

Phytoplankton Bloom Phenomena in the North Atlantic Ocean and Arabian Sea

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ABSTRACT

We review bio-optical and physical data from three mooring experiments, the Marine Light-Mixed Layers program in spring 1989 and 1991 in the Iceland Basin (59°N/21°W), and the Forced Upper Ocean Dynamics Experiment in the central Arabian Sea from October 1994 to October 1995 (15.5°N/61.5°E). In the Iceland Basin, from mid-April to mid-June in 1989, chlorophyll-a concentrations are sensitive to small changes in stratification, with intermittent increases early in the record. The spring increase occurs after 20 May, coincident with persistent water column stratification. In 1991, the bloom occurs two weeks earlier than in 1989, with a background of strong short-term and diurnal variability in mixed layer depth and minimal horizontal advection. In the Arabian Sea, the mixing response to the northeast and southwest monsoons, plus the response to mesoscale eddies, produces four blooms over the annual cycle. The mixed layer depth in the Arabian Sea never exceeds the euphotic zone, allowing interactions between phytoplankton and grazer populations to become important. For all three mooring

30 experiments, changes in water column stratification are key in producing phytoplankton
31 blooms.

32

33 **1. Introduction**

34 There has been a resurgence of interest in Sverdrup's (1953) critical depth hypothesis in
35 recent years (e.g., Behrenfeld, 2010; Chiswell, 2011; Taylor and Ferrari, 2011), as
36 exemplified by a special session at a recent Ocean Sciences Meeting (February, 2014) and
37 by this issue of the Journal of Marine Science. The most recent contributions indicate an
38 on-going interest in what has been called 'one of the largest transient events on the
39 planet' (Gillis, 1991): the North Atlantic spring bloom is the rapid increase in
40 phytoplankton biomass that occurs in the spring months because of the interacting
41 conditions of high nutrients and the seasonal increasing solar irradiance.

42

43 There are few datasets that can be used to understand the initiation of the North Atlantic
44 spring bloom. In situ data on hydrographic structure and biological response are sparse
45 or non-existent for the winter or early spring, before seasonal re-stratification of the
46 water column. Here, we re-analyze observational data from two moorings deployed in
47 1989 and in 1991 in the Iceland Basin (59°N/21°W) (Fig. 1a) that help illuminate the
48 causes of the initiation of the North Atlantic spring bloom. We also use data from the
49 Forced Upper Ocean Dynamics Experiment in the Arabian Sea in 1994 to 1995 (15.5°N,
50 61.5°E) (Fig. 1b). The Arabian Sea is a monsoon-dominated regime, with two energetic
51 surface forcing seasons, and unlike the North Atlantic, has near-constant surface
52 irradiance over the year and a more modest seasonal temperature range (20-30°C). In

53 these respects the Arabian Sea serves as a useful contrast to the North Atlantic. The
54 moorings recorded both physical (e.g., irradiance, temperature) and biological (e.g.,
55 chlorophyll-a) properties at high temporal resolution (minutes, hours) allowing the
56 covariability between physical and biological properties to be examined as the structure
57 of the water column evolved from late winter to spring.

58

59 **2. Data Sources¹**

60 a. Marine Light – Mixed Layers, 1989, Iceland Basin

61 The first Marine Light – Mixed Layers (ML-ML) mooring experiment took place in spring,
62 1989 (hereafter, ‘MLML89’). The mooring was deployed from 15 April (year day, or yd
63 105) to 10 June (yd 161). Details of the mooring design, instrumentation, sensor
64 calibration, and methods can be found in Stramska and Dickey (1992) and Dickey et al.
65 (1994). Multi-Variable Moored Sensors (MVMS) (Dickey et al., 1994) were put at 10, 30,
66 50, 90, 150, and 250 meters. [See Table 1 for the array of sensors relevant to the data
67 discussed. Sensors at 70 and 110 m failed, and are not included in the analysis.] Bio-
68 optical observations at these depths included chlorophyll-a fluorescence, beam
69 attenuation coefficient, and photosynthetic active radiation (PAR). Temperature and
70 current velocities were measured at these same depths; salinity was measured only at 10
71 m and 250 m. Meteorological data were collected at a surface buoy. We do not have data
72 for all these variables for the full deployment period; but overall, the data retrievals were
73 good.

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¹ Much of the data are publicly available, with assistance from the authors, at <http://www.ldeo.columbia.edu/research/biology-paleo-environment/bioinfo>

75 b. Marine Light – Mixed Layers, 1991, Iceland Basin

76 The mooring for MLML91 collected data from 30 April (yd 120) until 6 September (yd
77 249), although we consider only data through yd 155. For MLML91, MVMS' were
78 deployed at 10, 30, 50, 70, and 90 m. Temperature sensors were placed approximately
79 every 20 m, starting at the surface, to 250 m (Table 1). Key references for background,
80 methods, instrumentation, and mooring design are Plueddemann et al. (1993),
81 Plueddemann et al. (1995), Stramska et al. (1995), and Weller et al. (1990).

82

83 c. Forced Upper Ocean Dynamics Experiment, 1994-1995, Arabian Sea

84 The mooring for the Arabian Sea was similar in design to that deployed in MLML91, with
85 MVMS units deployed at 10, 35, 65 and 80 m, and temperature sensors placed at depth
86 intervals of 5-10 m throughout the upper 250 m (Table 1). The mooring discussed here
87 (Fig. 1b) was at the central point of an array of 5 moorings, also part of the Forced Upper
88 Ocean Dynamics Experiment. There were two deployments of six months each, the first
89 from 16 October 1994 to 21 April 1995, and the second from 23 April to 25 October 1995.
90 Key references for this program are Rudnick et al. (1997), Dickey et al. (1998), Kinkade et
91 al. (2001) and Weller et al. (2002).

92

93 **3. Results**

94 a. MLML89

95 The time evolution of the temperature and chlorophyll-a distributions from the various
96 sampled depths are shown in Fig. 2. Temperature (Fig. 2a) is marked by periods of
97 intermittent stratification and isothermal layers, culminating in the springtime re-

98 stratification beginning about yd140 (May 20) and remains so for the remainder of the
99 record.

100

101 There are periods of stratification early in the series (April through early May), followed
102 by a few days where the sensors recorded nearly uniform temperatures, suggesting
103 vertical mixing. The water column then begins to re-stratify, punctuated by a shallow
104 mixing event (0-150 m) around May 18 (yd138). By yd145, the water column becomes
105 stratified and remains so for the remainder of the record. Winds and wind stress were
106 highly variable throughout the time series; the mixing event during May 18-May 20 (yd
107 138-140) was associated with wind stress event (see Dickey et al., 1994).

108

109 The chlorophyll-a record (Fig. 2b) essentially follows the temperature in character. There
110 are early periods of vertical stratification where chlorophyll-a increases at all depths
111 except the deepest (150 and 250 m), followed by a brief periods where chlorophyll values
112 are nearly uniform. After this, the spring increase is almost completely confined to the
113 shallowest three depths, 10, 30, and 50 m, with the values at 10 m exceeding for a short
114 time $3 \mu\text{g chlorophyll-a l}^{-1}$. The increases in near-surface chlorophyll-a near the end of the
115 record (late May, early June, where June 1 is yd 152) come at the expense of chlorophyll-a
116 deeper in the water column, an example of self-shading of the population (see Marra,
117 2004).

118

119 Contour plots are always an interpretation. The two episodic stratification events are
120 best viewed with time series plots of data from each of the depths to span the period

121 covering both events (Fig. 3). In both instances, the water column becomes progressively
122 warmer at all depths, and subsequently becomes more uniform in temperature,
123 producing near isothermal conditions as a function of depth. For the first event, the water
124 column becomes increasingly stratified until yd1113 (April 23), after which vertical mixing
125 occurs to 50 m, followed by a near-isothermal water column over the full range of
126 observed depths (10-250 m) by April 28. Chlorophyll-a concentrations (Fig. 3b) increase
127 coincident with stratification of the upper ocean at April 17 down to 90 m, and then
128 declines as the upper 50 meters become isothermal, with roughly constant levels (within
129 $\pm 0.1 \mu\text{g chlorophyll-a l}^{-1}$) by April 28, from 10-150 m. Following this episode,
130 chlorophyll-a at the shallowest three depths (10, 30, 50 m) increases until the next time
131 the water column mixes in mid-May.

132

133 It is likely that the increases in chlorophyll-a in the near-surface records (10, 30, and 50
134 m) are from phytoplankton growth. Certainly, estimates of the increase in chlorophyll-a,
135 of about a doubling per day, are reasonable. Deeper than this the evidence is less clear. As
136 the water column becomes isothermal, chlorophyll-a concentration at all depths declines
137 with the exception of the two deepest depths (150 and 250m), perhaps because
138 concentrations at these depths increase as a result of mixing from above. Integrated over
139 10-250 m, however, chlorophyll-a declines by 30% from April 24-29 (data not shown).
140 The increases shown for the 150 and 250 m depths are not enough to compensate for the
141 losses shallower in the water column. The integrals may be inaccurate because of the
142 limited depth resolution, because chlorophyll had been mixed deeper than our deepest
143 sensor (250 m), or else because local currents (which were otherwise relatively weak

144 during this period: Dickey et al., 1994) play a role in the variability. [Unfortunately,
145 satellite altimetry data are unavailable for the time period of our experiment, and
146 satellite sea surface temperatures are not sufficiently resolved to identify mesoscale
147 features. See Dickey et al., 1994.] The water column becomes isothermal over all depths
148 by April 28 (yd118), and chlorophyll-a does as well. Nevertheless, temperature
149 differences of less than a few tenths of a degree between adjacent instruments allows an
150 increase chlorophyll-a concentrations by a factor of two during the first half of the record.

151

152 Daily PAR in the water column is shown in Fig. 4. It is fairly constant at 20-30 mol
153 photons $\text{m}^{-2} \text{d}^{-1}$ until the wind event (May 18-May 20), and declines thereafter as the
154 bloom develops. Daily-averaged net heat flux is also relatively constant over the course of
155 the experiment (see Dickey et al., 1994).

156

157 b. MLML91

158

159 The experiment in 1991 shows similar temperature variability, but very different
160 biological behavior than in 1989 (Fig. 5). We had the benefit of extensive sampling from
161 shipboard that we did not have in MLML89, and that allowed phytoplankton and
162 zooplankton species to be identified.

163

164 Compared to 1989, the major chlorophyll-a increase over the water column occurred
165 much earlier in 1991, and during the incipient stratification from yd125 to yd140 (May 5
166 - 20). The first stage of the bloom was interrupted by a storm (yd128-130, May 8 - 10),

167 as recorded at the mooring's surface meteorological sensors (Plueddemann et al., 1995).
168 Shipboard observations during two ML-ML cruises, covering the period May 7 -24 (see
169 Marra et al., 1995), identified the bloom occurring then as *Phaeocystis pouchetii*, an
170 unusual occurrence south of Iceland. After a major storm on May 22 (yd142), with >70
171 knot (35 m s^{-1}) winds, the *Phaeocystis* bloom was no longer observed. The water column
172 quickly stratified under clear skies (total daily irradiance of $60 \text{ mol photons m}^{-2} \text{ d}^{-1}$:
173 Marra et al., 1995), and diatoms and copepods appeared in the surface layers (Cowles
174 and Fessenden, 1995). Chlorophyll-a levels had declined by nearly an order of magnitude
175 after the storm, and recovered only partially thereafter. At this time, nitrate
176 concentrations in the upper water column ranged from 6-8 μM (Marra et al., 1995), and
177 therefore nutrients were likely not a factor in the development of the bloom.

178

179 c. Forced Upper Ocean Dynamics Experiment, Arabian Sea, 1994-1995

180

181 The Arabian Sea mooring experiment extended an entire year, from October 1994 to
182 October 1995. The seasonal cycle is dominated by monsoons, with within-season effects
183 from mesoscale eddies. The moored observations captured both the northeast
184 (December 1994 - February 1995) and southwest (June-August 1995) monsoons. The
185 seasonal cycle in the vertical structure of temperature and areal chlorophyll-a (Fig. 6) is
186 reported in Kinkade et al. (2001), compiled from both mooring and shipboard data. This
187 is the most complete description of the physical setting and biological response
188 determined for the Arabian Sea. In the one-dimensional sense, convective cooling occurs
189 during the northeast monsoon, as the winds from Asia blow over the northern Arabian

190 Sea (Fischer et al., 2002). The result is vertical, convective, mixing resulting in deepening
191 the mixed layer. In contrast, the mixing during the southwest monsoon is mechanical, and
192 caused by wind stress at the surface. In neither case, however, does the mixing extend to
193 great depth, never reaching > 100 m, and perhaps never exceeding the euphotic zone
194 (Barber et al. 2001). The biological response to the southwest monsoon was greater than
195 for the northeast monsoon, but for reasons that are not clear (Barber et al., 2001).

196

197 It is possible that enhanced activity at the mesoscale, associated with the southwest
198 monsoon, provides a source of nutrients (Keen et al., 1997) to support the greater
199 productivity during the southwest monsoon. Fischer et al. (2002), using data from the
200 array of 5 moorings and data from Sea Soar tows, show that cool, and nutrient-rich water
201 is brought from the coastal upwelling regime out to the central mooring. In the fall of
202 1994, there is evidence of two mesoscale eddies at the mooring location (Fig. 6a), one
203 when the mooring was first deployed, and the second about a month later. In these cases
204 the increase in chlorophyll-a is associated with relatively shallow mixed layers, and the
205 strong temperature contrast with depth suggests an enhanced nutrient supply that
206 supports surface productivity. Over the year, there are four episodes of phytoplankton
207 increases observed in the Arabian Sea, two occurring as a result of each of the monsoon
208 periods, and two associated with the passage of mesoscale eddies in the fall of 1994.

209

210 **4. Discussion**

211 Sverdrup's (1953) critical depth hypothesis is a mathematical derivation of ideas
212 presented earlier by Gran and Braarud (1935) and Riley (1942). These earlier authors

213 recognized that since autotrophic respiration (R) was, perhaps, 20% of gross production
214 (G), the phytoplankton cells could be mixed to five times the depth of the compensation
215 irradiance (i.e., where $G = R$) without incurring a population decline (see Mills, 1989).
216 Conceptually, this is sound. The balance between net production by phytoplankton and
217 the vertical mixing that they were subject to, should determine the fate of phytoplankton
218 populations in a variety of environments, and especially in the late winter North Atlantic
219 where Sverdrup (1953) tested his ideas. The North Atlantic, unlike many other ocean
220 regimes, convectively mixes during the winter, producing mixed layers >600 m deep
221 (Robinson et al., 1979; Monterey and Levitus, 1997), and in effect, resetting the seasonal
222 production cycle.

223

224 The critical depth hypothesis has achieved wide acceptance, becoming a core tenet of
225 biological oceanography, but it has had few, if any, actual tests. Later, there were
226 critiques. Smetacek and Passow (1990) note that actual critical depths may exceed
227 wintertime mixing depths, anyway, and further that phytoplankton communities are
228 capable of overwintering (see, e.g., Backhaus et al., 2003). They also remind that
229 stratification occurs from the surface, and that there are difficulties in accounting for
230 other losses (e.g., zooplankton grazing). They conclude that the critical depth hypothesis
231 serves little practical value. Behrenfeld (2010) takes a similar position, noting that there
232 is significant biomass during the winter months in the boreal North Atlantic, and
233 highlights the balance between growth and grazing. This balance would be disrupted by
234 seasonal re-stratification in favor of a spring bloom. Behrenfeld's (2010) evidence,
235 however, comes from satellite ocean color estimates of chlorophyll-a, which can be

236 subject to error during winter at high latitudes, and carries no information on vertical
237 structure.

238

239 While we do not have wintertime data, our records for the North Atlantic are from before
240 the seasonal re-stratification of the water column, and indicate the initiation of the spring
241 bloom. We cannot provide a direct test of the critical depth hypothesis, however we can
242 point to factors that are important for bloom initiation. For the Arabian Sea, we show
243 how mixed layer dynamics affect chlorophyll-a for a situation where the mixed layer
244 never exceeds the critical depth.

245

246 a. North Atlantic

247

248 For both the 1989 and 1991 programs, the moorings provide a good realization of the
249 temporal dynamics of bloom development for the Iceland Basin in the North Atlantic.
250 Dickey et al. (1994) note that current velocities during the re-stratification period were
251 weak and barotropic. For the 1991 experiment, Plueddemann et al. (1995) show that the
252 springtime evolution of the temperature structure in the upper water column could be
253 successfully simulated by a one-dimensional mixed layer model. Thus, for 1989 and 1991,
254 we conclude that we observed the temporal evolution of the spring bloom.

255

256 It is clear from the time series observations of temperature and chlorophyll-a that
257 stratification of the water column instigated the bloom. In 1989, there are two early,
258 incipient episodes of stratification that elicit a chlorophyll-a response, followed by

259 vertical mixing (isothermal periods). When the water column becomes persistently
260 stratified, chlorophyll-a increases in the top 150 m. In both 1989 and 1991, there are
261 periods of weak stratification interspersed with water column mixing, the result of wind
262 events (1989) or major storms (1991), with the stratified periods associated with strong
263 increases in chlorophyll-a. The timing of the blooms in 1989 and 1991 are within a week
264 or so of each other. The seasonal re-stratification (Figs. 2, 5) begins at yd135-140 (May
265 15-20) in 1989, and yd140-143 (May 20-23) in 1991. Stramska and Dickey (1993) have
266 pointed out that chlorophyll-a is very sensitive to water column stratification. A vertical
267 difference of a few tenths of a degree Celsius within the top layers is enough to stabilize
268 the surface layers and thereby allow increases in chlorophyll-a. Similarly, mixing those
269 top layers disperses the chlorophyll-a.

270

271 We can try to interpret our data using the critical depth idea, to make an inference as to
272 what happens earlier in the season (prior to our observations), and whether it is a useful
273 guide for the initiation of the bloom. We use a simplification of Sverdrup's (1953)
274 equation, presented in Lalli and Parsons (1993), whereby

275

$$276 \quad Z_{cr} = E(0)/(k \bullet E_c).$$

277

278 Here, Z_{cr} is the critical depth, $E(0)$ is the solar irradiance penetrating the surface, k is the
279 vertical diffuse attenuation coefficient for PAR, and E_c the compensation irradiance. The
280 1989 observations show that $E(0)$ is about 20-30 mol photons $m^{-2} d^{-1}$ (Fig. 4), with a
281 vertical diffuse attenuation coefficient of 0.06 m^{-1} . Marra (2004) has estimated that E_c for

282 the data from 1989 is 0.1-0.3 mol photons $\text{m}^{-2} \text{d}^{-1}$ based on the time and depth changes in
283 chlorophyll-a at the height of the bloom. If we assume a surface irradiance of 10 mol
284 photons $\text{m}^{-2} \text{d}^{-1}$ to represent times earlier in the year than our observations, then the
285 critical depth will have a calculated range from 600 to 1600 m. This depth range is large,
286 but comparable to, or greater than, prior estimates of the late-winter mixed layer depths
287 (Monterey and Levitus, 1997; see also Smetacek and Passow, 1990). The depth range also
288 suggests that the critical depth will be greater than the mixed layer depth for the
289 observations in MLML89 and MLML91. As the season progresses, and $E(0)$ increases, the
290 critical depth will increase accordingly. We do not have data for the late winter
291 population of phytoplankton, however, it is unlikely that the balance between
292 photosynthesis and respiration, as suggested by Sverdrup (1953) is a primary factor to
293 the initiation of the bloom.

294

295 Thus, we conclude that seasonal re-stratification of the water column drives the change
296 in chlorophyll-a concentration in surface waters from midwinter lows to springtime
297 highs. The change in stratification may be episodic, and the spring bloom may accordingly
298 proceed episodically, and be spatially heterogeneous. We recognize that there are other
299 processes at work that we cannot account for with our dataset. For example, it is possible
300 that there is advection of more stratified water past the mooring. However 1-D models
301 (Plueddemann et al., 1995 Stramska and Dickey, 1994, Stramska et al., 1995) support the
302 notion that air-sea interaction processes could explain the change in water stratification
303 that we observe. Recently, Mahadevan et al. (2012) suggest that mixed layer 'slumping'
304 produces stratification that can affect the timing of the spring bloom. Our data do not

305 contradict that assessment. Our data support the view that it is a decrease in mixing
306 intensity followed by development of water column stratification that is the driver for the
307 changes in chlorophyll-a concentration observed.

308

309 Another control on biomass (although not on physiology) is grazing. Sverdrup (1953)
310 considered this as a significant loss term in his analysis of the spring bloom, and it has
311 been considered so in subsequent treatments (e.g., Platt et al., 1991; Behrenfeld, 2010).
312 Behrenfeld (2010) regards grazing as a strong control of overwintering phytoplankton
313 populations in the North Atlantic. Although we don't have the appropriate measurements
314 to assess the impact of grazing for the 1989 dataset, there are shipboard observations for
315 the MLML91 experiment, for the period of May 17-24 (yd137-144). During the
316 *Phaeocystis* bloom, there are few mesozooplankton, but they become abundant after the
317 seasonal stratification, after May 23 (Cowles and Fessenden, 1995). Despite that
318 abundance, these authors calculate that the grazing impact by copepods after 23 May is
319 about 5% of primary production. The impact of grazing by microzooplankton is higher,
320 but as well, only after the onset of persistent stratification (Gifford et al., 1995). To be
321 sure, there are not data earlier in the season than this, but the sequence suggests that
322 grazing only becomes important after the bloom is established, and plays little role in the
323 initial stages. Finally, incubation estimates of primary production (^{14}C assimilation and
324 ΔO_2), are roughly comparable to changes in in situ biomass, calculated from diurnal
325 changes in beam attenuation (Marra et al., 1995). This is further evidence that grazing is
326 not a large loss process during the spring bloom in the North Atlantic.

327

328 b. Arabian Sea

329 In the Arabian Sea, unlike in the North Atlantic, there is not a dominant annual cycle.
330 Incoming shortwave radiation, expressed as PAR, averages $\sim 30 \pm 5$ mol photons $m^{-2} d^{-1}$
331 over the year, with summertime values dropping significantly below that because of
332 cloudiness (Marra et al., 1998; Weller et al., 1998). Sea surface temperature lacks a
333 summer peak, instead showing cooling during both the summer and winter (Weller et al.,
334 1998) from the effects of the two monsoons. The Arabian Sea, like most other ocean
335 areas, also has mesoscale variability and advective transports that are at times significant
336 contributors to local water column properties. Like many tropical areas (e.g., Lukas and
337 Lindstrom, 1991), vertical mixing in the Arabian Sea does not extend deeper than 100 m
338 (Dickey et al., 1998; Weller et al., 2002). Because of these attributes, we can expect that
339 biological interactions, most prominently zooplankton grazing on phytoplankton, will be
340 relevant to the dynamics in the Arabian Sea.

341

342 One way to ascertain the importance of grazing is to compare photosynthetic carbon
343 assimilation in incubations (Barber et al., 2001) with the variability of chlorophyll-a from
344 the moored fluorometers (Fig. 6a). The biggest change in in situ chlorophyll-a occurs with
345 the southwest monsoon. The data are scattered towards the end of the monsoon period,
346 but the realized increase in chlorophyll-a from July to September (~ 90 days) is 0.3-0.5
347 $mg m^{-2} d^{-1}$. Assuming a (high) carbon:chlorophyll-a ratio of 100, this daily increase
348 amounts to 30-50 $mg C m^{-2} d^{-1}$, about a factor of 20 smaller than the daily photosynthetic
349 carbon assimilation measured during the southwest monsoon (Barber et al., 2001). Still,
350 zooplankton grazing does not completely regulate phytoplankton biomass in the Arabian

351 Sea; biomass variations are observed. The argument made in Marra and Barber (2005)
352 and also in Marra and Moore (2009) is that the monsoons affect more strongly micro-
353 grazer populations than the phytoplankton, since the phytoplankton are not mixed to
354 depths greater than their capacity for positive production. Thus, in the Arabian Sea, we
355 have a situation where vertical mixing, may apply more to heterotrophs than autotrophs.

356

357 In conclusion, our observations from moorings in the Iceland Basin (1989, 1991) support
358 the idea that water column stratification is the trigger for the spring phytoplankton
359 bloom in the North Atlantic. Prior to the observed springtime increase, calculations of the
360 critical depth suggest that the phytoplankton are capable of maintaining themselves
361 through the winter, with perhaps small episodic increases of biomass with transient
362 stratification events. The Arabian Sea serves as a useful contrast to the North Atlantic, in
363 that vertical mixing does not extend beneath the euphotic zone (and certainly not to a
364 critical depth). Thus, for the Arabian Sea, vertical mixing affects grazer populations to a
365 greater extent than the phytoplankton.

366

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471 the Arabian Sea during 1994-1995, *Deep Sea Research II*, 49, 2195-2230.

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480 Figure Captions

481 Fig. 1. (a) Site of the Marine Light – Mixed Layers ('ML-ML') mooring experiments in
482 1989, 1991. (b) Map showing the site of the Forced Upper Ocean Dynamics
483 Experiment in 1994-1995. The mooring data comes from the marker designated as
484 'WHOI' at 15.5°N/61.5°E (see Rudnick et al., 1997).

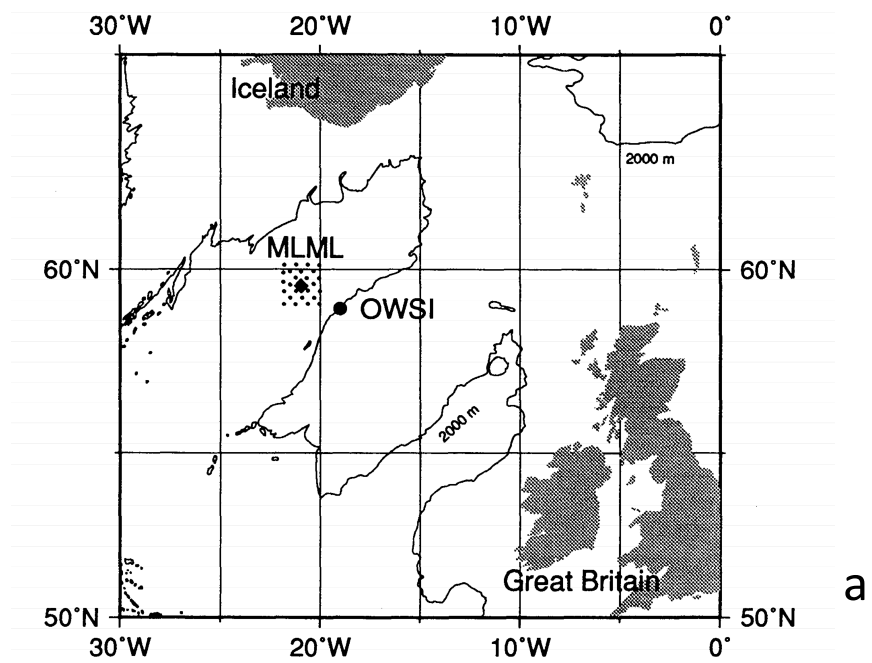
485 Fig. 2. Contour plots of (a) temperature (°C) and (b) chlorophyll-a ($\mu\text{g l}^{-1}$) for the MLML89
486 mooring experiment.. There are no data from 110 m for chlorophyll-a.. Time is
487 presented as calendar dates (mm/dd) in (a) and as yeardays in (b).

488 Fig. 3. Time series from individual depths for mid-April to mid-May in Fig. 2 with (a)
489 temperature (°C) and (b) chlorophyll-a ($\mu\text{g l}^{-1}$). The data have been smoothed, for
490 clarity. The unsmoothed data are shown in Marra (2004).

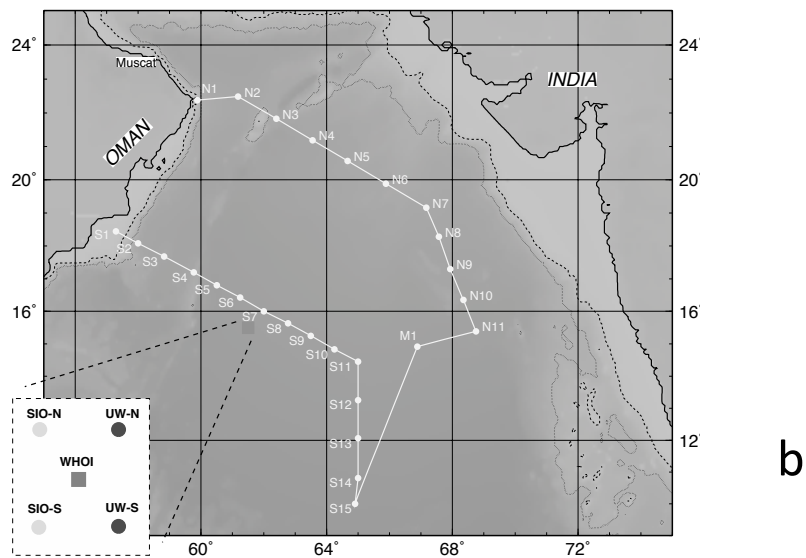
491 Fig. 4. Daily PAR in the water column. $E(0)$ is calculated as the intercept from the linear
492 regression of PAR (z).

493 Fig. 5. Contour plots of (a) temperature (°C) and (b) chlorophyll-a ($\mu\text{g l}^{-1}$) for the MLML91
494 springtime data. There is more detail in the temperature plot because of the higher
495 sampling resolution with depth. Depths for temperature sensors begin at the surface,
496 and then with ~ 20 m spacing until 310 m. Depths for chlorophyll-a sensors are 10, 30,
497 50, 70, and 90 m. Time is presented as calendar dates (mm/dd) in (a) and as yeardays
498 in (b).

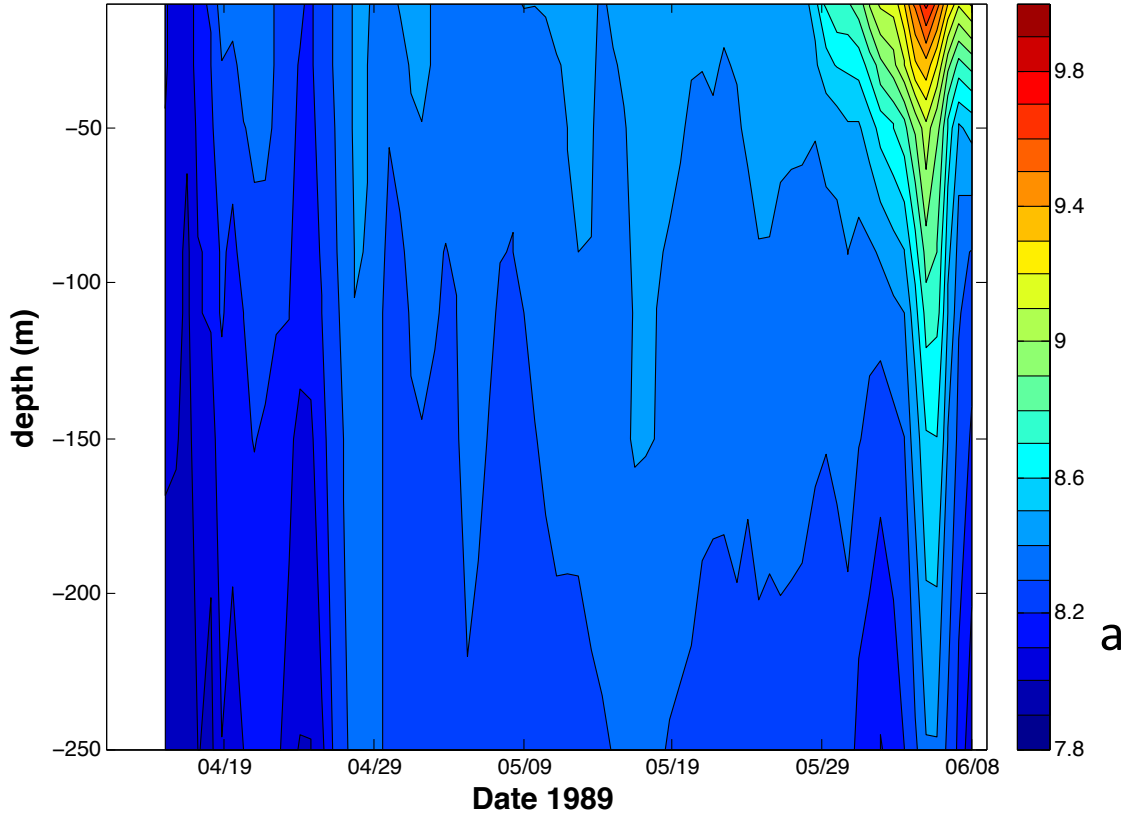
499 Fig. 6 (a) Contour plot of temperature overlain with areal chlorophyll-a (C_{tot} , mg m^{-2})
500 from the mooring (open circles) and from shipboard hydrographic casts (filled
501 circles), and (b) mixed layer depth. Mixed layer depths were estimated as a
502 temperature difference of 0.1°C from the surface temperature.



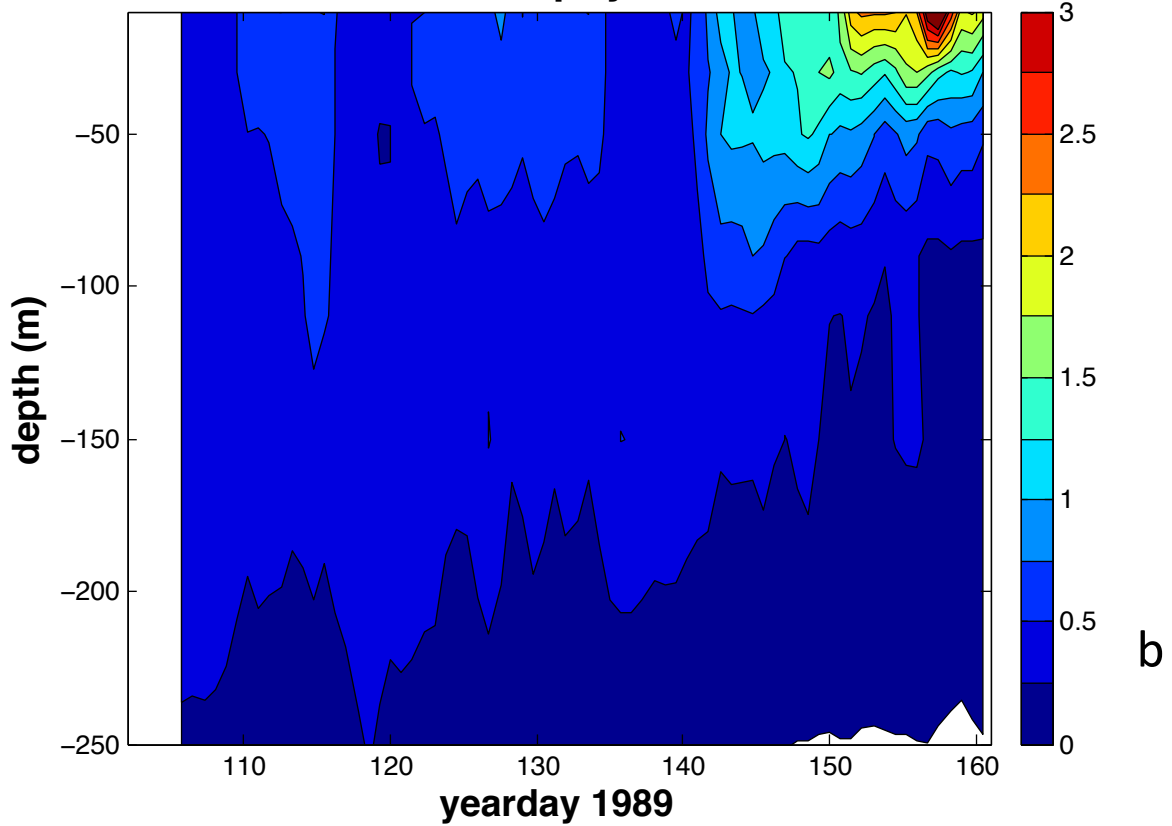
THE ARABIAN SEA EXPEDITION, 1995

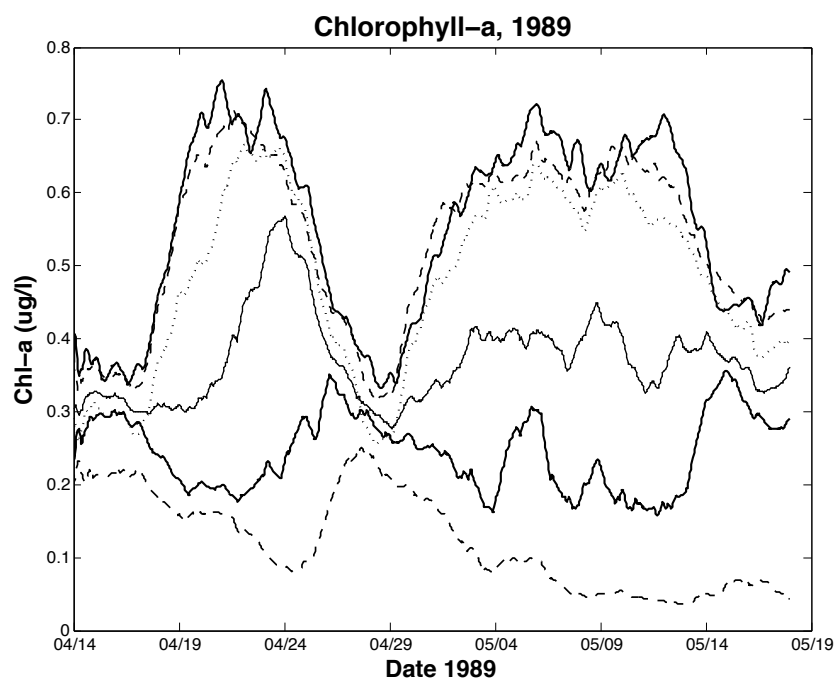
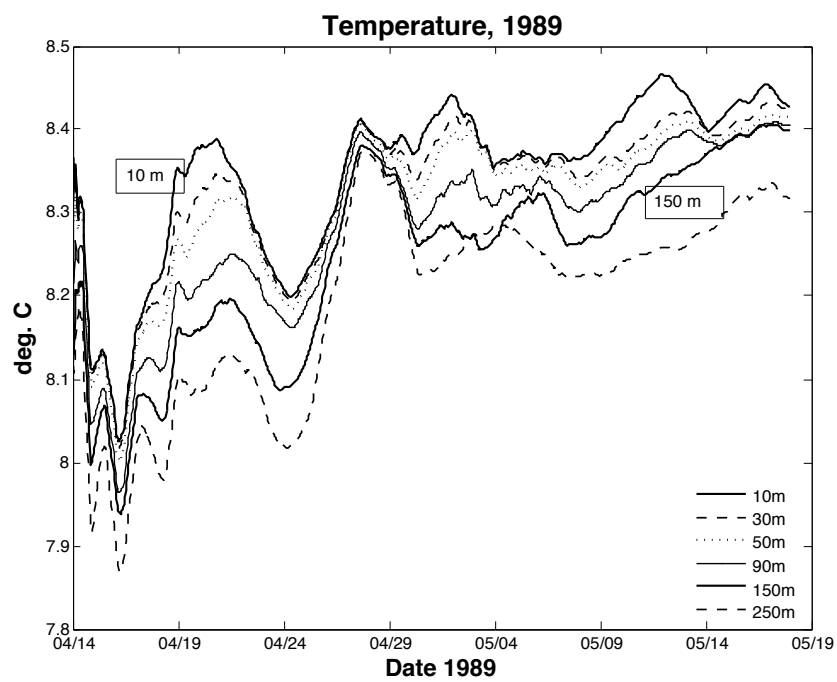


Temperature

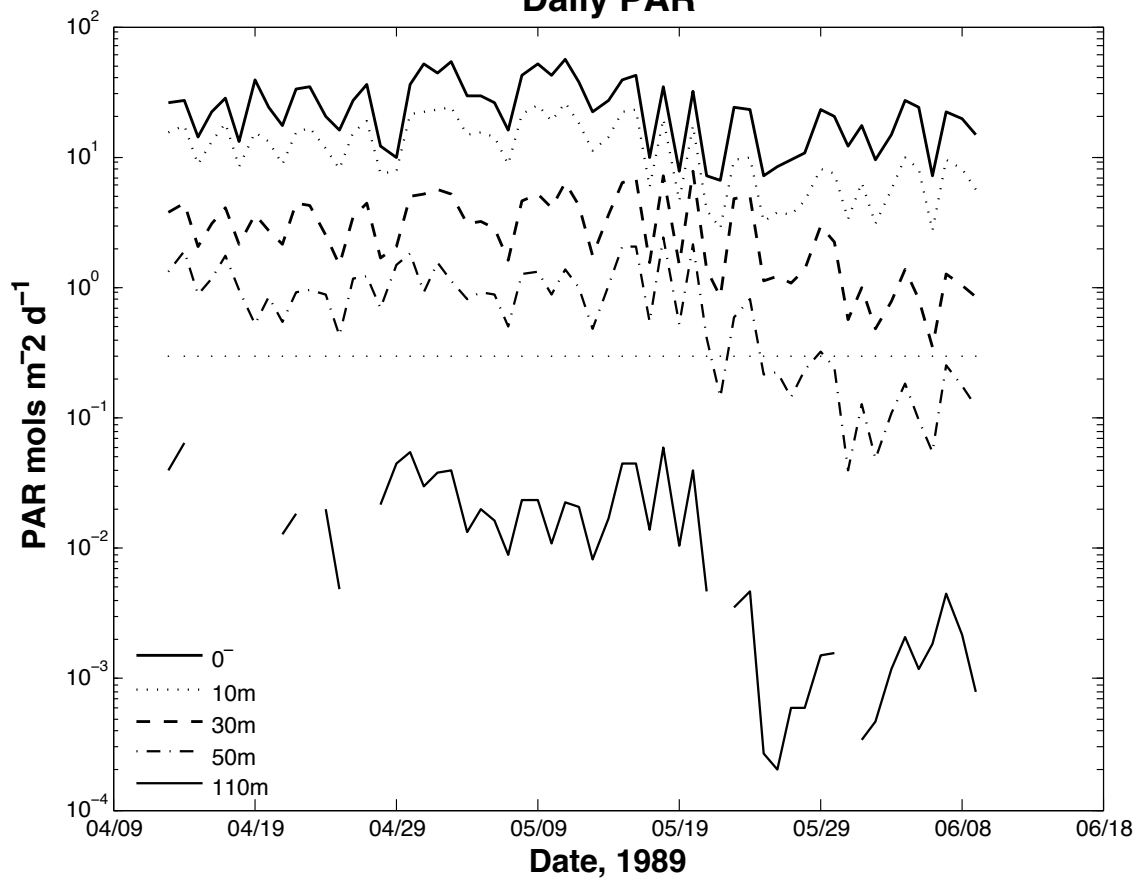


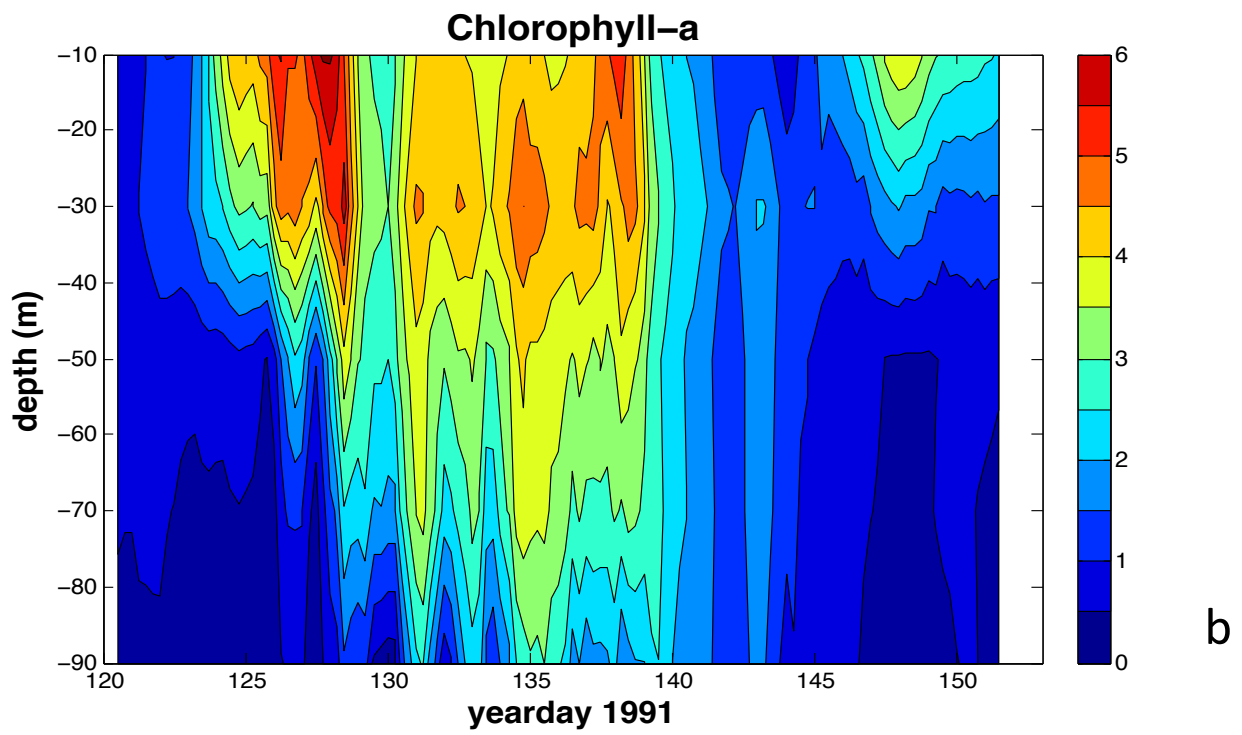
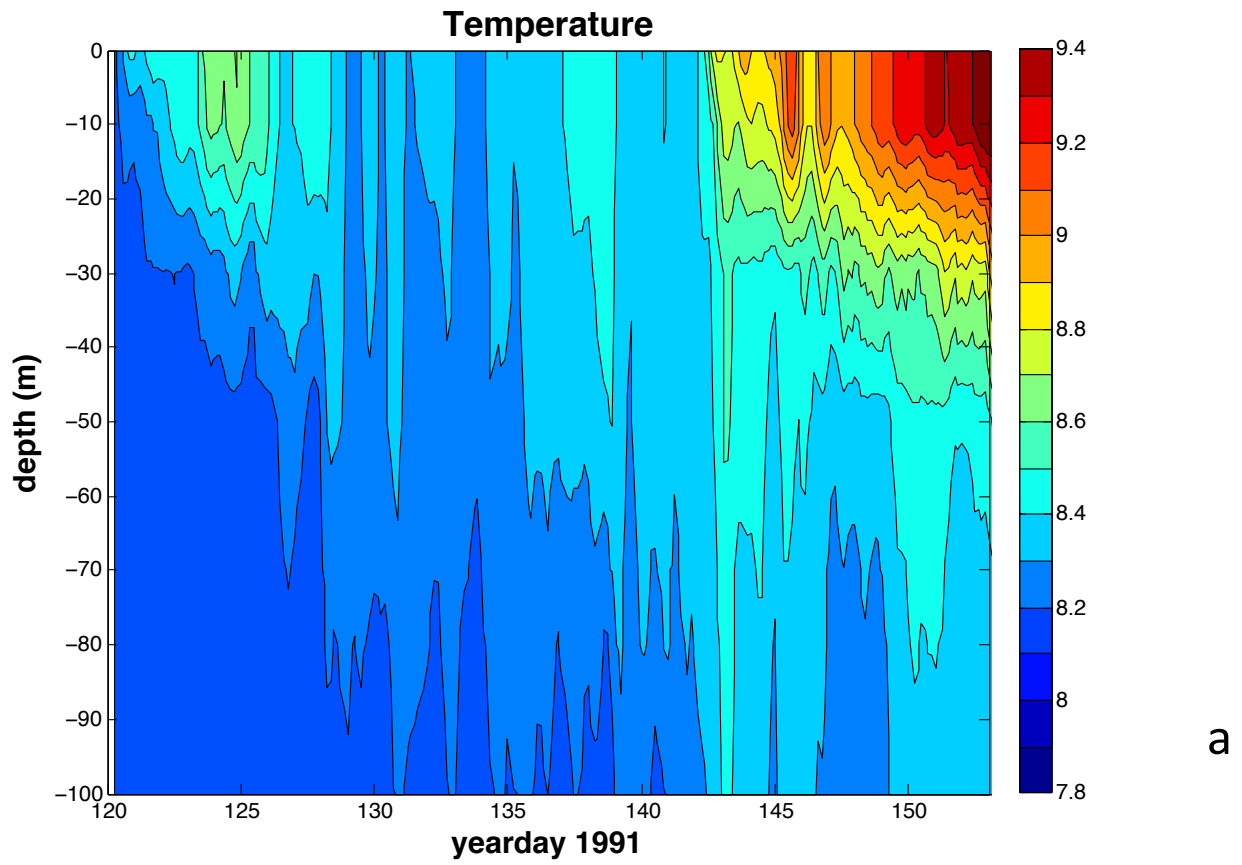
Chlorophyll-a

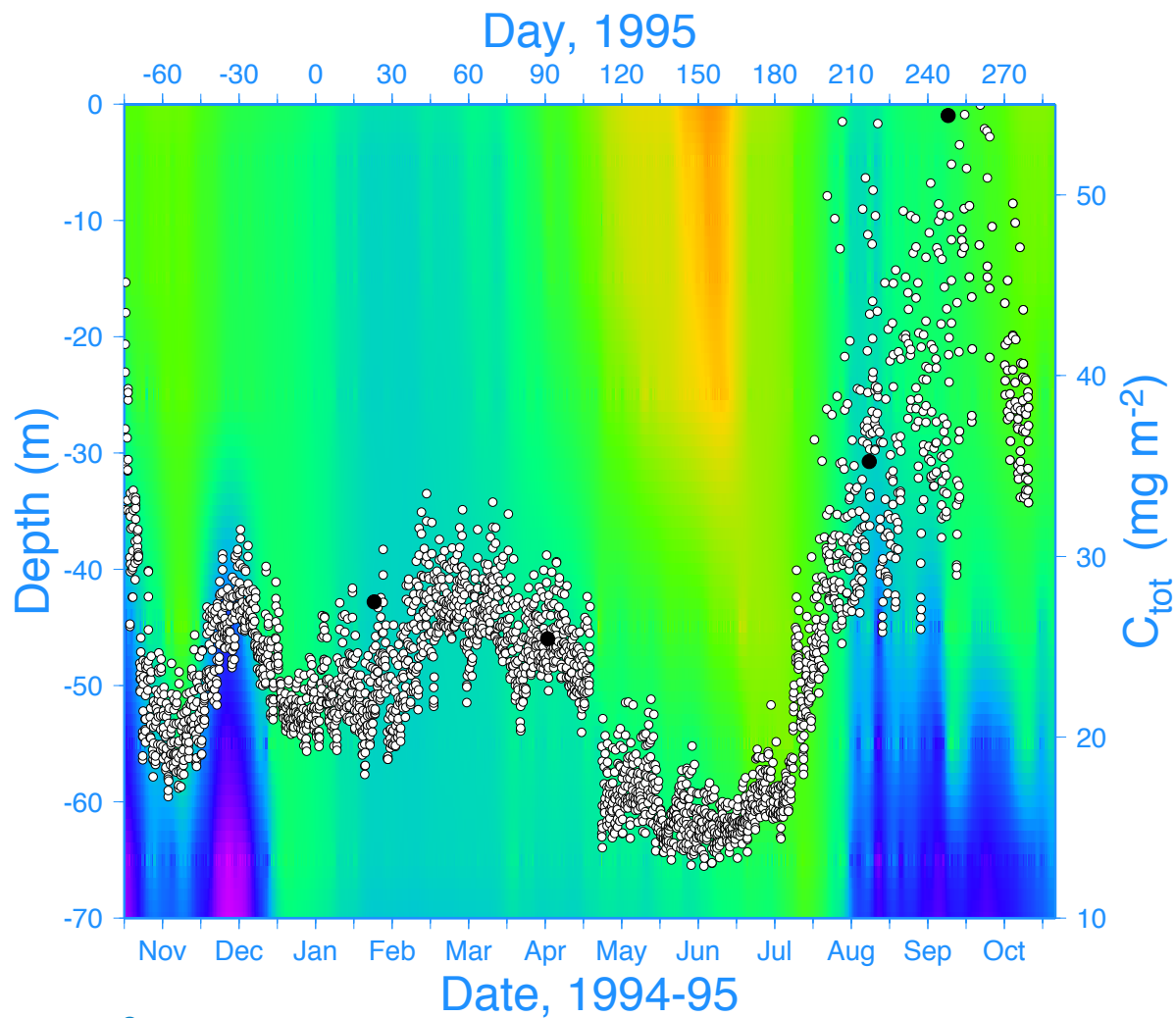




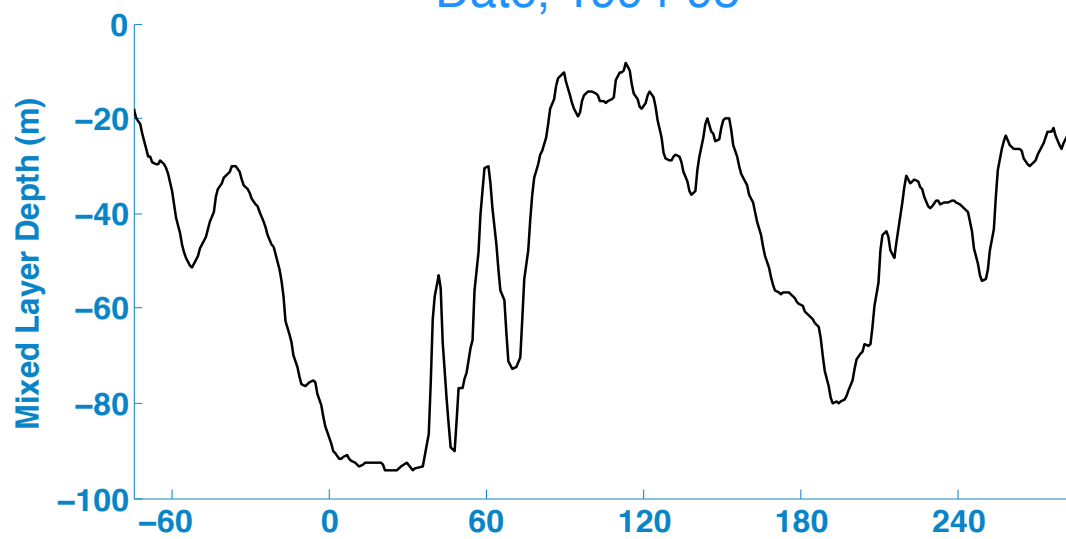
Daily PAR







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