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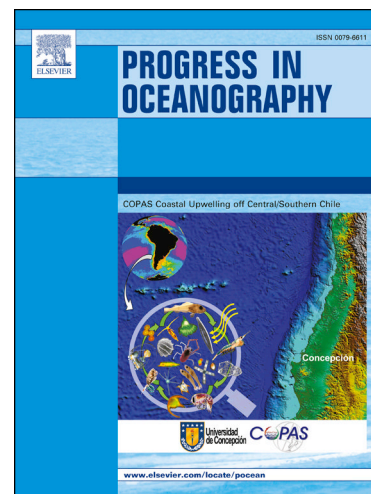
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1 Observations of vertical mixing in autumn and its effect
2 on the autumn phytoplankton bloom

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14 **Abstract**

15 This work examines the seasonal cycle of density structure and its influence on primary pro-
16 duction in a temperate shelf sea, with a particular focus on the breakdown of stratification in
17 autumn. We do this by combining new, high resolution observations of water column struc-
18 ture, meteorological forcing, nitrate and chlorophyll fluorescence collected between March
19 2014 and July 2015 on the North West European Shelf.

20 Our results challenge the generally accepted assumption that convection dominates over wind
21 driven mixing resulting in seasonal breakdown of stratification. Furthermore we found, that
22 vertical mixing in autumn not only transformed the vertical density structure but also the
23 vertical structure of chlorophyll biomass and surface nutrients. The subsurface chlorophyll
24 maximum was eroded and a vertically homogeneous profile of chlorophyll biomass established
25 itself above the pycnocline. This increased mixing also led to replenishment of surface nitrate
26 concentrations, which supported an autumn phytoplankton bloom. While the significance
27 of phytoplankton blooms in autumn has previously not been well quantified, we argue that

28 these can act as a significant contributor to the seasonal drawdown of carbon.

29 **Keywords**

30 Seasonal cycle, breakdown of stratification, SML dynamics, primary production, autumn
31 phytoplankton bloom, critical depth, long-term observations, North-West European shelf

32 **Highlights**

- 33 • We present new observations of a full seasonal cycle of vertical density structure and
34 its control on the seasonal cycle of primary production in a temperate shelf sea.
- 35 • Wind mixing appears to be the dominant SML deepening process.
- 36 • Surface mixed layer deepening in autumn replenishes surface nutrient concentrations,
37 which fuels an autumn phytoplankton bloom.
- 38 • We show that Sverdrup's critical depth hypothesis can be used to predict the shut-down
39 of primary production in autumn.
- 40 • The autumn phytoplankton bloom has the capacity to significantly contribute to the
41 seasonal drawdown of atmospheric CO₂.

42 1 Introduction

43 Continental shelves are known to be highly energetic and biologically productive regions.
44 Despite only covering $\sim 10\%$ of the ocean surface area, they perform a disproportionately
45 important role within the global carbon cycle (Liu, 2010). They support up to a third of
46 all oceanic primary productivity (Wollast, 1998; Bauer *et al.*, 2013), and at least 40 % of
47 oceanic particulate organic carbon (POC) is sequestered on continental margins of depth $<$
48 200 metres (Muller-Karger *et al.*, 2005; Dunne *et al.*, 2007; Regnier *et al.*, 2013). Temperate
49 shelf seas have also been highlighted as being substantial sinks for atmospheric CO_2 (Thomas
50 *et al.*, 2004; Borges *et al.*, 2005; Cai *et al.*, 2006; Cai, 2011).

51 Away from the influence of fresh river input near the coast, seasonal changes in the vertical
52 water column structure of temperate shelves are dictated by the competition between the
53 stratifying influence of solar irradiance and de-stabilising vertical mixing processes (Simpson
54 and Hunter, 1974; Garrett *et al.*, 1978; Simpson and Bowers, 1984). Tidal bed stress, wind
55 stress at the surface and convective mixing all make varying contributions to vertical mixing
56 (Pingree *et al.*, 1976; Simpson and Bowers, 1984). The water column structure evolves from
57 one that is fully mixed during the winter months, into a two-layer system during the spring
58 and summer, when the seasonal increase in heat input outcompetes the ability of the tides
59 and wind to break down the near surface stratification that additional heating promotes.
60 A loss of heat from the ocean to the atmosphere during the autumn (convection) triggers
61 the breakdown of stratification and a return to fully mixed conditions (Pingree *et al.*, 1976;
62 Townsend *et al.*, 2015). This seasonal cycle of stratification has a significant role to play in
63 determining the light and nutrients available to phytoplankton throughout the year (Gowen
64 *et al.*, 1995; Ji *et al.*, 2008; Sharples *et al.*, 2013; Holt *et al.*, 2014).

65 The influence the vertical structure has on primary production can be best understood
66 when assessing its constituents and their roles separately. In a simplified two-layer system
67 typical of a summer stratified shelf sea these constituents are the surface mixed layer overlying
68 the pycnocline region, which itself connects the surface to the bottom mixed layer. The
69 surface mixed layer (SML) is an ubiquitous feature of almost all oceans and describes the
70 topmost layer of the ocean in contact with the atmosphere and is assumed to be fully mixed

71 by wind, wave and/or convective processes. Its variations in depth have strong implications
72 for the exchange of gases, heat and freshwater between the atmosphere and the ocean (e.g.
73 de Boyer Montégut *et al.*, 2004; Belcher *et al.*, 2012; Seguro *et al.*, 2017) but also for biological
74 production (Sharples, 1999; Taylor and Ferrari, 2011; Brody and Lozier, 2014). In fact,
75 the SML constitutes a major control on primary productivity as it impacts on the vertical
76 distribution of phytoplankton and their exposure to nutrients and light (e.g. Sverdrup,
77 1953; Franks, 2014). The bottom mixed layer (BML) is only found in shallow seas, where
78 tidal mixing is strong enough to homogenise density gradients (Pingree and Griffiths, 1977;
79 Pingree *et al.*, 1982). While the BML is usually nutrient replete it is beyond the euphotic
80 zone. Both the surface and bottom mixed layer are connected by the pycnocline region,
81 which is characterised by the strongest density gradient. Here, the diapycnal transport
82 of momentum, heat and tracers (such as nutrients) between the SML and BML occurs,
83 however this exchange can be restricted by the density gradient within the pycnocline region.
84 Identifying the key processes controlling the vertical density structure is therefore critical to
85 physical and biological oceanography.

86 The transition from well-mixed to stratified conditions is typically associated with a spring
87 phytoplankton bloom that depletes the nutrient concentrations in the surface, an event that
88 has received considerable attention and one that makes the most important contribution to
89 annual primary production (e.g. Townsend *et al.*, 1994; Rees *et al.*, 1999; Sharples *et al.*, 2006;
90 Liu, 2010). During the following summer months, the majority of phytoplankton biomass
91 adapts to survive in low light conditions and becomes concentrated within a sub-surface
92 chlorophyll maximum (SCM) at the base of the pycnocline, in order to take advantage of
93 vertical flux of nutrients from bottom waters (Hickman *et al.*, 2012; Williams *et al.*, 2013;
94 Davis *et al.*, 2014). Receiving much less attention in the literature however is the autumnal
95 bloom in phytoplankton, which has been observed in most temperate and subpolar oceans
96 (Longhurst, 1995; Findlay *et al.*, 2006; Behrenfeld, 2010; Song *et al.*, 2010; Martinez *et al.*,
97 2011).

98 The classical view suggests that autumn blooms are caused by the deepening of the SML
99 at the end of summer (Findlay *et al.*, 2006; Song *et al.*, 2010). The SML is increased by

100 a combination of shear driven mixing due to wind stress acting on the sea surface during
101 storms for example, and convective overturning of the water column due to cooling of the sea
102 surface. The deepening of the SML subsequently leads to replenishment of nutrients to the
103 euphotic layer by entraining them from below the pycnocline (Pingree *et al.*, 1976; Findlay
104 *et al.*, 2006). For a bloom to occur, light levels need to remain high enough during the
105 deepening to support photosynthesis, despite the increase in SML resulting in phytoplankton
106 receiving less light on average. The deepening of the SML has also been linked to the dilution
107 of grazers, which can further promote phytoplankton growth by decoupling phytoplankton
108 biomass from grazing pressure by zooplankton (Smayda, 1957; Landry and Hassett, 1982;
109 Martinez *et al.*, 2011; Behrenfeld, 2010).

110 Owing to their small surface signature, short duration and spatial and temporal variability
111 (Colebrook and Robinson, 1961; Hu *et al.*, 2011; Chiswell, 2011; Song *et al.*, 2011), autumn
112 blooms are less well studied than their spring counterparts or the summer SCM, although
113 arguably some of these characteristics can also be attributed to the spring bloom (Thomas
114 *et al.*, 2003; Chiswell, 2011; Song *et al.*, 2011). While observations of the occurrence and
115 strength of autumn blooms have been documented extensively (e.g. Thomas *et al.*, 2003;
116 Aiken *et al.*, 2004; Henson *et al.*, 2009; Chiswell, 2011; Chiswell *et al.*, 2013), its significance
117 within the seasonal cycle of primary production is not well quantified.

118 In this paper our aim is to investigate the transition of vertical water column structure
119 from summer to autumn, and its effect on the inorganic nutrients and chlorophyll biomass.
120 We do this by combining long-term, high resolution observations of water column structure,
121 inorganic nutrient concentrations, chlorophyll-a fluorescence and meteorological forcing, over
122 the entire seasonal cycle observed in a temperate shelf sea. We will investigate the dominate
123 mechanisms deepening the SML in autumn and estimate their relative contributions. We
124 will further study an autumn phytoplankton bloom that was supported by the deepening of
125 the SML and the subsequent resupply of nutrients to the euphotic layer. Finally, we will
126 estimate the autumn bloom's contribution to the annual primary production of a temperate
127 shelf sea and aim to establish the role the autumn bloom plays within the seasonal cycle.

128 Improving our understanding of the significance these events play within the seasonal

129 cycle is of fundamental importance to better represent global carbon budgets and predict the
130 response of temperate shelf seas to future climate change.

131 **2 Data collection and processing**

132 In this paper we present new measurements of unprecedented detail spanning 17 months
133 (March 2014 – July 2015), which were collected in a temperate shelf sea on the North-West
134 European Shelf as part of the UK Shelf Sea Biogeochemistry (SSB) programme (Sharples
135 *et al.*, issue). A long-term mooring array in the Celtic Sea collected measurements of full-
136 depth water column structure (Wihsgott *et al.*, 2016) and dynamics, surface inorganic nutri-
137 ent concentrations, surface chlorophyll-a fluorescence and meteorological forcing. This long-
138 term mooring array consisted of a temperature-salinity logger mooring, a bottom mounted,
139 upward looking acoustic current profiler, a SmartBuoy, maintained by Centre for Environ-
140 ment, Fisheries and Aquaculture Science (Cefas) and an Ocean Data Acquisition System
141 (ODAS) buoy maintained by the UK Met Office.

142 In order to get a greater appreciation of the depth variation of biogeochemical variables
143 and to put the autumn bloom event into context, we also incorporate full-depth profiles of
144 CTD, chlorophyll-a fluorescence and inorganic nutrient samples collected during nine process
145 cruises supporting this field campaign. Their names and dates can be found in Table 1.

146 All observations presented here were taken at the centre of the Celtic Sea (CCS), at a
147 nominal location of 49.4°N and 8.6°W, in a mean water depth of 145.4 m. This location
148 is shown by the white triangle in Figure 1. The colours in Figure 1 represent the sea sur-
149 face temperatures (SST) [°C] during summer 2014. Away from coastal boundaries, warmer
150 SSTs represent seasonally stratified regions and colder SSTs the year-round vertically mixed
151 regions. As can be seen from the relatively warm SSTs surrounding CCS in Figure 1, the ob-
152 servations were taken in the seasonally stratifying part of the Celtic Sea, well away from any
153 tidal mixing fronts. The site was located centrally on the continental shelf, approximately
154 120 km northeast of the continental shelf break and approximately 200 km south-west from
155 the British Isles.

Cruise name	Dates
DY008	18 th March – 13 th April 2014
JC105	15 th June – 24 th June 2014
DY026a	03 rd August – 15 th August 2014
DY026b	15 th August – 25 th August 2014
DY018	09 th November – 03 rd December 2014
DY021	01 st March – 26 th March 2015
DY029	01 st April – 30 th April 2015
DY030	04 th May – 25 th May 2015
DY033	11 th July – 03 rd August 2015

Table 1: SSB process cruises. Here, DY stands for RRS Discovery and JC for RRS James Cook.

156 2.1 CTD profiles and bottle samples

157 During each cruise a Seabird *9plus* Conductivity-Temperature-Depth (CTD) and a CTG
 158 Aquatracka fluorometer mounted on a 24-bottle rosette system collected vertical profiles of
 159 temperature, salinity, and chlorophyll-*a* fluorescence (referred to as Chl *a* for the rest of this
 160 paper). While Chl *a* is not a direct measure of cell abundance, it is used in this paper as a
 161 proxy for chlorophyll biomass.

162 The raw 24 Hz profiles were extracted, filtered and corrected for thermal inertia using SeaBird
 163 data processing Software (Seasave V 7.23.2). The data were subsequently screened and
 164 anomalous data removed, averaged onto a 1 db grid and calibrated against samples of Chl *a*
 165 concentration and salinity.

166 Water samples between the surface and near bed were collected on most CTD casts and
 167 analysed on board for dissolved inorganic nutrients using a Bran and Luebbe segmented flow
 168 colorimetric auto-analyser following classical analytical techniques as described in Woodward
 169 and Rees (2001). Our focus in this paper is on nitrate (NO₃) plus nitrite (NO₂), referred
 170 to as nitrate hereafter. Clean sampling and handling techniques were employed during the
 171 sampling and manipulations within the laboratory, and where possible carried out according

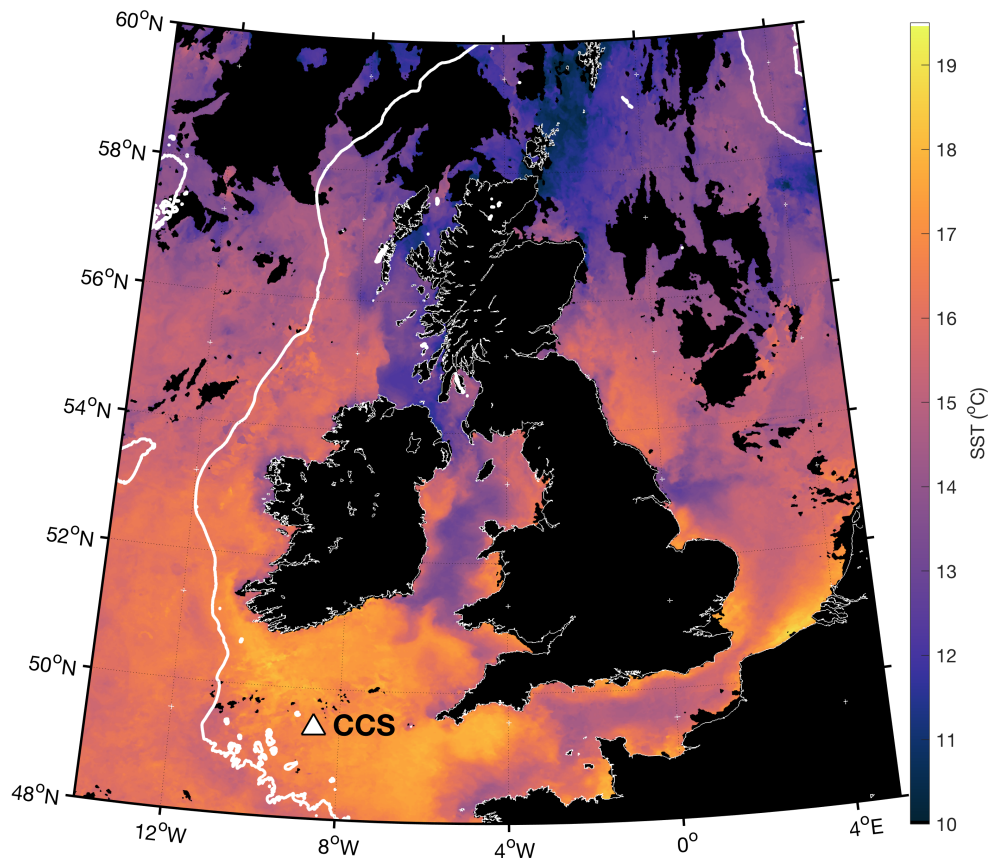


Figure 1: Sea surface temperature (SST) [$^{\circ}\text{C}$] around the British Isles during summer 2014. The white triangle marks the location of the central Celtic Sea (CCS) mooring array location. The thick, white line denotes the 200 metre bathymetry contour, which marks the edge of the NW European continental shelf. This satellite image is a 1 week median SST composite, 25th June - 1st July 2014, courtesy of NEODAAS Plymouth Marine Laboratory, UK.

172 to the International GO-SHIP nutrient manual recommendations (Hydes *et al.*, 2010). All
 173 samples were analysed as soon as possible after sampling from the CTD Rosette. Nutrient
 174 reference materials (KANSO Japan) were run each day to check analyser performance and
 175 to guarantee the quality control of the final reported data. The typical uncertainty of the
 176 analytical results was between 2-3%, and the limits of detection for nitrate was $0.02 \mu\text{mol}$
 177 l^{-1} .

178 2.2 Mooring observations

179 The full-depth (10-15 m to sea bed) temperature-salinity (TS) mooring monitored the
180 evolution of the vertical water column structure from March 26th 2014 to July 25th 2015
181 (Wihsgott *et al.*, 2016). It was designed to capture the vertical structure of the whole water
182 column and had a vertical resolution of 2.5 metres in the pycnocline and 5 - 20 metres
183 resolution in the surface and bottom layer. The instruments' temporal sampling resolution
184 was 5 minutes. After recovery all instruments were calibrated against the ship's CTD data (a
185 SBE 9plus). At each time step, 8 instruments on the mooring took coincident measurements
186 of temperature, conductivity and pressure throughout the water column. To construct full
187 water column profiles of salinity we used a similar method to Hopkins *et al.* (2014) and fitted
188 a salinity surface as a function of all simultaneous observations of salinity, temperature and
189 time. Delaunay triangulation was then used to evaluate salinity for all available temperature
190 measurements. Potential density, ρ [kg m^{-3}], was derived using the Gibbs-SeaWater (GSW)
191 Oceanographic Toolbox (McDougall and Barker, 2011).

192 To complement the near-surface observations of the TS mooring, we also used temperature
193 data collected by instruments suspended from a SmartBuoy, maintained by the Centre for
194 Environment, Fisheries and Aquaculture Science (Cefas) and an Ocean Data Acquisition
195 Systems (ODAS) buoy, maintained by the Met Office, at CCS. Over the observational period
196 their setup varied but for the majority of the time, sensors were located between 0.3 - 7.5
197 metres below the sea surface.

198 A bottom mounted, upward facing 150 kHz FlowQuest acoustic current profiler (ACP)
199 recorded horizontal velocities throughout the whole water column (Wihsgott *et al.*, 2018).
200 The ACP had a vertical resolution of 2 metres and a 2.5 minute temporal resolution. The
201 current measurements were corrected for time varying magnetic declination, which is the
202 angle between magnetic and true north. Furthermore, the top 14 metres of velocity data were
203 removed owing to spurious readings near the sea surface due to side lobe contamination. A
204 battery failure after the 6th May further resulted in loss of data until a new instrument had
205 been deployed on 9th June 2014.

206 All TS chain measurements were linearly interpolated onto a 5 minute x 2.5 metres reso-
207 lution grid.

208 2.2.1 Mixed layer estimates

209 Mixed layer depth estimates were derived using profiles of potential density collected at
210 the CCS mooring site. Here we define the depth of the surface mixed layer (SML) as a
211 density change of $+0.02 \text{ kg m}^{-3}$ relative to the value at 10 metres depth, and the depth of
212 the bottom mixed layer (BML) was defined as a density change of -0.02 kg m^{-3} relative to
213 the value closest to the bed.

214 2.3 Cefas SmartBuoy

215 In addition to near surface temperature sensors, the Cefas SmartBuoy sensor package
216 also consisted of a Seapoint Chlorophyll Fluorometer (SCF) [mg m^{-3}] and a quantum photo-
217 synthetically active radiation (PAR) [$\mu\text{E m}^{-2} \text{ s}^{-1}$] meter (LiCor Inc., USA). The data were
218 stored using the ESM2 data logger, which was configured to sample for 10 min at 1 Hz
219 every 30 min as outlined in Kröger *et al.* (2009); Hull *et al.* (2016). In order to correct for
220 instrument drift, the SCF was standardised to arbitrary fluorometry units using fluorescent
221 sulphate microspheres (FluoSpheres, Thermo Fisher Scientific Inc.) after each deployment
222 at the Cefas laboratories. In order to omit artefacts due to non-photochemical quenching,
223 only Chl *a* data that were collected when $\text{PAR} < 10 \mu\text{E m}^{-2} \text{ s}^{-1}$ (i.e. hours of darkness)
224 were included in the analysis.

225 The Cefas SmartBuoy also took measurements of nitrate concentration [$\mu\text{mol l}^{-1}$] at the
226 sea surface. Samples were collected using automated water samplers operated by pumping
227 samples into polyethylene bags pre-injected with 5 ml of 1.4 g l^{-1} mercuric chloride (HgCl_2 in
228 ultrapure water) as a preservative. On return to shore bag samples were then filtered using 0.2
229 μm pore size Whatman Cyclopore polycarbonate filters and analysed using a Skalar SAN plus
230 segmented flow autoanalyser, by standard spectrophotometric methods (Kirkwood, 1996).

231 2.4 Meteorological observations and heat flux calculations

232 The hourly observations of wind speed, w [m s^{-1}], relative humidity, r_h [%], air temper-
233 ature, T_a [$^{\circ}\text{C}$], mean sea level pressure, p [hPa] and air density, ρ_a [kg m^{-3}] recorded by the
234 Met Office ODAS buoy provided the majority of the meteorological data. We complement

235 these observations with shortwave radiation, Q_{sw} [$W\ m^{-2}$] and total cloud cover [%] data
 236 from the extended-range reanalysis European Reanalysis (ERA)-Interim product of gridded
 237 meteorological fields (Dee *et al.*, 2011) from the European Centre for Medium-Range Weather
 238 Forecasts (ECMWF). This product integrates observations to model the atmospheric fields
 239 across the globe to give 3 hourly datasets with 80 km spatial resolution. The time series
 240 used here has been interpolated onto the CCS mooring location. In order to verify the model
 241 data, they were compared to observations of the Met Office buoy and the overall fit for the
 242 wind speed was found to be good ($R^2 = 0.9097$).

243 With the combined data the net heat flux, Q_{net} [$W\ m^{-2}$] (Figure 2a), into the ocean was
 244 calculated as the sum of all in- and outgoing heat fluxes:

$$Q_{net} = Q_{sw} + Q_{lw} + Q_{sen} + Q_{lat}, \quad (1)$$

245 where Q_{sw} is the shortwave, Q_{lw} is the longwave, Q_{sen} is the sensible and Q_{lat} is the evaporative
 246 heat flux. Here, following the convention of the ECMWF fields, all vertical fluxes are defined
 247 to be positive downwards. Except for Q_{sw} , which was obtained from the ECMWF reanalysis
 248 ERA-Interim product, all other heat fluxes were calculated following Gill (1982).

249 3 Results

250 This section will present the high-resolution, long-term observational data introduced above
 251 to provide an overview of the physical conditions that prevailed at CCS throughout the
 252 17-month observational campaign of the SSB programme. The length of the observational
 253 campaign provided an excellent opportunity to focus particularly on the seasonality, and the
 254 chance also to compare recurring events in 2014 and 2015.

255 3.1 The seasonal cycle at CCS

256 In general, meteorological conditions intuitively displayed a strong seasonal cycle, most evi-
 257 dent in the Q_{sw} (solar irradiance) and thus Q_{net} , which formed a key component of boundary
 258 forcing. The seasonal cycle of Q_{net} , had maxima during June during both 2014 and 2015 and

259 was at a minimum during December - January 2014/2015 (Figure 2a). Daily averaged Q_{net}
260 reveals the ocean to be gaining heat between the end of March until the end of September
261 2014 and losing heat from October 2014 to March 2015. This periodicity was less evident in
262 wind speeds, which despite displaying winter maxima were highly variable throughout the
263 observations and provided a constant source of energy with minimum monthly averages of
264 around 7 m s^{-1} during summer 2014 (not shown). Winds were predominantly coming from
265 the southwest. The impacts of meteorological seasonality is clearly evident in the vertical
266 density structure, ρ [kg m^{-3}] provided by the TS mooring at CCS (Figure 2b) and will be
267 explored in more detail in the following sections.

268 3.1.1 Onset of stratification in spring 2014

269 When the TS mooring was first deployed on March 26th 2014, the water column was still
270 vertically mixed from the previous winter. During the first days of the observations the very
271 top layers of the sea surface stratified during the day with a top-bottom density, ρ , difference
272 of 0.01 kg m^{-3} , however this could not be sustained throughout the diurnal cycle. On
273 March 30th 2014 Q_{net} became predominantly positive (heat gain by the ocean) and supplied
274 more buoyancy than was dissipated by wind and tidal mixing. This marked the onset of
275 spring stratification. In the following days stratification continued to strengthen until April
276 26th 2014, when a strong low-pressure system passed overhead the mooring site. Wind
277 speeds exceeding 18 m s^{-1} and significant wave heights briefly reaching 9 metres (not shown)
278 deepened the SML by 20 metres (Figure 2b). Following the storm, re-stratification of the
279 subsurface layers took place until the water column resembled a typical summer density
280 structure (Figure 2b). The depth of the SML throughout summer 2014 was on average 20
281 metres. Along with the heat gain at the sea surface through direct heat exchange with the
282 atmosphere, the temperature of the bottom boundary layer also increased by $1.9 \text{ }^\circ\text{C}$ between
283 April and December 2014 due to heat transfer through the pycnocline (Figure 2d).

284 **3.1.2 Breakdown of stratification - convection vs wind forcing during autumn**
 285 **2014**

286 In October 2014 Q_{net} turned predominantly negative and wind speeds increased compared
 287 to the summer months (Figure 2a & c, average wind speeds of 8.8 m s^{-1} during October -
 288 December compared to average wind speeds of 6.75 m s^{-1} during July - September). This led
 289 to deepening of the SML depth and marked the beginning of the breakdown of stratification
 290 in 2014 (arrows in Figure 2).

291 During this period negative heat fluxes rarely occurred in isolation from strong wind
 292 forcing at CCS. In order to determine whether the breakdown of stratification was driven by
 293 shear driven processes caused by wind stress or convective mixing due to buoyancy reduction
 294 initiated by negative heat fluxes, the Obukhov length scale, L_{OB} [m] (Obukhov, 1946) was
 295 used to examine this competition:

$$L_{\text{OB}} = -\frac{u_*^3}{\kappa B_0} \quad (2)$$

296 Here, u_* [m s^{-1}] is the friction velocity, $u_* = \left(\frac{\tau}{\rho_0}\right)^{1/2}$, where τ [N m^{-2}] is the wind stress,
 297 and $\rho_0 = 1026 \text{ kg m}^{-3}$ is the reference density. $\kappa = 0.41$ is the von Kármán constant,
 298 and B_0 [m^2s^{-3}] is the surface buoyancy flux. Considering that temperature is the dominant
 299 control on density in the Celtic Sea (Pingree *et al.*, 1976; Simpson and Hunter, 1974) we
 300 estimate B_0 to be directly proportional to Q_{net} using $B_0 = \frac{\alpha g}{c_p \rho_0} Q_{\text{net}}$. Here, α [$^{\circ} \text{C}^{-1}$]
 301 is the thermal expansion coefficient of seawater calculated using the GSW Oceanographic
 302 Toolbox (McDougall and Barker, 2011), $g = 9.81 \text{ m s}^{-2}$ is the acceleration due to gravity
 303 and $c_p = 3985 \text{ J kg}^{-1} \text{ }^{\circ}\text{C}^{-1}$ is the heat capacity of seawater. Similar to the observed and
 304 calculated heat flux terms introduced earlier, B_0 was defined to be positive downwards.

305 The $|L_{\text{OB}}|$ specifies the vertical extent over which either convection or mechanical stirring
 306 (at the boundary) is the dominant surface mixing mechanism (Taylor and Ferrari, 2011). If
 307 the water column is unstable due to strong surface cooling (negative Q_{net}) the L_{OB} is greater
 308 than 0 ($L_{\text{OB}} > 0$). In contrast, if the water column is vertically stratified due to positive heat
 309 fluxes the L_{OB} is less than 0 ($L_{\text{OB}} < 0$). Coupling the Obukhov length scale with the depth
 310 of the surface mixed layer, Brody and Lozier (2014) define three surface regimes controlling

311 the SML (Table 2) that we use here to help identify the contribution that convection and
 312 wind-mixing make to autumnal deepening of the SML. When the buoyancy flux is large and
 313 negative (the ocean is losing heat to the atmosphere), and wind speeds are low, convection
 314 is the dominant control on the SML depth (case 1, Table 2). In contrast, when wind speeds
 315 are moderate to large, the wind becomes the driver of surface mixing and SML deepening
 316 (case 2, Table 2). The sign of the Q_{net} and thus B_0 are irrelevant on this occasion. In case
 317 of a small positive net heat/buoyancy flux, which promotes stable stratification ($L_{\text{OB}} < 0$),
 318 the wind becomes the sole surface mixing mechanism by default. When the buoyancy flux
 319 is large and positive, stratification counteracts any surface mixing and SML deepening is
 320 suppressed (case 3, Table 2).

Convective mixing regime	$ L_{\text{OB}} < \text{SML}$
case 1	while $B_0 < 0$ and hence $Q_{\text{net}} < 0$
Wind mixing regime	$ L_{\text{OB}} > \text{SML}$
case 2	
Heat regime	$ L_{\text{OB}} < \text{SML}$
case 3	while $B_0 > 0$ and hence $Q_{\text{net}} > 0$
(stratification counteracts mixing)	

Table 2: Surface regimes controlling the SML

321 Using hourly data of observed wind speed, w , and net heat flux, Q_{net} , the L_{OB} was
 322 calculated for the entire time series. These hourly results of the L_{OB} were then compared to
 323 the SML (Figure 2b) and categorised accordingly for each day, using the criteria in Table 2.
 324 Subsequently, a relative contribution was attributed to each regime on a daily basis, e.g. if
 325 $|L_{\text{OB}}| > \text{SML}$ for 12 hours during 10th October 2014, then wind forcing was considered the
 326 dominant SML affecting mechanism during 50% of that day. To filter out some of the short
 327 term variability owing to sporadic events in heating and wind forcing, the daily contributions
 328 were smoothed using an 8 day running average (Figure 3a).

329 As might be expected from the observed Q_{net} (Figure 2a), the convective and heating

330 regime (cases 1 & 3 Table 2) displayed a clear seasonal cycle (Figure 3a), with convection
331 more dominant during winter, and heating in the summer months. While the wind regime
332 (case 2 Table 2), was less seasonal, it dominated throughout the observational campaign
333 (53% of the entire observational period). During the period of the active SML deepening
334 (2nd October - 31st December 2014, grey bar Figure 3a), the contribution of both wind and
335 convection (cases 1 & 2 Table 2) increased compared to the rest of the year, and the heating
336 regime (case 3 Table 2) was completely shut off at times. Despite several periods of sustained
337 surface cooling occurring during autumn 2014 (Figure 2a), the wind regime significantly
338 increased its control on the SML (two sample t -test: $p < 0.01$, t -test), being dominant
339 63% of the time the SML deepened (2nd October - 31st December 2014). Periods when the
340 convective regime was dominant accounted for 32% of this time, which coincided with low
341 wind speeds/stresses (Figure 3b-c). This represents a statistically significant increase of 8%
342 (two sample t -test: $p < 0.01$) compared to the whole observational period. Periods when
343 positive stratification counteracted wind mixing (case 3 Table 2) accounted for the least
344 amount of time during the SML deepening period, of 5%. While shear stresses due to wind
345 appear to be the dominant SML deepening mechanism, considerable variability between and
346 within days was observed. Figure 3b-d demonstrate this short-term variability by focusing
347 on a 2 week period in December 2014. The main sources of this variability was the diurnal
348 heat cycle and the relatively short duration of some wind events.

349 This is an interesting and potentially significant result as it challenges many previous as-
350 sumptions that convection is the dominant mechanism driving seasonal breakdown of stratifi-
351 cation in shelf seas (Edinger *et al.*, 1968; Nielsen and St. John, 2001; Townsend *et al.*, 2010), as
352 well as in open-ocean environments, (Kraus and Turner, 1967; Lacombe *et al.*, 1970; Marshall
353 and Schott, 1999; Taylor and Ferrari, 2011). While an attempt has been made to separate
354 the individual contributions from wind and convection, the observed mixing effects on the
355 density structure are difficult to distinguish as they both contribute to the same process of
356 deepening the SML. We note that the dependence of both the sensible and latent heat flux
357 (Q_{sen} , Q_{lat}) on the wind speed, w , ensures that the sum of all heat fluxes, Q_{net} , can never
358 act fully decoupled from the wind forcing. Furthermore, both convection and shear driven
359 mixing can aid each other to be more efficient at deepening the SML. Convection can act to

360 better connect surface mixing processes with the stratified interior by homogenising the sur-
361 face boundary layer, supporting further breakdown of seasonal stratification. Whereas wind
362 stress can aid convection by disrupting the thin viscous sublayer and thereby permitting a
363 more rapid transfer of heat through the sea surface.

364 During the winter months of January and February 2015 the water column was further
365 losing heat to the overlying atmosphere and eventually cooling down to approximately 10°C
366 (Figure 2d). While the water column was vertically fully mixed for most of the winter months,
367 periods of transient stratification did exist. These generally only lasted one day but could
368 occur for up to 5 consecutive days but the stratification only manifested itself in the top 10
369 metres of the water column.

370 On March 26th 2015 the buoyancy input of the positive heat flux became strong enough
371 to overcome the wind and tidal mixing and the water column began to re-stratify. While the
372 timing of the onset of stratification is similar to 2014, the rate at which stratification was
373 strengthening was lower during 2015. This resulted in the water column being less strongly
374 stratified at any time during 2015 compared to the previous year (Figure 2d, Figure 4a).
375 At the end of the observational period in July 2015 the difference in top-bottom density
376 difference was 0.75 kg m⁻³ less than observed in July 2014 (Figure 4a).

377 In summary, the observed evolution of water column structure was typical for a seasonally
378 stratifying shelf sea, such as the Celtic Sea. Here, the change in vertical water column
379 structure is predominantly a vertical exchange process driven by the competition of buoyancy
380 input versus stirring at the boundaries i.e. sea surface/bed (Simpson and Hunter, 1974;
381 Garrett *et al.*, 1978; Simpson and Bowers, 1984). The buoyancy input was supplied by Q_{net}
382 at the sea surface, whereas wind and tides were supplying stirring powers to mix gradients
383 near the sea surface and sea bed.

384 **3.2 Seasonal cycle of chlorophyll-a and inorganic nitrate concen-** 385 **trations**

386 The seasonal cycle of primary production in the Celtic Sea is, like in other seasonally strati-
387 fying shelf sea regions, tightly coupled to the change in vertical water column structure (Tett
388 *et al.*, 1993; Thomas *et al.*, 2003; Hu *et al.*, 2011; Sharples *et al.*, 2013). The long-term
389 observations of surface Chl *a* and nitrate shown in Figure 4c-d demonstrate a clear response
390 to the physical events described above. At the end of winter in March 2014, before stratifi-
391 cation was fully established (Figure 4a), Chl *a* concentrations were low ($< 1 \text{ mg m}^{-3}$) and
392 nitrate concentrations were high ($\sim 9 \text{ } \mu\text{mol l}^{-1}$) throughout the water column. As spring
393 stratification became established a spring phytoplankton bloom was initiated, which peaked
394 on April 11th 2014 with surface Chl *a* concentrations of up to 6.2 mg m^{-3} . Consequently
395 the available nitrate in the surface mixed layer (SML) became quickly depleted and con-
396 centrations dropped to $\sim 2.5 \text{ } \mu\text{mol l}^{-1}$. During the following summertime stratified period,
397 the diapycnal transport of momentum, heat and tracers is restricted due to suppressed tur-
398 bulent motions at the pycnocline. Thus the resupply of inorganic nutrients from the dark,
399 nutrient rich bottom waters to the well-lit, nutrient depleted surface waters is inhibited. The
400 resulting nutrient limitation, and potentially also an increased impact of grazers, led to a
401 decrease in the surface population and the demise of the spring phytoplankton bloom. The
402 secondary peak in surface nitrate concentration around April 26th 2014 was induced by a
403 strong storm event described above. Here, strong wind and waves deepened the SML by 20
404 metres (Figure 2b) and thereby entrained dissolved nutrients from the BML, raising surface
405 nitrate concentrations to $6.9 \text{ } \mu\text{mol l}^{-1}$. Subsequently a secondary phytoplankton bloom was
406 initiated, with surface Chl *a* concentration of up to 9 mg m^{-3} that peaked on May 4th 2014.

407 On May 12th the SmartBuoy platform drifted away from its location and hence no surface
408 nitrate and Chl *a* observations were available from CCS until June 19th 2014. At this time
409 the vertical profiles of density, nitrate and Chl *a* resembled that of a typical shelf sea summer
410 profile as also observed in other shelf seas e.g. (Williams *et al.*, 2013; Townsend *et al.*, 2015;
411 Du *et al.*, 2017). Compared to the spring phytoplankton bloom at the surface, the biomass
412 peak had been shifted to the interior of the water column to the SCM. In all coincident, full
413 depth profiles of CTD, nitrate and Chl *a* at CCS, the SCM was located within the base of
414 the pycnocline and in the vicinity of the nitracline. Here, turbulence from tidal and internal

415 mixing mechanisms, for example internal waves, together with the strong nutrient gradient
416 (the nitracline) caused an upward flux of nutrients that sustained this biomass peak (Williams
417 *et al.*, 2013; Lee *et al.*, 2016; Du *et al.*, 2017). Peak concentrations of Chl *a* within the SCM
418 were variable (average 2.06 ± 0.92 mg Chl *a* m^{-3} ; $n=9$), while Chl *a* concentrations within
419 the SML were uniformly low (average 0.31 ± 0.1 mg Chl *a* m^{-3} ; $n=9$).

420 The breakdown of stratification commenced in early October 2014 due to increased wind
421 mixing and, to a lesser extent, also surface cooling (Figure 3a). While this resulted in a
422 deepening of the SML and sharpening of the pycnocline (Figure 3c), it also transformed the
423 vertical structure of chlorophyll biomass and inorganic nutrients. Figure 5 illustrates the
424 change in vertical structure between summer (Figure 5a-c) and autumn (Figure 5d-f): The
425 deepening of the mixed layer resulted in entrainment of nutrients from below the pycnocline,
426 which increased surface nitrate concentrations by 2.1 ± 0.1 $\mu\text{mol l}^{-1}$ (Figure 4d). This
427 increase is seen over the entire SML (Figure 5b & e). The deepening also led to the erosion
428 of the SCM and a vertically homogenous profile of chlorophyll biomass was established above
429 the pycnocline (Figure 5c & f). Simultaneously we observed an increase in surface Chl *a*
430 concentrations of up to 2.2 mg m^{-3} (Figure 4c), which could be indicative of an autumn
431 phytoplankton bloom driven by the resupply of nutrients replenished by SML deepening.
432 Surface light levels were low during this period, and less than half of spring and summer
433 PAR levels (Figure 4b).

434 Surface Chl *a* concentrations dropped to winter background levels of < 1 mg m^{-3} around
435 December 13th 2014 and stayed low during the mixed period. While nitrate data were unus-
436 able between October 16th 2014 and March 20th 2015 due to problems with the preservative,
437 pre bloom nitrate concentrations of ~ 7.5 $\mu\text{mol l}^{-1}$ were observed during the DY021 February
438 process cruise.

439 The phytoplankton spring bloom that followed the onset of stratification in 2015, was
440 significantly stronger in magnitude compared to 2014, with peak surface Chl *a* concentrations
441 of up to 11 mg m^{-3} (Figure 4c). In general, the 2015 bloom had several peaks and hence
442 the main bloom event was less well defined compared to 2014. Following the bloom Chl *a*
443 concentrations within the SML, surface values dropped back to low summer values (average

444 0.16 ± 0.05 mg Chl *a* m^{-3} ; $n=40$). Peak Chl *a* concentrations within the SCM in the
445 following summer were again variable (average 1.05 ± 0.41 mg Chl *a* m^{-3} ; $n = 40$).

446 4 Discussion

447 We have presented observations of the evolution of vertical water column structure through-
448 out the seasonal cycle of 2014 and 2015, and showed a clear response of Chl *a* and nitrate
449 to these events. We find that the deepening of the SML depth in autumn 2014, which was
450 mostly driven by wind mixing, replenished inorganic nutrient concentrations in the surface
451 layer. Simultaneously, we observed the erosion of the summer SCM peak by homogenising
452 the vertical chlorophyll biomass profile over the entire SML. We will now consider whether
453 the observed increase in Chl *a* during the autumn is linked to in-situ phytoplankton growth
454 as a result of replenishment of nutrients, or simply a redistribution of the subsurface phyto-
455 plankton community. We will also examine the role that light availability plays terminating
456 the autumn bloom. Finally, using the well resolved time series of water column structure
457 and changes in nutrient concentrations throughout the year, we make an estimate of the
458 contribution to new production, i.e. the proportion of primary production that is supported
459 by nitrate (Dugdale and Goering, 1967), made by the autumn bloom and compare this to
460 estimated and measured rates of productivity during the spring and summer months.

461 4.1 In-situ growth in autumn

462 The depth integrated Chl *a* biomass can be used to help determine whether a phytoplankton
463 population is actively growing in response to additional resource availability (light or nu-
464 trients), or whether changes in Chl *a* concentration are simply redistributed due to vertical
465 mixing of the water column. Figure 6 shows the seasonal cycle of depth integrated chlorophyll
466 biomass during the stratified periods of 2014 and 2015. For each CTD cast at CCS this was
467 calculated by taking the depth integral from the surface to the top of the BML. In most ver-
468 tical profiles of Chl *a* we found evidence of photochemical quenching during daytime CTDs
469 in the near surface. To avoid underestimating the depth integrated chlorophyll biomass we
470 extrapolated Chl *a* values from the SML depth to the near surface using nearest neighbour

471 extrapolation for all daytime CTDs. This led to an average increase of 4% compared to using
472 non-corrected profiles of Chl *a*.

473 In order to estimate depth integrated biomass from surface Chl *a* concentrations, recorded
474 by the SmartBuoy, we assumed a homogeneous profile of Chl *a* throughout the SML as
475 observed during DY018 (Figure 5f). We then calculated the depth integral from the surface
476 to the SML depth, and hence this should be considered as a minimum estimate of chlorophyll
477 biomass.

478 As might be expected, the highest observed values of up to 186 mg m^{-2} were found during
479 the spring bloom cruise (DY029) in 2015. In contrast to this, the summer values (JC105,
480 DY026a/b, DY030 and DY033) were relatively low, yet variable (average $21.33 \pm 9.89 \text{ mg}$
481 $\text{Chl } a \text{ m}^{-2}$, $n = 55$), but similar in magnitude to values observed by Hickman *et al.* (2012)
482 in the Celtic Sea. As soon as the vertical water column structure began to break down in
483 early October 2014, we observed a sharp increase in integrated chlorophyll biomass of up
484 to 90 mg m^{-2} compared to summer values (Figure 6). This increase is indicative of in-
485 situ growth fuelled by the resupply of inorganic nutrients to the euphotic layer from depth,
486 as opposed to redistribution of Chl *a*, and the availability of sufficient light to sustain an
487 autumnal phytoplankton bloom. Evidence of enhanced primary production during DY018
488 indicative of an autumn phytoplankton bloom was also found in other studies: García-Martín
489 *et al.* (2017) found evidence that the system at CCS turned net-autotrophic during DY018
490 thus acting as a sink of CO_2 due to primary production. Giering *et al.* (2018) observed a
491 secondary peak in the abundance of nauplii and copepodites (zooplankton), indicative of an
492 autumn phytoplankton bloom. Further evidence was also observed by Davis *et al.* (2018),
493 who noted increases of particulate organic carbon (POC) and particulate organic nitrogen
494 (PON) during DY018, similar to the signal they observed during the spring phytoplankton
495 bloom in 2015 (DY033).

496 4.2 Light limitation during autumn

497 As mentioned earlier the in-situ light levels during the autumn period were less than half
498 compared to those experienced during the spring and summer months (Figure 4b), yet clearly

499 sufficient for the onset of the autumn phytoplankton bloom (Figure 4c, Figure 6). Despite
500 this a change in phytoplankton production must have occurred, as we noticed the presence
501 of significant levels of nitrate concentrations of $2.1 \mu \text{ mol l}^{-1}$ on average throughout the SML
502 during DY018 (Figure 4d, Figure 5e). While biomass was increasing, phytoplankton did not
503 deplete the newly available nitrate pool to undetectable levels, which is normally the case
504 during spring and summer conditions (Figure 4c-d) when surface phytoplankton communities
505 are thought to be nitrogen (N) limited in the Celtic Sea (Pemberton *et al.*, 2004; Davis *et al.*,
506 2014; Williams *et al.*, 2013). The presence of nitrate within the SML during autumn is thus
507 an indication that primary production within the SML had shifted from N-limited production
508 during spring and summer to light limited production, which was also suggested by Poulton
509 *et al.* (2017) based on their observed phytoplankton turnover times.

510 We want to further study this light limitation by comparing the SML depth to the critical
511 depth, z_{cr} , the theoretical depth at which vertically integrated phytoplankton growth out-
512 weighs losses. The concept of z_{cr} was developed by Sverdrup in 1953 as part of his critical
513 depth theory (SCD) (Sverdrup, 1953), which predicts the onset of a phytoplankton bloom
514 when the actively turbulent layer shoals above the critical depth (Franks, 2014). As a re-
515 sult phytoplankton are no longer light limited, growth outweighs losses, and a bloom can
516 occur. This concept has been usually applied to study the mechanisms triggering the onset
517 of the spring phytoplankton bloom (Siegel *et al.*, 2002) but has recently received consider-
518 able debate regarding its validity (Behrenfeld, 2010; Taylor and Ferrari, 2011; Brody and
519 Lozier, 2014). Interestingly, Chiswell (2011) & Chiswell *et al.* (2015) proposed that the SCD
520 may actually apply in autumn and winter to determine the shut-off of primary production.
521 One of the SCD's main assumption regards an actively turbulent surface layer that ensures
522 equal light exposure, rather than a surface mixed layer that is defined by a fixed difference
523 in temperature/density to a near surface value (Franks, 2014). In contrast to most spring
524 conditions, during autumn the SML is approximately equal to the actively turbulent layer,
525 as the SML is being actively deepened, which homogenises the surface layer (Figure 5d-f).
526 We therefore use the SML depth as an indicator for the depth of the turbulent layer during

527 autumn. Values for z_{cr} were calculated using

$$\frac{1}{K z_{cr}} (1 - e^{-K z_{cr}}) = \frac{I_c}{I_0} \quad (3)$$

528 where $K = 0.1 \text{ m}^{-1}$ is the attenuation coefficient, I_c [$\text{mol m}^{-2} \text{ d}^{-1}$] is the compensation
 529 irradiance, where integrated losses and production balances, and I_0 [$\text{mol m}^{-2} \text{ d}^{-1}$] is the
 530 surface irradiance. Here, we calculated z_{cr} for $I_c = 1.24 \text{ mol m}^{-2} \text{ d}^{-1}$ a value obtained by
 531 Siegel *et al.* (2002) for an open ocean zonal average between 45-50° N, and $I_c = 3.03 \text{ mol}$
 532 $\text{m}^{-2} \text{ d}^{-1}$ a value observed by Langdon (1988) for a coastal dinoflagellate. We also compare
 533 these to z_{cr} values calculated for the Celtic Sea by Pingree *et al.* (1976).

534 As might be expected, all variants of z_{cr} show a clear seasonal cycle with deepest values
 535 during summer and shallowest during winter (Figure 7a), which is in good agreement with
 536 the magnitude of surface irradiance (Figure 4b). While the values calculated by Pingree
 537 *et al.* (1976) clearly show a stronger response to the seasonal cycle, the timings at which z_{cr}
 538 becomes shallower/deeper than the SML are similar to the values calculated by us. Since we
 539 cannot draw conclusion from the SML depth versus z_{cr} outside the autumn period we want to
 540 focus on Figure 7b-c. During the first half of the autumn bloom period the SML is shallower
 541 than the critical depth ($\text{SML} < z_{cr}$) and surface Chl *a* concentrations increase (Figure 7b-c).
 542 Throughout November the SML approaches z_{cr} . The SML is deeper than z_{cr} ($\text{SML} \geq z_{cr}$)
 543 from around mid November 2014 onwards, which coincides with depth integrated chlorophyll
 544 biomass (Figure 6) and surface Chl *a* concentrations steadily decreasing to winter background
 545 levels of $< 1 \text{ mg m}^{-3}$ (Figure 4c, Figure 7c). This observed relationship does suggest that the
 546 SCD might be applicable to winter conditions and can be used to predict the shut-down of
 547 the autumn bloom, based on SML depth and surface irradiance values. Using these criteria
 548 to determine the shut-down of the autumn phytoplankton bloom we can estimate the bloom
 549 to have taken place between early October to November 20th 2014, which results in a duration
 550 of approximately 50 days.

551 4.3 Autumnal primary production

552 In order to assess the relative importance of primary production during the autumn bloom in
553 comparison to the contribution to the annual budget during the spring and summer months
554 we make an estimate of new (gross) primary production based on the fraction of new nitrate
555 supplied during the SML deepening that was taken up by phytoplankton.

556 Between summer and autumn the SML deepened from an average 21 m to 52 m (Fig-
557 ure 5a, d). This would have entrained 31 m of bottom water with a nitrate concentration of
558 $9.2 \pm 0.1 \mu\text{mol l}^{-1}$ (Figure 5e). Distributing this over the 52 m autumn mixed layer gives
559 a concentration of $5.5 \mu\text{mol l}^{-1}$. Knowing that in November only $2.1 \pm 0.1 \mu\text{mol l}^{-1}$ were
560 observed in the surface layer (Figure 4d, Figure 5e), we assume that phytoplankton took up
561 $3.4 \pm 0.1 \mu\text{mol l}^{-1}$ during the autumn bloom event. Using the elemental ratio of carbon (C)
562 and nitrogen (N) found in phytoplankton we can convert the amount of utilised nitrate into
563 an estimate of new, gross primary production. The C:N ratio of primary production has been
564 shown to vary across a range of timescales, environmental conditions and between different
565 phytoplankton groups (eg Geider and La Roche, 2002; Sterner, 2015; Moreno and Martiny,
566 2018). On average it tends to be close to the Redfield ratio, 106:16 (Redfield, 1934), which
567 has more recently been revised to be 117:14 (Anderson and Sarmiento, 1994). Unfortunately,
568 seasonally resolved observations of the C:N ratio were not available, but Humphreys *et al.*
569 (2018) derived C:N ratios that span from spring - summer for each year of the SSB field
570 campaign. For spring-summer 2014 Humphreys *et al.* (2018) found a C:N ratio of 117:13.0,
571 which suggests a C rich production compared to Redfield. Observations by Davis *et al.* (2018)
572 also suggest the production was C-rich compared to Redfield. They found that the compo-
573 sition of dissolved organic matter (DOM), which is a direct product of primary production,
574 comprised $93 \pm 1\%$ of the total organic matter (TOM) during DY018 and, both pools, DOM
575 and TOM, were reported to be C-rich compared to Redfield, with a C:N ratio of 12.5 ± 1.5
576 and 11.3 ± 1.2 , respectively (Davis *et al.*, 2018). Throughout the observational campaign the
577 C:N stoichiometry of the TOM pool showed little seasonal variability overall. The average
578 ratios were comparable to previous studies in the Celtic Sea and other shelf seas that are
579 characterised by nitrate limited production and thus the carbon and nitrate pools appeared
580 to be closely coupled throughout (Davis *et al.*, 2018 and references therein). In the absence of
581 a cruise or season specific C:N ratio we thus assumed that the phytoplankton during autumn

582 maintained the same C:N ratio as in spring and summer 2014 of 117:13.0 (Humphreys *et al.*,
583 2018). In order to then derive the nitrate-supported C fixation we multiplied the converted
584 amount of C by its molecular weight of 12 g mol^{-1} and obtained an estimate of $19.1 \pm 0.3 \text{ g C}$
585 m^{-2} . Hence throughout a duration of 50 days, the autumn phytoplankton bloom potentially
586 supported $382 \pm 6 \text{ mg C m}^{-2} \text{ d}^{-1}$ of new production.

587 In order to put the autumn phytoplankton bloom into context with other events during the
588 seasonal cycle we calculated the equivalent new production rates for each season (Figure 8). As
589 before, we use the observed C:N ratios by Humphreys *et al.* (2018) who found C:N ratios of
590 117:13.0 and 117:12.2 for spring-summer 2014 and 2015, respectively.

591 For spring values we calculated new primary production rates based on the initial nitrate
592 concentrations within the SML prior to the bloom and the average SML at the beginning
593 of the bloom. The initial nitrate concentrations were simply defined as the pre-bloom con-
594 centrations of nitrate, these were $8 \pm 0.1 \mu\text{mol l}^{-1}$ in 2014 (DY008) and $7 \pm 0.1 \mu\text{mol l}^{-1}$
595 (DY021) in 2015 (Figure 4d). Due to increased solar radiation and thus increased stratifica-
596 tion the SML generally shoals throughout spring and summer (Figure 2a-b). We therefore
597 decided to use the average SML during the onset of the spring phytoplankton bloom in both
598 years as this generally sets the depth over which nutrients will become depleted. Here we
599 found average SML depths of 30 and 29 m for the spring period 2014 and 2015, respectively
600 (Figure 5b). The new (gross) primary production was then derived using the observed C:N
601 ratios of 117:13.0 (Humphreys *et al.*, 2018) as $25.9 \pm 0.1 \text{ g C m}^{-2}$ for the spring phyto-
602 plankton bloom of 2014. While for the 2015 spring phytoplankton bloom we used the C:N
603 ratio of 117:12.2 (Humphreys *et al.*, 2018) and obtained an estimate of $23.4 \pm 0.3 \text{ g C m}^{-2}$.
604 In order to obtain the daily production rates for each spring bloom event its duration had
605 to be defined first. Using a 32 year-long record of monthly averaged data collected by a
606 Continuous Plankton Recorder (CPR) at a shelf site in the Celtic Sea Joint *et al.* (2001)
607 suggested a period of 2 months (April - May) for the spring phytoplankton bloom. This
608 agrees well with our observations of overall increased surface Chl *a* concentrations during
609 April-May 2014 and 2015 (Figure 4c). It could be argued, that in 2014 the spring phyto-
610 plankton bloom actually concluded with the onset of the spring storm in late April 2014,

611 which initiated a secondary peak in surface Chl *a* due to replenishment of surface nitrate
612 (Figure 2b, Figure 4c-d). However we believe this is unlikely to occur every year and thus
613 apply the commonly used duration of 60 days, which suggests rates of 432 ± 2 and 390 ± 5
614 $\text{mg C m}^{-2} \text{ d}^{-1}$ of new production during spring 2014 and 2015, respectively.

615 During summer months surface nutrients are depleted (Figure 4a) and hence new primary
616 production within the SCM depend on diapycnal nutrient fluxes from the BML, which is the
617 product of the vertical diffusivity at the base of the pycnocline, K_z [$\text{m}^2 \text{ s}^{-1}$], times the vertical
618 nitrate gradient $\frac{\Delta N}{\Delta z}$ [mmol m^{-4}]. Here, ΔN is the difference in nitrate within the SML and
619 BML, and Δz is the thickness of the nitracline. Due to the relatively low vertical resolution
620 of discrete bottle samples, especially compared to physical data (Figure 5a-c), deducing the
621 thickness of the nitracline from discrete data points would have resulted in an underesti-
622 mate of the nitrate gradient. Instead, we followed methods by Sharples *et al.* (2001), who
623 defined the thickness of the nitracline between the depth of the SCM peak and the BML
624 depth derived from CTD profiles. Using this method we found the nitracline thickness, Δz ,
625 to vary between 4.0 and 8.0 metres during both DY026 (summer 2014) and DY033 (summer
626 2015). Using the average thickness of 5.5 metres during DY026 resulted in a vertical nitrate
627 gradient, $\frac{\Delta N}{\Delta z}$, of 1.7 mmol m^{-4} in summer 2014. Similarly, using the average thickness of 6.0
628 metres during DY033 results in a vertical nitrate gradient of 1.4 mmol m^{-4} in summer 2015.
629 By assuming a typical value for K_z (at the base of the pycnocline) of $1 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ during
630 both summers (Townsend, 1991; Benitez-Nelson *et al.*, 2000; Sharples *et al.*, 2001, 2009) we
631 obtained estimates of gross primary production rates of 158 ± 1 and $139 \pm 4 \text{ mg C m}^{-2} \text{ d}^{-1}$
632 in 2014 and 2015, respectively. As already mentioned by Townsend (1991), the amount of
633 new production is extremely sensitive to the chosen value of K_z , and in reality the nitrate flux
634 will vary with time in response to changes in tidal, wind and internal mixing (Sharples, 2008;
635 Burchard and Rippeth, 2009; Williams *et al.*, 2013). The current estimates are thus based on
636 being supported by a background vertical flux of nitrate at the base of the thermocline. Our
637 calculations thus do not reflect any short lived injections due to sporadic turbulent events
638 and should be considered long-term estimates. Nevertheless, our rates for summer production
639 agree with rates previously found in other temperate shelf seas (Townsend, 1991; Sharples
640 *et al.*, 2001; Williams *et al.*, 2013).

641 By defining the summer regime as the period where new production is predominantly sus-
642 tained by diapycnal nutrient fluxes, hence the time between spring bloom and autumnal
643 deepening, suggests a duration of approximately 120 days (June - September), which is sim-
644 ilar to previous estimates in temperate shelf seas (Hickman *et al.*, 2012).

645 For ease of comparing our estimates of production rates among each other and with other
646 studies, we summarised them in Table 3 & Figure 8. The error bounds presented here take,
647 where applicable, account of uncertainties (1 standard deviation) in the SML, BML & SCM
648 depths as well as nitrate concentrations within the SML & BML.

649
650 Our results confirm the widely held view that the spring phytoplankton bloom is the
651 dominant event fixing carbon in the seasonal cycle of primary production (e.g. Townsend
652 *et al.*, 1994; Rees *et al.*, 1999; Sharples *et al.*, 2006; Liu, 2010). The spring phytoplankton
653 bloom in 2014 was characterised by the highest production rate of $432 \pm 2 \text{ mg C m}^{-2} \text{ d}^{-1}$
654 (Table 3 & Figure 8a) within the observational period. During the observational campaign
655 the production rates were lowest during the summer, sustaining 45 and 36% of the spring
656 production in 2014 and 2015, respectively. The overall reduced production in 2015, compared
657 to 2014, was potentially caused by a reduced nitrate inventory (Figure 4d, Davis *et al.*, 2018;
658 Humphreys *et al.*, 2018) and overall weaker stratified conditions in summer 2015 compared to
659 summer 2014 (Figure 4a), which could result in a less effective diapycnal flux of nutrients into
660 the euphotic layer during the summer months. We were surprised to see the rate of carbon
661 production during autumn 2014 ($382 \pm 6 \text{ mg C m}^{-2} \text{ d}^{-1}$) was of similar magnitude to that of
662 the following spring phytoplankton bloom 2015 ($390 \pm 5 \text{ mg C m}^{-2} \text{ d}^{-1}$), which suggests that
663 the autumn phytoplankton bloom could act as a significant contributor to carbon fixation
664 within the seasonal cycle.

665 Comparing our estimates to in-situ measurements of net primary productivity (NPP) at
666 CCS by Poulton *et al.* (2017) shows some overlap in autumn 2014 (mean of 436 mg C m^{-2}
667 d^{-1} , range of $222\text{-}563 \text{ mg C m}^{-2} \text{ d}^{-1}$). Since our values (Table 3 & Figure 8a) reflect the
668 potential new production supported by the injection of new nitrate the relative agreement
669 between our estimate and the NPP estimates by Poulton *et al.* (2017) suggests that a large

Season	Gross primary production [mg C m ⁻² d ⁻¹]
Spring 2014	432 ± 2
Summer 2014	158 ± 1
Autumn 2014	382 ± 6
Spring 2015	390 ± 5
Summer 2015	139 ± 4

Table 3: Carbon fixation rates (new production) [mg C m⁻² d⁻¹] at CCS

670 fraction of the primary production during the autumn bloom was new rather than regenerated
671 (approximately 88%). This is clearly higher than the estimated f -ratios proposed by Joint
672 *et al.* (2001) that ranged between 0.25-0.39 throughout September and October using data
673 sets obtained in the Celtic Sea. Joint *et al.* (2001) assumed f -ratios to increase during winter
674 months to up to 0.5 during January and February. Taking an f -ratio of 0.4 and 382 mg
675 C m⁻² d⁻¹ of new production suggests 955 mg C m⁻² d⁻¹ of total production, which is
676 evidently higher than the maximum observed NPP rates found by Poulton *et al.* (2017). We
677 do however note that 50% of the CCS samples by Poulton *et al.* (2017) were taken after
678 our predicted shutdown of the autumn phytoplankton bloom due to insufficient light levels
679 using the SCD hypothesis (Figure 6 & Figure 7c). While it is feasible that production still
680 occurred, the decreasing trend in depth integrated chlorophyll biomass (Figure 6) and surface
681 Chl a (Figure 7c) beyond this point suggests that production occurred at a reduced rate.
682 These samples might therefore underrepresent the total production that took place during
683 the autumn phytoplankton bloom.

684 Whilst assumptions we made about the bloom duration and the depth of the SML are
685 justified based on the physical data presented here, we recognise that the C:N ratio of pri-
686 mary production is variable (eg Geider and La Roche, 2002; Sterner, 2015; Moreno and
687 Martiny, 2018). Despite using the best available estimate of in-situ C:N ratio at the time,
688 we acknowledge the need for further research to better constrain the autumn phytoplankton
689 bloom.

690 In addition to providing a third burst of primary production in the seasonal cycle of tem-
691 perate shelves, the autumn phytoplankton bloom potentially plays a critical role in exporting
692 carbon to the open ocean, which ultimately determines the efficiency of the continental shelf
693 pump (Thomas *et al.*, 2004; Chen and Borges, 2009; Barrón and Duarte, 2015). The autumn
694 bloom is triggered by an increase in convection and wind mixing that gradually deepen the
695 SML and ultimately restores a fully mixed water column. During the winter mixed period
696 there is a weak net off-shelf transport (Ruiz-Castillo *et al.*, 2018) that has the potential to
697 remove organic material fixed on the outer shelf during the autumn bloom to deep water.
698 During the spring and summer, when bottom water transport is more typically on-shelf
699 (Ruiz-Castillo *et al.*, 2018) removal of organic matter is less likely. The carbon fixed during
700 the autumn bloom, just before the water column fully mixes may therefore constitute an
701 important fraction of the carbon removed annually from the shelf.

702 5 Conclusion

703 This paper examined newly collected, long-term observational data of full-depth density, Chl
704 *a* and nitrate profiles collected during the continuous 17 months observational campaign
705 of the UK Shelf Sea Biogeochemistry programme. We observed an entire seasonal cycle of
706 vertical density structure and its control on the seasonal cycle of primary productivity in
707 a temperate shelf sea. The focus of this paper was the transition of vertical water column
708 structure from summer to autumn, and its effect on the inorganic nutrients and chlorophyll
709 biomass.

710 In an attempt to investigate the relative contributions to the vertical density structure from
711 wind mixing, heating and convection, the Obukhov length scale (L_{OB} , Equation 2) was used,
712 as it represents a balance between wind stress and buoyancy fluxes. The concept of Brody
713 and Lozier (2014) provided a useful framework for this work (Table 2). Wind mixing (case
714 2 conditions) was shown to be the dominant control on density structure making the largest
715 contribution for 53% of the time. This influence was found to further increase during October
716 - December 2014 during the breakdown of stratification, wind being the dominant control for
717 63% during this period. This is a potentially significant result since convection is typically

718 thought to dominate SML deepening in autumn. We also observed that SML deepening
719 during this period eroded an established SCM, whilst replenishing surface concentrations
720 of nitrate. A subsequent increase in surface Chl *a* concentrations suggested in-situ growth,
721 which was confirmed by examining depth integrated chlorophyll biomass. The presence
722 of detectable nitrate concentrations within the surface layer also suggested that primary
723 production had shifted to become light limited.

724 Building on the comprehensive understanding of water column dynamics and long-term
725 time series of surface nitrate and Chl *a* we have investigated the role the autumn phyto-
726 plankton bloom plays within the seasonal cycle and estimated its contribution to the annual
727 primary production. We propose that the autumn bloom has the potential to act as a signif-
728 icant contributor to carbon fixation within the seasonal cycle. While the approach to winter
729 appeared to have been a key time for shelf water to be exported into the NE Atlantic (Ruiz-
730 Castillo *et al.*, 2018), which could make the autumn productivity particularly important,
731 further research is required to establish whether this may then contribute to the export of
732 carbon into the deep ocean.

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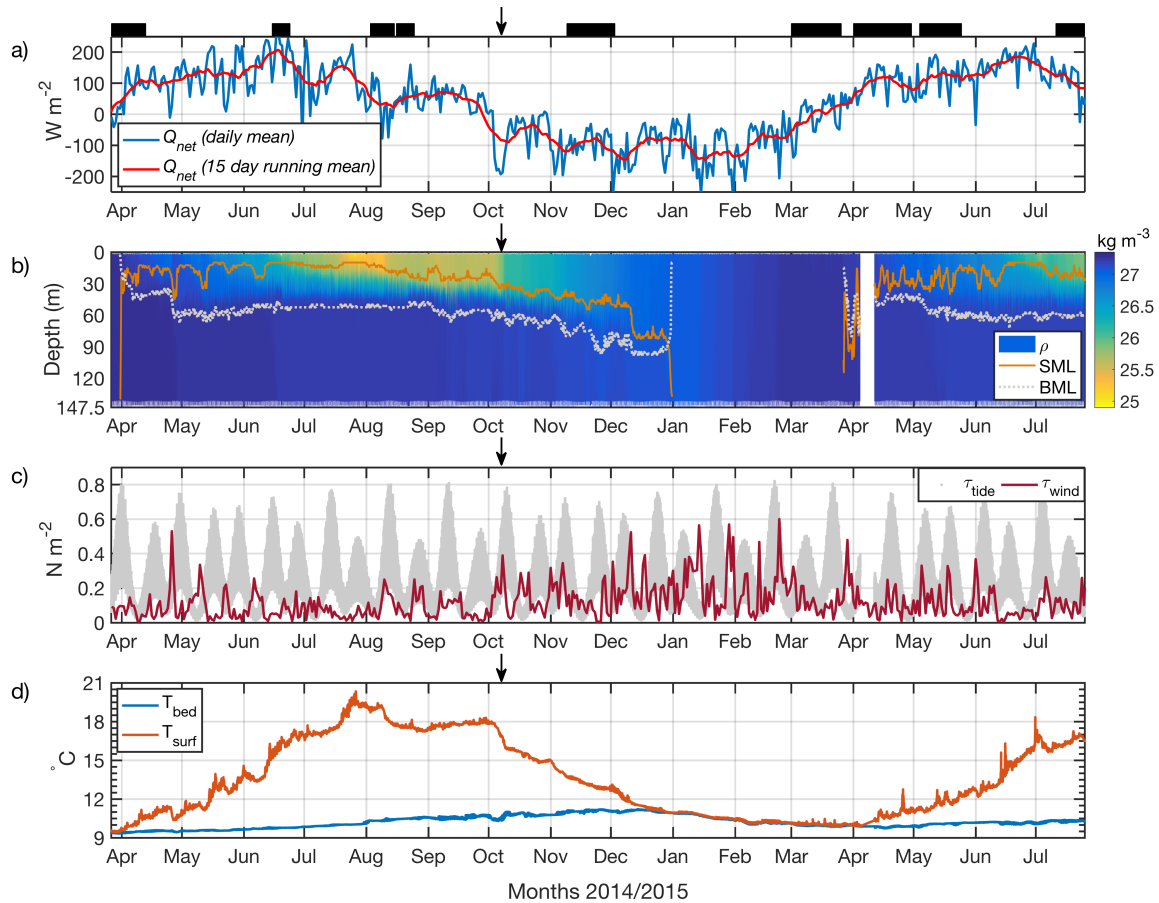


Figure 2: Physical environment: a) Q_{net} [$W m^{-2}$] (blue - daily averaged, red - 15 day running average). The black bars above denote the cruise dates (Table 1). b) full depth observations of ρ [$kg m^{-3}$], overlaid are the SML (solid orange) and BML (dotted grey). c) daily averages of wind (red) and hourly averages of tidal (grey) stresses [$N m^{-2}$] acting on the sea surface and bed, respectively. d) Evolution of near bottom (blue) and near surface (red) temperature [$^{\circ}C$].

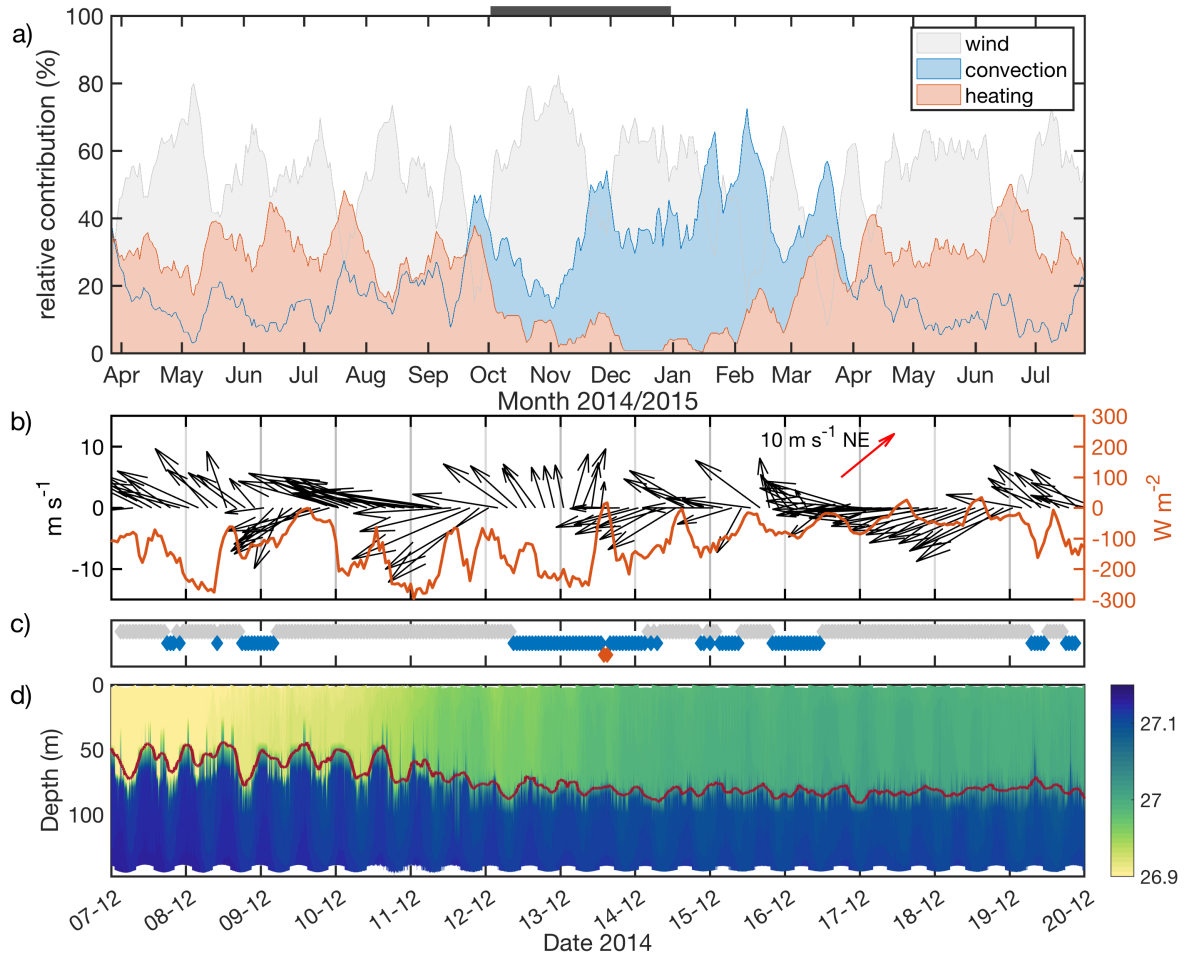


Figure 3: Dominant controls on SML: a) 8 day running average of proportional control on SML: wind (grey), convective (blue) and heat (red) regime. The grey bar above marks the SML deepening period, October 2nd - December 31st 2014. b) Observed w [m s^{-1}] and wind direction (black) and Q_{net} [W m^{-2}] (orange) c) Dominant surface regimes controlling the SML: wind (grey), convective (blue) and heat (red) d) Observed ρ [kg m^{-3}] with overlaid SML depth [m] (red) during a 2 week period in December 2014.

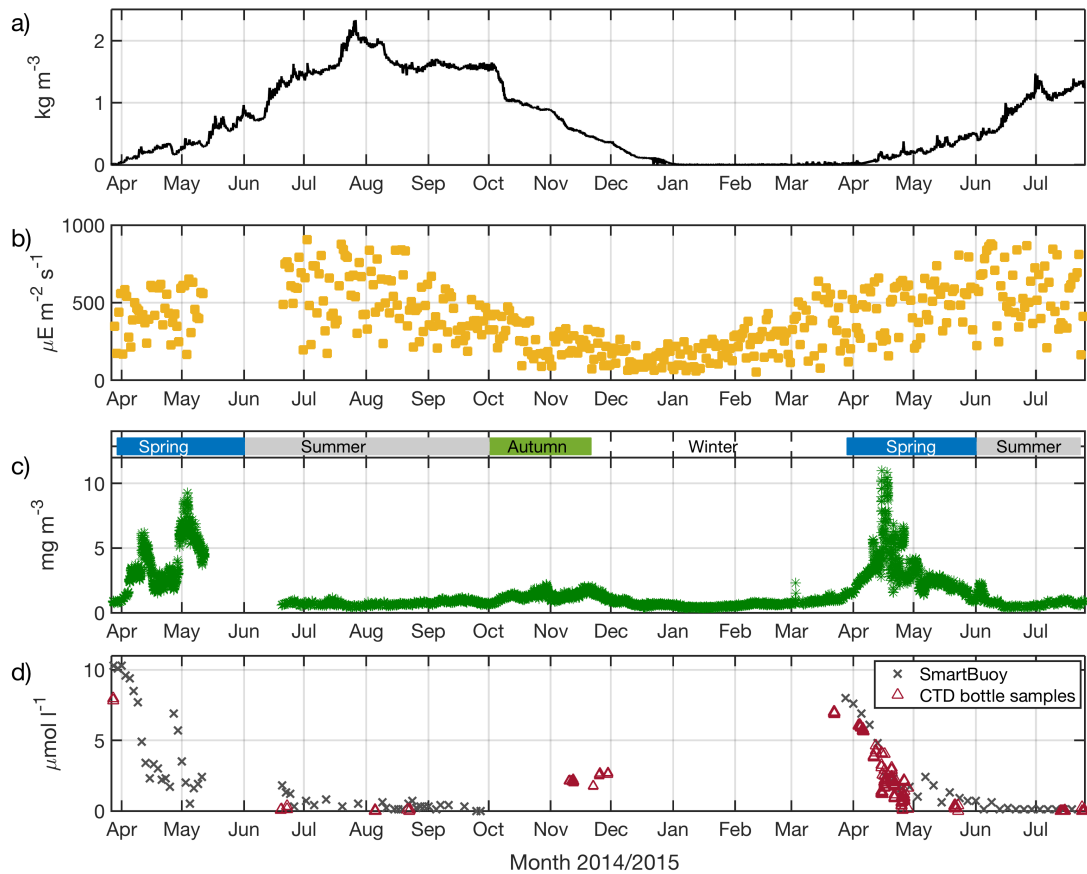


Figure 4: Combined physical and biogeochemical observations: a) top-bottom ρ difference [kg m^{-3}]. b) daily averaged PAR [$\mu\text{E m}^{-2} \text{s}^{-1}$]. c) surface Chl *a* [mg m^{-3}]. The bars above mark the duration of each seasonal regime. d) surface nitrate concentration [$\mu\text{mol l}^{-1}$].

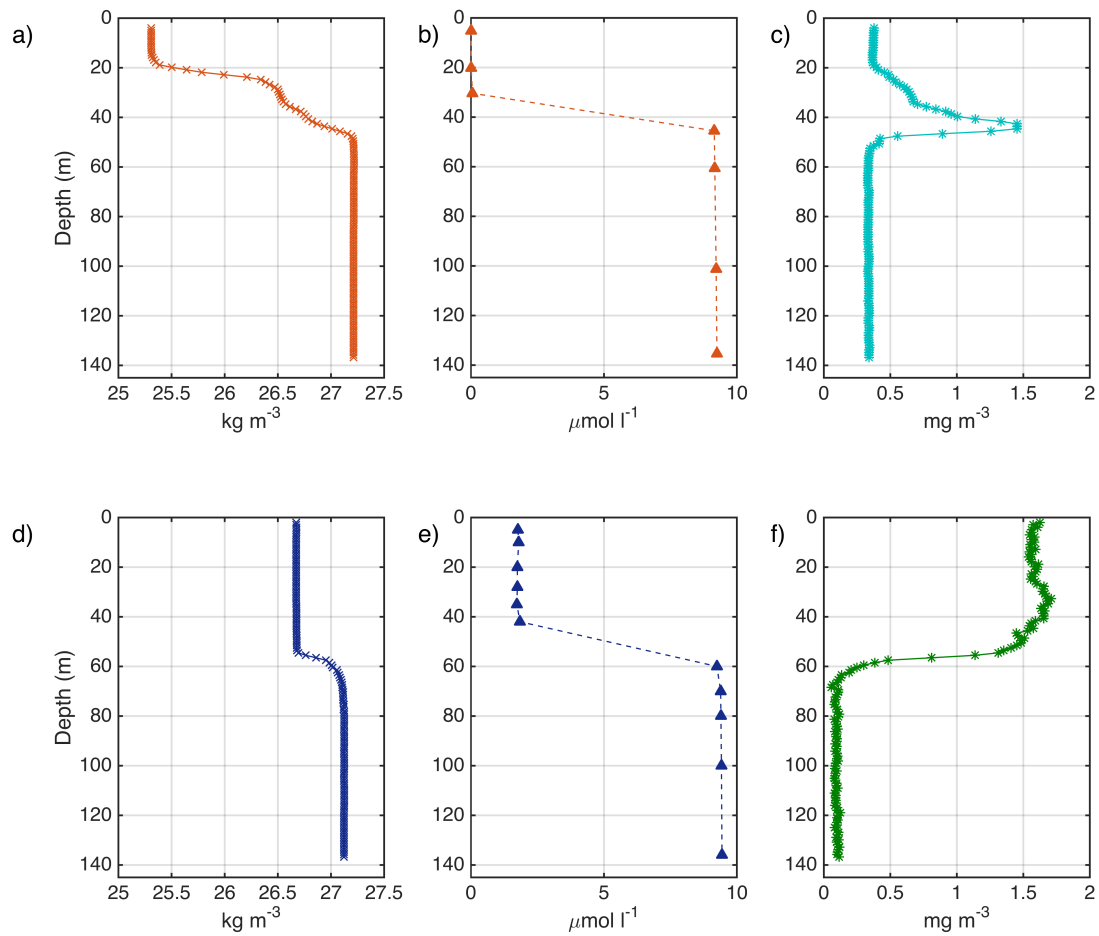


Figure 5: Vertical profiles during a)-c): summer (DY026a/b) and d)-f): autumn (DY018). a) & d) potential density [kg m^{-3}]. b) & e) nitrate [$\mu\text{mol l}^{-1}$]. c) & f) Chl *a* [mg m^{-3}].

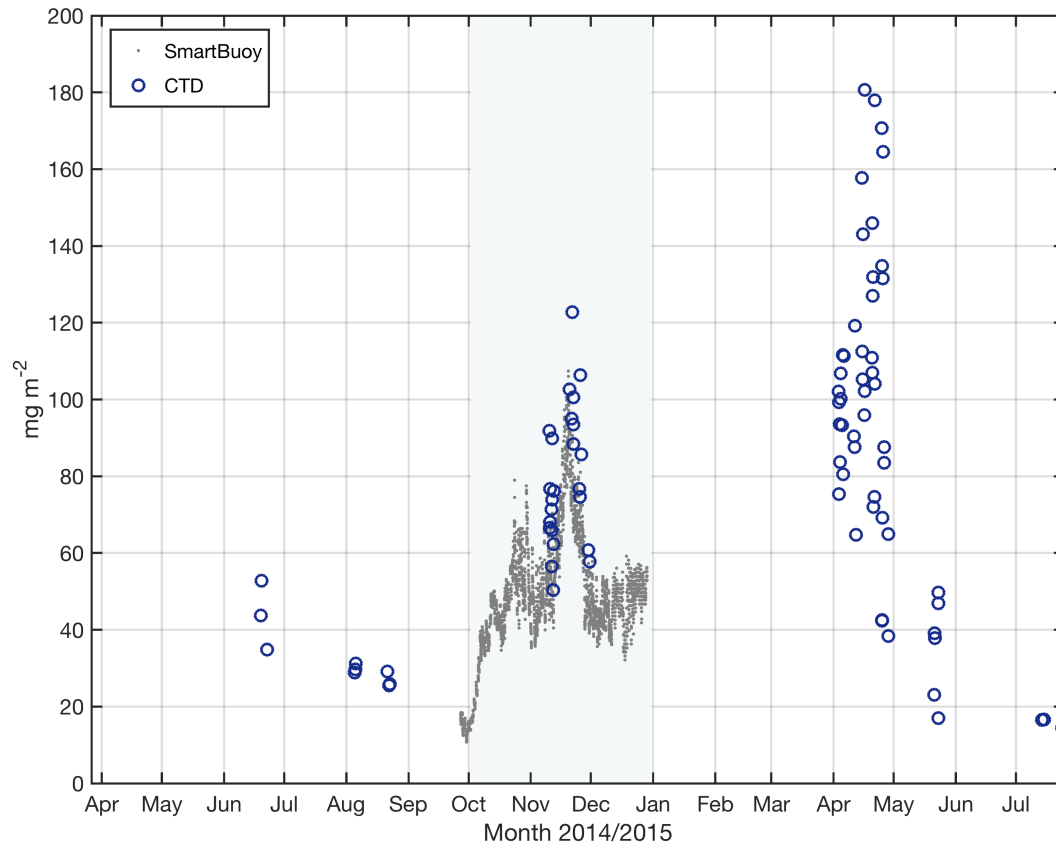


Figure 6: Depth integrated Chl *a* biomass [mg m^{-2}]. Markers denote the SmartBuoy platform and CTD derived values during the stratified periods of observations. The shaded area denotes the time of active SML deepening (October 2nd - December 31st 2014). For comparison we also included SmartBuoy data before the breakdown of stratification started.

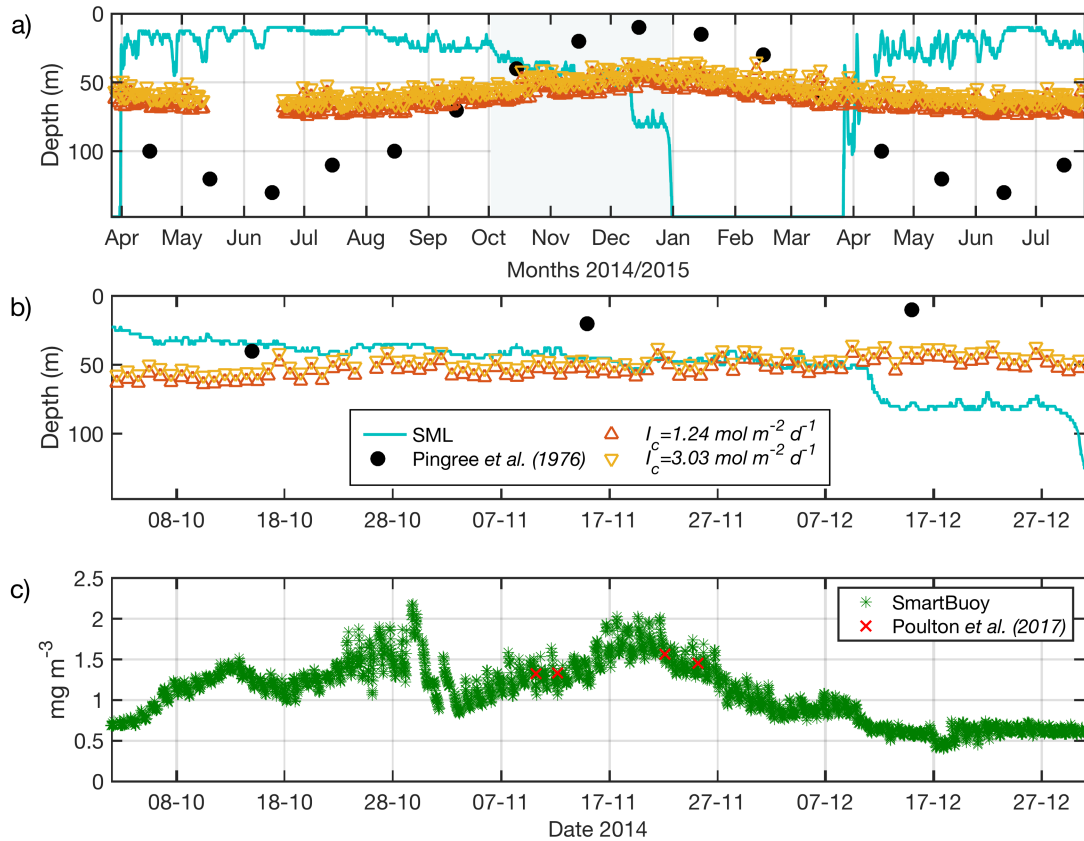


Figure 7: Critical depth hypothesis. a) Seasonal cycle of SML depth [m] (turquoise) compared to calculated values of z_{cr} using $I_c = 1.24 \text{ mol m}^{-2} \text{ d}^{-1}$ (orange), $I_c = 3.03 \text{ mol m}^{-2} \text{ d}^{-1}$ (yellow) and z_{cr} by Pingree *et al.* (1976) (black). The shaded area marks the time of active SML deepening (October 2nd - December 31st 2014). b) same as a) but focused on autumn period. c) surface Chl *a* fluorescence [mg m^{-3}] observed by SmartBuoy (green) and CTD bottle samples (red) by Poulton *et al.* (2017) during autumn period.

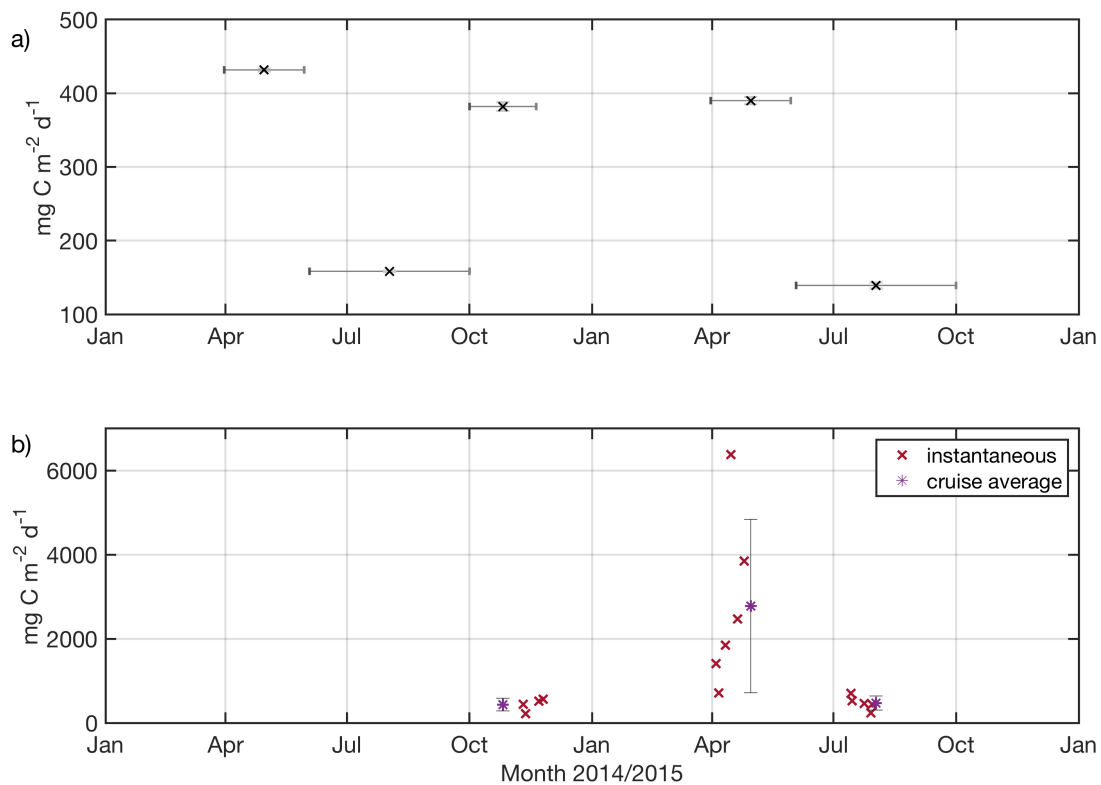


Figure 8: Rates of primary production [$\text{mg C m}^{-2} \text{d}^{-1}$] at CCS. a) gross (new) production, here horizontal bars show approximate duration of each seasonal state. b) instantaneous (red crosses) and cruise averages (purple stars) of net primary production obtained by Poulton *et al.* (2017). Vertical bars in both panels denote error estimates (1 standard deviation).

Highlights

- We present new observations of a full seasonal cycle of vertical density structure and its control on the seasonal cycle of primary production in a temperate shelf sea.
- Wind mixing appears to be the dominant SML deepening process.
- Surface mixed layer deepening in autumn replenishes surface nutrient concentrations, which fuels an autumn phytoplankton bloom.
- We show that Sverdrup's critical depth hypothesis can be used to predict the shut-down of primary production in autumn.
- The autumn phytoplankton bloom has the capacity to significantly contribute to the seasonal drawdown of atmospheric CO₂.