



1 **Physical mechanisms for biological carbon uptake during the onset of the spring**
2 **phytoplankton bloom in the northwestern Mediterranean Sea (BOUSSOLE site)**

3 Liliane Merlivat¹, Michael Hemming², Jacqueline Boutin¹, David Antoine^{3, 4}, Vincenzo
4 Vellucci⁵, Melek Golbol⁵, Gareth A. Lee⁶, Laurence Beaumont⁷

5 ¹ Sorbonne Université, CNRS/IRD/MNHN, LOCEAN, IPSL, Paris, France

6 ² Coastal and Regional Oceanography Lab, School of Mathematics and Statistics, University
7 of New South Wales, Sydney, New South Wales, Australia

8 ³ Remote Sensing and Satellite Research Group, School of Earth and Planetary Sciences,
9 Curtin University, Perth, WA 6845, Australia

10 ⁴ Sorbonne Université, CNRS, Laboratoire d'Océanographie de Villefranche, Villefranche sur
11 Mer 06230, France

12 ⁵ Sorbonne Université, CNRS, Institut de la Mer de Villefranche, Villefranche sur Mer 06230,
13 France

14 ⁶ Centre for Ocean and Atmospheric Sciences, School of Environmental Sciences, University
15 of East Anglia, Norwich Research Park, Norwich NR4 7TJ, UK

16 ⁷ Division Technique INSU-CNRS, 92195 Meudon CEDEX, France

17

18 Correspondence to: liliane.merlivat@gmail.com

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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



28 phytoplankton growth is concomitant with the start of significant biological activity detected
29 by a sudden decrease in dissolved inorganic concentrations derived from measurements in the
30 upper 10 m of the water column. A rapid reduction in wind stress following high-wind events
31 is observed at the same time. A resulting shallow mixing layer favors carbon uptake by
32 phytoplankton lasting a few days. Simultaneously, the air-sea net heat flux switches from
33 negative to positive because of changes in the latent air-sea heat flux, which is proportional to
34 the wind speed. This results in an increased thermal stratification of the ocean's surface layers.
35 In 2016, glider data show that the mixing layer is significantly shallower than the mixed layer
36 at the onset of the phytoplankton bloom. We conclude that decreases in the mixing and mixed
37 layer depths lead to the onset of the phytoplankton bloom due to the relaxation of wind speed
38 following storms. We estimate net daily community production in the mixing layer over
39 periods of 3 days between 2016 and 2019 as between 38 mmol C m^{-2} and $191 \text{ mmol C m}^{-2}$.
40 These results have important implications on the oceanic carbon cycle and biological
41 productivity estimates in the Mediterranean Sea in a scenario of climate-driven changes of the
42 wind regimes.

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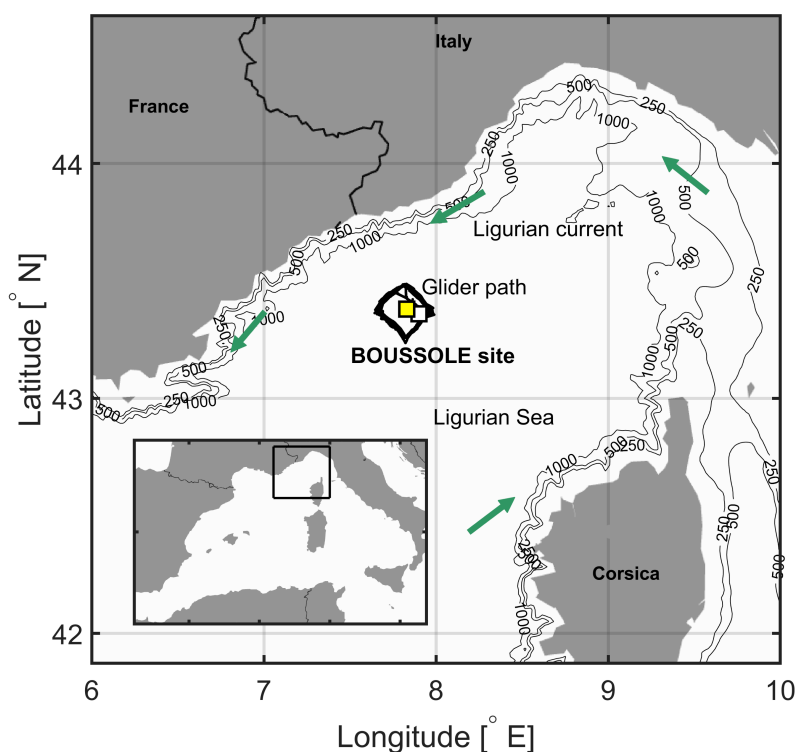
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45 Introduction

46 Investigating the biological and physical mechanisms contributing to changes in air-sea CO_2
47 fluxes and affecting carbon export from the surface to the interior ocean is essential to
48 quantitatively estimate the effects of future environmental change. The magnitude of the
49 ocean sink for atmospheric CO_2 can be assessed through direct estimates of air-sea net CO_2
50 fluxes. It depends on the gradient of CO_2 partial pressure, $p\text{CO}_2$, between the atmosphere and
51 the surface ocean [Merlivat et al., 2015]. $p\text{CO}_2$ in the ocean mixed surface layer depends on
52 the relative contribution of dynamic transport, thermodynamics and biogeochemistry. The
53 biological carbon pump lowers $p\text{CO}_2$ and Dissolved Inorganic Carbon (DIC) at the sea
54 surface, hence a decrease in surface DIC, after the impact of physical processes have been
55 eliminated, essentially indicates the beginning of the phytoplankton spring growth. Various
56 conceptual frameworks have been proposed to explain the onset of spring blooms [Behrenfeld,
57 2010; Sverdrup, 1953]. The timing of phytoplankton spring bloom initiation depends in
58 particular on the atmospheric forcing, through wind stress and heat fluxes [Chiswell, 2011;
59 Taylor and Ferrari, 2011; Brody and Lozier, 2015; Enriquez and Taylor, 2015, Rumyantseva



60 et al, 2019]. Thus, to better understand which fundamental physical processes control the
61 timing of the onset of the bloom, in situ measurements are needed at the appropriate hourly to
62 daily timescale. It is important to draw attention to the atmospheric forcing timescale. In
63 biogeochemical ocean modeling studies, typically built on atmospheric models, reanalysis or
64 satellite data, time resolution of at best a few days are required [Mahadevan et al, 2012].



65
66 Figure 1. The area of the northwestern Mediterranean Sea showing the French and Italian
67 Rivièras, the island of Corsica, the main current branches (green arrows) in the Ligurian Sea
68 and the location of the BOUSSOLE buoy (white square) and the AZUR meteorological buoy
69 (yellow square). The black line indicates the glider's path in March-April 2016.

70 In the northwestern Mediterranean Sea, the BOUSSOLE buoy (Bouée pour l'acquisition de
71 Séries Optiques à Long Terme; Antoine et al., 2006, 2008a,b) is located in the central area of
72 the Liguro-Provençal basin (Fig.1), where the dominant ocean currents are generally weak
73 ($<20 \text{ cm s}^{-1}$) [Millot, 1999; Antoine et al., 2008b; Niewiadomska et al., 2008]. The main
74 processes that govern the pCO_2 variations at the sea surface on an annual scale, after removal



75 of the temperature effect, are vertical convective mixing and biological production [Begovic
76 and Copin-Montégut, 2002; Copin-Montégut et al., 2004; Hood and Merlivat, 2001; Merlivat et
77 al., 2018]. Variability of atmospheric forcing over short timescales is greatest in winter or
78 early spring [Andersen and Prieur, 2000; Antoine et al., 2008b], generating deep convection
79 that brings nutrients to the surface layer [Marty et al., 2002; Pasqueron de Fommervault et al.,
80 2015].

81 The objective of this study is to examine the impact of atmospheric forcing on phytoplankton
82 bloom initiation in the northwestern Mediterranean Sea. To this aim we use concurrent hourly
83 data of seawater carbon chemistry measured at depths less or equal to 10 m at the
84 BOUSSOLE mooring [Merlivat et al, 2018] and atmospheric flux measurements at the Azur
85 meteorological buoy 3.1 nm (5.8 km) away from BOUSSOLE in March and April 2016 to
86 2019. Additional measurements of phytoplankton chlorophyll-a (Chla), temperature and
87 density were provided by an underwater glider deployed in March-April 2016. During this
88 time, the initiation of biological carbon uptake is revealed by the sudden decrease in DIC
89 measured at the BOUSSOLE buoy.

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

96

97 2 Material and methods

98 2.1- Data from the BOUSSOLE mooring

99 CARIOCA sensors [Merlivat and Brault, 1995] installed on the BOUSSOLE buoy (43°22'N,
100 7°54'E; Fig.1) monitored hourly partial pressure, $p\text{CO}_2$, at both 3 and 10 m alongside salinity
101 and temperature (SEABIRD Scientific SBE 37 MicroCAT CTDs). Protocols for calibration of
102 in situ measurements are described in [Merlivat et al, 2018]. CARIOCA measurements of
103 $p\text{CO}_2$ and temperature are combined with total alkalinity (Alk) derived from salinity [Merlivat
104 et al., 2018] to calculate DIC. The dissociation constants of Mehrbach et al., (1973) refitted by
105 Dickson and Millero (1987) are used, as recommended by Alvarez et al. (2014) for the



106 Mediterranean Sea. For the March-April months mean salinity is 38.3, corresponding with
107 mean Alk equal to $2571.4 \mu\text{mol kg}^{-1}$ [Copin et al, 2004]. The expected precision of the
108 CARIOCA pCO_2 , and DIC is $5 \mu\text{atm}$, and $3 \mu\text{mol kg}^{-1}$ respectively [Merlivat et al, 2018].

109 2.2- Glider observations

110 An iRobot Seaglider model 1KA (SG537) with an ogive fairing operated by the University of
111 East Anglia (UEA) [see Hemming et al., 2017] was deployed during the BOUSSOLE
112 monthly cruise (#169) [Golbol et al 2000] and recovered 1 month later with a ship of
113 opportunity. Diving to 1000 m, 147 dives were completed between March 7 and April 5 2016
114 repeatedly-sampling a square-shaped path surrounding the BOUSSOLE buoy (Fig. 1) with
115 corners pointing to cardinal directions. Out of 147 dives, a number of them contain data that
116 are within 10 km of the BOUSSOLE mooring, i.e. when it was sampling in the eastern corner
117 of the diamond path. The glider was equipped with a non-pumped SEABIRD Scientific SBE
118 9 CTD instrument package providing temperature and salinity, and a WET Labs Eco Puck
119 sensor providing relative phytoplankton Chla fluorescence, factory calibrated into Chla
120 concentration units (mg m^{-3}).

121 2.3- Azur meteorological buoy

122 The Azur meteorological buoy has been deployed since 1999 by the French weather forecast
123 Agency (Meteo-France) at $43^{\circ}23'N$, $7^{\circ}50'E$ (Fig. 1). It provides hourly measurements of
124 surface wind speed extrapolated to 10 m height, U_{10} (m s^{-1}), net long wave radiation, net short-
125 wave radiation, and the sensible and latent heat fluxes. These are needed to estimate the net
126 air-sea heat budget at the sea surface (<http://mistrals.sedoo.fr>). U_{10} is used to calculate wind
127 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
128 coefficient (1.4×10^{-3}). Photosynthetically Active Radiation (PAR, $\mu\text{mol quanta m}^{-2} \text{ s}^{-1}$) is
129 related to solar irradiance (i.e. short-wave flux), SW, following $\text{PAR} = 0.473 \text{ SW}$
130 [Papaioannou et al., 1993].

131 2.4- Estimation of mixed and mixing layer depths

132 The mixing layer depth, Z_{mx} , is the upper part of a mixed layer of uniform density where
133 active turbulence occurs [Brainerd and Gregg 1995]. Several recent studies have highlighted
134 the significance of this layer to understand the phytoplankton phenology of phytoplankton
135 blooms [Brody and Lozier, 2015; Taylor and Ferrari, 2011]. Autonomous profiling floats
136 equipped with bio-optical sensors have also highlighted the importance of mixing rather than



137 mixed layers to characterize the onset and the development of a phytoplankton bloom. Wind-
138 driven and buoyancy-driven regimes control the characteristics and depth of active mixing
139 [Brody and Lozier, 2014, 2015, Ramuyantseva et al, 2015, Lacour et al, 2019, Pellichero et al,
140 2020]. Wind-driven mixing dominates in the presence of weak buoyancy forcing when heat
141 fluxes are small. In that case, the dominant mixing length scale is equal to the Ekman length
142 layer depth scale.

143 The daily mean mixed layer depth, Z_{ml} is derived for 2016 from glider vertical profiles of
144 temperature and salinity using the algorithm provided by Holte and Talley, (2009). To
145 estimate Z_{mx} , τ is used to calculate the Ekman layer depth, Z_{ek} , $Z_{ek}=\gamma u_*/f$, where γ is an
146 empirical constant (0.5), f is the Coriolis parameter (10^{-4} s^{-1}) over the BOUSSOLE area, and
147 u_* is the turbulent friction velocity, $u_* = \sqrt{\frac{\tau}{\rho_w}}$ (m s^{-1}), with ρ_w the density of the surface sea
148 water [Lacour et al, 2019]. When heat fluxes are large and positive, Enrikez and Taylor
149 (2015) proposed to express the mixing depth Z_{mx} in terms of the surface forcing, i.e. the
150 surface stress and a stabilizing surface buoyancy flux B_0 , as follows:

$$151 \quad \frac{1}{Z_{mx}^2} = \frac{f^2}{(C_3 u_*)^2} + \frac{f B_0}{(C_4 u_*^2)^2} \quad (1)$$

152 With $C_3=1$, $C_4=0.57$, $B_0=Q \lambda g /c_p \rho_0$, where Q is the net surface heat flux, λ the thermal
153 expansion coefficient ($1.65 \cdot 10^{-4} \text{ }^\circ\text{C}^{-1}$), g the gravitational acceleration (9.81 m s^{-2}), c_p the
154 specific heat of water ($4 \cdot 10^3 \text{ J kg}^{-1} \text{ }^\circ\text{C}^{-1}$) and ρ_0 the pure water density (1000 kg m^{-3}).

155 2.5-Estimation of euphotic zone depth

156 The euphotic depth, Z_{eu} , is generally derived as the depth where PAR is 1% of its surface
157 value. Z_{eu} can be estimated from Chla at the surface using the method described by Lee et al.
158 (2007) based on the equation of Morel and Berthon (1989):

$$159 \quad Z_{eu} = 34 (\text{Chla})^{-0.39} \quad (2)$$

160 For the period between 2017 and 2019, GlobColour merged Chla products were used, which
161 are based on satellite observations with a resolution of 25 km, and a binning period of 8 days
162 (<http://www.oceancolour.org>). For 2016, glider surface Chla was used to derive Z_{eu} instead
163 of satellite measurements. We excluded glider Chla between 05:00 and 20:00 UTC affected
164 by quenching, and we applied a depth-constant offset to glider measurements using Chla
165 water samples available on March 7.



166 2.6 – Estimation of net community production

167 Biological Net Community Production rates, NCP ($\text{mmol m}^{-2} \text{ day}^{-1}$), are calculated for
168 separate 3-day periods in March-April, 2016 to 2019 from DIC concentrations derived from
169 CARIOCA measurements, assuming that measurements in the top 10 m of the water column
170 are representative of an homogeneous mixing layer [Boutin and Merlivat, 2009; Merlivat et al,
171 2015; Pelichero et al, 2020]. As current velocities are expected to be small [Millot, 1999;
172 Antoine et al., 2008; Niewiadomska et al., 2008], and the focus of this study is on the
173 atmospheric drivers of bloom initiation, the effect of horizontal advection at the BOUSSOLE
174 buoy is assumed negligible over the short time periods used to estimate NCP. In the study
175 area, vertical mixing events are identified by an increase in DIC resulting from upward
176 movement of high DIC Levantine Intermediate Water (LIW) at approximately 200 m depth
177 [Copin-Montégut and Bégovic, 2002]. In this study, decreases in DIC concentrations over
178 days to weeks have been identified. For these identified periods, biological production and
179 air-sea exchange are the dominant processes responsible for daily changes in DIC. We neglect
180 the eddy diffusion term as the DIC gradient at the base of the mixing layer is very weak, and
181 entrainment is not expected as changes in DIC are observed during periods of stratification.

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

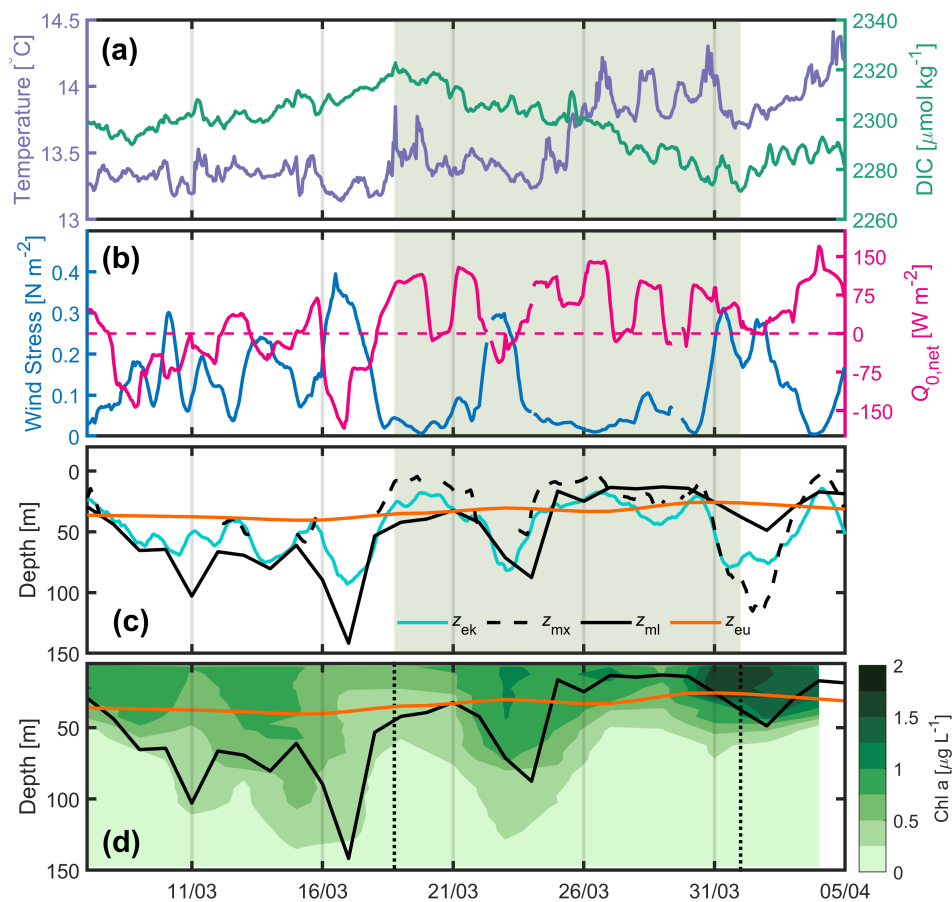
$$184 \quad \text{NCP} = \left(\frac{\Delta \text{DIC}_{\text{int}}}{\Delta t} \right)_{\text{bio}} = \left(\frac{\Delta \text{DIC}_{\text{int}}}{\Delta t} \right)_{\text{meas}} - k \times s \times (\text{pCO}_{2\text{atm}} - \text{pCO}_{2\text{sw}}) \quad (3)$$

185 The first term in equation 3, $\left(\frac{\Delta \text{DIC}_{\text{int}}}{\Delta t} \right)_{\text{meas}}$, is the daily variation of DIC_{int} calculated over 3
186 days from the date identified as the start of phytoplankton bloom (t_0), whilst the remaining
187 part corresponds to DIC variability due to air-sea CO_2 exchange, where k is the gas transfer
188 velocity [Wanninkhof, 2014], s is the solubility of CO_2 in seawater [Weiss, 1974], and
189 $\text{pCO}_{2\text{sw}}$ and $\text{pCO}_{2\text{atm}}$ are respectively the partial pressures of CO_2 , in μatm , in sea water and in
190 air [Merlivat et al, 2018].

191

192 3 Results

193 3.1- Description of the 2016 time-series



194

195 Figure 2. 2016 (a) Hourly DIC (green) and SST(purple) at the BOUSSOLE site, (b) hourly
196 wind stress (blue) and net heat flux (pink) at the Azur buoy site. The pink dotted line indicates
197 the change of sign of the net heat flux from negative to positive values, (c) daily depths of the
198 mixed layer (black), mixing layer (dotted black), Ekman layer (cyan) and euphotic layer
199 (orange) at the BOUSSOLE site. The green background on panels (a), (b), and (c) indicates
200 the onset period of the bloom (18 March to 1st April). (d) Glider-derived depths of the mixed
201 layer (black curve, same as in (c)), euphotic layer (orange), and nighttime 2-day binned
202 chlorophyll concentration (green coloring) retrieved between March 7 and April 5.

203 The temporal evolution of physical and biogeochemical parameters between March 7 and
204 April 5 2016 is displayed in Fig. 2. It shows an initial increase in DIC until March 18 (Fig. 2a),



205 resulting from the vertical mixing of surface waters with the DIC-enriched underlying LIW
206 [Copin-Montégut and Bégovic, 2002], followed by a decrease in DIC that corresponds to an
207 increase of surface Chla (Fig 2d). Temperature increases intermittently over the same time
208 period (Fig. 2a), with a sharp increase on March 18 and 25. Before March 18 frequent
209 atmospheric fronts associated with wind forcing between 0.2 and 0.5 N m⁻² were observed
210 (Fig. 2b). $Q_{0,net}$ was negative most of the time leading up to March 18, but mostly positive
211 afterwards (Fig. 2b). A clear anticorrelation was observed between τ and $Q_{0,net}$ (Fig. 2b). The
212 mixed-layer depth derived from the glider measurements (< 20 km away from the Boussole
213 site) deepened to > 70 m at times between March 8 and March 18, consistent with the passage
214 of frequent atmospheric fronts. From March 18 to April 5, Z_{ml} shoaled from ~ 45 m to ~10 m,
215 excluding two short periods around March 24 and April 2 when atmospheric fronts were
216 passing (Fig. 2c). From March 18 to 31, Z_{mx} was for a considerable time either shallower than,
217 or at similar depth to Z_{ml} and Z_{eu} (Fig. 2c). From March 18 to April 5, an increase in glider
218 Chla was observed in the upper layer (Fig. 2d) as the result of the biomass accumulation
219 within the water column following the initial growth of phytoplankton [Behrenfeld, M. and E.
220 Boss, 2014]. The same was observed in the Southern Ocean [Pelicherro et al, 2019].

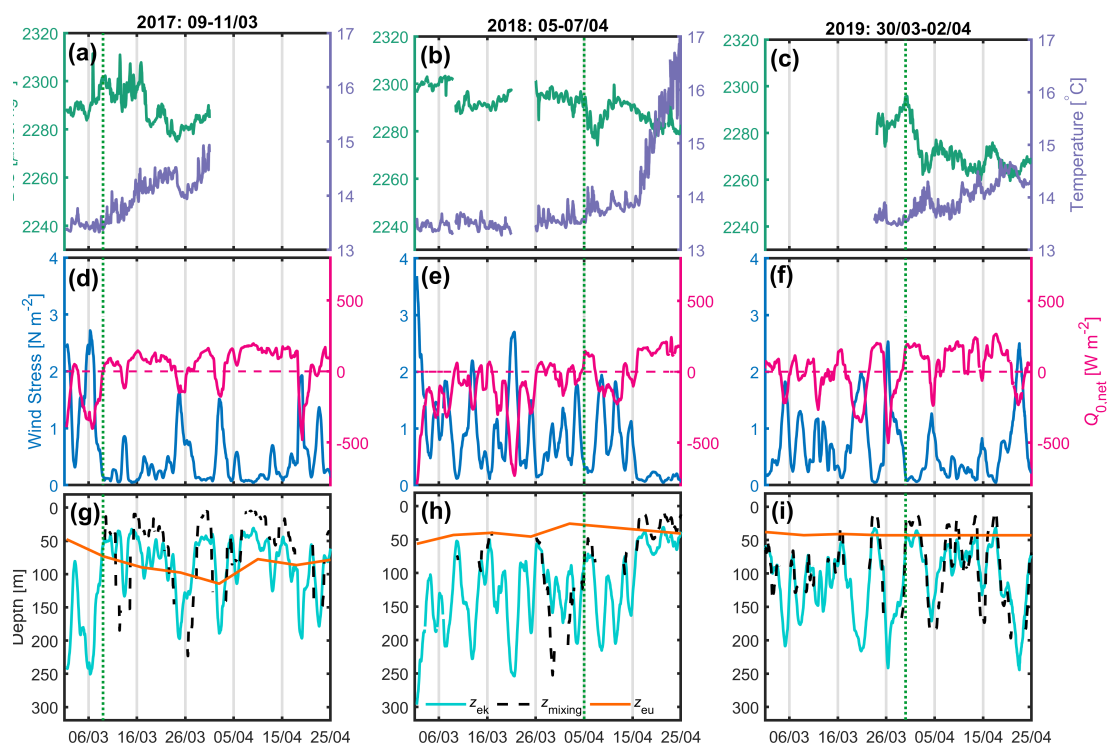
221 3.2- Description of the 2017-2018-2019 time-series

222 Only measurements at the BOUSSOLE and Azur buoys were available between 2017 and
223 2019. Here we focus on the periods between March 1 and April 25. In 2017, from March 9,
224 after the passage of a very active atmospheric front (τ up to 2.7 N m⁻²), a decrease in DIC
225 accompanied by an increase in temperature was observed for a period of 14 days despite a
226 few short bursts of vertical mixing, typically lasting less than 1 day (Fig.3a, d, g). At the same
227 time, the net surface heat flux switched from negative to positive most of the time. Over this
228 14-day period, Z_{mx} was predominantly shallower than the Z_{eu} at times when DIC was
229 decreasing.

230 In 2018 (Fig. 3b, e, h) and 2019 (Fig. 3c, f, i), the weather records throughout the month of
231 March are very similar, showing a succession of very strong wind events at the start of the
232 investigated time periods, with τ values of up to 3.7 N m⁻² and Z_{ek} values as deep as 300 m.
233 On April 5, 2018, following a period of high wind stress resulting in the formation of a deep
234 mixing layer (~250 m), a decrease in DIC and an increase in SST were observed over 3 days.
235 At the same time, the net heat flux oscillated around zero. In 2019, CARIOCA data were only
236 available after March 24. On March 30, 3 days after high winds, a decrease in DIC and an



237 increase in temperature lasted 3 days, while simultaneously the net heat flux reversed from
238 negative to positive.
239



240
241 Figure 3. (a, b, c) 2017, 2018 and 2019 DIC (green) and SST (purple) at the BOUSSOLE site,
242 (d, e, f) wind stress (blue) and net heat flux (red) at the Azur buoy site. The red dotted line
243 indicates the change of sign of the net heat flux from negative to positive values, (g, h, i)
244 depths of the mixing layer (dotted black), Ekman layer (cyan) and euphotic layer (orange) at
245 the BOUSSOLE site. The vertical dotted green line indicates the onset of the bloom.

246 3.3- NCP

247 NCP was estimated (Eq. 3) for 3-day periods over the four time-series between 2016 and
248 2019 (Table 1). Because we compute NCP only during periods when Z_{ml} and Z_{mx} shoal and
249 the air-sea CO_2 flux is very small, the measured total daily changes of DIC, $dDIC/dt$
250 (correlation coefficient, $r^{*2} > 0.70$), is expected to represent the biological consumption of



251 DIC. In 2016, 2017, and 2019, daily biological carbon uptake fluxes are between 38 and 191
 252 $\text{mmol m}^{-2} \text{d}^{-1}$. PAR varies between 63 ± 3 and 113 ± 2 $\text{mol photons m}^{-2} \text{d}^{-1}$ for the four
 253 considered periods (Table 2).

	T_0	$d\text{DIC}/dt^a$	r^{*2b}	$\Delta\text{DIC} / \Delta t^c$	r^{*2d}	air-sea flux CO_2^e	NCP ^f
		$\mu\text{mol kg}^{-1}\text{d}^{-1}$		$\text{mmol m}^{-2}\text{d}^{-1}$		$\text{mmol m}^{-2}\text{d}^{-1}$	$\text{mmol m}^{-2} \text{d}^{-1}$
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

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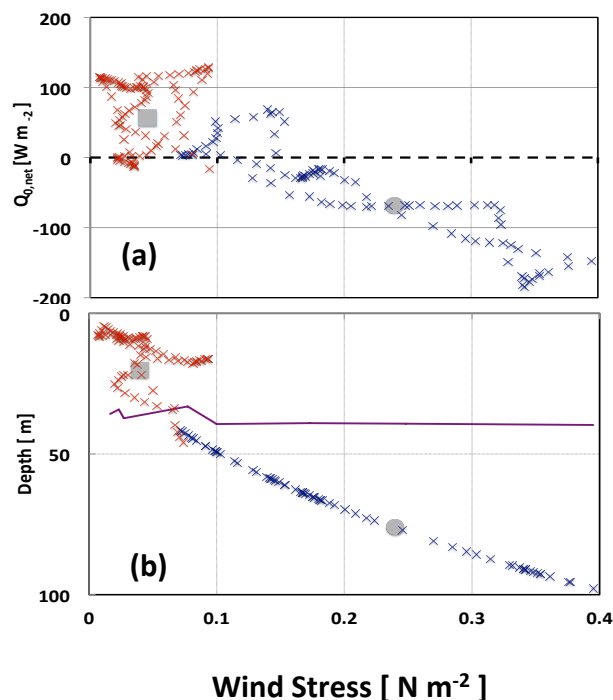
255 Table 1. Net community production computed in the mixing layer over 3 days (NCP) after the
 256 bloom onset (T_0). (a, b) variation of DIC at the measurement depth and linked correlation
 257 coefficient; (c, d) change of integrated DIC over Z_{mx} and linked correlation coefficient ;(e)
 258 CO_2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

261

262 4 Discussion

263 4.1 Analysis of the physical drivers of the bloom initiation over the 4 years, 2016-2019

264 In 2016, the biological consumption of carbon indicated by the decrease in DIC from March
 265 18 attests the initiation of spring phytoplankton growth (Fig. 2a). Between 2 successive days,
 266 March 17 and 18, a rapid decrease in daily wind stress from 0.24 to 0.04 N m^{-2} results in a
 267 change of sign of the net heat flux from -68 to $+56 \text{ W m}^{-2}$ (Fig.4a) and a shoaling Z_{mx} , which
 268 is for the most part shallower than Z_{eu} (Fig.4b) after March 18.



269

270 Figure 4. Changes of physical parameters (hourly values) at the onset of the 2016 bloom
271 during 2 consecutive periods of 4 days, March 14-17 (blue) and March 18-21 (red) as a
272 function of wind stress (a) net surface heat flux (b) depths of the mixing layer. Grey circles
273 and grey squares indicate mean values respectively on March 17 and March 18. The purple
274 line shows the euphotic layer depth (March 14-21).

275 This illustrates the very fast change in heat flux and mixing layer thickness resulting from the
276 decrease in wind stress. The main drivers to explain the initiation of near surface spring
277 phytoplankton blooms proposed by [Taylor and Ferrari, 2011, Brody-Lozier, 2014, 2015] are
278 a decrease in wind strength, a positive surface heat flux, and a decrease in the dominant
279 mixing length scale. Our results indicate that a decrease in τ is a common component for these
280 last two mechanisms (Fig.4). The variability of $Q_{0,\text{net}}$ is strongly controlled by the variability
281 of the latent heat flux proportional to the wind speed. From March 17 to 18, the latent heat
282 flux changed by 84 W m^{-2} representing nearly 2 thirds of the total change of Q_0 , with the
283 remainder originating from an increased short-wave flux. Hence, $Q_{0,\text{net}}$ increased as τ
284 decreased. In the same way Z_{mix} , which is related to τ and $Q_{0,\text{net}}$, decreased. This highlights the
285 major contribution of wind stress relaxation in controlling the net surface heat flux, and



286 subsequent stratification leading to shallower Z_{mx} . Due to the seasonal increase of surface
287 irradiance in March-April, as well as the availability of nutrients following vertical mixing
288 with intermediate nutrient-enriched waters [Begovic and Copin-Montégut, 2002], the
289 conditions for the onset of the bloom are met.

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

300 4.2- Biological carbon uptake

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

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



318 bloom-triggering atmospheric conditions are met, such as seasonal light increase and
319 availability of nutrients.

	T_0	Wind stress τ_0 N m^{-2}	Net heat flux Q_0 W m^{-2}	PAR $_0$ $\text{mol photons m}^{-2}\text{d}^{-1}$	$Z_{\text{mixing } 0}$ m	$Z_{\text{Euphotic } 0}$ m
2016	March 18	0.04 +/-0.02	56+/-36	78+/-3	20+/-9	47
2017	March 9	0.12+/-0.02	45+/-21	63+/-3	28+/-12	71
2018	April 5	0.30+/-0.24	127+/-12	88+/-6	62+/-9	29
2019	March 30	0.10+/-0.02	183+/-9	113+/-2	20+/-5	43

320

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

323

324 5- Concluding remarks

325 Around March and April in the northwestern Mediterranean Sea, the seasonal change in
326 lighting conditions and atmospheric forcing are the essential parameters that control the
327 triggering of near-surface phytoplankton blooms. These mechanisms have been highlighted
328 through the coupling of high-frequency physical and biogeochemical in situ observations. For
329 years 2016 to 2019, a bloom started in March or early April, following a period of intense
330 mixing. The rapid decrease in surface wind speed entailed a switch from negative to positive
331 air-sea net heat fluxes, essentially due to changes in the latent air-sea heat flux proportional to
332 wind speed. These analysis support the hypothesis that decreases in the depth of active mixing,
333 a result of the transition from buoyancy-driven to wind-driven mixing, control the timing of
334 the spring bloom.

335

336 Data availability



337 BOUSSOLE data (2016-2019) are available in the SEANOE data base (Merlivat Liliane,
338 Boutin Jacqueline (2020). Mediterranean Sea surface CO₂ partial pressure and temperature
339 data. SEANOE. <https://doi.org/10.17882/56709>

340 Meteorological buoy Azur: <http://mistrals.sedoo.fr>

341 SChl, 8-day, 4 kmx4 km resolution, level 3 mapped ocean colour product distributed by the
342 European Space Agency, available at <http://www.ocean-colour.org>

343 All glider data will be archived at the British Oceanographic Data Centre
344 (BODC, https://www.bodc.ac.uk/data/bodc_database/gliders/) prior to article acceptance.

345 Author contributions: VV and MG were respectively in charge of the BOUSSOLE mooring
346 buoy maintenance and monthly water sampling. LB was responsible for the laboratory
347 preparation and calibration of the CARIOCA sensors. GAL led the UEA glider mission in
348 March-April 2016. DA and VV provided data and funding through the BOUSSOLE project
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361

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478

479 Figures and tables

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

484 Figure 2. 2016 (a) Hourly DIC (green) and SST(purple) at the BOUSSOLE site, (b) hourly
485 wind stress (blue) and net heat flux (pink) at the Azur buoy site. The pink dotted line indicates
486 the change of sign of the net heat flux from negative to positive values, (c) daily depths of the
487 mixed layer (black), mixing layer (dotted black), Ekman layer (cyan) and euphotic layer
488 (orange) at the BOUSSOLE site. The green background on panels (a), (b), and (c) indicates
489 the onset period of the bloom (18 March to 1st April). (d) Glider-derived depths of the mixed
490 layer (black curve, same as in (c)), euphotic layer (orange), and nighttime 2-day binned
491 chlorophyll concentration (green coloring) retrieved between March 7 and April 5.

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

497 Figure 4. Changes of physical parameters (hourly values) at the onset of the 2016 bloom
498 during 2 consecutive periods of 4 days, March 14-17 (blue) and March 18-21 (red) as a
499 function of wind stress (a) net surface heat flux (b) depths of the mixing layer. Grey circles
500 and grey squares indicate mean values respectively on March 17 and March 18. The purple
501 line shows the euphotic layer depth (March 14-21).

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



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

510