0.473 SW [Papaioannou et al., 1993].

2.4- Estimation of mixed and mixing layer depths

137 The mixing layer depth, Z_{mx}, is the upper part of a mixed layer of uniform density where
138 active turbulence occurs [Brainerd and Gregg 1995]. Several recent studies have highlighted
139 the significance of this layer to understand the phytoplankton phenology of phytoplankton
140 blooms [Brody and Lozier, 2015; Taylor and Ferrari, 2011]. Autonomous profiling floats
141 equipped with bio-optical sensors have also highlighted the importance of mixing rather than
142 mixed layers to characterize the onset and the development of a phytoplankton bloom. Wind-
143 driven and buoyancy-driven regimes control the characteristics and depth of active mixing
144 [Brody and Lozier, 2014, 2015, Ramuyantseva et al, 2015, Lacour et al, 2019, Pellichero et al,
145 2020]. Wind-driven mixing dominates in the presence of weak buoyancy forcing when heat
146 fluxes are small. In that case, the dominant mixing length scale is equal to the Ekman length
147 layer depth scale.

148 The daily mean mixed layer depth, Z_{ml} is derived for 2016 from glider vertical profiles of
149 temperature and salinity using the algorithm provided by Holte and Talley, (2009) estimated
150 using the potential density algorithm.

151 To estimate Z_{mx}, τ is used to calculate the Ekman layer depth, Z_{ek}, $Z_{ek}=\gamma\ u_*/f$, where γ is an
152 empirical constant (0.5), f is the Coriolis parameter (10^{-4} s^{-1}) over the BOUSSOLE area, and
153 u_* is the turbulent friction velocity, $u_* = \sqrt{\dfrac{\tau}{\rho_w}}$ (m s^{-1}), with ρ_w the density of the surface sea
154 water [Lacour et al, 2019]. When heat fluxes are large and positive, Enrikez and Taylor
155 (2015) proposed to express the mixing depth Z_{mx} in terms of the surface forcing, i.e. the
156 surface stress and a stabilizing surface buoyancy flux B_0, as follows:

157
$$\frac{1}{Z_{mx}^2} = \frac{f^2}{(C_3\ u_*)^2} + \frac{f\ B_0}{(C_4\ u_*^2)^2} \qquad (1)$$

158 With C_3 =1, C_4 =0.57, B_0 = Q λ g /c_p ρ_0, where Q is the net surface heat flux, λ the thermal
159 expansion coefficient (1.65 10^{-4} °C^{-1}), g the gravitational acceleration (9.81 m s^{-2}), c_p the
160 specific heat of water (4 10^{-3} J kg^{-1} °C^{-1}) and ρ_0 the pure water density (1000 kg m^{-3}).

161 2.5-Estimation of euphotic zone depth

162 The euphotic depth, Z_{eu} , is generally derived as the depth where PAR is 1% of its surface
163 value. Z_{eu} can be estimated from Chla at the surface using the method described by Lee et al.
164 (2007) based on the equation of Morel and Berthon (1989):

165
$$Z_{eu} = 34\ (Chla)^{-0.39} \qquad (2)$$

166 For the period between 2017 and 2019, GlobColour merged Chla products were used, which
167 are based on satellite observations with a resolution of 25 km, and a binning period of 8 days
168 (http://www.oceancolour.org). For 2016, glider surface Chla was used to derive Zeu instead
169 of satellite measurements. We excluded glider Chla between 05:00 and 20:00 UTC affected
170 by sunlight–induced fluorescence quenching, and we applied a depth-constant offset to glider
171 measurements using Chla water samples available on March 7.

172 2.6-Estimation of average mixing layer irradiance

173 We calculate the average mixing layer irradiance, I, function of the incident surface
174 irradiance, PAR, mixing layer depth, h, and the diffuse attenuation coefficient, K_d, estimated
175 from surface chlorophyll-a concentrations (Venables and Moore, 2010).

$$K_d = 0.05 + 0.057 \ Chla^{0.58}$$

176 $$I = \frac{PAR}{K_d h} (1 - e^{-K_d h}) \quad (3)$$

177

178 2.7 – Estimation of net community production

179 Biological Net Community Production rates, NCP (mmol m^{-2} day^{-1}), are calculated for
180 separate 3-day periods in March-April, 2016 to 2019 from DIC concentrations derived from
181 CARIOCA measurements, assuming that measurements in the top 10 m of the water column
182 are representative of an homogeneous mixing layer [Boutin and Merlivat, 2009; Merlivat et al,
183 2015; Pelicherro et al, 2020]. In the study area, current velocities are expected to be small
184 [Millot, 1999; Antoine et al., 2008; Niewiadomska et al., 2008], hence to estimate NCP, we
185 ascertain that the contribution of horizontal advection was negligible. Vertical mixing events
186 are identified by an increase in DIC resulting from upward movement of high DIC Levantine
187 Intermediate Water (LIW) at approximately 200 m depth [Copin-Montégut and Bégovic,
188 2002]. We isolated time periods when local physical processes were largely one-dimensional
189 in order to study changes in biological and chemical parameters that occurred during rapid
190 transitions from deep mixing to intermittent stratification. In 2016, over a four-day period,
191 March 18-21, the diurnal cycle of DIC values characterized by a maximum in the morning
192 followed by a minimum at the end of the day indicates the onset of organic matter formation.
193 The eddy diffusion term is negligible as the DIC gradient at the base of the mixing layer is
194 very weak, entrainment is not expected as changes in DIC are observed during periods of

195 stratification. For the identified periods, biological production and air-sea exchange are the
196 dominant processes responsible for daily changes in DIC.

197 NCP is thus estimated from temporal variation of the DIC integrated over Z_{mx} (DIC_{int}), and
198 corrected for the air-sea CO_2 flux contribution [Pelicherro et al, 2020]:

199 $$NCP = \left(\frac{\Delta DIC_{int}}{\Delta t}\right)_{bio} = \left(\frac{\Delta DIC_{int}}{\Delta t}\right)_{meas} - k \times s \times (pCO_{2atm} - pCO_{2sw}) \quad (4)$$

200 The first term in equation 4, $\left(\frac{\Delta DIC_{int}}{\Delta t}\right)_{meas}$, is the daily variation of DIC_{int} calculated over 3
201 days from the date identified as the start of phytoplankton bloom (t_0), whilst the remaining
202 part corresponds to DIC variability due to air-sea CO_2 exchange, where k is the gas transfer
203 velocity [Wanninkhof, 2014], s is the solubility of CO_2 in seawater [Weiss, 1974], and
204 pCO_{2sw} and pCO_{2atm} are respectively the partial pressures of CO_2, in µatm, in sea water and in
205 air [Merlivat et al, 2018].

206 3 Results

207 3.1- Description of the 2016 time-series
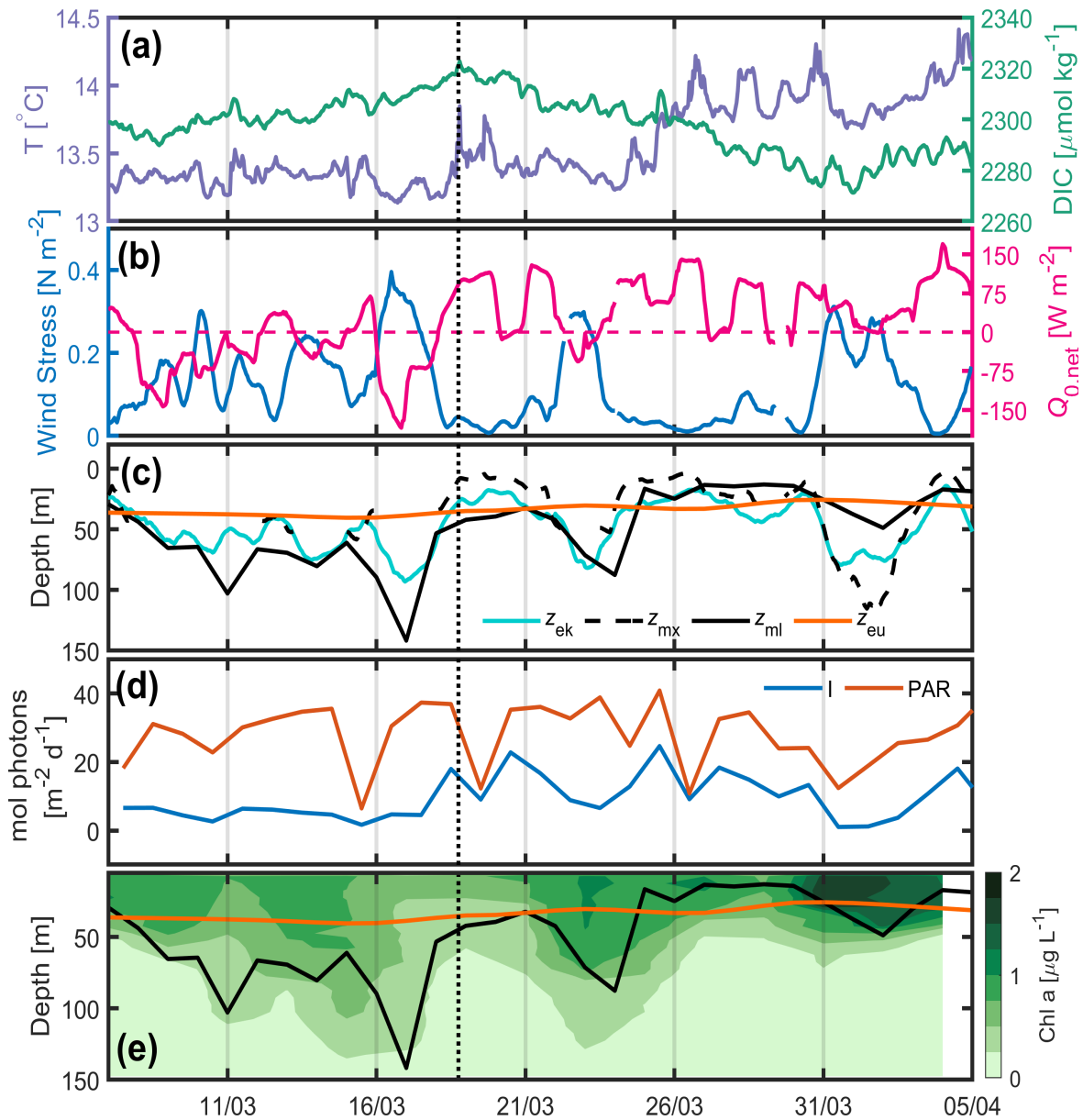
Figure 2. 2016 (a) Hourly DIC (green) and SST(purple) at the BOUSSOLE site, (b) hourly wind stress (blue) and net heat flux (pink) at the Azur buoy site. The pink dotted line indicates the change of sign of the net heat flux from negative to positive values, (c) daily depths of the mixed layer (black), mixing layer (dotted black), Ekman layer (cyan) and euphotic layer (orange) at the BOUSSOLE site. The black dotted line on panels (a), (b), (c) (d) and (e) indicates the onset time of the bloom (March 18). (d) PAR (orange) and I, average mixing layer irradiance (blue), (e) Glider-derived depths of the mixed layer (black curve, same as in (c)), euphotic layer (orange), and nighttime 2-day binned chlorophyll concentration (green coloring) retrieved between March 7 and April 5.

218   The temporal evolution of physical and biogeochemical parameters between March 7 and

219   April 5 2016 is displayed in Fig. 2. It shows an initial increase in DIC until March 18 (Fig.

220   2a), resulting from the vertical mixing of surface waters with the DIC-enriched underlying

221   LIW [Copin-Montégut and Bégovic, 2002], followed by a decrease in DIC that corresponds

222   to an increase of surface Chla (Fig 2e). Temperature increases intermittently over the same

223   time period (Fig. 2a), with a sharp increase on March 18 and 25. Before March 18 frequent

224   atmospheric fronts associated with wind forcing between 0.2 and 0.5 N m^{-2} were observed

225   (Fig. 2b). $Q_{0,net}$ was negative most of the time leading up to March 18, but mostly positive

226   afterwards (Fig. 2b). A clear anticorrelation was observed between τ and $Q_{0,net}$ (Fig. 2b). The

227   mixed-layer depth derived from the glider measurements (< 20 km away from the Boussole

228   site) deepened to > 70 m at times between March 8 and March 18, consistent with the passage

229   of frequent atmospheric fronts. From March 18 to April 5, Z_{ml} shoaled from ∼ 45 m to ∼10 m,

230   excluding two short periods around March 24 and April 2 when atmospheric fronts were

231   passing (Fig. 2c). From March 18 to 31, Z_{mx} was for a considerable time either shallower than,

232   or at similar depth to Z_{ml} and Z_{eu} (Fig. 2c). The start of the increase in PAR from March 15

233   precedes the increase in irradiance by 3 days as a result of the decrease of the mixing layer

234   depth observed only after March 18 (Fig. 2d**).**

235   From March 18 to April 5, an increase in glider Chla was observed in the upper layer (Fig.

236   2e) as the result of the biomass accumulation within the water column following the initial

237   growth of surface phytoplankton [Behrenfeld, M. and E. Boss, 2014]. The same was observed

238   in the Southern Ocean [Pelicherro et al, 2019].

239   3.2- Description of the 2017-2018-2019 time-series

240   Only measurements at the BOUSSOLE and Azur buoys were available between 2017 and

241   2019. Here we focus on the periods between March 1 and April 25. In 2017, from March 9,

242   after the passage of a very active atmospheric front (τ up to 2.7 N m^{-2}), a decrease in DIC

243   accompanied by an increase in temperature was observed for a period of 14 days despite a

244   few short bursts of vertical mixing, typically lasting less than 1 day (Fig.3a, d, g). At the same

245   time, the net surface heat flux switched from negative to positive most of the time. Over this

246   14-day period, Z_{mx} was predominantly shallower than the Z_{eu} at times when DIC was

247   decreasing.

248   In 2018 (Fig. 3b, e, h) and 2019 (Fig. 3c, f, i), the weather records throughout the month of

249   March are very similar, showing a succession of very strong wind events at the start of the

250 investigated time periods, with τ values of up to 3.7 N m^{-2} and Z_{ek} values as deep as 300 m.

251 On April 5, 2018, following a period of high wind stress resulting in the formation of a deep

252 mixing layer (~250 m), a decrease in DIC and an increase in SST were observed over 3 days.

253 At the same time, the net heat flux oscillated around zero. In 2019, CARIOCA data were only

254 available after March 24. On March 30, 3 days after high winds, a decrease in DIC and an

255 increase in temperature lasted 3 days, while simultaneously the net heat flux reversed from

256 negative to positive.

257
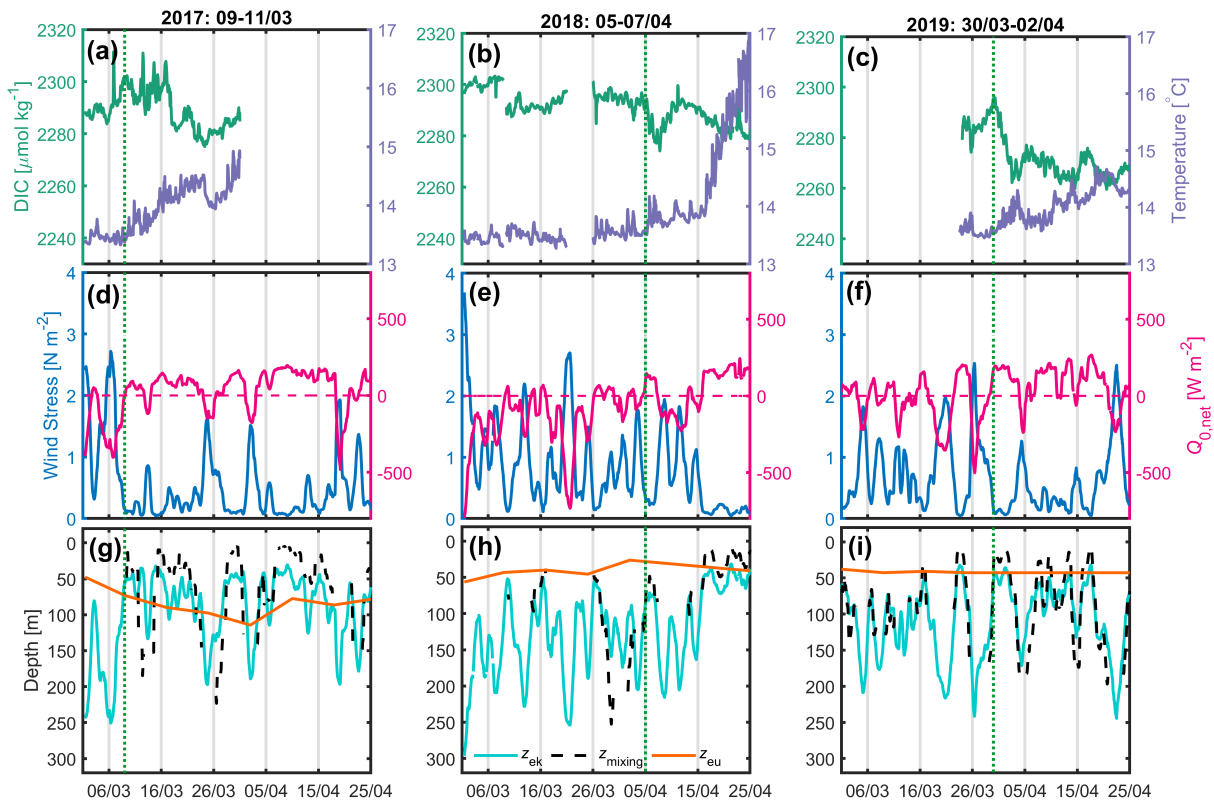
258

259 Figure 3. (a, b, c) 2017, 2018 and 2019 DIC (green) and SST (purple) at the BOUSSOLE site,

260 (d, e, f) wind stress (blue) and net heat flux (red) at the Azur buoy site. (d, e, f) The red dotted

261 line indicates the change of sign of the net heat flux from negative to positive values, (g, h, i)

262 depths of the mixing layer (dotted black), Ekman layer (cyan) and euphotic layer (orange) at

263 the BOUSSOLE site. The vertical dotted green line indicates the onset of the decrease of DIC.

264 For the period between 2017 and 2019, only satellite chlorophyll-a concentrations with a

265 binning period of 8 days were available, which is too large to calculate the average mixing

266 irradiance as it is highly variable on a daily scale (Fig 2.d).

11

267 3.3- NCP

268 NCP was estimated (Eq. 3) for 3-day periods over the four time-series between 2016 and
269 2019 (Table 1). Because we compute NCP only during periods when Z_{ml} and Z_{mx} shoal and
270 the air-sea CO_2 flux is very small, the measured total daily changes of DIC, dDIC/dt
271 (correlation coefficient, r^{*2} >0.70), is expected to represent the biological consumption of
272 DIC. In 2016, 2017, and 2019, daily biological carbon uptake fluxes are between 38 and 191
273 mmol m^{-2} d^{-1}. PAR and I vary respectively from 30 to 53 and 13 to 25 mol photons m^{-2} d^{-1} for
274 the four considered periods (Table 2).

	T_0	dDIC/dt [a]	r^{*2} [b]	ΔDIC / Δt [c]	r^{*2} [d]	air-sea flux CO_2 [e]	NCP [f]
		μmol kg^{-1}d^{-1}		mmol m^{-2}d^{-1}		mmol m^{-2}d^{-1}	mmol m^{-2} d^{-1}
2016 10m	March 18	-3.0	0.80	-37	0.68	1	38
2017 3m	March 9	-4.5	0.84	-188	0.86	3	191
2018 3m	April 5	-6.5	0.70			17	
2019 10m	March 30	-8.3	0.93	-163	0.63	4	167

275

276 Table 1. Net community production computed in the mixing layer over 3 days (NCP) after the
277 bloom onset (T_0). (a, b) variation of DIC at the measurement depth and linked correlation
278 coefficient; (c, d) change of integrated DIC over Z_{mx} and linked correlation coefficient ;(e)
279 CO_2 flux from the atmosphere to the ocean; (f), absolute value of calculated NCP is
280 considered in order to provide positive values throughout the presentation and discussion in
281 the manuscript.

282

283 4 Discussion

284 4.1 Analysis of the physical drivers of the bloom initiation over the 4 years, 2016-2019

285 4.1.1 In 2016, the consumption of carbon indicated by the decrease in DIC from March 18
286 attests the initiation of the formation of phytoplankton biomass (Fig. 2,). Between 2

287   successive days, March 17 and 18, a rapid decrease in daily wind stress from 0.24 to 0.04 N

288   m^{-2} results in a change of sign of the net heat flux from -68 to +56 W m^{-2} (Fig.4a) and a

289   shoaling Z_{mx}, which is for the most part shallower than Z_{eu} (Fig.4b) after March 18.

290
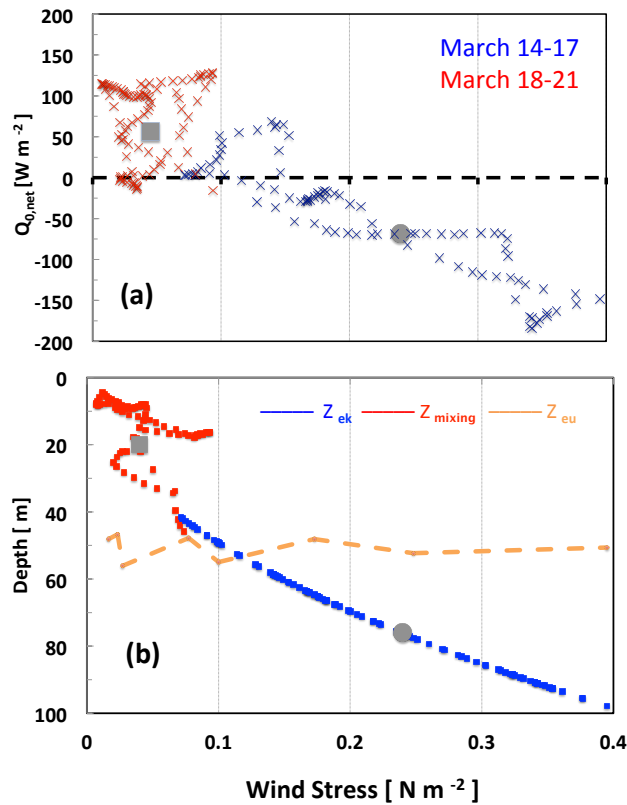
291   Figure 4. Changes of physical parameters (hourly values) at the onset of the 2016 bloom

292   during 2 consecutive periods of 4 days, March 14-17 (blue) and March 18-21 (red) as a

293   function of wind stress (a) net surface heat flux (b) depths of the Ekman, mixing and euphotic

294   layer. Grey circles and grey squares indicate mean values respectively on March 17 and

295   March 18. For comparison, the orange line shows the euphotic layer depth (March 14-21).

296   This illustrates the very fast change in heat flux and mixing layer thickness resulting from the

297   decrease in wind stress. The main drivers to explain the initiation of near surface spring

298   phytoplankton blooms proposed by [Taylor and Ferrari, 2011, Brody-Lozier, 2014, 2015] are

299   a decrease in wind strength, a positive surface heat flux, and a decrease in the dominant

300   mixing length scale. Our results indicate that a decrease in τ is a common component for these

301   last two mechanisms (Fig.4). The variability of $Q_{0,net}$ is strongly controlled by the variability

302   of the latent heat flux proportional to the wind speed. From March 17 to 18, the latent heat

13

303 flux changed by 84 W m^{-2} representing nearly 2 thirds of the total change of Q_0, with the
304 remainder originating from an increased short-wave flux. Hence, $Q_{0,net}$ increased as τ
305 decreased. In the same way Z_{mx}, which is related to τ and $Q_{0,net}$, decreased. This highlights the
306 major contribution of wind stress relaxation in controlling the net surface heat flux, and
307 subsequent stratification leading to sallower Z_{mx}

308 4.1.2 Over the 4 years, 2016-2019, during the months of March and April, at the very
309 beginning of spring, the conditions required for initiating the spring bloom are very similar
310 although the range of wind stress values is smaller in 2016 (Fig. 2, Fig. 3). The wind stress
311 values are low following storm events. For all years except 2018, Z_{mx} is smaller than Z_{eu}. The
312 average atmospheric and oceanographic conditions during the onset of the bloom are shown
313 in Table 2. The decrease of wind-driven mixing is then a dominant physical mechanism,
314 namely mixed layer shoaling which best predicts the timing of the spring bloom [Brody
315 Lozier, 2014, 2015]. Likewise, in the Southern Ocean, Pelicherro et al (2020) showed that a
316 bloom is initiated only when the wind-driven mixed layer decreases, even if the net heat flux
317 has been positive for some months.

4.2 DIC, Chla, Irradiance at the ocean surface in 2016

319 Between 18 and 21 March, a diurnal cycle in DIC is observed (Figure 5a) characterized by a
320 maximum in the morning followed by a minimum at the end of the day, indicating the onset
321 of organic matter formation. The decrease in DIC is accompanied by an increase of glider
322 surface Chla on 18 March and a simultaneous increase in the average mixing layer irradiance
323 (Fig. 5 b). It is worth noting that the Chla maximum does not occur until 31 March, 13 days
324 after the decrease in DIC. This maximum is the one identified in most cases by the Chla
325 satellite measurements with a binning period of 8 days. It follows that satellite data cannot
326 accurately characterize the conditions that prevail for the onset of a bloom, as these are
327 dictated by a great temporal variability of atmospheric forcings. Due to the seasonal
328 increase of surface irradiance in March-April, as well as the availability of nutrients following
329 vertical mixing with intermediate nutrient-enriched waters [Begovic and Copin-Montégut,
330 2002], the conditions for the onset of a bloom are met.

331

332
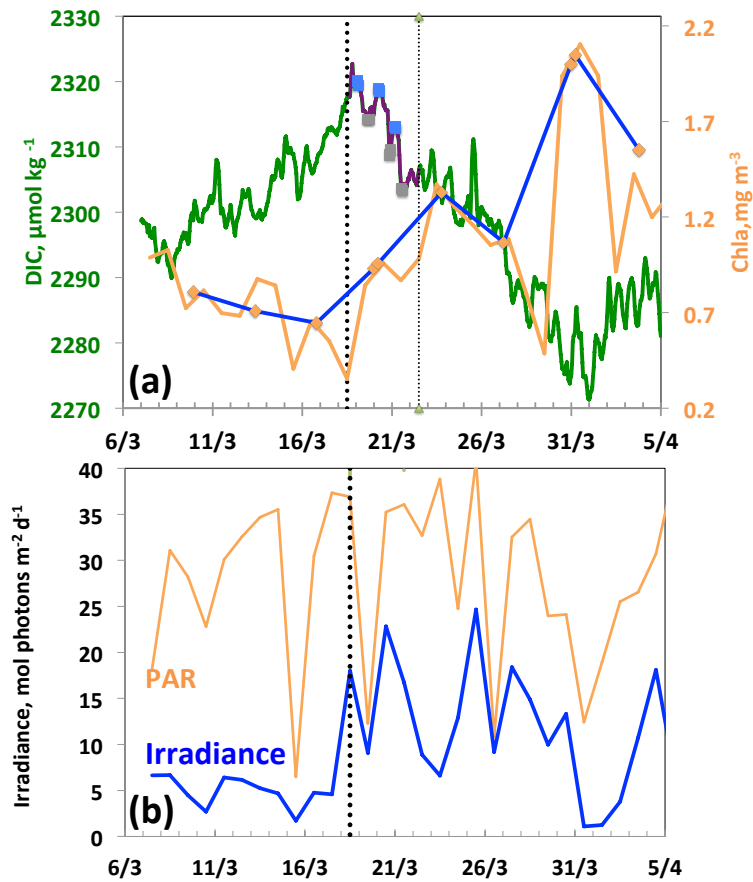
Figure 5. From March 7 to April 5, (a) DIC and Chla. The purple line and the squares (blue,morning; grey, evening) indicate the 3 days biological diurnal DIC changes during the period considered to compute NCP The blue and orange lines indicate the surface Chla when the glider was at a distance of less than 5 km (blue) and less than 20 km (orange) respectively from the Boussole buoy. (b) PAR and I average mixing layer irradiance. The vertical dotted black line indicates the onset of the bloom on March 18.

4.3- Biological carbon uptake

In 2016, 2017, and 2019, at the onset of the bloom Z_{mx} is shallower than Z_{eu} (Table 2). Two factors account for the observed large range of NCP variability at the time of triggering the bloom: dDIC/dt and Z_{ml}. In 2016 and 2017, DIC decreased over a period of 14 days, whereas it lasted for 3 days in 2018 and 2019, as a result of short-lived high wind events ($\tau > 1$ N m^{-2}). On April 2, 2018, a high surface phytoplankton Chla, 1.9 mg m^{-3}, was measured by satellite (Globcolour data) suggesting a strong biological signal. However, as Z_{eu} is significantly shallower than Z_{mx} (26 m vs. 65m; Fig. 3h), it would be incorrect to estimate NCP integrated over Z_{mx} using the DIC gradient measured at 3 m depth and likewise to compute the average mixing layer irradiance.

15

At the nearby DyFAMed site (Dynamique des Flux Atmospheriques en Mediterranee), using a 20-year O_2 time series, Coppola et al. (2018) estimated an annual net community carbon production equal to 7.1 mol m^{-2} yr^{-1}, or 19.4 mmol m^{-2} d^{-1}. Additionally, using determinations of primary production from ^{14}C in-situ incubations and of carbon export fluxes from sediment traps at DyFAMed, Marty and Chiaverini (2002) calculated average daily primary productions between 19.6 and 53 mmol m^{-2} d^{-1}. These estimates may have underestimated the annual NCP values as they were calculated from monthly observations while phytoplankton production varies on a daily timescale. Our DIC measurements highlight the large short-term variability of carbon consumption over a few days in March-April when bloom-triggering atmospheric conditions are met, such as a seasonal light increase and the availability of nutrients.

| | T_0 | Wind stress | Net heat flux | PAR | Z_{mixing} | $Z_{Euphotic}$ | Irradiance |
		N m^{-2}	W m^{-2}	mol photons m^{-2}d^{-1}	m	m	mol photons m^{-2}d^{-1}
2016	March 18	0.04 +/0.02	56+/-36	37+/-3	20+/-9	47	18
2017	March 9	0.12+/-0.02	45+/-21	30+/-3	28+/-12	71	13
2018	April 5	0.30+/-0.24	127+/-12	42+/-6	62+/-9	29	
2019	March 30	0.10+/-0.02	183+/-9	53+/-2	20+/-5	43	25

Table 2. 2016-2019. Atmospheric drivers of the onset of the bloom: average values and standard deviation on the day of the start of the bloom, T_0.

5- Concluding remarks

Around March and April in the northwestern Mediterranean Sea, the seasonal change in lighting conditions and atmospheric forcing are the essential parameters that control the triggering of near-surface phytoplankton blooms. These mechanisms have been highlighted through the coupling of high-frequency physical and biogeochemical in situ observations. For years 2016 to 2019, a bloom started in March or early April identified by a decrease in surface DIC, following a period of intense mixing. The rapid decrease in surface wind speed entailed a switch from negative to positive air-sea net heat fluxes, essentially due to changes in the

373 latent air-sea heat flux proportional to wind speed. These analysis support the hypothesis that
374 decreases in the depth of active mixing, a result of the transition from buoyancy-driven to
375 wind-driven mixing, control the timing of the surface spring bloom.

376 We have shown that the onset of surface phytoplankton growth, identified in 2016 by a
377 simultaneous initial decrease in DIC and an increase in surface Chla, precedes by a few days
378 the surface chlorophyll maximum as detected by satellite measurements with a binning period
379 of 8 days. This time span does not allow to identify precisely the contribution of atmospheric
380 drivers to trigger the onset of the growth of phytoplankton as it occurs on a daily basis.

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382

383 Data availability

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542

543   Figures and tables

544   Figure 1. The area of the northwestern Mediterranean Sea showing the French and Italian
545   Rivieras, the island of Corsica, the main current branches (green arrows) in the Ligurian Sea
546   and the location of the BOUSSOLE buoy (white square) and the AZUR meteorological buoy
547   (yellow square). The black line indicates the glider's path in March-April 2016.

548   Figure 2. 2016 (a) Hourly DIC (green) and SST (purple) at the BOUSSOLE site, (b) hourly
549   wind stress (blue) and net heat flux (pink) at the Azur buoy site. The pink dotted line indicates
550   the change of sign of the net heat flux from negative to positive values, (c) daily depths of the
551   mixed layer (black), mixing layer (dotted black), Ekman layer (cyan) and euphotic layer
552   (orange) at the BOUSSOLE site. The black dotted line on panels (a), (b), (c) (d) and (e)
553   indicates the onset time of the bloom (18 March). (d) PAR (orange) and average mixing layer
554   irradiance (blue), (e) Glider-derived depths of the mixed layer (black curve, same as in (c)),
555   euphotic layer (orange), and nighttime 2-day binned chlorophyll concentration (green
556   coloring) retrieved between March 7 and April 5.

557   Figure 3. (a, b, c) 2017, 2018 and 2019 DIC (green) and SST (purple) at the BOUSSOLE site,
558   (d, e, f) wind stress (blue) and net heat flux (red) at the Azur buoy site. The red dotted line
559   indicates the change of sign of the net heat flux from negative to positive values, (g, h, i)
560   depths of the mixing layer (dotted black), Ekman layer (cyan) and euphotic layer (orange) at
561   the BOUSSOLE site. The vertical dotted green line indicates the onset of the bloom.

562   Figure 4. Changes of physical parameters (hourly values) at the onset of the 2016 bloom
563   during 2 consecutive periods of 4 days, March 14-17 (blue) and March 18-21 (red) as a
564   function of wind stress (a) net surface heat flux (b) depths of the Ekman, mixing and euphotic
565   layer. Grey circles and grey squares indicate mean values respectively on March 17 and
566   March 18. For comparison, the orange line shows the euphotic layer depth (March 14-21).

567   Figure 5. From March 7 to April 5, (a) DIC and Chla. The purple line and the squares
568   (blue,morning; grey, evening) indicate the 3 days biological diurnal DIC changes during the
569   period considered to compute NCP The blue and orange lines indicate the surface Chla when
570   the glider was at a distance of less than 5 km (blue) and less than 20 km (orange) respectively
571   from the BOUSSOLE buoy. (b) PAR and I average mixing layer irradiance. The vertical
572   dotted black line indicates the onset of the bloom on March 18.

573   Table 1. Net community production computed in the mixing layer over 3 days (NCP) after the

574   bloom onset (T_0). (a, b) variation of DIC at the measurement depth and linked correlation

575   coefficient; (c, d) change of integrated DIC over Z_{mx} and linked correlation coefficient ;(e)

576   CO_2 flux from the atmosphere to the ocean; (f), absolute value of calculated NCP is

577   considered in order to provide positive values throughout the presentation and discussion in

578   the manuscript.

579   Table 2. 2016-2019. Atmospheric drivers of the onset of the bloom: average values and

580   standard deviation on the day of the start of the bloom, T_0.

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