



Observations of open-ocean deep convection in the northwestern Mediterranean Sea: Seasonal and interannual variability of mixing and deep water masses for the 2007-2013 Period

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1 OBSERVATIONS OF OPEN-OCEAN DEEP CONVECTION IN THE NORTHWESTERN
2 MEDITERRANEAN SEA : SEASONAL AND INTERANNUAL VARIABILITY OF MIXING
3 AND DEEP WATER MASSES FOR THE 2007-2013 PERIOD

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20 Key points

- 21 - Continuous monitoring of 5 consecutive winters of deep convection that reached the
22 seabed, found at a depth of 2300m
- 23 - Observation of the overlapping of the deep convection phases (vertical mixing /
24 restratification; spreading/preconditioning)
- 25 - Formation of warmer and saltier new deep water mass “vintages” after each event
26 of deep convection

27

28 Abstract

29 *We present here a unique oceanographic and meteorological dataset focus on*
30 *the deep convection processes. Our results are essentially based on in situ data*
31 *(mooring, research vessel, glider, and profiling float) collected from a multi-platform*
32 *and integrated monitoring system (MOOSE: Mediterranean Ocean Observing System*
33 *on Environment), which monitored continuously the northwestern Mediterranean Sea*
34 *since 2007, and in particular high-frequency potential temperature, salinity and*
35 *current measurements from the mooring LION located within the convection region.*

36 *From 2009 to 2013, the mixed layer depth reaches the seabed, at a depth of*
37 *2330m, in February. Then, the violent vertical mixing of the whole water column lasts*
38 *between 9 and 12 days setting up the characteristics of the newly-formed deep*
39 *water. Each deep convection winter formed a new warmer and saltier “vintage” of*
40 *deep water. These sudden inputs of salt and heat in the deep ocean are responsible*
41 *for trends in salinity ($3.3 \pm 0.2 \cdot 10^{-3}/\text{yr}$) and potential temperature (3.2 ± 0.5*
42 *$\cdot 10^{-3} \text{ }^{\circ}\text{C}/\text{yr}$) observed from 2009 to 2013 for the 600-2300m layer.*

43 *For the first time, the overlapping of the 3 “phases” of deep convection can be*
44 *observed with secondary vertical mixing events (2-4 days) after the beginning of the*
45 *restratification phase, and the restratification/spreading phase still active at the*
46 *beginning of the following deep convection event.*

47

48

49

50

51 Index terms

52 4572 Upper ocean and mixed layer processes

53 4504 Air/sea interactions

54 4262 Ocean observing systems

55 4277 Time series experiments

56 4283 Water masses

57 4243 Marginal and semi-enclosed seas

58

59 Keywords

60 Physical Oceanography; Ocean Observations; Dense Water Formation; Open Ocean

61 Deep Convection; Mixed Layer; Gulf of Lions; Deep water; Mediterranean Sea

62

63

64

65

67 1. Introduction

68

69 Open-ocean deep convection is a key process that transfers the heat and salt
70 content of the surface layer to the deep ocean and only takes place in a few regions
71 of the world's oceans. In addition to polar regions such as the Labrador [Lazier,
72 1973; Clarke and Gascard, 1983], the Greenland [Schott et al., 1993], the Weddell
73 and Ross Seas [Killworth, 1983], open-ocean deep convection can occur in mid-
74 latitude regions such as the East/Japan Sea [Kim et al., 2008] and the
75 Mediterranean Sea (Gulf of Lions, Adriatic Sea, Aegean Sea, [CIESM, 2009]). In the
76 Gulf of Lions area (northwestern Mediterranean), a basin-scale cyclonic circulation
77 drives a doming of isopycnals, which helps of the deepening of the mixed layer
78 during intense surface buoyancy loss associated with intense and dry northerly
79 winds (Mistral, Tramontane). The doming of the isopycnals due to the cyclonic gyre
80 and the high surface buoyancy loss can lead to deep mixing and to the formation of
81 the Western Mediterranean Deep Waters (WMDW). The pioneer experiment
82 dedicated to the study of open-ocean deep convection in the Gulf of Lions identified
83 three phases of convection (the preconditioning phase, the intense vertical mixing
84 phase, and the restratification phase with the spreading of the newly formed deep
85 water) [MEDOC Group, 1970]. Recently, several studies illustrate the role of deep
86 convection on transfers of biogeochemical components to the deep water such as
87 oxygen and inorganic and organic matter, contributing to the ventilation and the
88 “feeding” of the deep pelagic and benthic ecosystems [Stabholz et al., 2013;
89 Tamburini et al., 2013]. The deep vertical mixing also impacts the upper-ocean
90 through the supply of nutrients from the deep ocean to the euphotic layer [Severin et
91 al., 2014], and is a control parameter of the phytoplankton phenology [Lavigne et al.,
92 2013].

93 The significant interannual variability of the convective activity in the Gulf of
94 Lions [Mertens and Schott, 1998; L'Hévéder et al., 2012] leads also to an interannual
95 variability of thermohaline characteristics of the WMDW. The evolution of the WMDW
96 thermohaline characteristics formed by deep convection could explain the long-term

97 warming and salting trends observed in the deep layers of the Western
98 Mediterranean (WMED) by several authors [Bethoux et al., 1990, 1998; Leaman and
99 Schott, 1991; Rohling and Bryden, 1992; Krahmann and Schott, 1998; Rixen et al.,
100 2005; Vargas-Yanez et al., 2010a, 2010b]. The storage of heat, salt, and other
101 properties in the deep layers and their distribution over time and space are of crucial
102 importance for the evolution of this physical-biogeochemical coupled system over the
103 long term. It is crucial to understand and correctly model this phenomenon in the
104 context of climate change studies.

105 The deep convection phenomenon is very variable as illustrated by the
106 abrupt changes in WMDW characteristics and stratification, referred as the Western
107 Mediterranean Transition [CIESM, 2009], which occurred during winters 2004-2005
108 and 2005-2006 [López-Jurado et al., 2005; Schroeder et al., 2006, 2008a; Font et
109 al., 2007; Smith et al., 2008, Zunino et al., 2012]. These changes were related to
110 intense deep water formation events that occurred in the Gulf of Lions region and in
111 the Ligurian subbasin. Moreover, different studies highlight that dense shelf water
112 formation and cascading through the canyons of the Gulf of Lions and Catalan
113 margin can be sometimes important in changing the WMDW stratification and the
114 bottom water mass characteristics, and especially in 2005 [Salat et al., 2005; Canals
115 et al., 2006; Durrieu de Madron et al., 2013; Puig et al., 2013b].

116 Three causes for such intense deep convection events were identified. First,
117 winter 2004-2005 was one of the coldest and driest winters of the last 40 years
118 [López-Jurado et al., 2005]. Second, Gasparini et al. [2005] showed the appearance
119 of a θ -S anomaly in the Eastern Mediterranean Deep Waters (EMDW) flowing
120 westward through the Sicily channel to the deep Tyrrhenian subbasin, linked to
121 remarkable injection of heat and salt due to the Eastern Mediterranean Transient
122 (EMT, [Klein et al., 1999; Lascaratos et al., 1999; Malanotte-Rizzoli et al., 1999]). The
123 propagation of this θ -S increase from the Eastern Mediterranean could have induced
124 a warming and salting of the intermediate layers of the northwestern Mediterranean
125 [Schroeder et al., 2010]. Modeling studies [Herrmann et al., 2010] moderate the role
126 of the EMT: a high volume of newly-formed waters in 2005 but a number of weak
127 deep convection events before winter 2005 could have been effective in inducing a
128 more pronounced Levantine Intermediate Water (LIW) layer, as well as enhanced

129 heat and salt content at intermediate depths. Grignon et al., [2010] suggested that
130 even a normal winter would have led to an intense deep convection in 2004/2005
131 due to a low pre-winter stratification. Moreover, the high interannual and decadal
132 variability of the potential temperature (θ) and the salinity (S) of Atlantic surface
133 water in the Gulf of Lions [Vargas-Yanez et al., 2010b] can also influence the
134 intensity of deep convection in these years [Rixen et al., 2005].

135 The intense winter mixing of 2005 is just an example of the deep convection
136 variability. Different observational and modelling works on Open-Ocean Deep
137 Convection were conducted since the 70s (see [Marshall and Schott, 1999] for a
138 review), and the main drivers of the deep convection variability are not yet well
139 defined. Part of it can be explained by the absence of long-term observatory system,
140 and by the fact that we do not know enough the role of the processes involved,
141 particularly at the small scales, which need to be accurately represented in modeling
142 studies. In this article, we will address some of the open questions related to the
143 variability of the deep ocean convection: What are the spatio-temporal scales
144 associated with deep convection? How varies deep convection on an interannual
145 basis? Are the increasing trends in potential temperature and salinity of the WMDW
146 related to deep convection?

147 We present the recent results on deep convection observed from seven years
148 (2007-2013) of hydrological and hydrodynamical measurements of the LION mooring
149 line, the MF-LION meteorological buoy and Conductivity-Temperature-Depth (CTD)
150 data collected from research cruises, gliders and profiling floats (Section 2). The
151 methods used to estimate the mixed layer depth, and the heat, salt, and buoyancy
152 contents are presented in Section 3. The results mainly derived from observations at
153 the LION mooring line are described in Section 4:

154 Among the six years of data available, the Nov 2009-Dec 2010 period
155 is described in detail in Section 4.1. During this period, observations provided an
156 invaluable description of the three overlapping, but dominating successively, phases
157 of this representative case of deep convection. In this section, we highlight the
158 variability of the mixed layer which can go from 150m to 2300m in a couple of
159 months, the intense vertical mixing events and the changes in the θ - S characteristics
160 of the water column.

161 The temporal and spatial characteristics of the different winters from 2008 to
162 2013 are presented in Section 4.2. In addition, to present the temporal and spatial
163 scales, we also discuss their similarities/differences and assess the role of
164 atmospheric forcing and water column stratification on the triggering of deep
165 convection.

166 The strong buoyancy losses and the mixing of the water column directly impact
167 the thermohaline characteristics of the newly-formed deep water. In Section 4.3, we
168 discuss the 6-year evolution of their θ -S properties, together with the temporal scales
169 associated with the evolution of the heat, salt, and buoyancy contents during the
170 different phases of the deep convection are presented in Section 4.3. Mooring data
171 and CTD profiles from summer cruise clearly indicate how the successive events of
172 deep convection produce different "vintages" of WMDW, inducing a warming and a
173 salt increase of the deep (1000-2300m) northwestern Mediterranean.

174

175 **2. Data**

176 **2.1 LION mooring line**

177 A multi-platform and integrated monitoring system was setup in the framework of
178 the Mediterranean Ocean Observing System for the Environment (MOOSE,
179 <http://www.moose-network.fr>) to continuously monitor the deep water formation
180 processes in the Gulf of Lions since 2007. Due to the important role of deep
181 convection on climate and ecosystems, this observing system constitutes the multi-
182 year observational system of the international and multidisciplinary programs HyMeX
183 (Hydrological cycle in the Mediterranean Experiment; [Drobinski et al., 2014]) and
184 MERMeX (Marine Ecosystem Response in the Mediterranean Experiment;
185 [MERMEX Group, 2011]). MOOSE combines eulerian observatories and
186 autonomous mobile platforms (gliders, profiling floats): 1) to observe the long-term
187 physical and biological evolution of the NW Mediterranean Sea in the context of the
188 climate change and anthropogenic pollution (metals, oil, hormones, organic
189 compounds, plastics); 2) to detect and identify long-term environmental anomalies;
190 3) to build efficient indicators of the status of the pelagic planktonic ecosystem in the
191 Northwestern Mediterranean.

192 The LION mooring [Testor et al., 2016] is located in the center of the convection zone
193 at 42.03°N, 4.68°E (figure 1) where the seabed is found at a depth of 2300m. The
194 mooring position is about where the center of the deep convection area is supposed
195 to be [MEDOC Group, 1970; Leaman and Schott, 1991; THETIS Group, 1994].
196 There were seven deployments (we refer in the following as LIONX to indicate each
197 deployment, where X, from 1 to 7, indicates the different deployment) during which
198 the line was gradually equipped between September 2007 and July 2013 with initially
199 8 and finally up to 26 instruments. The evolution of the instrumented line over the
200 years is presented in table 1. Details about the instruments (sampling, data gap) and
201 the calibration are given in appendix A.

202

203 **2.2 Météo-France Meteorological Buoy data**

204 The Météo-France meteorological buoy MF-LION, located at 42.06°N 4.65°E (i.e.
205 4-4.5 kilometers from the deep mooring location) provides atmospheric observations
206 of surface pressure, wind, air temperature, humidity, radiative fluxes, and sea
207 surface temperature and salinity. In addition, a 250m instrumented line was installed
208 below the surface buoy since November 2009 and is currently equipped with 20
209 temperature sensors between 5m and 250m depth below the sea surface. The
210 evolution of the sensors on the line is shown in table 2. These NKE sensors measure
211 ocean temperature with a precision of 0.1°C, actual depth with a precision of 0.5 m
212 and are calibrated after every recovery and before every release. Two additional
213 SeaBird SBE37 CTDs were installed at 2m depth since September 2011 and at
214 120m depth since January 2013 (table 2).

215

216 **2.3 CTD Profile Data**

217 From 2007 to 2013, CTD profiles from research vessels were collected during
218 several oceanographic cruises (DOCONUG2007, DOCONUG2008, 42N5E,
219 MOOSE-GE2010, CASCADE, MOOSE-GE2011, MOOSE-GE2012, DOWEX2012,
220 DEWEX2013, MOOSE-GE2013, and DOWEX2013) that took place in the
221 northwestern Mediterranean. In total, 385 CTD stations were carried out in the Gulf

222 of Lions area (between 41°N/44°N and 3°E/7°E) and more particularly we use 69
223 CTD casts located at less than 30 km from the LION mooring (figure 1).

224 During all the cruises, pressure, temperature and conductivity were measured
225 with a CTD Sea-Bird SBE 911plus. Water samples were collected and analyzed with
226 a Guildline Autosol salinometer to calibrate the CTD conductivity sensor. The final
227 accuracy is estimated to be ± 0.004 for salinity and $\pm 0.001^{\circ}\text{C}$ for potential
228 temperature.

229 For the period 2007-2013, gliders [Testor et al., 2010] were intensively used in
230 the framework of several European and national projects (see Everyone's Gliding
231 Observatories (EGO) <http://www.ego-network.org>). Glider profiles were considered
232 as vertical and were checked with quality control procedures defined in the EU FP7
233 GROOM project. Repeated glider surveys over the Gulf of Lions were conducted
234 between 2007 and 2013, with 38 deployments between January 2007 and August
235 2013 corresponding to 11860 profiles (3410 profiles corresponding to 29
236 deployments were made at less than 30km around the LION mooring, figure 1). The
237 calibration of the gliders CTD relies on nearby CTD casts from research vessels and
238 the LION mooring line data (see more details in Bosse et al., [2015]).

239 In addition to the shipborne and the glider CTD casts, about 2500 Argo profiles
240 were collected during the same period (2007-2013), 32 were in the vicinity of the
241 mooring (figure 1). We used in this analysis only the data flagged as "good" by the
242 Coriolis Data Center (<http://www.coriolis.eu.org/>).

243

244 **2.4 Miscellaneous Data (ERA-Interim Atmospheric reanalysis,** 245 **Ocean Color images)**

246 2.4.1 ERA-Interim reanalysis

247 In this study, we are interested to discuss the daily evolution of the mixed layer
248 depth compared to the surface buoyancy and heat losses. In addition, we also want
249 a flux-product to compare different winters with each other. We choose to use the
250 ERA-Interim reanalysis essentially because of these two aspects: the chronology of
251 the flux at a daily scale and the long-term temporal homogeneity [Herrmann et al.,

252 2011]. Moreover, the four components of the net heat flux are calculated in a
253 consistent way, at the same time.

254 Based on the European Centre for Medium-Range Weather Forecasts (ECMWF)
255 numerical weather prediction model, the ERA-Interim reanalysis data set contains
256 consistent atmosphere and sea surface analyses for the period starting in 1979 [Dee
257 et al, 2011]. The reanalysis makes use of the ECMWF Integrated Forecast System at
258 T255 spectral resolution (80 km horizontal resolution) with 91 vertical levels. We
259 considered here the daily fields of the air-sea fluxes (downward and upward short-
260 wave radiation, downward and upward long-wave radiation, latent heat flux, sensible
261 heat flux, total precipitation, and evaporation) in order to compute the daily net heat
262 flux (Q_{net}) and the net freshwater flux out of the ocean.

263

264 2.4.2 Ocean Color images

265 The 8-day Level 3 standard mapped images of MODIS Aqua surface chlorophyll
266 concentration at 4 km resolution were obtained from the NASA web site
267 (<http://oceancolor.gsfc.nasa.gov/>) for the 2007-2013 period.

268

269 3. Methods

270 3.1 Estimation of the Mixed Layer Depth

271 Since the buoy oceanographic sensors and the deep mooring instruments are
272 not the same, the resolution and the accuracy of the different sensors are also
273 different (see tables 1 and 2). Initially, observations of the deep mooring line and the
274 buoy sensors are merged, and a linear interpolation in time and along the vertical
275 coordinate results in potential temperature profiles gridded on 1m bins, every 30min.
276 Because of the different accuracy of the sensors in the upper and in the deep parts
277 of the water column and the weak deep stratification (from a depth of 1000m to the
278 bottom), we adopt a double criterion to estimate the mixed layer depth.

279 The Mixed Layer Depth (MLD) is calculated for the first 300m of the upper water
280 column using a potential temperature criterion $\Delta\theta_1 = 0.1^\circ\text{C}$ and a reference level at
281 10m. If the MLD calculated with the $\Delta\theta_1$ criterion is deeper than 300m, we used a

282 second criterion $\Delta\theta_2 = 0.01^\circ\text{C}$ and a reference level at 300m to define the MLD. This
283 second criterion is required to define a more accurate MLD, as potential temperature
284 gradients at the base of the ML can be smaller than $d\theta_1=0.1^\circ\text{C}$ in winter when the
285 mixing exceed 1000m depth (see appendix B for details).

286 There was no instrument below the MF-LION surface buoy before November
287 2009, only the sea surface temperature sensor at 1m depth could be used in the
288 MLD calculation. Due to the low accuracy of this sensor we had to calculate the MLD
289 in the first 300m using a criterion $\Delta\theta_1 = 0.6^\circ\text{C}$ and a reference level at 1m. If the MLD
290 is deeper than 300m, we used as for the 2009-2013 period, a secondary criterion
291 $\Delta\theta_2 = 0.01^\circ\text{C}$ and a reference level at 300m depth to calculate the MLD below 300m
292 depth (table 3; appendix B for more details).

293

294

295

296 **3.2 Estimation of the Heat and Salt Contents of the Water Column**

297 The vertical resolution of CTD measurements below the MF-LION buoy and
298 along the LION mooring line enables heat or salt budget calculations for specific
299 water masses. Merging data from both sources, there are 5 to 17 additional potential
300 temperature levels in the first 150m since 2009 and 2 additional salinity levels (at 2m
301 depth since 2011 and 120m depth since 2013), compared to the number of sensor
302 levels on the LION mooring line alone.

303 Observation levels have been distributed throughout the water column to reflect
304 an optimized sampling along the vertical of climatological profiles. Since 2009 with
305 20 potential temperature levels and 10 salinity levels, distances between potential
306 temperature levels are about 50 m in the upper 700m of the water column and about
307 250m deeper down. Distances between salinity levels are about 250 m throughout
308 the water column. Before budget calculations, we performed a linear interpolation of
309 potential temperature and salinity versus depth to obtain 1m resolution profiles from
310 the surface (2m) to the seabed (2300m) every 30 minutes.

311 To quantify the error in heat or salt budget in a specific layer due to this linear
312 interpolation, we use a reference database created from 1m-binned CTD profiles

313 (from research vessels, gliders and profiling floats) carried out in a 30km radius
314 (typical scale of the mesoscale processes in summer in the deep convection area)
315 centered on the mooring.

316 We estimate the error due to the vertical interpolation, for each different
317 configuration of the deep mooring and surface buoy arrays (shown in section 4.3.1).
318 By taking into account the data availability of the merged mooring-buoy line, we get
319 33 different configurations between September 2007 and July 2013 for the daily
320 data. For each layer and for each time-step (daily), a reference profile is chosen as
321 the most stratified profile in a specific time window (2 months for the 1-200m, 200-
322 600m and the 600-1000m layers; 6 months for the deepest layers: 600-2300m and
323 1000-2300m). If there is no profile available during this time window, we choose the
324 most stratified profile from a climatology created from the 6 years of data, using the
325 same time window but on climatological month. We calculate the error in potential
326 temperature, salinity and potential density as the difference between the high
327 resolution profile and the same profile subsampled at the mooring vertical resolution
328 and linearly interpolated.

329

330 **3.3. Estimation of the Integrated Buoyancy Content of the Water** 331 **Column**

332 The influence of the water column stratification on the deep convection
333 occurrence can be explored using a 1D approach, where only the surface buoyancy
334 flux drives overturning and deepens the mixed layer (i.e., when the surface buoyancy
335 flux is positive, buoyancy is removed from the ocean, the water gets denser, and
336 deepening of the mixed layer convection arises). Grignon et al. [2010] and L'Hévéder
337 et al. [2012] used this approach in the northwestern Mediterranean Sea. Lilly et al.,
338 [1999] followed the same assumptions to study the mixed layer property evolution in
339 a θ -S space in the Labrador Sea.

340 In our study, this 1D approach is used during the period going from the end of the
341 preconditioning period (November) to the beginning of the restratification
342 (corresponding in our case to end of February). At that time of the year, the mooring
343 area is dominated by vertical mixing and away for the boundary circulation where
344 the horizontal currents are known to be intense and mostly oriented along the

345 bathymetry (f/H contours). Thus, the main lateral contribution is due to eddies.
346 Following a method described in Lilly and Rhines [2002] and Lilly et al., [2003] to
347 detect eddies from mooring currents time-series, Houpert [2013] detected more than
348 30 eddies between October 2009 and July 2012. Over this whole time period, eddies
349 are present about 10% of the time at the mooring. Some of those eddies could have
350 been identified as Submesoscale Coherent Vortices (SCVs) described by
351 McWilliams [1985]. These SCVs induce the propagation of columnar
352 buoyancy perturbations typically of $0.05\text{-}0.15 \text{ m}^2.\text{s}^{-2}$ as discussed in Bosse et al.
353 [2016]. The eddy-induced buoyancy flux would then represent about 0.5-1.4% of the
354 total integrated buoyancy flux over the winter (typically $1.15 \text{ m}^2.\text{s}^{-2}$). Therefore, eddy
355 lateral buoyancy fluxes can be considered, at the first order, as negligible effect on
356 the buoyancy content budget of the deep convection zone.

357 In this work, we will further examine into details IS(1000) the columnar buoyancy
358 of the first 1000 m and compare it to the integrated surface buoyancy losses from
359 November 1st to the end of February. The methodology used to calculate this
360 stratification index is detailed in Appendix C.

361

362 **4. Results and Discussion**

363 **4.1 Water Column Variability and Transients in the Deep Convection** 364 **Area in 2009/2010**

365 Among the six-year time series available, a specific focus on the Nov 2009-July
366 2010 period is made. For the first time, the MF-LION surface buoy was equipped
367 with subsurface temperature sensors. The LION mooring was fully instrumented and
368 for the first time monitored the preconditioning, the intense vertical mixing and the
369 spreading/restratification phases. As for the winters of deep convection (2009, 2011,
370 2012, 2013), intense deep vertical mixing is observed during winter 2010, as
371 revealed by the complete homogenization of the water column during the month of
372 February 2010 (figures 2a,b,c), and the strong vertical currents (figure 3c).

373

374 4.1.1 Evolution of the θ -S characteristics of the mixed layer and the 375 water column during a deep convection event

376 The deepening of the mixed layer

377 In September, the monthly net heat flux is close to 0 W.m^{-2} . The strong heat loss
378 starts to increase later in October and November. The negative heat fluxes (figure
379 3d) result in gradually cooling surface layers (figure 2a), and in mid-December the
380 first 200m of the water column is fully mixed (figure 2a), indicating a mixed layer
381 reaching 200m. At the end of December and beginning of January the upper sensor
382 (at 170m, dark blue line in figures 2b and 2c) records a potential temperature
383 (salinity) dropping rapidly to 13.05°C (38.4). This different water mass can be seen
384 more clearly on a θ -S diagram, where it has a very distinct signature, colder and
385 fresher (figure 4b1), compared to the “classical” Atlantic Water (AW) that can be
386 found usually in this area in the surface layer (figure 4a1). The arrival of this new
387 water mass corresponds to an increase in the horizontal velocities, with
388 intensification in the upper part of the water column (see current meter at 250m in
389 figure 3b). This could be the signature of mode waters, the Winter Intermediate
390 Water (WIW, Millot [1999]) formed around the mixed patch and being advected into
391 the deep convection area, where the surface waters are usually saltier.

392 After this first period of strong cooling (drop of $\approx 0.3^{\circ}\text{C}$ at 170 m), the mixing goes
393 deeper with strong vertical currents detected at 250m ($\approx 15 \text{ cm.s}^{-1}$) and 500m depth
394 ($\approx 5 \text{ cm.s}^{-1}$) from January 9th to January 12th (figure 3c), associated with strong daily
395 heat losses ($\sim -800 \text{ W.m}^{-2}$, figure 3d). The warm and salty LIW is therefore entrained
396 by the vertical mixing, resulting in a decrease in potential temperature and salinity at
397 300m and 500m depths (see blue and cyan lines in figures 2b,c). At the same time,
398 the heat and salt content of the LIW is vertically homogenized within the surface
399 layer and this induces an increase of potential temperature and salinity in the mixed
400 layer above (figures 2a,b, dark blue line).

401 From January 15th to January 23rd, the atmospheric forcing and the vertical
402 mixing weaken with less intense vertical velocities (figure 3c) and potential density in
403 the upper-ocean layer stops increasing (figure 3a). Between January 23rd and
404 January 26th, a sharp increase in potential temperature (0.2°C) and salinity (0.1) at
405 170m depth indicates the entrainment of LIW in the mixed layer. At that time, the

406 MLD reaches 800m (figure 3e). The deepening of the mixed layer, not related to a
407 parallel increase of the vertical velocity, could indicate that this signal corresponds to
408 the advection of a homogeneous water column in the area rather than a local
409 formation in a 1D framework.

410 From January 27th to January 29th, the mixed layer deepens more than 500m.
411 The event is associated with an increase in surface heat loss and vertical velocities
412 (more than 5 cm.s⁻¹ at 250, 500 and 1000m depth). The vertical mixing increases
413 until the MLD reaches 2000m on February 1st (figure 3e). Vertical mixing stops for a
414 couple of days on February 3rd, at the same time that atmospheric heat loss
415 decreases (figure 3c,d). However on February 5th, an important increase in potential
416 temperature (0.04°C) and salinity (0.01) is recorded at 2300m depth (figures 2d,e),
417 which indicates that the mixed layer, saltier and warmer than the deep waters,
418 reaches the seabed. At that time, the potential density of the bottom water is still the
419 same 29.111 kg.m⁻³ (figure 2f). When the mixed layer reaches the deep layer, the
420 colder and fresher deep waters associated with old WMDW (figure 4a2, 4b2) are
421 transformed, through this heat and salt input coming from the mixed layer, into a new
422 water mass (figure 4c2). At that time, no sign of violent mixing at great depths can be
423 found, as shown by weak vertical velocities measured at 1000m depth (figure 3c).

424 **Deep Convection down to the Bottom**

425 On February 10th very strong heat losses (daily values exceed -600 W.m⁻²)
426 induce violent vertical mixing, with the strongest vertical currents recorded during this
427 winter at 250m, 500m and 1000m depth (figure 3c,d). Vertical speeds sporadically
428 exceed +10 cm.s⁻¹ and -15 cm.s⁻¹.

429 From February 10th to February 13th, the potential temperature of the whole
430 water column decreases by more than 0.02° (figures 2d, 4c2,d2), due to the strong
431 heat losses and to the fact that the deepening of the mixed layer is constrained by
432 the seafloor. During the same period, the salinity does not present any detectable
433 change. The winter buoyancy fluxes are largely dominated by the heat fluxes
434 [Grignon et al., 2010], so the salinity of the newly-formed deep water is mostly set by
435 the salt content of the water column before the deep convection, while its potential
436 temperature results from a combination of the initial heat content of the water column
437 and of the surface heat fluxes.

438 On February 16th, one day after the end of the violent vertical mixing phase,
439 horizontal currents increase, reaching 30-38cm.s⁻¹ between 170m and 1000m deep
440 and 20cm.s⁻¹ at 2300m (figure 3b). This increase of the horizontal velocities occurs
441 after the water column reached its highest density level of the whole year (figure 3a),
442 and can be seen as an adjustment of the ocean: once the atmospheric forcing
443 stopped, the potential energy gained by the water column during the whole vertical
444 mixing period is converted into kinetic energy.

445 **End of the Deep Convection and Horizontal Homogenization**

446 After the last event of strong vertical mixing (March 12th), the water column
447 stratification progressively increases (figure 3a). Thanks to surface heat gain (figure
448 3d) and advection of light waters near the surface [Leaman and Schott, 1991; Schott
449 et al., 1996; Herrmann et al., 2008], vertical gradients of potential temperature,
450 salinity and density increase inside the water column. Potential temperature
451 difference between the surface and 200m depth starts to be higher than 1°C on April
452 20th (figure 2a). May 1st, the potential temperature difference between the surface
453 and 200m depth reaches 3°C (figure 2a). From mid-April to the end of May, the
454 potential temperature stratification in the upper 200m fluctuates between 0.005°C.m⁻¹
455 and 0.015°C.m⁻¹ (figure 2a). The scale of variability associated with these changes
456 in the upper-ocean stratification (figure 2a) is typically of 5~10 days and are similar to
457 the scale of variability observed in the horizontal velocities at all levels (figure 3b).
458 This suggests an important role of submesoscale and mesoscale circulation features
459 such as eddies during the restratification phase [Houpert 2013; Bosse et al., 2016].
460 Similar eddies were also observed in other places such as the Labrador Sea [Lilly
461 and Rhines 2002; Lilly et al., 2003; Hatun et al., 2007; De Jong et al., 2012] or the
462 Irminger Sea [Fan et al., 2014], where they play an important role in the
463 restratification of the deep convection area.

464 The Levantine Intermediate Water (between 200m and 500m deep) is easily
465 distinguishable in the Gulf of Lions, because of its maximum in salinity. On a θ -S
466 diagram, this water mass can be found between a potential density anomaly of 29.05
467 kg.m⁻³ and 29.10 kg.m⁻³, as a maximum salinity and a local maximum in potential
468 temperature (figure 4a1). The LIW is completely mixed during deep convection
469 events (figures 4d1, 2b, 2c), but, as soon as the atmospheric forcing stops (the net

470 heat flux becomes positive), the strong horizontal mixing associated with mesoscale
471 features (figure 3b), brings back the LIW into the deep convection area, such as on
472 March 17th (figure 4g1, 2b, 2c).

473

474 4.1.2 Vertical mixing and convective plumes

475 The LION mooring offers for the first time a complete year of observation of
476 the deep convection process. The significant vertical velocities observed during the
477 intense vertical mixing phase exceed $+10 \text{ cm.s}^{-1}$ and -15 cm.s^{-1} , and are in
478 agreement with previous observations of convective plumes of horizontal scale of
479 1km characterized by downward vertical velocities up to 10 cm.s^{-1} . These small scale
480 features were first observed using isobaric floats [Voorhis and Webb, 1970], then
481 using moored ADCPs [Schott and Leaman, 1991; Schott et al., 1996]. More recently,
482 downward vertical velocities up to 10 cm.s^{-1} were indirectly inferred using gliders
483 [Merckelbach et al., 2010; Frajka-Williams et al., 2011, Bosse et al., 2014]. These
484 plumes act at short temporal and spatial scales and result in the mixing of the
485 properties over the convection area.

486 If the spatial (single-point) and temporal (30 mins) sampling of the current
487 meters on the mooring is not enough to resolve the convective plumes (diameter of
488 800-1000m in the Gulf of Lions [Marshall and Schott 1999]), the θ -S sensors of the
489 mooring seem to give us some clues about the presence of convective plumes which
490 may have decelerated and don't have vertical velocities anymore. This detection is
491 possible particularly at the beginning of the intense vertical mixing phase, when the
492 area is not completely mixed, and the old deep water can still be distinguished from
493 the newly-formed deep waters. On February 7th and February 8th, 2 isolated peaks in
494 potential temperature and salinity can be distinguished at 2300m (figure 2d,e). The
495 length of these peaks is respectively 12h and 6h, while the mean horizontal
496 velocities associated with these events are respectively 6 cm.s^{-1} , and 14 cm.s^{-1} ,
497 without associated vertical velocities. These high-frequency events seen in the
498 potential temperature and salinity bihourly time series (figure 2d,e) are certainly
499 related to the presence of small-scale plumes, already observed in Greenland Sea
500 [Schott et al., 1993] and in the Gulf of Lions [Schott and Leaman, 1991; Schott et al.,
501 1996], which act as a mixing agent [Send and Marshall, 1995; Marshall and Schott

1999]. Since in these cases there is not vertical mixing synchronous with the appearance of these θ -S anomalies, the structures detected are not "active" in the vertical mixing. In addition to their important role in diapycnal mixing, these small-scale structures play certainly an important role in the homogenization of the mixed patch, through isopycnal mixing. Setting up specific experiments to study the scales and the energy associated with these vertical and horizontal small-scale processes could allow us to better understand how the vertical/diapycnal mixing and the horizontal/isopycnal mixing occur in the deep convection area.

4.1.3 Secondary Event of Deep Vertical Mixing

From February 18th to March 8th, the net heat flux from the atmosphere to the ocean is close to zero and even occasionally positive, and consequently the vertical mixing stopped. High horizontal currents start to advect stratified waters into the mooring area (figures 3a and 4e). The last strong winter cooling event occurs from March 9th to March 12th, associated with an increase in vertical currents down to, at least 1000m depth, where the deepest current meter showing intense vertical motion is located (figure 3c). Based on a θ -S diagram, the water column was homogeneous down to about 1750m depth (figure 4f).

Restratification processes seem to dominate as soon as the atmospheric forcing stops to trigger active mixing. Strong net surface buoyancy losses can trigger secondary events of deep vertical mixing if they are superior to the buoyancy content of the water column, highlighting the overlapping of the intense vertical mixing phase and the restratification phase.

4.1.4 Role of the atmospheric forcing on setting up the θ /S characteristics of the newly-formed deep water

In winters 2004-2005 and 2005-2006, abrupt changes, referred as the Western Mediterranean Transition [CIESM, 2009], were observed in the Western Mediterranean Deep Water characteristics and stratification [López-Jurado et al., 2005; Schroeder et al., 2006, 2008a; Font et al., 2007; Smith et al., 2008, Zunino et al., 2012]. In November 2009, the deep stratification of the water column in the Western Mediterranean still indicates the presence of 3 distinct deep water masses

534 (figure 4a2). In this subsection, we will present the effect of the surface net heat flux
535 on the characteristics of the newly-formed deep waters during the intense vertical
536 mixing phase of a deep convection event. A detailed analysis of the evolution of the
537 deep waters in the Gulf of Lions from 2007 to 2013 is made in subsection 4.3.3.

538 The post-2005 newly-formed deep waters (nWMDW), which are saltier and
539 denser than the “old” deep waters (pre-2005, oWMDW), are essentially found below
540 1300m. Lowermost potential temperature and salinity observed in the 750/1200m
541 layer (figures 4 a2, b2, g2, h2, i2), correspond to the “old” WMDW θ/S characteristics
542 that can be found in other part of the basin. On February 10th 2010, the potential
543 temperature at 2300m depth increased by 0.03°C once the mixed layer reaches the
544 seabed (see figures 2d, 3e, 4c2). The consecutive ten days of strong heat loss
545 (figure 3d) decrease the mixed layer potential temperature by 0.02°C (figures 2d,
546 4d2). During this 10-day period, there are no perceptible salinity changes (figures 2e,
547 4d2), and the potential density increases by 0.004 kg.m⁻³ (figures 2f, 4d2). During
548 this period of 10 days of intense vertical mixing, the evaporation does not seem to
549 play a significant role in setting up the characteristics in salinity of the newly-formed
550 deep waters.

551 These observations are in agreement with previous studies which related the
552 characteristics of the newly formed water mass to the frequency and the intensity of
553 the surface forcing [Artale et al., 2002; Grignon, 2009] and to the heat and salt
554 content of the pre-winter oceanic state [Grignon et al., 2010; Herrmann et al., 2010;
555 Schroeder et al., 2010]. Furthermore, the surface heat losses during this 10-day
556 period are very important in setting the potential temperature and density of the
557 newly-formed deep water. The heat content of the water column, together with the
558 surface net heat flux, is responsible for the evolution of the potential temperature of
559 the mixed layer during the whole deepening phase. However, once the mixed layer
560 reaches the seabed, the surface net heat losses will be the main driver in the
561 decrease of the mixed layer potential temperature and in setting up the final potential
562 temperature and density of the newly-formed deep water.

563

564 **4.2 Interannual Variability of the recent Deep Convections Events**

565 We derive and compare here the temporal and spatial characteristics of the
566 various winter convection events. We then further examine the role of atmospheric
567 forcing and water column stratification for six consecutive winters.

568

569 **4.2.1 Mixed layer and multiple events of vertical mixing**

570 Characteristics of five successive winters of open-ocean deep convection (2009
571 to 2013) are summarized in the table 4. The chronology is different according to the
572 events, mainly due to the interannual variability of the heat losses and the stability of
573 the water column (discussed in section 4.2.4). For example, in 2011, the vertical
574 mixing penetrates below the LIW one month earlier than in 2009, 2010, 2012 and
575 2013 (figure 5). On December 26th, 2010, the MLD reaches already 1200m (figure
576 5). The water column is homogenized (decrease in potential temperature and salinity
577 in figures 6a,b and increase in potential density in figure 7a on the 170m-, 300m- and
578 700m-Microcat), while significant vertical velocity superior at 15cm.s^{-1} are recorded
579 at 150, 250 and 500m depth, (figure 7c).

580 Some common features can be pointed out from the analysis of the five events of
581 deep convection observed between 2009 and 2013. Due to the surface heat and
582 buoyancy losses, the mixed layer starts to deepen in September and reaches the
583 base of the AW (150/200m deep) generally in December. The time needed for the
584 mixed layer to deepen from the base of the AW down to the bottom is between 1 and
585 2 months (table 4). Compared to the other winters, heat losses in the beginning of
586 February 2012 are very strong (the mean net heat flux from February 1st to February
587 15th is about -440 W.m^{-2}) and the MLD deepens by 2000m (from the LIW to the
588 bottom), reaching the seabed in only one week (figure 5). This deepening of the MLD
589 is associated with important vertical velocities (15cm.s^{-1}) that vary at high-frequency
590 and are consistent with past observations of vertical mixing in plumes of horizontal
591 scale of $O(1\text{km})$ (order of magnitude of 1km) carried out by Voorhis and Webb
592 [1970], Schott and Leaman [1991], and Schott et al. [1996]. A more detailed analysis
593 discussing the water column stratification and the surface buoyancy losses for this
594 specific event is made in section 4.2.4.

595 Once the mixed layer reaches the seabed, a jump in potential temperature
596 (figure 6d), salinity (figure 6e) and potential density (figure 6f) can be observed on
597 the near bottom Microcat data. The timing of the deep vertical mixing can be seen on
598 all the current meters except the deepest one, located 30m above the seabed in the
599 bottom boundary layer. All the deep convection winters present a common feature
600 with a violent vertical mixing phase of the whole water column lasting between 9 and
601 12 days, which lead to a potential temperature decrease of the whole water column
602 between 0.015°C and 0.03°C (figure 6d, table 4). After the MLD reached the bottom,
603 we don't see significant changes in salinity of the deep water during the 10/14-day
604 period of vertical mixing (figure 6e), which is in agreement with Grignon et al. [2010]
605 who show that the haline component of the buoyancy flux is negligible compared to
606 the thermal component. The salinity of the newly-formed deep water is therefore
607 mostly set by the salt content of the water column before convection, while the
608 potential temperature of the newly-formed deep water results from the combination
609 of the initial heat content and the surface heat fluxes, as discussed for the 2010
610 winter in section 3.1.4.

611 All winters from 2009 to 2013 present a secondary vertical mixing period that
612 generally occurs in March after the main convective event, when the restratification
613 of the water column has already begun. This is the first time that these secondary
614 vertical mixing events are observed, thanks to the high temporal resolution
615 monitoring from full water column mooring. These short events (2-4 days) happen
616 when the restratification has already begun. By that time, the water column is weakly
617 stratified and a 3-6 days period of strong heat losses (typical duration of a winter
618 storm) is enough to destabilize the upper water column leading to vertical mixing that
619 can easily reach great depths.

620

621 **4.2.2 Winter Intermediate Water**

622 Another interesting feature that emerges from the analysis of the multi-year time
623 series is the presence of WIW advected at the mooring location before the deep
624 mixing period in winters 2008, 2009 and 2010, and for the whole year in 2011, 2012
625 and 2013. This modal water mass can be detected thanks to its lower salinity (below
626 36.45, figure 6c) and lower potential density (below 29.05 kg.m⁻³, figure 7a). Due to

its lower density, the WIW increases the stratification in the deep convection area. Fresher WIW could definitely affect the depth of the deep convection, or even inhibit it. This modal water mass is generally formed around the deep convection area from the Balearic Sea to the Ligurian Sea [Millot, 1999]. Recently Juza et al. [2013] show that the continental shelf of the Gulf of Lions can be an important source of WIW. These results could change our vision of the links between the open-ocean and the shelf, and particularly could highlight an important coupling of the two areas that has to be taken into consideration particularly by modelling simulations on deep convection and its interannual variability.

4.2.3 Horizontal extent of the Deep Convection Area

Satellites can provide estimates of the sea surface chlorophyll-a (chl-a) based on ocean color images. These data provide essential information about the extension of the deep water formation zone. During wintertime, the Gulf of Lions is often characterized by low surface chl-a (figure 8). This low chl-a patch indicates that the growth rate of the phytoplankton may be reduced in the euphotic layer, due to the active vertical mixing (figure 7c), and/or that the phytoplankton is diluted in the whole mixed layer, that can reach 2300m in February (figure 5, table 4). No clear relationship exists between the low chl-a patch and the mixed layer depth. Here we propose to use glider data to determine surface chl-a thresholds delimiting the area of open-ocean deep convection in the Gulf of Lions.

In February 2012, a glider crossed the chl-a front at the date of the satellite image shown in figure 8f. The figure 8h represents an enlargement of the satellite image and the glider trajectory within a 3-day interval centered on February 22nd. The transition between the mixed conditions, characterized by low chl-a, and the stratified conditions occurs around the 1500m isobath (figure 8i). In this transition area, the chl-a estimated by the satellite ranges between $\sim 0.15 \text{ mg.m}^{-3}$ and $\sim 0.25 \text{ mg.m}^{-3}$, and the MLD, estimated by the glider, reaches at least 1000m. Furthermore, the potential temperature measured by the glider is close to the newly formed deep waters ($\sim 12.92^\circ\text{C}$). The glider cannot provide information below 1000m, so we cannot detect the point where the MLD start reaching the seabed.

658 However, the LION mooring was deployed for all these winters in the center of the
659 deep convection, offering us a detailed timing of the evolution of the MLD.

660 The area of intense dilution ($<0.15\text{mg.m}^{-3}$) is not present in 2007 and 2008
661 (figures 8a,b). The weakness of the deep convection is also confirmed by glider data.
662 Indeed, 5 gliders deployed in the Gulf of Lions during the EGO2007 experiment from
663 January 2007 to May 2007 recorded a maximal mixed layer of $\sim 400\text{m}$ depth [Testor
664 et al., 2007]. In winter 2008, another swarm experiment took place in the Gulf of
665 Lions (EGO2008) with 6 gliders deployed in the Gulf of Lions from January to April.
666 That year, a MLD of 1000m depth was sporadically recorded mid-March by gliders
667 $\sim 50\text{ km}$ westward of the LION mooring location (near $42.1^\circ\text{N} / 4.1^\circ\text{E}$), while the MLD
668 was not deeper than 700m depth at the LION mooring. These data seem in good
669 agreement with the satellite image for the 8-day period going from the 2nd to the 9th
670 of February 2008 (figure 8b), where the low chl-a patch defined by a criterion
671 $<0.25\text{mg.m}^{-3}$ but $>0.15\text{mg.m}^{-3}$ extends westward of the mooring location. This might
672 suggest that deep open-ocean convection is not homogeneous everywhere in the
673 Gulf of Lions and could be triggered earlier, further west than the LION mooring,
674 close to $42.1^\circ\text{N} / 4.1^\circ\text{E}$.

675 Between winters 2009 and 2013, the area of intense dilution ($<0.15\text{mg.m}^{-3}$)
676 varies between $13\,000\text{ km}^2$ in 2010 to a maximal value of $23\,580\text{ km}^2$ in 2013 (figure
677 8, table 4). A criterion of surface chl-a $<0.15\text{mg.m}^{-3}$ seems to be more relevant to
678 estimate the maximal horizontal extension of the deep convection area when cloud-
679 free satellite images are available at the same period as the mooring records
680 significant vertical mixing of the water column.

681 For winter 2009, 2010 and 2012, deep convection produces new WMDW
682 significantly denser than the year before (figure 6e). We estimate the formation rate
683 of the new class of density for each of these 3 winters, by using a mean depth of the
684 mixed patch of 2200m and an area of intense dilution defined by the contour
685 0.15mg.m^{-3} (delimiting the deep convection area). The results are summarized in
686 table 4 and range from: 1.14 Sv in 2009 (for the density class $29.114\text{-}29.116\text{ kg.m}^{-3}$),
687 0.91 Sv in 2010 (for the density class $29.116\text{-}29.119\text{ kg.m}^{-3}$) and 1.25 Sv in 2012 (for
688 the density class $29.119\text{-}29.126\text{ kg.m}^{-3}$). The deep water formation rate in the
689 northwestern Mediterranean Sea has the same order of magnitude as the other

690 components of the circulation such as the volume transport associated with the
691 Northern Current [Millot, 1999]), and the inflow [Bryden 1994; Candela 2001] and
692 outflow [Tsimplis and Bryden 2000; Candela 2001; Soto-Navarro et al. 2010] at the
693 strait of Gibraltar. Our results agree well with other published estimates of the deep
694 water formation rate in the northwestern Mediterranean Sea based on observations
695 or models (0.3-2.4 Sv) [Tziperman and Speer 1994; Herrmann et al., 2008;
696 Schroeder et al., 2008a, L'Heveder et al., 2012; Durrieu de Madron et al., 2013;
697 Somot et al., 2016; Waldman et al., 2016]. In this study, our estimates are likely to
698 underestimate the formation rate of the new deep water due to different factors:

699 1) The upper limit for the chl-a criterion which is subjective and may vary from
700 one year to another. Here we choose a minimal value of 0.15mg.m^{-3} , a larger value
701 of 0.25mg.m^{-3} would still be acceptable (compared to the MLD observed by glider)
702 and would double our deep water formation area and dense water formation rate
703 estimate;

704 2) They are only based on one instantaneous cloud-free image per year that
705 could miss a part of the newly-formed deep water already exported outside of the
706 deep convection area.

707

708 4.2.4 Atmospheric Forcing and Water Column Stratification

709 The index of stratification IS(1000) from 2008 to 2013 is shown in figure 9a (red
710 line) with the error represented by the dotted purple lines. The higher the IS(1000)
711 index value, the larger the integrated buoyancy losses must be to ensure convection
712 down to 1000m depth. Errors in our calculation of IS are non-negligible for the first
713 years (September 2007 to April 2009), with an order of $0.5\text{m}^2.\text{s}^{-2}$. The stratification
714 index IS(1000) varies between $0\text{ m}^2.\text{s}^{-2}$ in winter, when the water column is
715 completely mixed, to $2\text{ m}^2.\text{s}^{-2}$ at the end of the summer, when the stratification of the
716 water column is maximal. As the number of instruments on the mooring increase,
717 errors on IS decrease (from $0.1\text{ m}^2.\text{s}^{-2}$ in 2010 to $0.02\text{ m}^2.\text{s}^{-2}$ or less in 2013).

718 The different heat fluxes (latent, sensible, radiative) for winter 2008 to winter
719 2013 are presented together in figure 9b. Mean heat fluxes are calculated from
720 November 1st of the previous year to the end of February of the year under

721 consideration. The strongest winter in terms of heat and integrated buoyancy losses
722 is 2012 (196 W.m^{-2} , $1.35 \text{ m}^2\text{s}^{-2}$), while the weakest is 2008 (135 W.m^{-2} , $0.94 \text{ m}^2\text{s}^{-2}$).
723 The net heat flux during wintertime is mainly dominated by the latent heat flux,
724 between 75% (2010) and 86% (2008) of the total net heat flux, while the sensible
725 heat flux represents only 17% (2008) to 24% (2010) of the total heat flux. The
726 radiative heat flux is close to zero (between 1% and 3%).

727 IS(1000) is calculated for November 1st. As mentioned in the methods section,
728 this index indicates how much the water column is stratified at the end of the
729 preconditioning period, before the beginning of the deep vertical mixing phase. For
730 all winters (2008 to 2013), we found that IS(1000) is inferior to the surface buoyancy
731 losses integrated from November 1st to the end of February, indicating that
732 atmospheric forcing is strong enough to explain that deep convection went down to
733 1000m (figure 9c). Another point is that 2011 appears to be the winter during which
734 the water column is the least stratified at the beginning of November. This may
735 explain why deep convection occurs in 2011, while the mean winter heat losses are
736 comparable to 2004, a year without any evidence of newly-formed deep water on
737 historical CTD profiles [Puig et al., 2013b].

738 As discussed in section 4.2.1, the heat losses in the beginning of February 2012
739 are very strong (figure 5). The surface buoyancy losses integrated from February 1st
740 to February 15th are about $0.35 \text{ m}^2\text{s}^{-2}$, corresponding to a net heat flux of -440 W.m^{-2} .
741 On February 1st, the stratification index IS(1000) has a value of $0.45 \pm 0.10 \text{ m}^2\text{s}^{-2}$. By
742 taking into account the $0.10 \text{ m}^2\text{s}^{-2}$ uncertainty on our calculation of IS(1000), we see
743 that the surface buoyancy losses integrated from February 1st to February 15th can
744 explain by themselves why the MLD deepened by 2000m at the beginning of
745 February 2012. For the other winters, the mixed layer went from the LIW layer to the
746 bottom in a period ranging between 2.5 weeks and 1 month, while in 2012 it took
747 only 5 days.

748

749 **4.3 Modification of water masses characteristics**

750 We analyze in this section the evolution of the heat and salt content of the water
751 column and the thermohaline characteristics of the deep water masses from 2007 to
752 2013.

753 4.3.1 Evolution of the Heat and Salt Content of the Water Column

754 Figures 10, 11 and 12 show the mean potential temperature, salinity and
755 potential density for the main water masses of the WMED: the Atlantic Water (AW, 0-
756 200m), the Levantine Intermediate Water (LIW, 200-600m) and the deep waters
757 (WMDW, 600-2300m). A distinction is made for the deep waters between 600 and
758 1000m (oWMDW) and between 1000 and 2300m (nWMDW), as θ -S diagrams show
759 since 2005 the appearance of newly-formed deep water masses below 1000m
760 ([López-Jurado et al., 2005; Salat et al., 2006; Schroeder et al., 2006, Puig et al.,
761 2013b], figure 13).

762 The surface layer 0-200m

763 The daily averaged data at the mooring (dark line) are represented together with
764 data from CTD casts from research vessels, gliders and profiling floats (red dots).
765 Errors due to the limited number of vertical levels for the mooring are also indicated
766 as cyan lines. We can clearly see a seasonal cycle in the mean potential
767 temperature of the surface layer (1-200m) varying from 13.0 ± 0.2 °C in winter to $17 \pm$
768 2 °C in summer. The higher error during summertime (figure 10) is mainly due to the
769 low number of instruments in this layer, the strong surface gradient in potential
770 temperature associated with the summer thermocline is not captured correctly. In
771 winter the surface layer is homogeneous in potential temperature and thus the errors
772 are negligible. After the addition of temperature sensors under the LION buoy in
773 November 2010 (table 2), the errors, associated with the vertical interpolation of the
774 potential temperature records, become negligible in both winter and summer. Since
775 a conductivity sensor has been installed on the LION surface buoy only from
776 September 2011, mean salinity and density are not available for the layer 1-200m or
777 1-2300m for the 2007-2011 period. The decrease in the error on the mean salinity
778 and potential density seen in 2011 and 2012 is due to the winter mixing, which
779 homogenize the upper ocean.

780 The intermediate layer (200-600m)

781 The heat and salt content variability of the intermediate layer (200-600m) at the
782 LION mooring site is mainly driven by the deep convection events with a ~ 0.3 °C
783 decrease in potential temperature (figure 10) and ~ 0.05 decrease in salinity (figure
784 11) in winter, when the deep convection reaches the seabed. The small number of

785 conductivity sensors (see table 2) explains why before April 2008, the error on the
786 mean salinity is larger than the error on the mean potential temperature.

787 No clear trend of warming of the 200-600m layer appears for the 2007-2013
788 period. After the intense vertical mixing period of the deep convection reaches the
789 bottom, the restratification of the 200-600m layer to its pre-convection potential
790 density level (29.09 kg.m^{-3}) takes between 5 and 6 months (figure 12).

791 **The deep layer (600-2300m)**

792 The spreading of the newly-formed deep waters can be seen in the 1000-2300m
793 layer, in figure 12. At the end of the intense vertical mixing period, the potential
794 density reaches a local maximum. Once the intense atmospheric forcing has
795 stopped, the potential density at the mooring location begins to decrease, indicating
796 the spreading of the newly-formed deep waters and their mixing with the older deep
797 waters found in the surrounding area. It takes between ~4 months (in 2012) and ~6
798 months (in 2010) for the deep layers to dissipate 50% of their potential density
799 increase due to bottom-reaching convection. The absence of deep convection in
800 2007 and 2008 explain why the important seasonal variability of the mean potential
801 density of the 1000-2300m layer, $O(0.01 \text{ kg.m}^{-3})$, is seen only since 2009. This
802 seasonal variability in the mean potential density of the deep layer is due to new
803 deep (i.e. dense) water formation, the spreading of this new deep water and the
804 restratification of older deep water.

805 The warming and salting of the deep layers (600-2300m) cannot be clearly seen
806 for the 2007-2008 period due to a too large error on the heat and salt content.
807 However, in September 2008, the addition of instruments on the line considerably
808 reduces the error on the estimation of the heat and salt content of the deep layers
809 (figure 10,11). The formation of warmer and saltier deep waters after each deep
810 convection event can be seen as an injection of heat and salt in the deep ocean.
811 From October 2009 (after the 6-month observational gap) to July 2013, linear trends
812 are estimated fitting a straight line by means of least squares, for the potential
813 temperature, salinity and potential density of the deep water. The slope of the fit
814 represents a trend or a mean annual change. The 95% confidence intervals for the
815 trends are calculated using a T-student test. All the trends in this study are significant
816 (slope different from 0) as their P-values are lower than 0.001.

817 Between October 2009 and July 2013, the linear trends, calculated with a 95%
818 confidence interval, in the deep layer corresponding to the nWMDW (1000-2300m
819 are: $3.0 \pm 0.3 \cdot 10^{-3} \text{ }^{\circ}\text{C/yr}$ for the potential temperature (figure 10), $3.6 \pm 0.2 \cdot 10^{-3}$
820 /yr for the salinity (figure 11) and $2.2 \pm 0.2 \cdot 10^{-4} \text{ kg.m}^{-3}/\text{yr}$ for the potential density
821 (figure 12). The increase in heat (salt) for the 600-2300m layer is about 3.2 ± 0.5
822 $\cdot 10^{-3} \text{ }^{\circ}\text{C/yr}$ (resp. $3.3 \pm 0.2 \cdot 10^{-3} / \text{yr}$). Vargas-Yanez et al., found trends of $1.2 \pm$
823 $1.0 \cdot 10^{-3} / \text{yr}$ for the salinity and $0.28 \pm 0.23 \text{ W.m}^{-2}$ for the heat content,
824 corresponding to a trend in potential temperature of $1.1 \pm 0.9 \cdot 10^{-3} \text{ }^{\circ}\text{C/yr}$ (using a
825 mean potential density of 1029 kg.m^{-3} , a mean potential temperature of 12.8°C , and
826 a mean salinity of 38.4).

827 We found a warming and salting rate for the 2009-2013 period three times higher
828 than the one previously established by Vargas-Yanez et al. [2010b] in the Gulf of
829 Lions for the 1952-1995 period. A first explanation for these different trends lies in
830 the different periods and the different methods used: Vargas-Yanez used a 43-year
831 reconstructed time series from CTD casts, while in this study we used a 4-year time
832 series from mooring data. Observations of deep convection were sparse before the
833 2000s, therefore we cannot be sure that five consecutive years of deep convection,
834 such as the one presented in this work, did not happen in the past.

835 Every deep convection event induces an injection of heat and salt in the deep
836 layer due to the vertical mixing of the LIW with the fresher and colder deep water.
837 Different factors can explain the formation of warmer and saltier deep water
838 compared to the 43-year period studied by Vargas-Yanez, such as changes in the
839 stratification of the surface and intermediate layers. For the same atmospheric
840 forcing, a less stratified AW/LIW layer would allow the vertical mixing to reach the
841 bottom more quickly, and the continuous buoyancy losses would continue
842 decreasing the temperature of the whole water column, leading to colder deep water.
843 The reasons for these changes in the AW/LIW stratification are still not clearly
844 understood. The changes in the θ -S characteristics of the AW and LIW can be due to
845 increasing temperature and salinity in the inflowing AW at Gibraltar Strait [Millot
846 2007], changes in large-scale atmospheric patterns [Josey et al., 2011], and an
847 accumulation of heat and salt in the intermediate layer due to the absence of deep

848 convection during the 90s [Herrmann et al., 2010] and the propagation of the EMT to
849 Western Mediterranean Sea [Schroeder et al., 2010].

850

851

852

853 **4.3.2 Variability and Transients in the Bottom Layer**

854 **Stepwise increases and long-term θ -S evolution of the nWMDW**

855 The stepwise increase in bottom waters characteristics can be noticed for five
856 consecutive years (2009-2013) as positive jumps in θ and S (figures 6d,e table 4),
857 resulting from deep convection events that reach the bottom in February. The
858 formation of these warmer and saltier deep waters can be seen as an injection of
859 heat and salt in the deep ocean, leading to the observed trends (detailed in the
860 previous subsection) in the potential temperature, salinity and potential density of the
861 deep water.

862 If the long-term increase in salinity and potential density of the deep layers
863 seems to be strongly related to the recurrent formation of salty deep waters, the
864 warming trend of the nWMDW can be reduced by very strong events of deep
865 convection when the mixed layer depth reaches the bottom and the strong buoyancy
866 loss continues to decrease the potential temperature of the mixed layer and
867 therefore of the newly-formed deep water by $O(0.01)$ degree. During this period of
868 intense vertical mixing (9-12days, table 4) the salinity remains relatively constant, but
869 the potential temperature of the bottom waters undergoes a significant decrease
870 (figures 6d,e) ranging from 0.012°C in winter 2011 to 0.032°C in winter 2012 (table
871 4). This potential temperature decrease induces an increase in the potential density
872 of the newly formed deep waters between 0.003 kg.m^{-3} in winter 2011 to 0.009 kg.m^{-3}
873 in winter 2012 (figure 6f).

874 Another deep mooring part of the HydroChange program monitors the evolution
875 of the deep waters in the Gulf of Lions since autumn 2006 (located at $45^{\circ}\text{N}/5^{\circ}\text{E}$, at
876 2400 m depth, about 30kms east of the LION mooring location). From data collected
877 at this mooring, Schroeder et al., [2013] also noticed the abrupt positive jumps in θ
878 and S occurring in February 2009, 2010 and 2011, and the subsequent decrease in

879 potential temperature. The two previous winters (2007 and 2008) show rather stable
880 near-bottom θ and S at $12.869 \pm 0.005^\circ\text{C}$ and 38.479 ± 0.002 . We interpret this
881 decrease in potential temperature of the mixed layer, already seen by Schroeder et
882 al. [2013], as the effect of intense net heat losses after that the MLD has reached the
883 seabed. The absence of an increase in the salinity of the bottom waters at the same
884 time can be explained by the dominance of the heat fluxes on the wintertime
885 buoyancy fluxes [Grignon et al., 2010]. As already discussed in section 4.1., the
886 salinity of the newly-formed deep waters is mostly set by the salt content of the water
887 column before convection, while its potential temperature results from a combination
888 of the initial heat content of the water column and of the surface heat fluxes.

889 **θ - S inhomogeneity of the new WMDW formed by deep convection**

890 In March 25th 2010, the potential temperature of the bottom waters undergoes a
891 sharp decrease ($\approx 0.02^\circ\text{C}$, figure 2d), resulting in an increase in the potential density
892 (0.002 kg.m^{-3} , figure 2f). At that moment, the water column is already stratified and
893 the atmospheric forcing only cannot explain this potential temperature change (figure
894 3d). Then the potential temperature progressively increases until reaching an
895 equilibrium state in August 2010 (figure 2d). No similar signal is recorded by the
896 conductivity sensors. The high salinity values of the deep waters measured after
897 February 2010 have never been recorded since the first deployment of the mooring
898 in November 2007 (figure 6e). This secondary newly-formed deep water mass is
899 characterized by a colder potential temperature, a similar salinity, and a higher
900 potential density. This new WMDW, observed from March 25th 2010, is therefore
901 distinct from the main one observed between February 18th and March 25th 2010
902 (figure 2d, e). The linear decrease in potential density between April and August
903 2010 observed on the near bottom Microcat (figure 2e) is associated with a period of
904 intense horizontal currents (figure 3b). This is certainly the indication of the mixing
905 between these two different newly-formed deep waters, as the salinity remains
906 almost constant from April to August.

907 This colder newly-formed deep water is detected one month after the last vertical
908 velocities are recorded at 1000m depth on February 18th (figure 3c). As both new
909 water masses have the same salinity (set by the salt content of the water column
910 before convection, as described in section 4.1), this new water mass should have

911 been formed within the deep convection zone. The only differences between the two
912 new water masses are their temperature, indicating that the two formation sites could
913 have experienced slightly different atmospheric conditions and/or the pre-convection
914 water column could have different θ -S characteristics. Lacombe et al., [1985] already
915 suggested heterogeneity inside the mixed patch with different water-mass formed in
916 different parts of the cyclonic gyre.

917 Considering similar buoyancy fluxes acting in two locations characterized by
918 different stratifications, the deepening of the mixed layer will be faster where the
919 stratification is lower. Such local modification of the stratification could be induced by
920 the presence of (sub) mesoscale eddies that help the mixed layer deepening
921 [L'Herminier et al, 1999; Legg and McWilliams, 2001; Bosse et al, 2015].
922 Furthermore, a shallower bottom depth could allow the MLD to reach the seabed
923 earlier. In this case, the entire homogeneous water column could be exposed to
924 severe heat losses for a longer period of time. This would finally lead to a stronger
925 potential temperature decrease and to the formation of colder and denser deep
926 waters.

927 **WMDW formed by dense shelf water cascading**

928 Another remarkable signal can be noticed between March 2012 and May 2012. It
929 is characterized by a sharp decrease of the bottom potential temperature and salinity
930 and high-frequency fluctuations in potential temperature and salinity (figures 6d,e).
931 This high-frequency variability is not seen in the potential density time series (figure
932 6f), and indicates two distinct water masses characterized by different potential
933 temperature and salinity (discussed more in details in the next section, 4.3.3). The
934 origin of the bottom waters characterized by a lower potential temperature and
935 salinity has been identified as the result of dense shelf water cascading [Durrieu de
936 Madron et al., 2013]. The peculiar atmospheric conditions of winter 2012 indeed
937 triggered a massive formation of dense waters over the continental shelf that
938 overflowed the shelf edge and cascaded down to the bottom of the Gulf of Lions
939 basin.

940

941 4.3.3 Evolution of the Western Mediterranean Deep Water(s)

942 Although the open-ocean deep convection is the main mechanism for the
943 renewal of the WMDW, dense shelf water cascading can have an influence on
944 WMDW, as it was first suggested by Béthoux et al. [2002], and sketched in Figure 1
945 of Puig et al [2013a]. Puig et al. [2013b] highlight the persistence in the deep waters
946 since winter 2005 of a θ -S anomaly related to dense shelf water cascading. The
947 coexistence of deep waters formed by deep convection and by dense shelf water
948 cascading can be clearly seen on a θ -S diagram with a V-shape linking the denser
949 deep waters formed by cascading and the deep waters formed by open-ocean
950 convection. Several authors reported this shift in deep waters characteristics since
951 2005 [Lopez-Jurado et al., 2005; Canals et al., 2006; Schroeder et al., 2006, 2008b;
952 Smith et al., 2008; CIESM, 2009].

953 In 2007, CTD stations made in the Gulf of Lions clearly show 3 different deep
954 water masses (figure 13a) and an inverse V-shape structure below the Levantine
955 Intermediate Waters (>600m). One can clearly see the stacking of three different
956 water masses (indicated by a θ -S extremum) with the “old” WMDW (oWMDW,
957 between 800 and 1400m), the “new” WMDW formed by deep convection in winters
958 2004/2005 and 2005/2006 ($\text{nWMDW}_{2005/06-OC}^{(2007)}$, between 1400 and 1900m) and the
959 “new” WMDW formed in winter 2004/2005 by dense shelf water cascading
960 ($\text{nWMDW}_{2005-CS}^{(2007)}$, between 2000m and the bottom). We use superscript in order to
961 define the year when the deep waters are observed. Thermohaline characteristics of
962 the deep waters change from year to year due the permanent turbulent mixing,
963 which tends to homogenize the deep layers and progressively fade the signal of the
964 different deep water masses. For example, in 2005 the θ -S characteristics of the
965 bottom waters ($\text{nWMDW}_{2005-CS}$) recorded in the Gulf of Lions by Schroeder et al.
966 [2006] were close to $\theta=12.76^\circ\text{C}$ and $S=38.46$, while two years later the observed θ -S
967 characteristics of this water mass are less pronounced ($\theta = 12.87^\circ\text{C}$ and $S = 38.48$).
968 The year to year evolution of a specific deep water mass can be tracked on θ -S
969 diagrams in particular during years without wintertime deep convection events.

970 In 2008 (figure 13b), CTD casts reveal the fading of the deep waters signal,
971 especially for the deep water mass found between 1400m and 1900m. The potential
972 temperature of the $\text{nWMDW}_{2005/06-OC}^{(2008)}$ decreases by 0.02°C in one year, while the

973 salinity decreases about several thousandths (same order of magnitude as our
974 confidence in the calibration of the CTD).

975 In July 2009, the deep waters are homogenized with a less pronounced V-shape
976 structure. This year a new thermohaline anomaly appears on the θ -S diagram. On
977 the last 300m of the water column, a warmer water mass can be seen (figure 13c).
978 As presented in section 4.2, 2009 is a year of bottom-reaching deep convection
979 (figure 5) characterized by an increase in the potential temperature and the salinity of
980 the bottom waters after the deep convection event (figures 6d, e). The range of near-
981 bottom salinity and potential temperature recorded at 2300m depth just at the end of
982 the vertical mixing is shown by the dark gray dots in figure 13c. These potential
983 temperature and salinity correspond to the characteristics of the newly-formed deep
984 water of 2009 ($nWMDW_{2009}$) at the time of its formation, while CTD casts from July
985 reveal less marked bottom water characteristics, a consequence of their mixing with
986 "older" deep waters.

987 The detection of newly-formed deep water on a θ -S diagram is relatively
988 straightforward in the months following the deep convection period (appearance of a
989 new extremum in the θ -S diagram). But identification of the older "vintage" of deep
990 water masses is more difficult on CTD profiles carried out in summer when the deep
991 water masses have started to mix with each other. However, a local maximum in
992 potential temperature between 1000 and 1400m depth can be noticed in the θ -S
993 diagram in 2009 (figure 13c), suggesting the presence of a water mass with θ -S
994 characteristics close to $nWMDW_{2005/06-0C}^{(2008)}$.

995 In June 2010, CTD casts still show the inverse V-shape structure between 800
996 and 1800m, but a second thermohaline structure appears below 1800m, also under
997 the shape of an inverse V. One can relate this second V-shape with the particular
998 characteristics of the newly-formed deep waters of 2010. As discussed in the
999 previous section (4.3.2), two different deep water masses $nWMDW_{2010-I}$ and
1000 $nWMDW_{2010-II}$ could have been formed in winter 2010 (indicated by the two dark
1001 rectangles in figure 13d). These two water masses can be seen on the temporal
1002 evolution of the bottom potential temperature in March 2010 (figure 2d). The similar
1003 salinity and the different potential temperature explain why during the summer cruise
1004 a second inverse V-shape structure can be seen at the end of θ -S diagram tail. A

possible explanation for the decrease of the vertical gradient of θ in the bottom layer in June compared to March can be found in the (isopycnal) mixing that constantly tends to homogenize the deep waters, during that time the horizontal currents are particularly important (figure 7b).

In winter 2011, deep convection reaches the seabed (figures 5 and 7c). Warmer and saltier deep waters are produced (figure 13e). This new bottom water mass ($nWMDW_{2011}$) is also observed on CTD casts performed in June 2011.

The first winter of concomitant open-ocean deep convection and dense shelf water cascading since 2006, is observed in winter 2012 [Durrieu de Madron et al., 2013], In addition to the mooring data, where a significant drop of potential temperature and salinity can be noticed in April 2012, the presence of the dense waters of shelf origin can clearly be seen on CTD casts performed in July 2012 (figure 13f). A new inverse V-shape structure appeared in the θ -S diagram below 1600m, similar to the deep stratification after winter 2005. This thermohaline anomaly extends over more than 600m on CTD casts made close to the mooring location. The shape of the anomaly can be related to the new deep waters formed in winter 2012 by open-ocean deep convection ($nWMDW_{2012-OC}$) and by cascading from the shelf ($nWMDW_{2012-CS}$) as indicated by the gray dots in figure 13f. Extrema in θ and S associated with the dense shelf waters are respectively $\theta=12.80^{\circ}\text{C}$ and $S=38.479$.

In winter 2013, no deep water formed by dense shelf water cascading is detected. After the deep convection event of February 2013, there is no sudden freshening of the bottom layer (figure 6), as in March-April 2012. Winter 2013 is a winter of deep convection with a mixed layer reaching the seabed (figure 5) but the newly-formed deep water $nWMDW_{2013}$ cannot be distinguished from the $nWMDWs_{2012}$ by using only θ -S diagrams of the 2013 summer CTD (figure 13g). The changes in potential density and in salinity of the bottom layer, after 2013 deep convection event, are the smallest since the first recording of deep convection in 2009 (figure 6e, table4). For winter 2013 the salinity of the bottom water increases by 0.004, and the potential density by 0.003 kg.m^{-3} . As for all the previous years, the deep layer is not uniform. A "smooth" inverse V-shapes can be seen below 1600m

indicating that the nWMDW formed in 2012 still influence the deep stratification in the Gulf of Lions.

5. Conclusions

For the first time, an observing system monitored continuously the deep convection phenomena providing invaluable hydrological observations of six years of open-ocean deep convection (October 2007-July 2013) from surface moored buoy MF-LION and deep mooring LION located in the center of the open-ocean dense water formation zone of the Northwestern Mediterranean Sea. The first winter in 2008 is characterized by deep convection reaching ~1000m depth, and the next five winters (from 2009 to 2013) are characterized by deep convection reaching the seabed, found at a depth of 2300m. Using glider, moored and satellite observations we propose to estimate the maximal extent of the area of deep convection in the Gulf of Lions. We delimit the deep convection area by a surface chlorophyll criterion of 0.15mg.m^{-3} , ranging from about 13000 km² in winter 2010 to 24000 km² in winter 2013. Assuming a mean depth of 2200m, we use these surfaces to estimate a formation rate of the new denser class of deep water that appear in 2009, 2010 and 2012. This formation rate, associated with the higher density class observed in winter, range between 0.9 Sv in 2010 and 1.25 Sv in 2012.

The LION mooring is a unique observing platform to study the temporal evolution of the mixed layer in an area where the MLD can reach the seabed, found at a depth of 2300m. A secondary vertical mixing period generally occurs in March after the end of the main event of deep ocean convection. This short mixing phase (2-4days) happens right after the onset of the restratification. The water column is weakly stratified at that time and can be easily destabilized. A period between 3 to 6 days of buoyancy losses, due to intense and frequent wind bursts (Mistral/Tramontane), can lead to a fast deepening of the mixed layer (>1500m). The long restratification and spreading times, which can be observed after each winter of deep convection, suggest that the restratification/spreading phase is still active at the beginning of the next deep convection event. The overlapping of the three phases of the deep

convection process suggests a “memory effect” of the water column. Further studies need to be conducted to investigate the role of (sub-) mesoscale eddies for the restratification of the area, the spreading of the newly formed deep waters, and also for the preconditioning of the area.

Observations from the LION mooring confirm that the salinity of the newly formed deep waters is essentially controlled by the salinity content of the mixed layer, as already suggested by [Grignon et al., 2010]. However the potential temperature of the newly-formed deep water is driven by the heat content of the mixed layer and by the surface buoyancy loss during the deep vertical mixing. Once the mixed layer reaches the seabed, the strong surface buoyancy losses act to decrease the potential temperature of the mixed layer, and therefore of the newly-formed deep water by $O(0.01)$ degree.

These sudden inputs of salt and heat in the deep water, after each event of deep convection, are responsible for the salting and heating of the deep water. For the period 2009-2013, the mean annual change in salinity is $3.3 \pm 0.2 \cdot 10^{-3}$ /yr for the 600-2300m layer, while the trend in potential temperature is $3.2 \pm 0.5 \cdot 10^{-3}$ °C/yr for the 600-2300m layer. These trends indicate warming and salting of the deep layer (600-2300m) for the 2009-2013 period three times higher than the one previously established by Vargas-Yanez et al. [2010b] in the Gulf of Lions for the 1952-1995 period, which could be due to recent changes in the stratification of the surface and intermediate layers. These observations suggest an important role of recurrent events of deep convection in accelerating the warming and salting of the Western Mediterranean Deep Water.

Newly-formed deep waters have been detected after every winter of deep convection from 2009 to 2013. In winter 2010, we observed for the first time two distinct new deep water formed by deep convection during the same winter. They present a different potential temperature but a similar salinity, suggesting that both might have been formed within the cyclonic gyre but in different locations. In 2012, two new deep waters have also been detected at the mooring location. One was identified as a result of open-ocean deep convection, while the other is the result of

1098 dense shelf water cascading that occurred in winter 2012 [Durrieu de Madron et al,
1099 2013].

1100 Although the persistence of the different deep waters in the WMED could be
1101 partly studied with the large number of observations collected during the last decade,
1102 the LION mooring offers new insights into deep waters dynamics, a major element of
1103 the thermohaline circulation of the Mediterranean Sea. The seven-year of θ -S
1104 diagrams clearly shows the successive appearance of new deep water masses after
1105 each deep waters formation event reached the seabed. However, after each winter
1106 of deep convection, the newly-formed deep waters spread into the whole basin and
1107 progressively mix with the former deep waters. This explains the “attenuated”
1108 characteristics of the deep waters recorded during the summertime CTD-surveys,
1109 compared to winter extrema recorded at the LION mooring. Several water masses
1110 composing the WMDW in the western Mediterranean have been found to persist for
1111 several years since 2005 and might be related to a mixing in the deepest layers less
1112 intense than previously thought, or to an increasing volume of newly formed deep
1113 waters during the last years.

1114

1115

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1144 Appendices

1145 Appendix A: The LION mooring line: instruments and calibration

1146 RBR sensors TR-1050/1 and TR-1060 (temperature recorders manufactured by
1147 RBR (<https://rbr-global.com/>) were used between 150m and 700m depth until June
1148 2011, and were replaced afterward by SeaBird Temperature Logger SBE56. SeaBird
1149 Microcat SBE37 (conductivity-temperature-pressure recorders) were located all
1150 along the line between 150m depth and 2300m. Nortek Aquadopp were deployed
1151 since LION3 (September 2008) at 5 levels measuring horizontal and vertical
1152 currents, while there were only 2 Aanderaa RCM 9 at 1000m and 2320m depth
1153 during LION1 (Sep. 2007 / Mar. 2008) and 1 Aanderaa RCM 9 at 1000m LION2
1154 (Sep. 2008). Hence, during these two first deployments, we solely get the horizontal
1155 currents from the current meters. The vertical resolution was improved during LION3
1156 and since then, with 20 potential temperature records, 10 salinity records and 5
1157 current records spanning depths from 150 m to 2320 m (30 m above the seabed).
1158 Here, the Aquadopp temperature data were not used because of the low resolution
1159 and accuracy of the sensor (0.1°) and much more accurate temperature data were
1160 available from other sensors at about the same depths (several meters).

1161 The severe environmental conditions imposed the use of a subsurface mooring,
1162 although the record of the upper 150m heat content was prevented. However,
1163 surface data could be obtained by glider profiles crossing nearby ($<30\text{km}$) and by
1164 temperature records of the MF-LION meteorological buoy situated at $<4.5\text{km}$ from
1165 the deep mooring (see section 2.2). RBR 1050/1060 and SBE56 temperature
1166 recorders were set up with a 15 seconds sampling, while SBE37 CTD recorders and
1167 current meters had a sampling of 6 minutes and 30 minutes respectively. In order to
1168 have a consistent data set with 30-minutes sampling rate, we subsampled the time
1169 series recorded at a higher frequency.

1170 The recovery percentage was satisfactory both for current meters (93%),
1171 temperature sensors (93%) and CTD sensors (96%). In particular, due to a breaking
1172 at the base of the line during the recovery (April 2009), we lost the bottom Aquadopp
1173 and no current data were obtained at 2300m for the period from September 2008 to
1174 March 2009. During LION5, there were some battery issues on five SBE37 CTDs
1175 (165m, 1100m, 1300m, 1780m and 200m) that stopped recording in February 2011.

1176 A delayed recovery of the mooring in July 2012 caused battery issues for most of the
1177 current meters: the 150m and the 2320m current meters stopped recording on 11
1178 June 11th 2012, while the 250m and the 500m depth stopped on July 14th 2012.

1179 Since the recovery of LION4 (in June 2010), intercalibration of the moored
1180 instruments, after and before each deployment, is performed. Niskin bottles are
1181 removed from the shipboard Rosette, and then Microcat and RBR (or SBE56) are
1182 attached instead. We perform a hydrographic cast with a 20 minutes stop at 1000m
1183 depth, thus we can have a relative calibration of the moored instruments with the
1184 CTD probe SBE911plus. Post- and/or pre- cruise calibrations, together with in-lab
1185 analysis of salinity bottles with a Salinometer (Guildline Autosol) calibrated using
1186 standard water, provide an absolute accuracy of the measurements. Overall, for all
1187 deep mooring data used in this study, calibration error was lower than 0.005 for S,
1188 and 0.001°C for θ . No intercalibration was done from September 2007 to April 2009.
1189 At that time, comparisons with CTD stations were the only solution to detect possible
1190 biases in the conductivity measurements. In this study the conductivity corrections
1191 applied to the SBE37 CTD correspond to equivalent salinity corrections ranging from
1192 0.000 to 0.005 (details in [Houpert 2014]).

1193

1194 Appendix B: Mixed Layer Depth calculation

1195 We choose a criterion $\Delta\theta_1$ large enough to overcome the lower accuracy of the
1196 sensors attached below the surface buoy. From November 2009 to July 2013 we
1197 choose $\Delta\theta_1 = 0.1^\circ\text{C}$ and a reference level at 10m. Compared to the commonly used
1198 $\Delta\theta=0.2^\circ\text{C}$ criterion [De Boyer Montegut et al, 2004; D'Ortenzio et al., 2005], this
1199 0.1°C criterion reduces the difference between a MLD calculated with a potential
1200 temperature criterion and a MLD calculated on potential temperature–salinity profiles
1201 with an equivalent density criterion [Houpert et al., 2015]. Using this first criterion, a
1202 Mixed Layer Depth (MLD) was calculated for the first 300m of the upper water
1203 column.

1204

1205 A second criterion was required to define a more accurate MLD from the deep
1206 mooring data, as potential temperature gradients at the base of the ML can be

1207 smaller than $d\theta=0.1^{\circ}\text{C}$ in winter when the mixing exceed 1000m depth (figure A1a).
1208 A $\Delta\theta_1=0.1^{\circ}\text{C}$ is better adapted for the surface layer where the vertical gradient is
1209 stronger. For the deep layers, the criterion has to be small enough to distinguish the
1210 homogenized mixed layer from the underlying deep waters. After performing
1211 sensibility tests for different potential temperature criteria and regarding the accuracy
1212 of the temperature sensors used on the LION mooring, we defined a $\Delta\theta_2 = 0.01^{\circ}\text{C}$
1213 criterion and a reference level at 300m, corresponding to the depth of the first SBE37
1214 CTD of the deep LION mooring located below the LION buoy oceanographic
1215 sensors. If the MLD calculated with the $\Delta\theta_1$ criterion was deeper than 300m, then we
1216 used the second criterion $\Delta\theta_2$ to define the MLD, otherwise the MLD is calculated
1217 only with the $\Delta\theta_1$ criterion.

1218 Since there was no instrument below the MF-LION surface buoy before
1219 November 2009, we could only use the sea surface temperature sensor at 1m depth
1220 for the winter 2007-2008 and 2008-2009. Due to the low accuracy of this sensor and
1221 the diurnal variability, we tested different criteria. We select the smallest $\Delta\theta$ criterion
1222 that indicates the periods where the MLD reached 300m depth in agreement with the
1223 potential temperature time-series of the top of the LION mooring (150, 200, 230, 250,
1224 300m depth, more details in table 1). We choose a criterion of 0.6°C calculated
1225 between the potential temperature at 1m and the potential temperature at 300m: if
1226 the potential temperature difference between 1m and 300m is below 0.6°C , as for
1227 the 2009-2013 period, the MLD is defined by a second criterion $\Delta\theta_2 = 0.01^{\circ}\text{C}$ and a
1228 reference level at 300m depth.

1229 The potential temperature profiles used for the calculation of the MLD were
1230 linearly interpolated between the different mooring levels. We estimate the errors
1231 associated with the linear interpolation of the potential temperature profiles by using
1232 all the available CTD profiles from 2007 to 2013 (ship CTD, gliders, profiling floats)
1233 and by calculating for each profile the difference between the MLD calculated on the
1234 high resolution CTD profile and the MLD calculated on the same profile subsampled
1235 at the mooring vertical resolution and linearly interpolated (figure A1b, c). The
1236 improvements in the estimation on the MLD after the installation of the thermistor
1237 chain below the MF-LION buoy can be seen in figure A1b. After 2009, the errors on

1238 the MLD are about O(1m) in the first 100m, O(10m) from 100m to 400m depth and
1239 O(50m) below 400m depth.

1240 **Appendix C: A 1D-model to study the role of winter buoyancy** 1241 **losses and hydrographic preconditioning**

1242 Following Mertens and Schott [1998], we can express the surface buoyancy flux
1243 B, which depends on the heat and freshwater fluxes at the sea surface, as:

$$1244 \quad B = \frac{g}{\rho_0} \left(\frac{\alpha_\theta}{c_w} Q_{\text{net}} + \rho_0 \beta_S S (E - P) \right) \text{(Eq. 1)}$$

1245 where $g = 9.81 \text{ m.s}^{-2}$ is the acceleration due to gravity, $\rho_0 = 1000 \text{ kg.m}^{-3}$ is the
1246 density reference, $\alpha_\theta = 2 \times 10^{-4} \text{ .K}^{-1}$ and $\beta_S = 7.6 \times 10^{-4}$ are the thermal expansion and
1247 haline contraction coefficients, $c_w = 4000 \text{ J.K.g}^{-1} \text{ .K}^{-1}$ is the heat capacity of water,
1248 Q_{net} is the surface net heat loss, S is the sea surface salinity and $E - P$ represents
1249 the net freshwater flux.

1250 The buoyancy content (BC) of the water column is defined as:

$$1251 \quad \text{BC}(t) = \frac{g}{\rho_0} \int \rho(z,t) dz \text{(Eq. 2)}$$

1252 The conservation of the BC undergoing a buoyancy flux $B(t)$ between t and $t + dt$
1253 implies:

$$1254 \quad \text{BC}(t+dt) - \text{BC}(t) = \int B(t) dz \text{(Eq. 3)}$$

1255 Following Turner's formalism [Turner, 1973] and hypothesis (i.e. the vertical
1256 mixing is non-penetrative) proposed by [Lascaratos and Nittis, 1998], the buoyancy
1257 flux required to mix an initially stratified water column down to the depth h can be
1258 written as:

$$1259 \quad \frac{h \partial h}{\partial t} = \frac{B(t)}{N^2(h)} \text{(Eq. 4)}$$

1260 where N is the buoyancy frequency ($N^2 = -\frac{g}{\rho_0} \frac{\partial \rho}{\partial z}$), and $B(t)$ the surface
1261 buoyancy flux defined in Eq. 1.

1262 Integrating Eq. 4 gives:

$$1263 \quad \text{IS}(Z) = \int_0^Z h N^2(h) dh = \int_{t_0}^{t_1} B(t) dt \text{(Eq. 5)}$$

where $IS(Z)$ represents an index of stratification of the water column [Herrmann et al., 2010].

In this work, we calculated $IS(Z)$ using the interpolated mooring and buoy data (see section 3.2) and we evaluated the possible source of errors. Before November 2011, the stratification due to the salinity in the first 200m is unknown because of the absence of a conductivity sensor at the sea surface. To tackle this problem, we used then a constant value corresponding to the shallowest salinity measurement (at 170m depth). To evaluate the error due to this approximation, we calculate $IS(Z)$ using high resolution potential temperature and salinity profiles collected in a 30km radius from the mooring (from gliders, and research vessels). Then, we compared $IS(Z)$ to $IS_{S0}(Z)$, calculated with a constant salinity in the upper ocean. We also estimated the error due to the low vertical resolution by computing IS_{S0} using the same salinity and potential temperature profiles but subsampled at the depth of the sensors on the mooring (similar to the method presented in section 3.2).

Another potential source of error can come from biases in vertical density gradients induced by the intercalibration of the different instruments. As we estimated an error in the calculation of the potential density less than 0.005 kg.m^{-3} (due essentially to the calibration of the conductivity sensors), we propagated this error in the calculation of IS . For an integration to 1000m depth, the error due to the accuracy of the intercalibration of the instruments represents between 13% and 34% of the total error (shown in section 4.2.4). If we use a stratification index integrated down to 2300m, the errors due to the intercalibration of the instruments become too large (76% to 92% of the total error) to have a relevant use of a stratification index calculated using the mooring. The winter convection can be considered deep when it penetrates deeper than the LIW (the intermediate water mass, located around 400m depth) in the Northwestern Mediterranean Sea, so a stratification index $IS(1000)$ integrated down to 1000m depth is enough to identify deep convection event.

1292 Table 1: Details of the instruments shackled to the LION mooring line from 2007 to 2013.

\Deployment Name		LION1	LION2		LION3		LION4	LION5	LION6	LION7
Date		9 Sep. 2007 27 Mar. 2008	28 Mar. 2008 21 Sep. 2008		24 Sep. 2008 5 Apr. 2009		27 Oct. 2009 23 Jun. 2010	23 Jun. 2010 12 Jun. 2011	13 Jun. 2011 25 Jul. 2012	25 Jun. 2012 17 Jun. 2013
Temperature Sensor	Type	RBR TR-1060	RBR TR-1050/1060		RBR TR-1050/1060		RBR TR-1050/1060	RBR TR-1050/1060	Seabird SBE56	Seabird SBE56
	Nominal depth (m)	250-350-500-600	150-200-230-250-350-400-450-550-600-650		150-200-230-250-350-400-450-550-600-650		150-200-230-250-350-400-450-550-600-650	150-200-230-250-350-400-450-550-600-650	150-200-230-250-350-400-450-550-600-650	150-200-230-250-350-400-450-550-600-650
	Sampling	15 sec.	15 sec.		15 sec.		15 sec.	15 sec.	15 sec.	15 sec.
Conductivity - Temperature-Depth sensor	Type	Seabird SBE37SM	Seabird SBE37SM		Seabird SBE37SM		Seabird SBE37SM	Seabird SBE37SM	Seabird SBE37SM	Seabird SBE37SM
	Nominal depth (m)	170-700-1500-2300	170-300-500-700-1500-2300		170-300-500-700-850-1100-1300-1500-1750-2000-2300		170-300-500-700-850-1100-1300-1500-1750-2000-2300	170-300-500-700-850-1100-1300-1500-1750-2000-2300	170-300-500-700-850-1100-1300-1500-1750-2000-2300	170-300-500-700-850-1100-1300-1500-1750-2000-2300
	Sampling	6 min.	6 min.		3 min. / 6 min.		3 min. / 6 min.	3 min. / 6 min.	3 min. / 6 min.	3 min. / 6 min.
Current meter	Type	Aanderaa RCM9	RDI ADCP	Aanderaa RCM9	RDI ADCP	Nortek Aquadopp	Nortek Aquadopp	Nortek Aquadopp	Nortek Aquadopp	Nortek Aquadopp
	Nominal depth (m)	1000	2300	1000	2300	150-250-500-1000	150-250-500-1000-2300	150-250-500-1000-2300	150-250-500-1000-2300	150-250-500-1000-2300
	Sampling	30 min	30 min	30 min	30 min	30 min.	30 min.	30 min.	30 min.	30 min.

1293 Table 2: Details of the instrument mounted on the MF-LION surface buoy from 2009 to 2013

Date		13 Nov. 2009 2 Mar. 2010	2 Mar. 2010 5 Nov. 2010	5 Nov. 2010 17 Jun. 2013
NKE SP2T Temperature Sensor	Nominal depth (m)	10-20-50-100-200	10-200	5-10-15-20-25-30-35-40- 50-60-70-75-80-90-100- 120-150-175-200-250
	Sampling	5 min	5 min.	5 min.
SBE37 Seabird CTD	Nominal depth (m)	n.a.	n.a.	2 (since 28 Sep.2011), 120 (since 30 Jan. 2013)
	Sampling	n.a.	n.a.	10 sec.

1294 Table 3: Criterion used to calculate the mixed layer depth at the LION site..

	2007-2009 (no temperature data in the first 150m, except SST)	2009-2013 (temperature sensors between 1 and 200m)
MLD < 300m	$\Delta\theta_1=0.6^\circ\text{C}$, reference level: 1m	$\Delta\theta_1=0.1^\circ\text{C}$, reference level: 10m
MLD $\geq$ 300m.	$\Delta\theta_2=0.01^\circ\text{C}$, reference level: 300m	

1295 Table 4: Main scales associated with the different deep convection events from 2008
1296 to 2013.

	2008	2009	2010	2011	2012	2013
Mixed layer deepens below 150 m	17/12/07	15/12/08	17/12/09	15/12/10	08/01/12	08/12/12
WIW detected (S<38.42 at 150m)	21/12/07	30/12/08	09/01/10	16/12/10	03/02/12	All the year 2012
LIW mixed	02/04/08	08/01/09	23/01/10	19/12/10	07/02/12	25/01/13
Mixed Layer reaches the bottom	no (700m)	12/02/09	10/02/10	27/01/11	10/02/12	13/02/13
Bottom pot. temperature increase	x	0.040°	0.030°	0.022°	0.042°	0.032°
Bottom salinity increase	x	0.009	0.007	0.006	0.012	0.004
End of vertical mixing (until 1000m)	x	26/02/09	20/02/10	05/02/11	23/02/12	28/02/13
Pot. temperature decrease of the mixed layer once the bottom is reached	x	0.024°	0.019°	0.012°	0.032°	0.032°
Bottom pot. density increase at the end of the mixing (kg.m ⁻³)	x	0.0044	0.0062	0.0030	0.0090	0.0030
Max. deepening of mooring head	0	350m	550m	550m	550m	550m
Second vertical mixing event						
Time period	x	06/03/09-08/03/09	09/03/10-12/03/10	01/03/11-04/03/11	07/03/12-11/03/12	13/03/13-17/03/13
Maximal depth detected on current meters	x	> 1000m	> 1000m	>500m <1000m	> 1000m	> 1000m
Maximum of MLD	x	2000m	1500m	1700m	2300m	1800m
End of the 2 nd vertical mixing period	04/04/08	09/03/09	13/03/10	05/03/11	13/03/12	21/03/13
Area of low surface chl-a (<0.15mg.m ⁻³) in km ²	0	16 350	13 006	15 790	17 863	23 583
Area of low surface chl-a (<0.25mg.m ⁻³) in km ²	1 487	34 162	29 180	24 111	41 616	56 351
New density class of WMDW (kg.m ⁻³)	none	29.114-29.116	29.116-29.119	none	29.119-29.126	none
Formation rate associated with the new density class (in Sv)		1.14	0.91		1.25	

1297

1298

1299 **Figures**

1300 Figure 1: Mooring designs (top left panel) with the different water masses (AW:
1301 Atlantic Water, WIW: Winter Intermediate Water, LIW: Levantine Intermediate Water,
1302 (o/n)WMDW: (old/new) Western Mediterranean Deep Water). Map of the
1303 northwestern Mediterranean (top right panel). Map of all the CTD profiles position
1304 (research vessel, glider, profiling float) used in this study (bottom panel). On the top
1305 right panel, the general circulation patterns are shown in red with the Northern
1306 Current flowing along the shelf break from the Ligurian Sea to the Balearic Sea, and
1307 the North Balearic Front indicated as an arrow with dashed lines. Prevailing
1308 northwestern (Tramontane) and north (Mistral) winds are shown in blue. The dashed
1309 yellow line shows the area of Open-Ocean Deep Convection and the purple dashed
1310 line indicates the area of dense water on the shelf. The inset map indicates the
1311 different geographic regions described in the manuscript (TY: Tyrrhenian Sea, SC:
1312 Sicily channel which separates the Western Mediterranean Sea from the Eastern
1313 Mediterranean Sea, EMED).

1314 Figure 2: (a) Bihourly potential temperature from the sensors mounted below the
1315 surface buoy LION (from 2m to 200m), potential temperature (b), and salinity (c)
1316 recorded by Seabird Microcats from 170m to 2330m between November 2009 and
1317 July 2010. The near-bottom potential temperature (d), salinity (e), and potential
1318 density anomaly (f) are also presented with a separate vertical scale.

1319 Figure 3: Bihourly potential density (a) recorded by Seabird Microcats from 170m to
1320 2330m between November 2009 and July 2010, with horizontal (b) and vertical (c)
1321 currents recorded at 250, 500, 1000, and 2330m, (vertical currents are offset by 20
1322 cm.s^{-1}), (d) daily surface net heat flux estimated by ERA-Interim at the mooring
1323 location, and (e) mixed layer depth estimated from the mooring data (see section 3.1
1324 in the manuscript for a description of the chosen criterion).

1325 Figure 4: θ -S diagrams from all the mooring Microcats (1) with an enlargement on
1326 the deep waters (2). The nominal depth of the different instruments is indicated by
1327 the color marks (from 170m to 2300m), while gray marks correspond to the data of
1328 the past month. The slanting black lines are σ_θ contours. Particular stages of the
1329 evolution of the water column from December 2009 to November 2010 are
1330 described: the mixed layer deepening (a,b,c,d); the formation of WIW (b); the

formation of the new WMDW (d); the overlapping of the restratification phase of the deep layer (e) with a second event of vertical mixing (f); restratification of the LIW and WMDW after the winter mixing associated with mesoscale features (g,h); and the new stratification of the deep waters in October 2010 (i) prior to the 2011 deep convection (note the significant changes compared to November 2009 (a), before the 2010 deep convection event).

Figure 5: Mixed Layer Depth, potential temperature, and net atmospheric heat flux from ERA-Interim, from December to May for the 2007-2013 period. Horizontal gray dotted lines show the depth of the mooring instruments. The yellow thick line indicates the mixed layer depth (see section 3.1 in the manuscript for a description of criterion chosen). Note the absence of temperature sensors between 2 m and 150 m in 2007-2008 and 2008-2009, for these 2 periods only the SST sensor (1m depth) was used, which might explain the “too cold” sub surface layer during winter when the sea surface experiences strong cooling.

Figure 6: (a) Daily potential temperature from the sensors mounted below the surface buoy LION (from 2m to 200m), potential temperature (b), and salinity (c) recorded by Seabird Microcats from 170m to 2330m between November October 2007 and July 2013. The near-bottom potential temperature (d), salinity (e), and potential density anomaly (f) are also presented with a separate vertical scale.

Figure 7: Daily potential density (a) recorded by Seabird Microcats from 170m to 2330m between October 2007 and July 2013, with horizontal (b) and vertical (c) currents recorded at 250, 500, 1000, and 2330m,. Vertical currents are offset by 20 cm.s^{-1} .

Figure 8: Weekly surface chlorophyll images from MODIS for winter 2007 to 2013 (a to g). MODIS surface chlorophyll (h) on February 22nd, 2012. The trajectory of a glider is shown for a 3-day interval centered on that date. Circles represent stratified profiles, whereas stars are for profiles having a MLD greater than 1000m. (i) Surface Chlorophyll-a with the potential temperature profiles collected by the glider. The MLD from each glider profile (computed using a criterion at 0.02°C on the potential temperature and a reference at 10m) is superimposed..

Figure 9: Time series of the integrated buoyancy content of the first 1000m IS(1000) in red (a), with the associated error (dotted purple line). The black dots correspond to IS(1000) calculated for vertical profiles carried out close to the mooring (<30km); (b): Surface heat fluxes at the air-sea interface for the winter 2007-2008 to winter 2012-2013 from ERA-Interim (the integration period goes from November 1st to March 1st);

(c): Integrated surface buoyancy losses (blue) for the winter 2007-2008 to winter 2012-2013 from ERA-Interim (the integration period goes from November 1st to March 1st), with vertically integrated buoyancy content of the first 1000m of the water column estimated on the November 1st of each year using the mooring (red circle). The purple vertical lines indicate the associated error (see section 3.3 for more details).

Figure 10: Temporal evolution of the mean potential temperature (°C) of the water column and specific layers at the LION mooring. The error due to the vertical integration is represented by the blue line. The red dots correspond to the mean potential temperature calculated with vertical profiles close to the mooring (<30km). From October 2009 to July 2013, linear trends are estimated fitting a straight orange line by means of least squares. The slope of the fit represents the observed trend. 95% confidence intervals for the trends are calculated considering a T-student test.

Figure 11: Equivalent to figure 10 but for the mean salinity.

Figure 12: Equivalent to figure 10 but for the mean potential density (kg.m^{-3}).

Figure 13: θ -S diagrams of WMDW from CTD stations highlighting the interannual variability of the deep stratification and θ -S characteristics of the WMDWs from 2007 to 2013 in the deep convection area, with the formation of new WMDW in 2009, 2010, 2011, 2012 and 2013. The new WMDWs are labelled (see section 4.3.3 for explanation) and their wintertime θ -S characteristics, observed by the mooring, are indicated by a black dashed box. For visual reference, each year the θ -S characteristics of the WMDW of the previous winter are shown as a light gray box. The dark gray (resp. light gray) dots show the θ -S properties recorded by the 2300m (resp. 1500m) Microcat at the end of the vertical mixing phase. The differences in θ , S and σ_θ , between the CTD stations carried out in summer (in color) and the wintertime observations (in February) from the mooring instruments (light and dark gray dots), illustrate the effects of the isopycnal mixing on the WMDW θ -S characteristics. The isopycnal mixing of the WMDW is associated with the spreading of the newly-formed deep water and the restratification of older WMDW surrounding the deep convection zone. The colorbar shows the pressure on the different CTD stations carried out close (<30km) to the LION mooring: (a) DOCONUG2007; (b) DOCONUG2008; (c) 42N5E in 2009; (d) MOOSE_GE_2010; (e) MOOSE_GE_2011; (f) MOOSE_GE_2012; (g) MOOSE_GE_2013.

Figure A1: (a) Potential temperature profiles from March 4th 2011 carried out during the CASCADE cruise. The black triangles show the MLD calculated using a

1402 $\Delta\theta=0.1^{\circ}\text{C}$ with a reference level at 10m. The black circles show the MLD calculated
1403 using a double criterion: if the MLD, calculated using a $\Delta\theta_1=0.1^{\circ}\text{C}$ criterion with a
1404 reference level at 10m, is deeper than 300m, the MLD is defined by a second
1405 criterion: $\Delta\theta_2=0.01^{\circ}\text{C}$ with a reference level at 300m).

1406 The MLD errors due to the depth resolution of the mooring are estimated by
1407 using CTD profiles (from research vessels, gliders and profiling floats): for the depth
1408 resolution of the mooring after 2009 (b), and for the period before the deployment of
1409 the thermistor chain below the MF-LION buoy in 2009 (c). The MLD is calculated
1410 using the double potential temperature criterion defined in section 3.1. For each
1411 profile, the MLD error is calculated as the difference between the MLD calculated on
1412 a 1m-bin averaged profile and the MLD calculated on the same profile, but
1413 subsampled at the mooring levels and linearly interpolated. The distribution of errors
1414 is binned in different MLD classes and is plotted as box plots. The number of
1415 samples used for each MLD class is indicated by the italic number on top of each
1416 box. The left side and right side of each blue "box" are the 25th and 75th percentiles
1417 of the samples, respectively. The distances between the left and right of the box are
1418 the interquartile ranges. The red line in the middle of each box is the sample median.
1419 The black lines extending from the left and right of each box are drawn from the left
1420 or right of the box to the furthest observations within 1.5 times the interquartile range.
1421 Outliers are displayed as black crosses.

References

- Artale, V., D. Iudicone, R. Santoleri, et al. (2002), Role of surface fluxes in ocean general circulation models using satellite sea surface temperature: Validation of and sensitivity to the forcing frequency of the Mediterranean thermohaline circulation, *Journal of Geophysical Research*, 107(C8), 3120, doi:10.1029/2000JC000452.
- Bethoux, J. P., B. Gentili, J. Raunet, and D. Tailliez (1990), Warming trend in the western Mediterranean deep water, *Nature*, 347(6294), 660–662, doi:10.1038/347660a0.
- Bethoux, J. P., X. Durieu de Madron, F. Nyffeler, and D. Tailliez (2002), Deep water in the western Mediterranean: peculiar 1999 and 2000 characteristics, shelf formation hypothesis, variability since 1970 and geochemical inferences, *Journal of Marine Systems*, 33-34, 117–131, doi:10.1016/S0924-7963(02)00055-6.
- Bethoux, J.-P., B. Gentili, D. Tailliez, and J. Bethoux (1998), Warming and freshwater budget change in the Mediterranean since the 1940 s, their possible relation to the greenhouse effect, *Geophysical Research Letters* 25(7), 1023–1026, doi:10.1029/98GL00724.
- Bosse, A, P. Testor, G. Legland, et al. (2014), Vertical velocities associated with open-ocean deep convection in the northwestern Mediterranean Sea as indirectly observed by gliders, *EGU General Assembly Conference Abstracts*, pp. EGU2014–16009.
- Bosse, A., P. Testor, L. Mortier, et al. (2015), Spreading of Levantine Intermediate Waters by Submesoscale Coherent Vortices in the Northwestern Mediterranean Sea as observed with gliders, *Journal of Geophysical Research: Oceans*, 120 (3), pp.1599-1622. <http://dx.doi.org/10.1002/2014JC010263>
- Bosse, A., P. Testor, L. Houpert, et al. (2016), Scales and dynamics of submesoscale coherent vortices formed by deep convection in the northwestern Mediterranean Sea, in preparation for *Journal of Geophysical Research: Oceans*, *in revision for this volume*, doi: 2016JC012144
- Bryden, H. L., J. Candela, and T. H. Kinder (1994), Exchange through the Strait of Gibraltar. *Progress in Oceanography*, 33(3):201–248. doi: 10.1016/0079-6611(94)90028-0.
- Candela, J. (2001), Mediterranean water and global circulation. In J. C. Gerold Siedler and J. Gould, editors, *Ocean Circulation and Climate Observing and Modelling the Global Ocean*, volume 77 of *International Geophysics*, pages 419 – 429. Academic Press. doi: 10.1016/S0074-6142(01)80132-7.
- Canals, M., P. Puig, X. D. de Madron, S. Heussner, A. Palanques, and J. Fabres (2006), Flushing submarine canyons., *Nature*, 444(7117), 354–7, doi:10.1038/nature05271.
- CIESM (2009), *Dynamics of Mediterranean deep waters*, CIESM Work. Monogr., 38, 132.

1464 Clarke, R. A., and J.-C. Gascard (1983), The Formation of Labrador Sea Water.
 1465 Part I: Large-Scale Processes, *Journal of Physical Oceanography*, 13(10), 1764–
 1466 1778, [http://dx.doi.org/10.1175/1520-0485\(1983\)013<1764:TFOLSW>2.0.CO;2](http://dx.doi.org/10.1175/1520-0485(1983)013<1764:TFOLSW>2.0.CO;2)

1467 De Boyer Montegut, C., Madec, G., Fischer, et al., 2004. Mixed layer depth over
 1468 the global ocean: an examination of profile data and a profilebased climatology.
 1469 *Journal of Geophysical Research*, 109, 1–20.
 1470 <http://dx.doi.org/10.1029/2004JC002378>.

1471 Dee, D. P., Uppala, S. M., Simmons, et al. (2011), The ERA-Interim reanalysis:
 1472 configuration and performance of the data assimilation system. *Q.J.R. Meteorol.*
 1473 *Soc.*, 137: 553–597. doi: 10.1002/qj.828

1474 De Jong, M. F., Bower, A. S., and Furey, H. H. (2014). Two years of observations of
 1475 warm-core anticyclones in the Labrador Sea and their seasonal cycle in heat and
 1476 salt stratification. *Journal of Physical Oceanography*, 44(2), 427-444, doi:
 1477 10.1175/JPO-D-13-070.1

1478 D’Ortenzio, F., Iudicone, D., De Boyer Montegut, C., et al. (2005). Seasonal
 1479 variability of the mixed layer depth in the Mediterranean Sea as derived from in situ
 1480 profiles. *Geophysical Research Letters* 32, 2–5.
 1481 <http://dx.doi.org/10.1029/2005GL022463>.

1482 Drobinski P., V. Ducrocq, P. Alpert, et al. (2014) HyMeX: A 10-Year
 1483 Multidisciplinary Program on the Mediterranean Water Cycle. *Bulletin of the*
 1484 *American Meteorological Society*, 95, 1063–1082. doi: <http://dx.doi.org/10.1175/>

1485 Durrieu de Madron, X. et al. (2013), Interaction of dense shelf water cascading
 1486 and open-sea convection in the northwestern Mediterranean during winter 2012,
 1487 *Geophysical Research Letters*, 40(7), 1379–1385, doi:10.1002/grl.50331.

1488 Fan, X., Send, U., Testor, P., Karstensen, J., Lherminier, P. (2013), Observations of
 1489 Irminger Sea anticyclonic eddies. *Journal of Physical Oceanography*, 43(4), 805-823,
 1490 doi: 10.1175/JPO-D-11-0155.1

1491 Font, J., P. Puig, J. Salat, A. Palanques, and M. Emelianov (2007), Sequence of
 1492 hydrographic changes in NW Mediterranean deep water due to the exceptional
 1493 winter of 2005, *Scientia Marina*, 71(2), 339–346, doi:10.3989/scimar.2007.71n2339.

1494 Frajka-Williams, E., C. C. Eriksen, P. B. Rhines, and R. R. Harcourt (2011),
 1495 Determining Vertical Water Velocities from Seaglider, *Journal of Atmospheric and*
 1496 *Oceanic Technology*, 28(12), 1641–1656, doi:10.1175/2011JTECHO830.1.

1497 Gasparini, G. P., a. Ortona, G. Budillon, et al. (2005), The effect of the Eastern
 1498 Mediterranean Transient on the hydrographic characteristics in the Strait of Sicily
 1499 and in the Tyrrhenian Sea, *Deep Sea Research Part I*, 52(6), 915–935,
 1500 doi:10.1016/j.dsr.2005.01.001.

1501 Grignon, L. (2009), *Causes of the Interannual Variability of Deep Convection*,
 1502 University of Southampton.

1503 Grignon, L., D. A. Smeed, H. L. Bryden, and K. Schroeder (2010), Importance of
 1504 the variability of hydrographic preconditioning for deep convection in the Gulf of Lion,
 1505 NW Mediterranean, *Ocean Science*, 6(2), 573–586, doi:10.5194/os-6-573-2010.

1506 Hátún, H., Eriksen, C. C., and Rhines, P. B. (2007), Buoyant eddies entering the
 1507 Labrador Sea observed with gliders and altimetry. *Journal of Physical*
 1508 *Oceanography*, 37(12), 2838-2854, doi: 10.1175/2007JPO3567.1

1509 Herrmann, M., S. Somot, F. Sevault, C. Estournel, and M. Déqué (2008),
 1510 Modeling the deep convection in the northwestern Mediterranean Sea using an
 1511 eddy-permitting and an eddy-resolving model: Case study of winter 1986–1987,
 1512 *Journal of Geophysical Research*, 113(C4), 1–25, doi:10.1029/2006JC003991.

1513 Herrmann, M., F. Sevault, J. Beuvier, and S. Somot (2010), What induced the
 1514 exceptional 2005 convection event in the northwestern Mediterranean basin?
 1515 Answers from a modeling study, *Journal of Geophysical Research*, 115(C12), 1–19,
 1516 doi:10.1029/2010JC006162.

1517 Herrmann M., Somot S., Calmanti S., Dubois C., Sevault F. (2011),
 1518 Representation of daily wind speed spatial and temporal variability and intense wind
 1519 events over the Mediterranean Sea using dynamical downscaling : impact of the
 1520 regional climate model configuration, *Nat. Hazards Earth Syst. Sci.*, 11, 1983-2001,
 1521 doi:10.5194/nhess-11-1983-2011

1522 Houpert, L. (2013), Contribution to the Study of Transfer Processes from the
 1523 Surface to the Deep Ocean in the Mediterranean Sea Using in-situ Measurements.
 1524 PhD Thesis, Université de Perpignan via Domitia

1525 Houpert L. (2014), Technical report on the intercalibration of the sensors on the
 1526 LION mooring line, LOCEAN, Univ. Pierre et Marie Curie, Paris, 19pp,
 1527 http://lhoupert.fr/files/report/2014_intercalib_sensors_LION.pdf

1528 Houpert L., P. Testor, X. Durrieu de Madron, et al. (2015), Seasonal cycle of the
 1529 mixed layer, the seasonal thermocline and the upper-ocean heat storage rate in the
 1530 Mediterranean Sea derived from observations, *Progress in Oceanography*, 132, 333-
 1531 352, <http://dx.doi.org/10.1016/j.pocean.2014.11.004>.

1532 Josey, S. A., Somot, S. and Tsimplis, M. Impacts of atmospheric modes of
 1533 variability on Mediterranean Sea surface heat exchange. *Journal of Geophysical*
 1534 *Research* 116, C02032 (2011).

1535 Juza, M., Renault, L., Ruiz, S., and Tintoré, J. (2013). Origin and pathways of
 1536 Winter Intermediate Water in the Northwestern Mediterranean Sea using
 1537 observations and numerical simulation. *Journal of Geophysical Research: Oceans*,
 1538 118(12), 6621-6633.

1539 Killworth, P. D. (1983), Deep convection in the World Ocean, *Reviews of*
 1540 *Geophysics*, 21(1), 1, doi:10.1029/RG021i001p00001.

1541 Kim, K., K.-I. Chang, D.-J. Kang, Y., et al. (2008), Review of recent findings on
 1542 the water masses and circulation in the East Sea (Sea of Japan), *Journal of*
 1543 *Oceanography*, 64(5), 721–735, doi:10.1007/s10872-008-0061-x.

1544 Klein, B., W. Roether, B. B. Manca, et al. (1999), The large deep water transient
 1545 in the Eastern Mediterranean, *Deep Sea Research Part I*, 46(3), 371–414,
 1546 doi:10.1016/S0967-0637(98)00075-2.

1547 Krahmann, G., and F. Schott (1998), Longterm increases in western
 1548 Mediterranean salinities and temperatures: Anthropogenic and climatic sources,
 1549 *Geophysical Research Letters*, 25(22), 4209, doi:10.1029/1998GL900143.

1550 Lacombe, H., P. Tchernia, and L. Gamberoni (1985), Variable bottom water in the
 1551 Western Mediterranean basin, *Progress in Oceanography*, 14(1971), 319–338,
 1552 doi:10.1016/0079-6611(85)90015-1.

1553 Lascaratos A, Nittis K (1998) A high-resolution three-dimensional numerical study
 1554 of intermediate water formation in the Levantine Sea. *Journal of Geophysical*
 1555 *Research* 103(C9):18497–18512

1556 Lascaratos, A., W. Roether, K. Nittis, and B. Klein (1999), Recent changes in
 1557 deep water formation and spreading in the eastern Mediterranean Sea: a review,
 1558 *Progress in Oceanography*, 44(1-3), 5–36, doi:10.1016/S0079-6611(99)00019-1.

1559 Lavigne, H., F. D'Ortenzio, C. Migon, et al. (2013), Enhancing the comprehension
 1560 of mixed layer depth control on the Mediterranean phytoplankton phenology, *Journal*
 1561 *of Geophysical Research*, 118, 3416–3430, doi:10.1002/jgrc.20251.

1562 Lazier, J. R. N. (1973), The renewal of Labrador sea water, *Deep Sea Res.*
 1563 *Oceanogr. Abstr.*, 20(4), 341–353, doi:http://dx.doi.org/10.1016/0011-7471(73)90058-
 1564 2.

1565 Leaman, K. D., and F. A. Schott (1991), Hydrographic Structure of the
 1566 Convection Regime in the Gulf of Lions: Winter 1987, *Journal of Physical*
 1567 *Oceanography*, 21(4), 575–598, doi:10.1175/1520-
 1568 0485(1991)021<0575:HSOTCR>2.0.CO;2.

1569 Legg, S., and J. C. McWilliams (2001), Convective Modifications of a
 1570 Geostrophic Eddy Field, *Journal of Physical Oceanography*, 31(4), 874–891,
 1571 doi:10.1175/1520-0485(2001)031<0874:CMOAGE>2.0.CO;2.

1572 Lherminier, P., J.-C. Gascard, and D. Quadfasel (1999), The Greenland Sea in
 1573 Water 1993 and 1994: preconditioning for deep convection, *Deep Sea Research Part*
 1574 *II*, 46(6-7), 1199–1235, doi:10.1016/S0967-0645(99)00020-X.

1575 L'Hévéder, B., L. Li, F. Sevault, and S. Somot (2012), Interannual variability of
 1576 deep convection in the Northwestern Mediterranean simulated with a coupled
 1577 AORCM, *Climate Dynamics*, doi:10.1007/s00382-012-1527-5.

1578 Lilly, J. M., P. B. Rhines, M. Visbeck, et al. (1999), Observing Deep Convection in
 1579 the Labrador Sea during Winter 1994/95. *Journal of Physical Oceanography*,
 1580 29(8):2065–2098. doi: 10.1175/1520-0485(1999)029<2065:ODCITL>2.0.CO;2.

1581 Lilly, J. M. and P. B. Rhines (2002), Coherent Eddies in the Labrador Sea
 1582 Observed from a Mooring. *Journal of Physical Oceanography*, 32(2):585–598. doi:
 1583 10.1175/1520-0485(2002)032<0585:CEITLS>2.0.CO;2

1584 Lilly, J. M., P. B. Rhines, F. Schott, et al. (2003), Observations of the Labrador
 1585 Sea eddy field. *Progress in Oceanography*, 59(1):75–176. doi:
 1586 10.1016/j.pcean.2003.08.013.

1587 López-Jurado, J.-L., C. González-Pola, and P. Vélez-Belchí (2005), Observation
 1588 of an abrupt disruption of the long-term warming trend at the Balearic Sea, western
 1589 Mediterranean Sea, in summer 2005, *Geophysical Research Letters*, 32(24), 2003–
 1590 2006, doi:10.1029/2005GL024430.

1591 Malanotte-Rizzoli, P., B. B. Manca, M. R. D'Alcala, et al. (1999), The Eastern
 1592 Mediterranean in the 80s and in the 90s: the big transition in the intermediate and

1593 deep circulations, *Dynamics of Atmospheres and Oceans*, 29(2-4), 365–395,
1594 doi:10.1016/S0377-0265(99)00011-1.

1595 Marshall, J., and F. Schott (1999), Open-ocean convection: Observations, theory,
1596 and models, *Reviews of Geophysics*, 37(1), 1–64, doi:10.1029/98RG02739.

1597 McWilliams, J. C. (1985), Submesoscale, coherent vortices in the ocean,
1598 *Reviews of Geophysics*, 23(2), 165–182, doi:10.1029/RG023i002p00165.

1599 MEDOC Group (1970), Observation of Formation of Deep Water in the
1600 Mediterranean Sea, 1969, *Nature*, 227(5262), 1037–1040, doi:10.1038/2271037a0.

1601 Merckelbach, L., D. Smeed, and G. Griffiths (2010), Vertical Water Velocities
1602 from Underwater Gliders, *Journal of Atmospheric and Oceanic Technology* 27(3),
1603 547–563, doi:10.1175/2009JTECHO710.1.

1604 Mermex Group (2011), Marine Ecosystems Responses to climatic and
1605 anthropogenic forcings in the Mediterranean, *Progress in Oceanography*, 91(2), 97-
1606 166, ISSN 0079-6611, <http://dx.doi.org/10.1016/j.pocean.2011.02.003>.

1607 Mertens, C., and F. Schott (1998), Interannual Variability of Deep-Water
1608 Formation in the Northwestern Mediterranean, *Journal of Physical Oceanography*,
1609 28(7), 1410–1424, doi:10.1175/1520-0485(1998)028<1410:IVODWF>2.0.CO;2.

1610 Millot, C. (1999), Circulation in the Western Mediterranean Sea, *Journal of*
1611 *Marine Systems*, 20(1-4), 423–442, doi:10.1016/S0924-7963(98)00078-5.

1612 Millot, C. Interannual salinification of the Mediterranean inflow. *Geophysical*
1613 *Research Letters* 34, L21609 (2007).

1614 Puig, P., A. Palanques, and J. Martín, (2013a), Contemporary Sediment-
1615 Transport Processes in Submarine Canyons. *Annual review of marine science*, 1–25,
1616 doi: 10.1146/annurev-marine-010213-135037

1617 Puig, P. et al. (2013b), Thick bottom nepheloid layers in the western
1618 Mediterranean generated by deep dense shelf water cascading, *Progress in*
1619 *Oceanography*, 111, 1–23, doi:10.1016/j.pocean.2012.10.003.

1620 Rixen, M. et al. (2005), The Western Mediterranean Deep Water: A proxy for
1621 climate change, *Geophysical Research Letters*, 32(12), L12608,
1622 doi:10.1029/2005GL022702.

1623 Rohling, E. J., and H. L. Bryden (1992), Man-Induced Salinity and Temperature
1624 Increases in Western Mediterranean Deep Water, *Journal of Geophysical Research*,
1625 97(C7), 11191–11198, doi:10.1029/92JC00767.

1626 Salat, J., M. Emelianov, and J. L. López-Jurado (2006), Unusual extension of
1627 Western Mediterranean deep water formation during winter 2005, *Proc. 5th*
1628 *Asamblea Hispano-Portuguesa de Geodesia y Geofísica*, Univ. Sevilla, Sevilla,
1629 Spain.

1630 Schott, F., and K. D. Leaman (1991), Observations with Moored Acoustic Doppler
1631 Current Profilers in the Convection Regime in the Golfe du Lion, *Journal of Physical*
1632 *Oceanography*, 21(4), 558–574, doi:10.1175/1520-
1633 0485(1991)021<0558:OWMADC>2.0.CO;2.

1634 Schott, F., M. Visbeck, and J. Fischer (1993), Observations of vertical currents
 1635 and convection in the central Greenland Sea during the winter of 1988–1989,
 1636 *Journal of Geophysical Research*, 98(C8), 14401, doi:10.1029/93JC00658.

1637 Schott, F., M. Visbeck, U. Send et al., (1996), Observations of Deep Convection
 1638 in the Gulf of Lions, Northern Mediterranean, during the Winter of 1991/92, *Journal*
 1639 *of Physical Oceanography*, 26(4), 505–524, doi:10.1175/1520-
 1640 0485(1996)026<0505:OODCIT>2.0.CO;2.

1641 Schroeder, K., G. P. Gasparini, M. Tangherlini, and M. Astraldi (2006), Deep and
 1642 intermediate water in the western Mediterranean under the influence of the Eastern
 1643 Mediterranean Transient, *Geophysical Research Letters*, 33(21), L21607,
 1644 doi:10.1029/2006GL027121.

1645 Schroeder, K., A. Ribotti, M. Borghini, et al. (2008a), An extensive western
 1646 Mediterranean deep water renewal between 2004 and 2006, *Geophysical Research*
 1647 *Letters*, 35(18), L18605, doi:10.1029/2008GL035146.

1648 Schroeder, K., V. Taillandier, A. Vetrano, and G. P. Gasparini (2008b), The
 1649 circulation of the western Mediterranean Sea in spring 2005 as inferred from
 1650 observations and from model outputs, *Deep Sea Research Part I*, 55(8), 947–965,
 1651 doi:10.1016/j.dsr.2008.04.003.

1652 Schroeder, K., S. A. Josey, M. Herrmann, et al. (2010), Abrupt warming and
 1653 salting of the Western Mediterranean Deep Water after 2005: Atmospheric forcings
 1654 and lateral advection, *Journal of Geophysical Research*, 115(C8), 1–18,
 1655 doi:10.1029/2009JC005749.

1656 Schroeder, K. et al. (2013), Long-term monitoring programme of the hydrological
 1657 variability in the Mediterranean Sea: a first overview of the HYDROCHANGES
 1658 network, *Ocean Science*, 9(2), 301–324, doi:10.5194/os-9-301-2013.

1659 Send, U., and Marshall, J. (1995). Integral effects of deep convection, *Journal of*
 1660 *Physical Oceanography*, 25(5), 855–872.

1661 Severin, T., Conan, P., Durrieu de Madron, X. , et al. (2014). Impact of open-
 1662 ocean convection on nutrients, phytoplankton biomass and activity. *Deep Sea*
 1663 *Research Part I* 194, 62–71.

1664 Smith, R. O., H. L. Bryden, and K. Stansfield (2008), Observations of new
 1665 western Mediterranean deep water formation using Argo floats 2004–2006, *Ocean*
 1666 *Science*, 4(2), 133–149, doi:10.5194/os-4-133-2008.

1667 Somot, S., L. Houpert, F. Sevault, et al. (2016), Characterizing, modelling and
 1668 understanding the climate variability of the deep water formation in the North-
 1669 Western Mediterranean Sea, *Climate Dynamics*, doi:10.1007/s00382-016-3295-0

1670 Soto-Navarro, J., F. Criado-Aldeanueva, J. García-Lafuente, and A. Sánchez-
 1671 Román (2010), Estimation of the Atlantic inflow through the Strait of Gibraltar from
 1672 climatological and in situ data. *Journal of Geophysical Research*, 115(C10):C10023.
 1673 doi: <http://dx.doi.org/10.1029/2010JC006302>

1674 Stabholz M., X. Durrieu de Madron, A. Khripounoff, et al. (2013) Impact of open-
 1675 sea convection on particulate fluxes and sediment dynamics in the deep basin of the
 1676 Gulf of Lions. *Biogeosciences*, 10, 1097–1116.

1677 Tamburini C, M. Canals M, X. Durrieu de Madron, et al. (2013) Deep-Sea
1678 bioluminescence blooms after dense water formation at the ocean surface. PLoS
1679 ONE 8(7): e67523. doi:10.1371/journal.pone.0067523

1680 Testor, P. et al. (2007), European Gliding Observatories (EGO), Coriolis Newsl.,
1681 4, 11–12.

1682 Testor, P. et al. (2010), Gliders as a Component of Future Observing Systems, in
1683 Proceedings of the “OceanObs’09: Sustained Ocean Observations and Information
1684 for Society,” vol. 2, edited by J. Hall, D. E. Harrison, and D. Stammer, OceanObs’09,
1685 Venice, Italy.

1686 Testor P., Durrieu De Madron X., Mortier L., et al. (2016). LION observatory data.
1687 SEANOE. <http://doi.org/10.17882/44411>

1688 THETIS Group (1994), Open-ocean deep convection explored in the
1689 Mediterranean, Eos Transactions American Geophysical Union, 75(19), 217,
1690 doi:10.1029/94EO00893.

1691 Tsimplis, M. and H. Bryden (2000), Estimation of the transports through the Strait
1692 of Gibraltar. Deep Sea Research Part I, 47(12):2219–2242. doi:
1693 [http://dx.doi.org/10.1016/S0967-0637\(00\)00024-8](http://dx.doi.org/10.1016/S0967-0637(00)00024-8).

1694 Turner, J. (1973), Buoyancy effects in fluids, Cambridge University Press.

1695 Tziperman, E. and K. Speer (1994), A study of water mass transformation in the
1696 Mediterranean Sea: analysis of climatological data and a simple three-box model.
1697 Dynamics of Atmospheres and Oceans 21:53–82

1698 Vargas-Yanez, M., F. Moya, M. C. García-Martínez, et al. (2010a), Climate
1699 change in the Western Mediterranean Sea 1900–2008, Journal of Marine Systems,
1700 82(3), 171–176, doi:10.1016/j.jmarsys.2010.04.013.

1701 Vargas-Yanez, M., P. Zunino, A. Benali, et al. (2010b), How much is the western
1702 Mediterranean really warming and salting?, Journal of Geophysical Research,
1703 115(C4), C04001, doi:10.1029/2009JC005816.

1704 Voorhis, A. D., and D. C. Webb (1970), Large Vertical Currents Observed in a
1705 Winter Sinking Region of the Northwestern Mediterranean., Cahiers
1706 Oceanographiques, XXII(6), 571–580.

1707 Waldman, R., S. Somot, M. Herrmann, et al. (2016), Estimating dense water
1708 volume and its evolution for the year 2012-2013 in the North-western Mediterranean
1709 Sea: An observing system simulation experiment approach, Journal of Geophysical
1710 Research, this issue, 10.1002/2016JC011694

1711 Zunino, P., K. Schroeder, M. Vargas-Yáñez, et al. (2012) Effects of the Western
1712 Mediterranean Transition on the resident water masses: Pure warming, pure
1713 freshening and pure heaving, Journal of Marine Systems, 96–97, 15-23.

1714

Figure 1. Figure

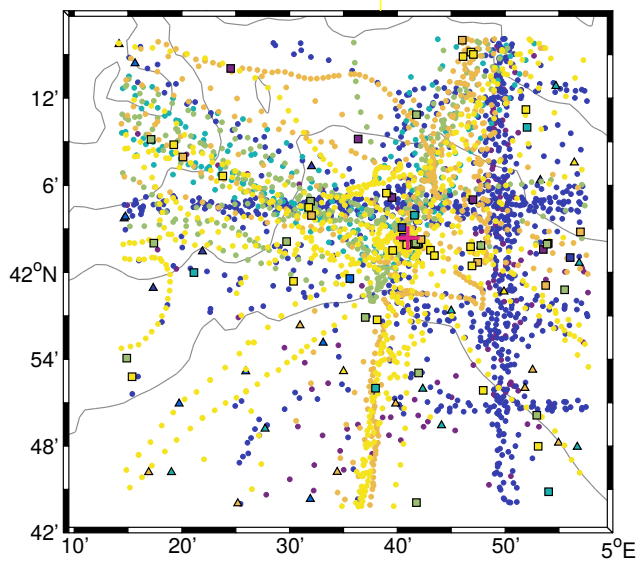
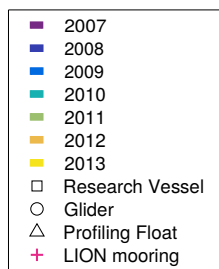
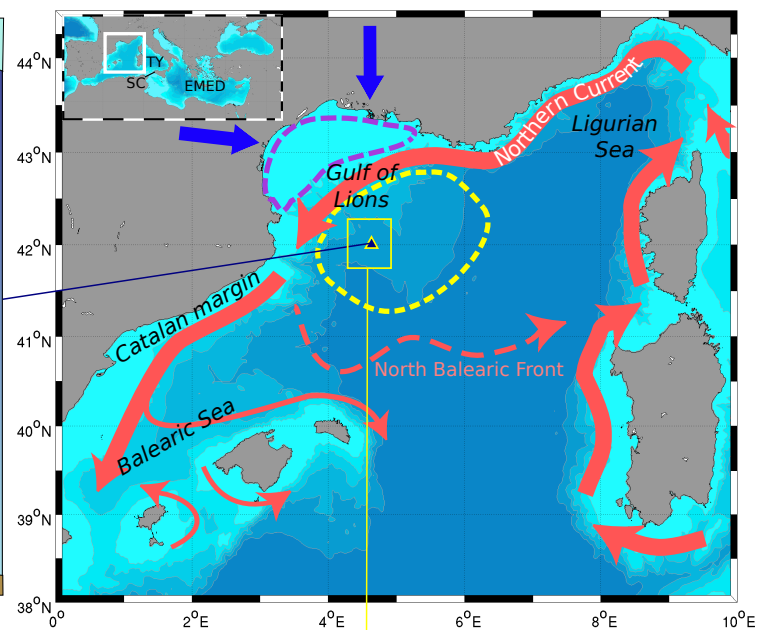
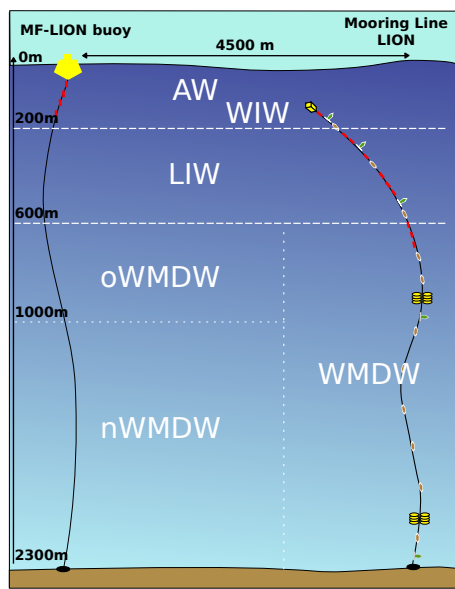


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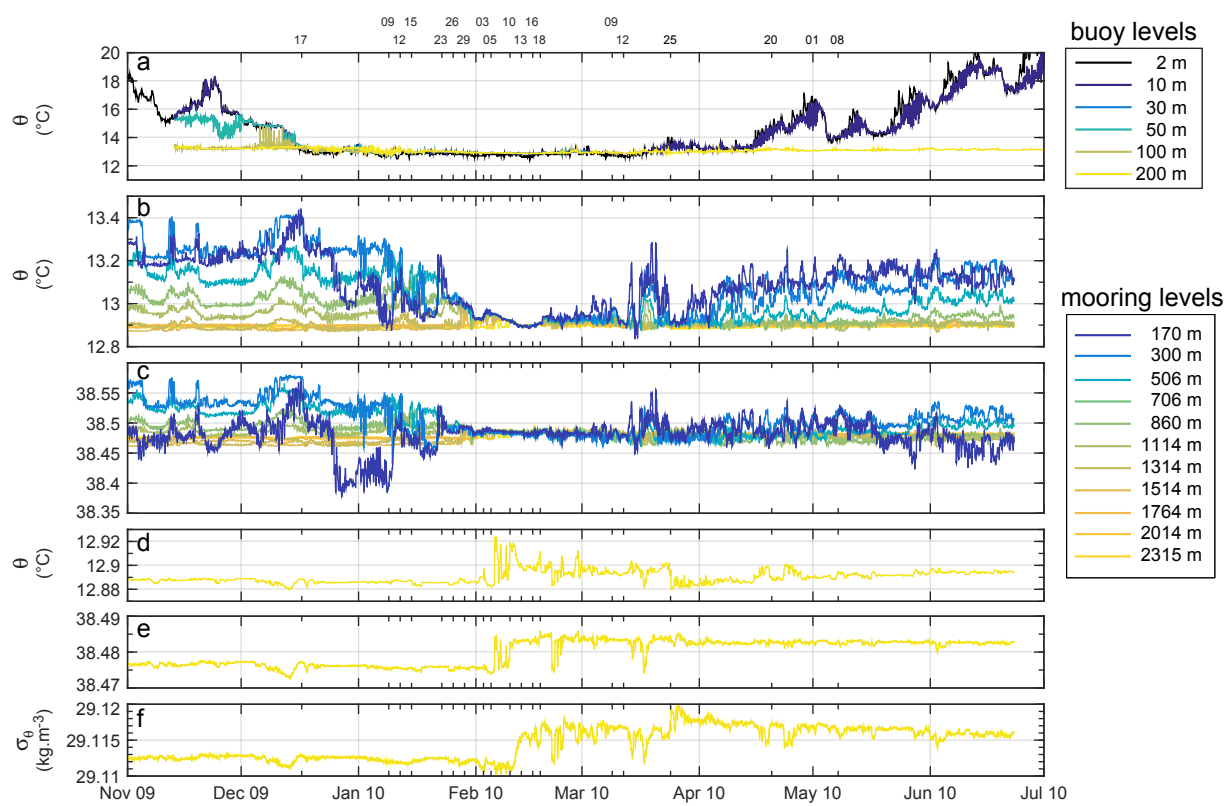


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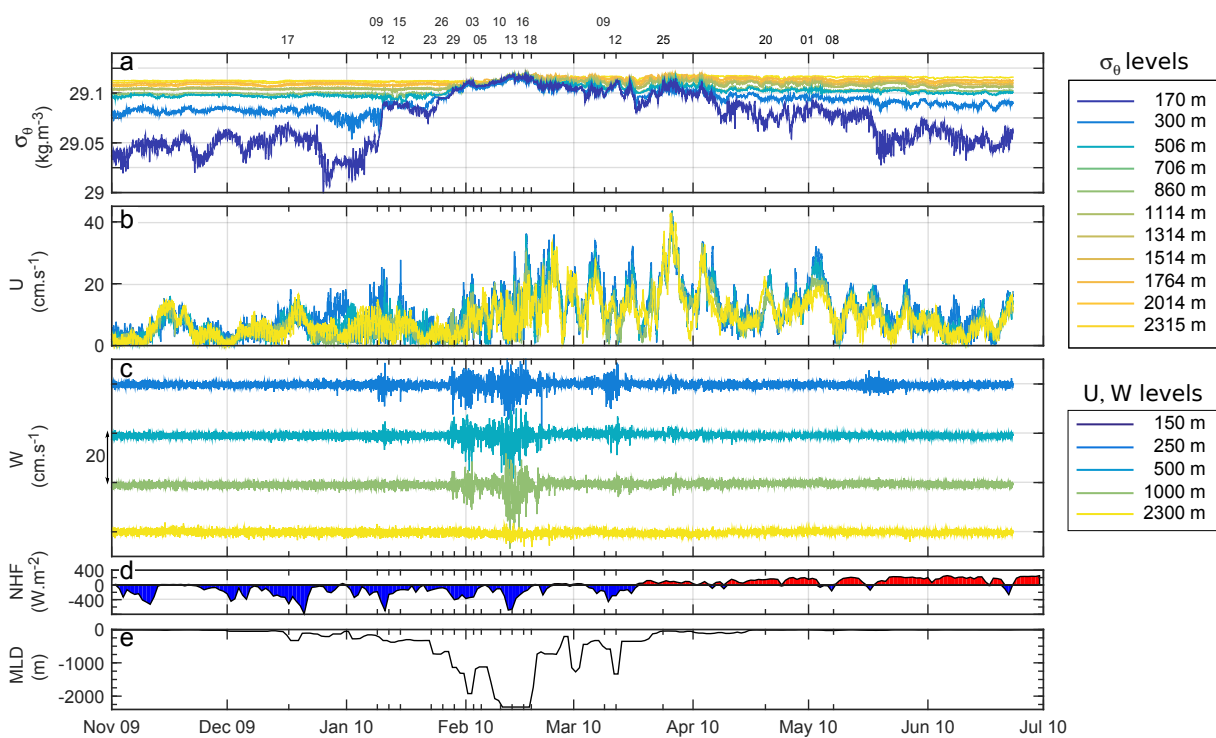


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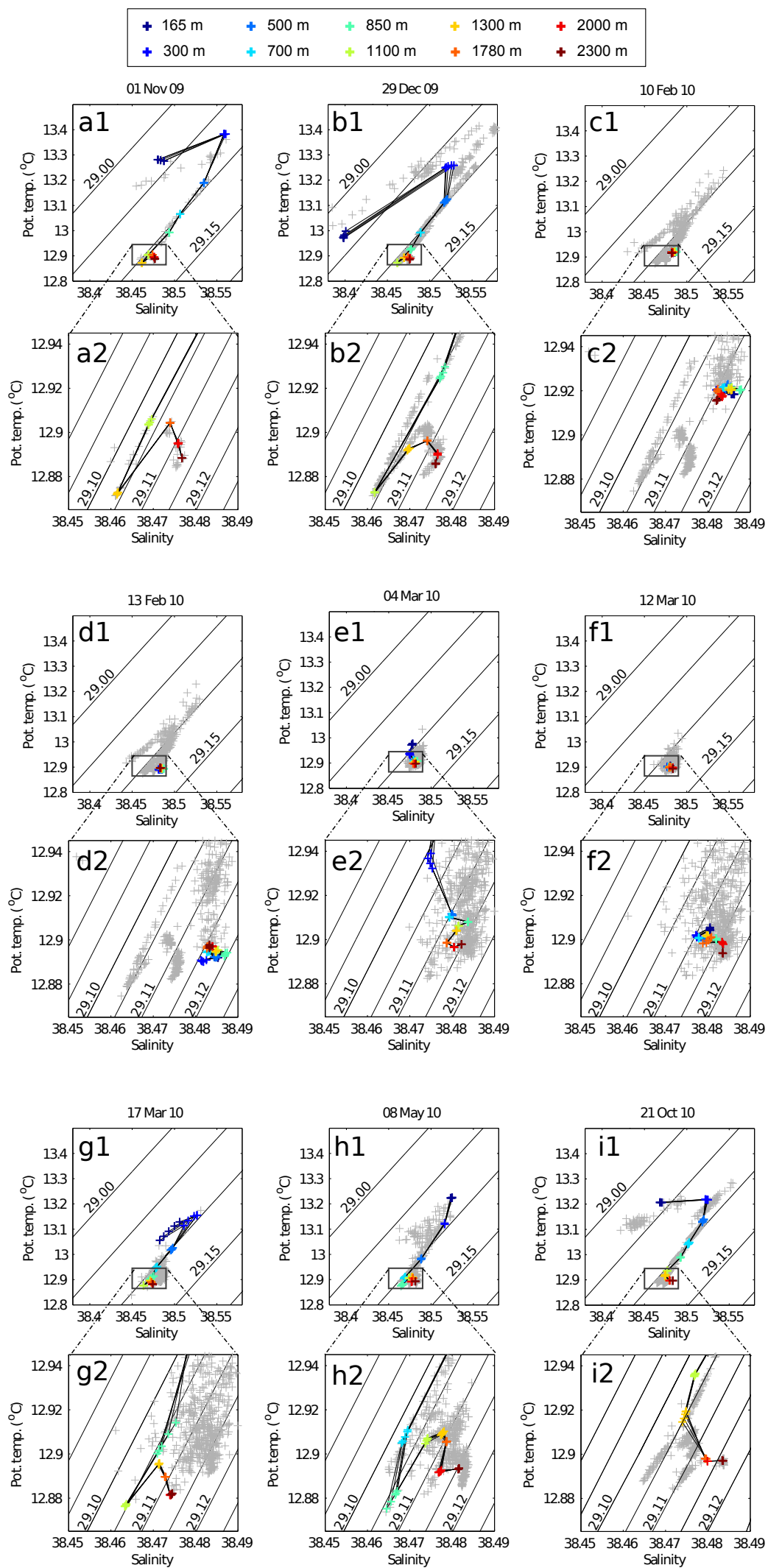


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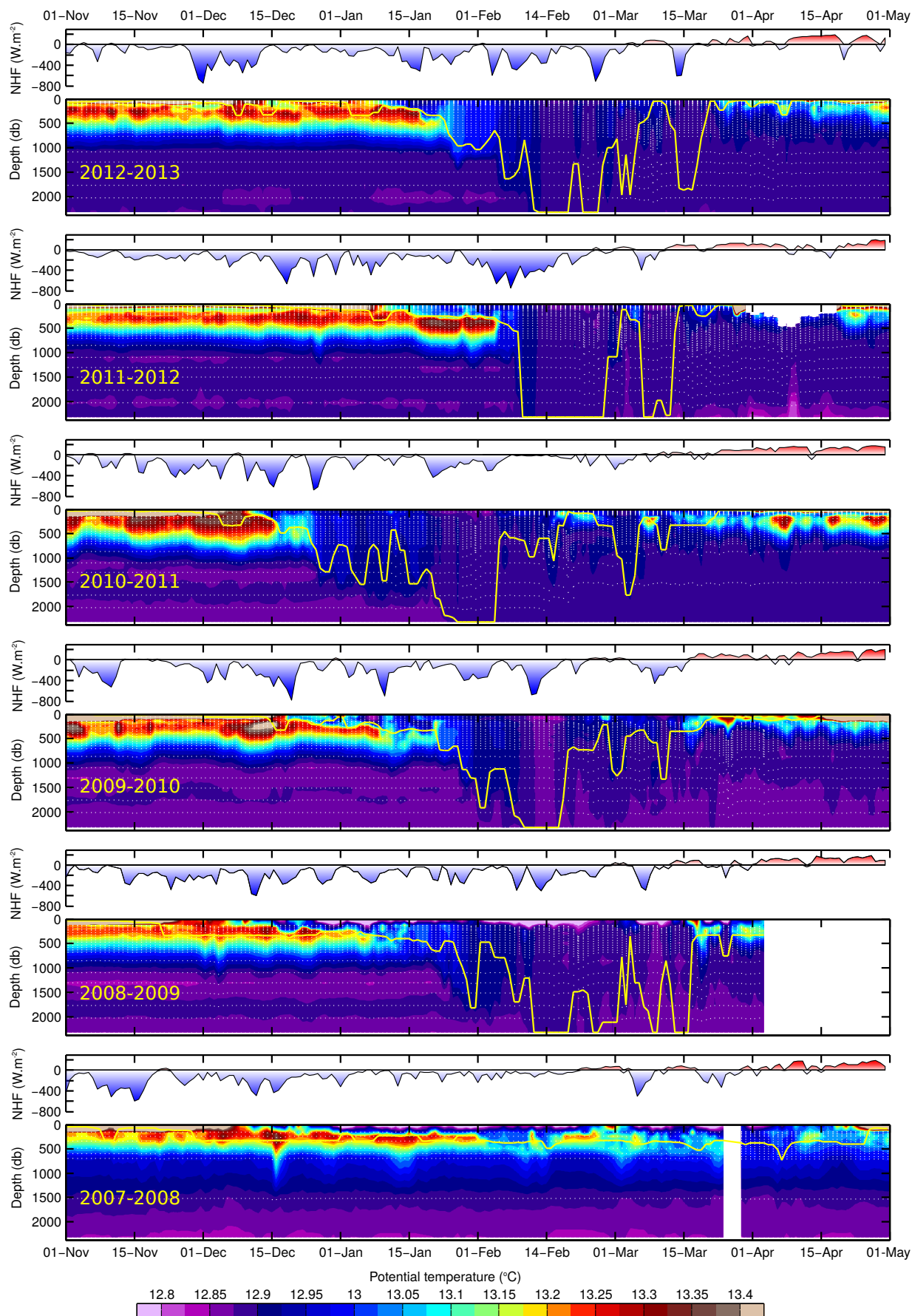


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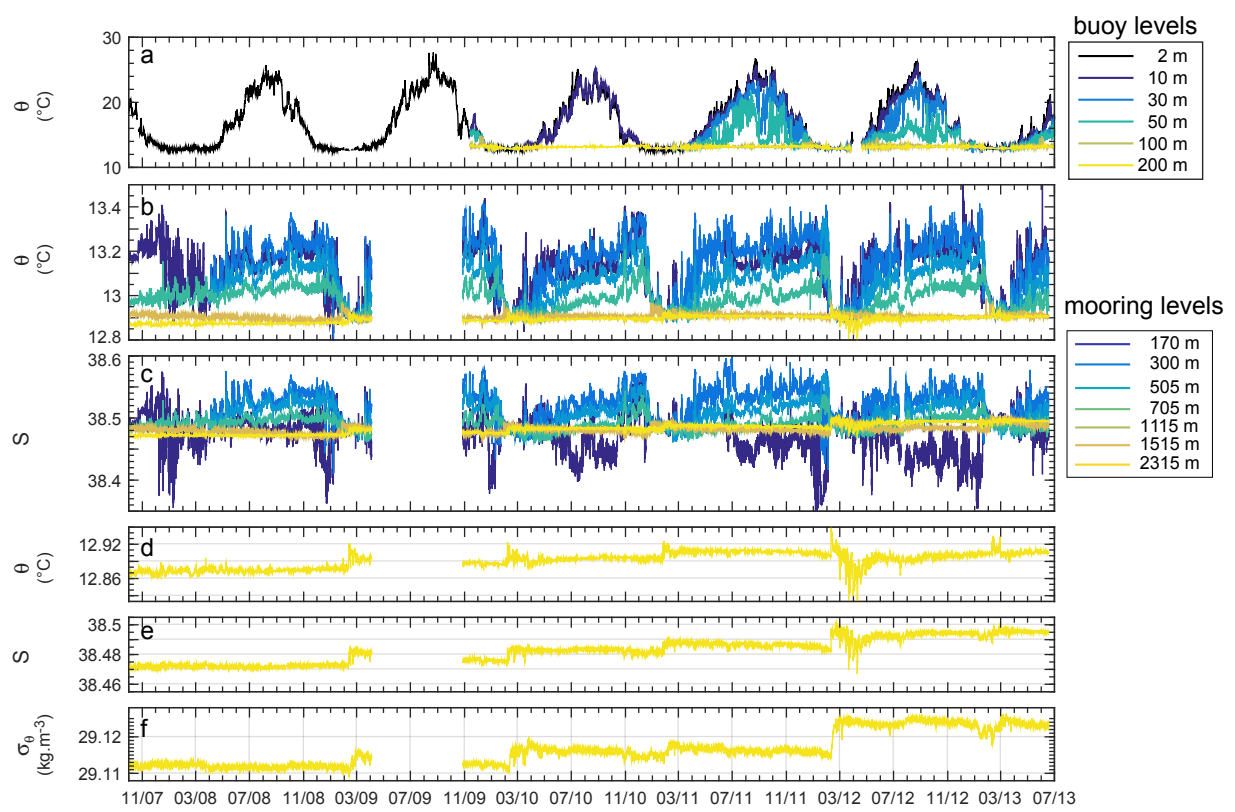


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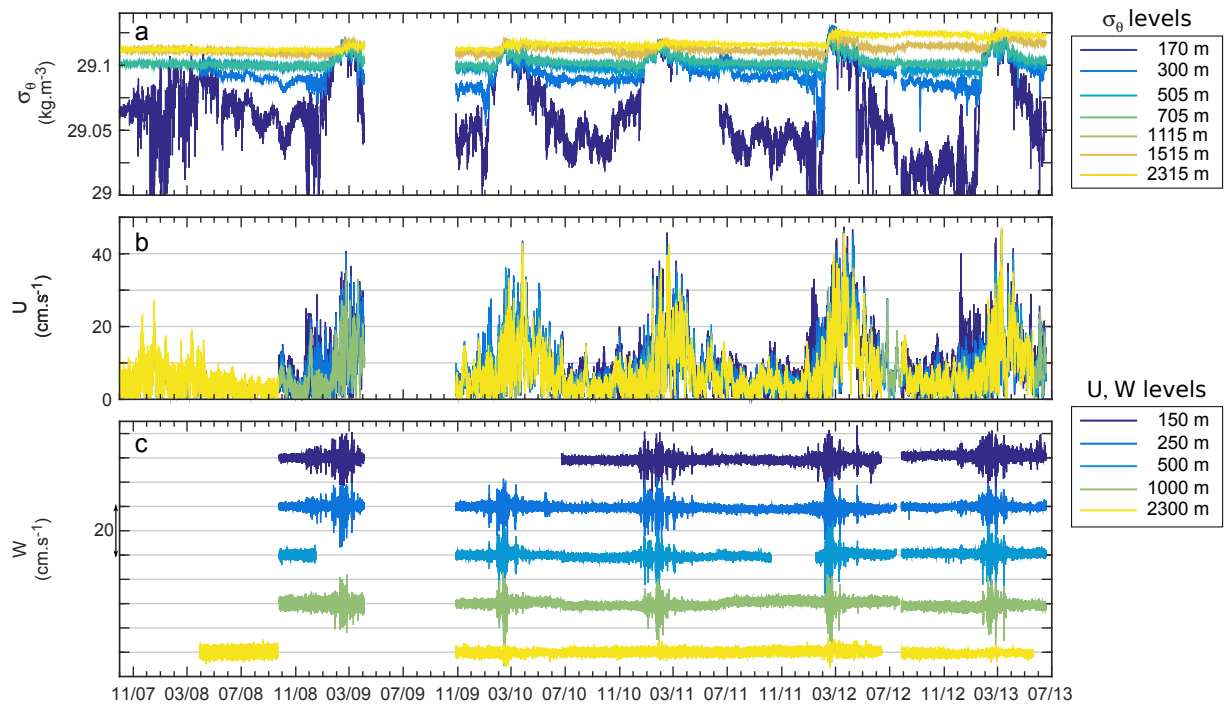


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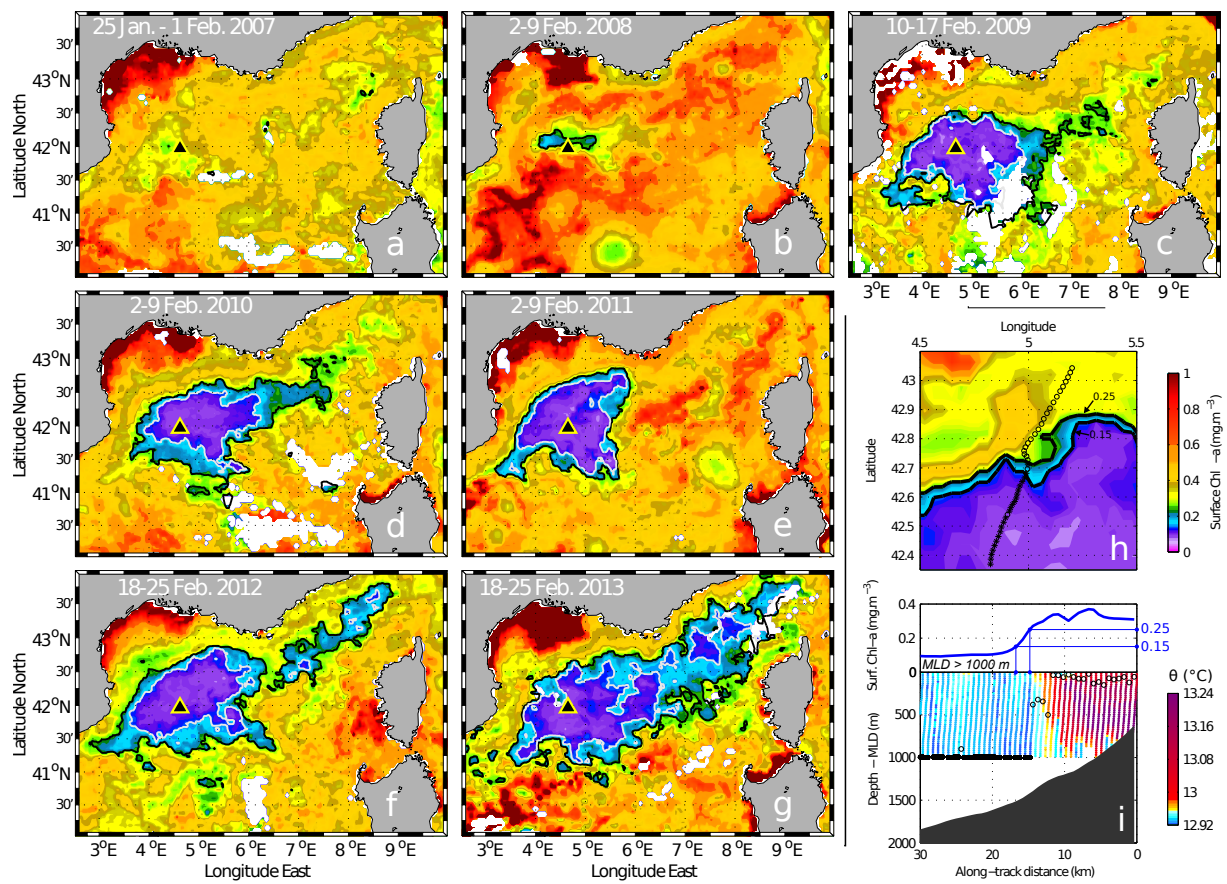


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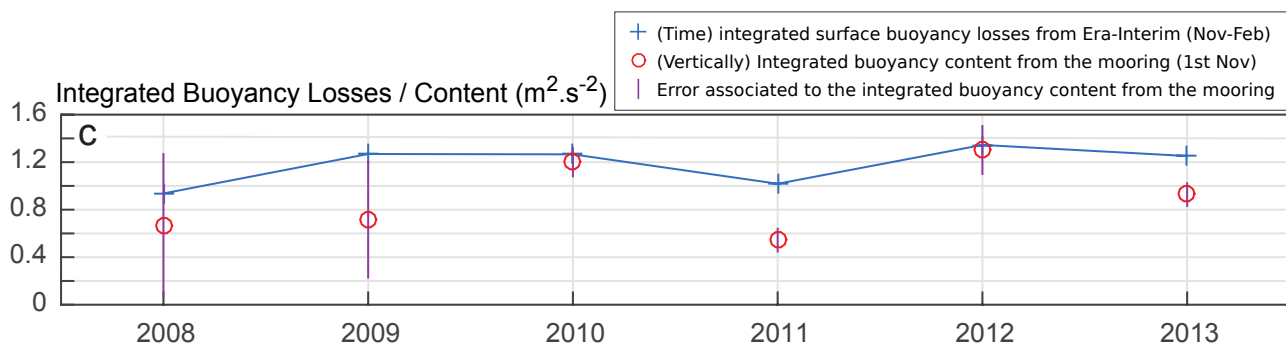
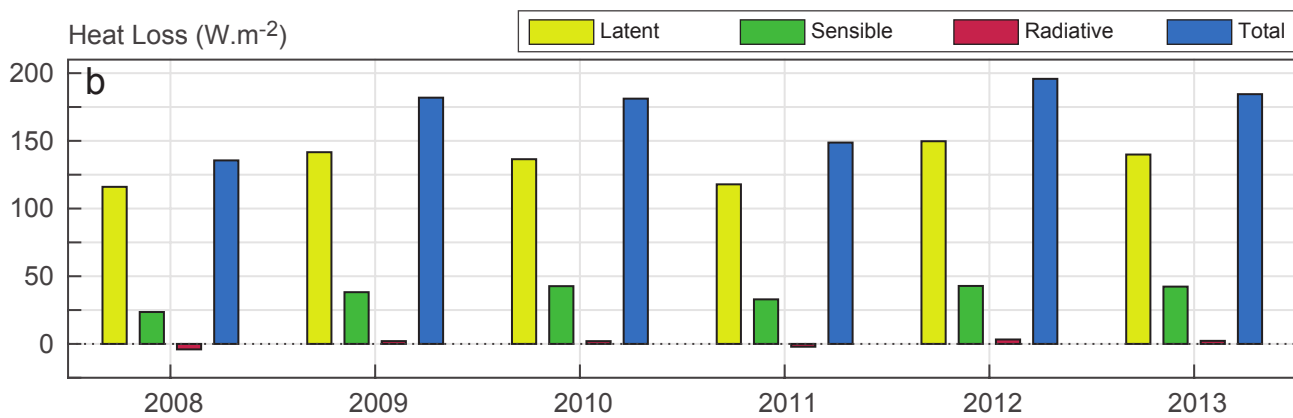
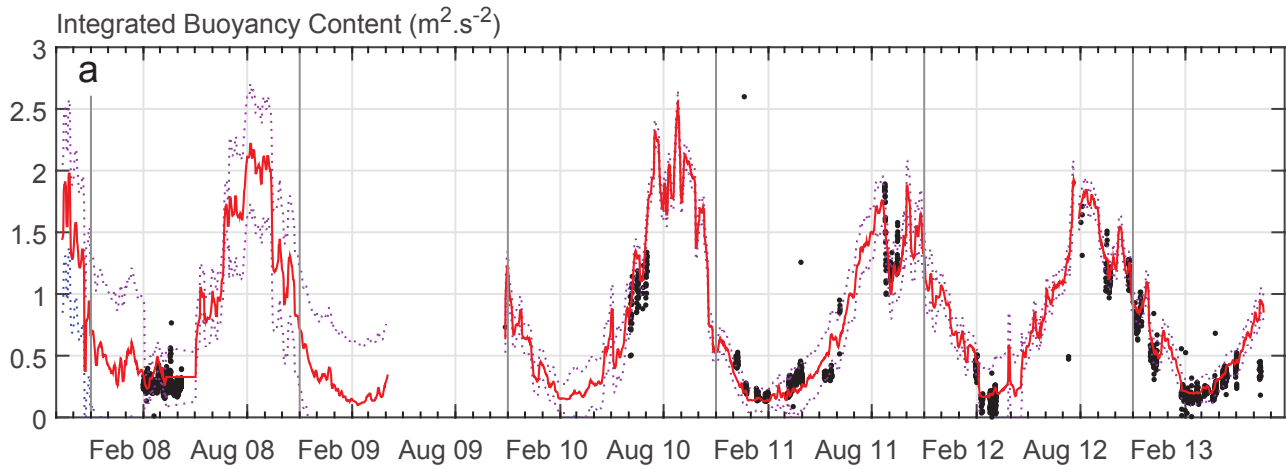


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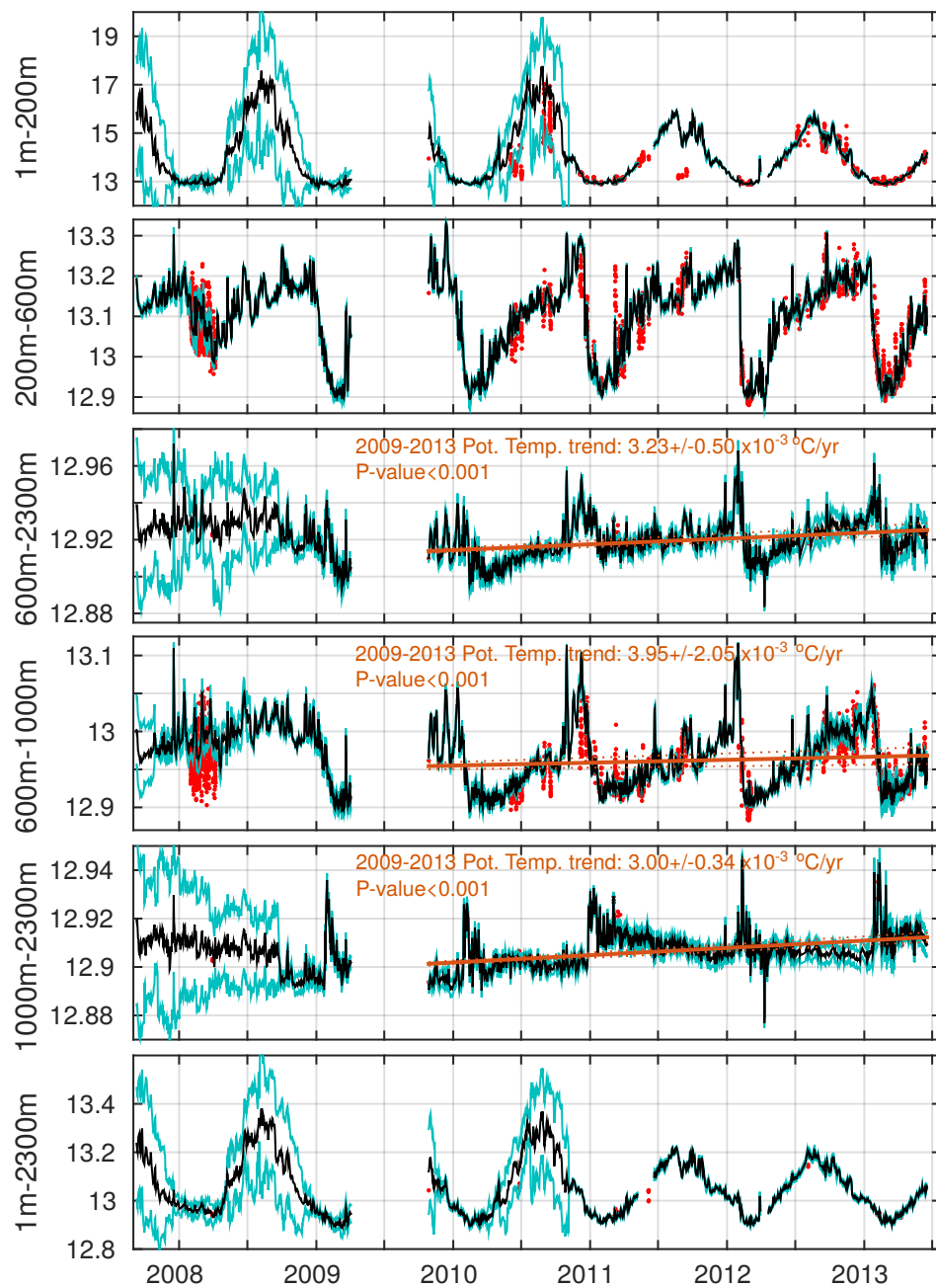


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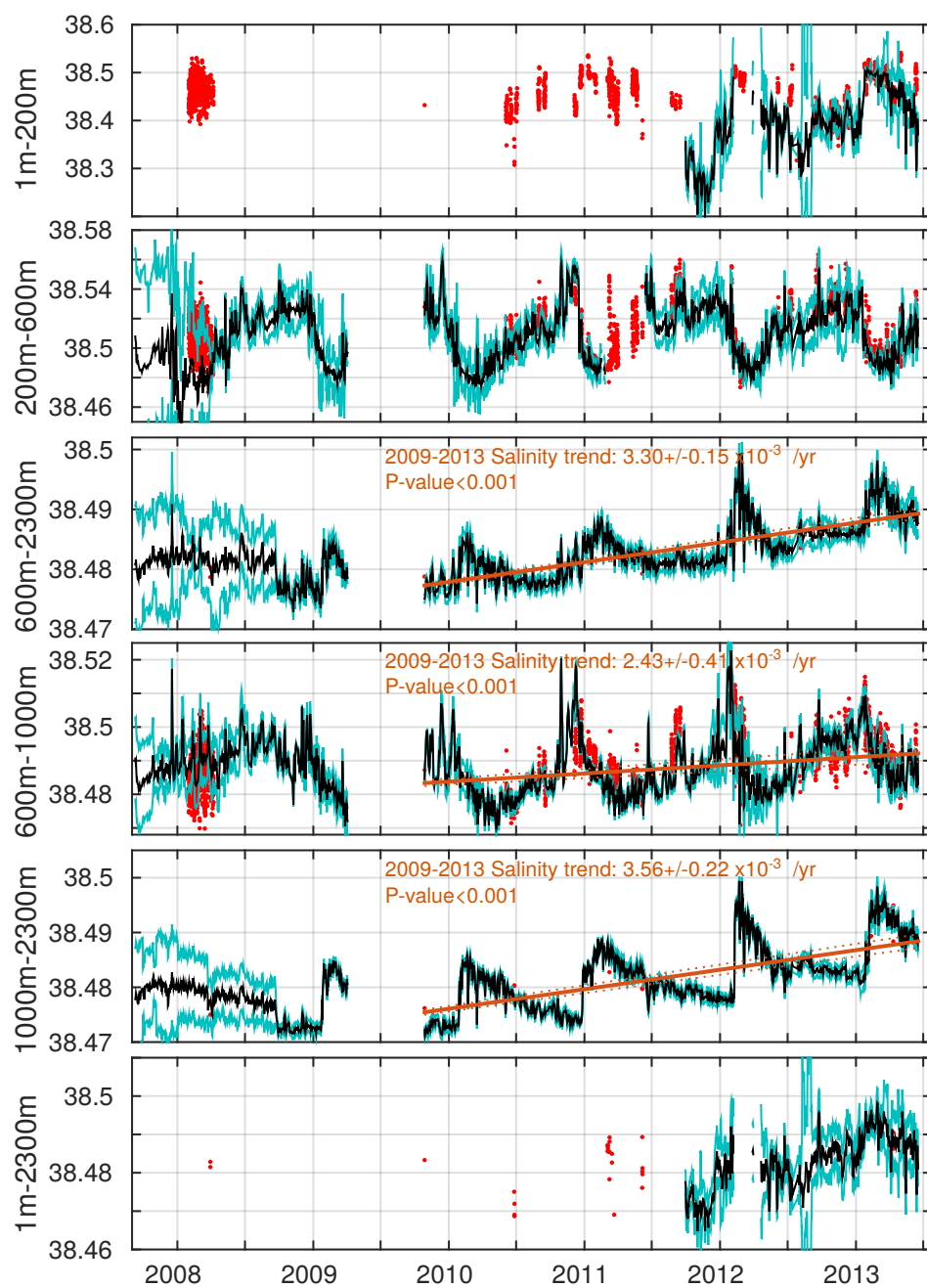


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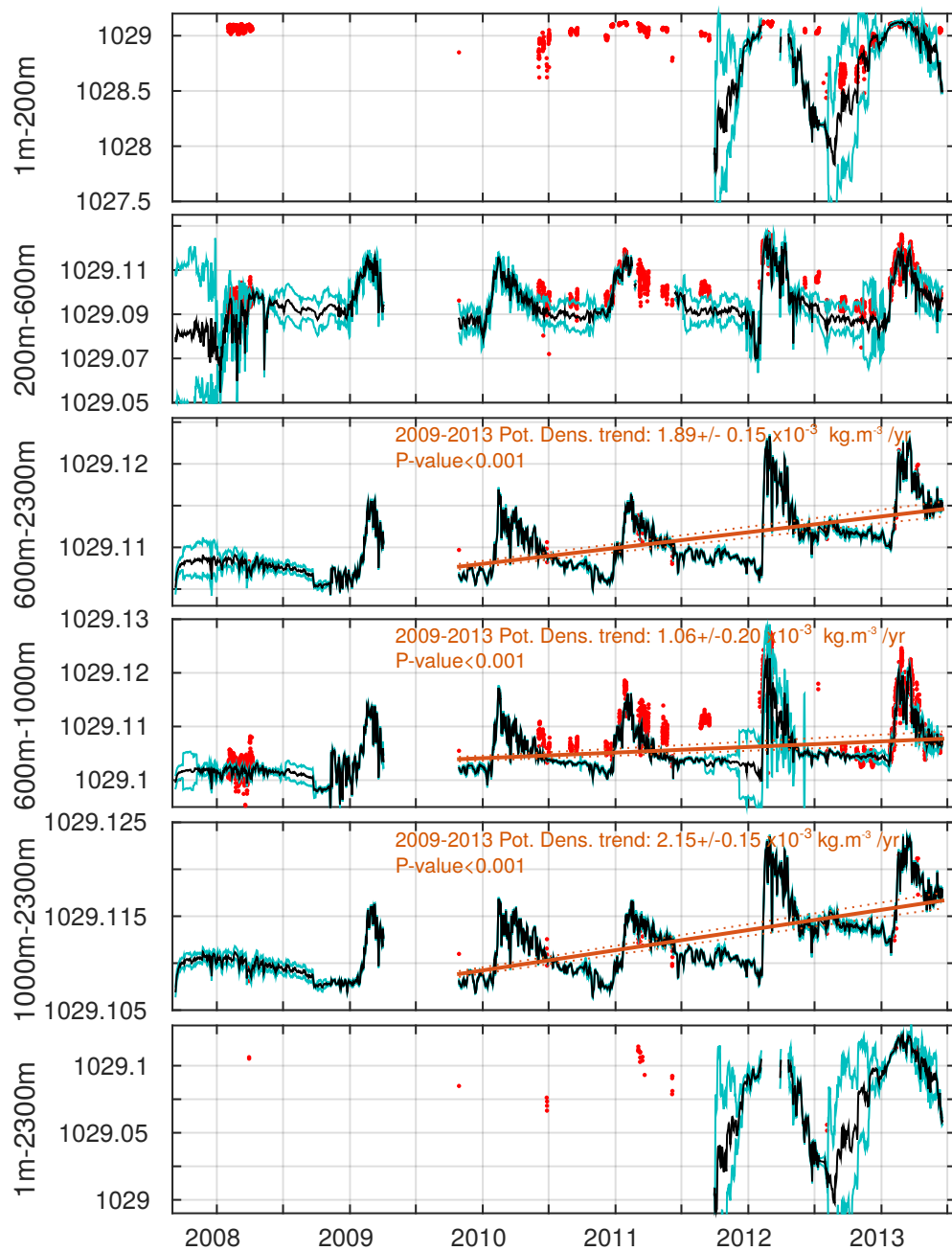


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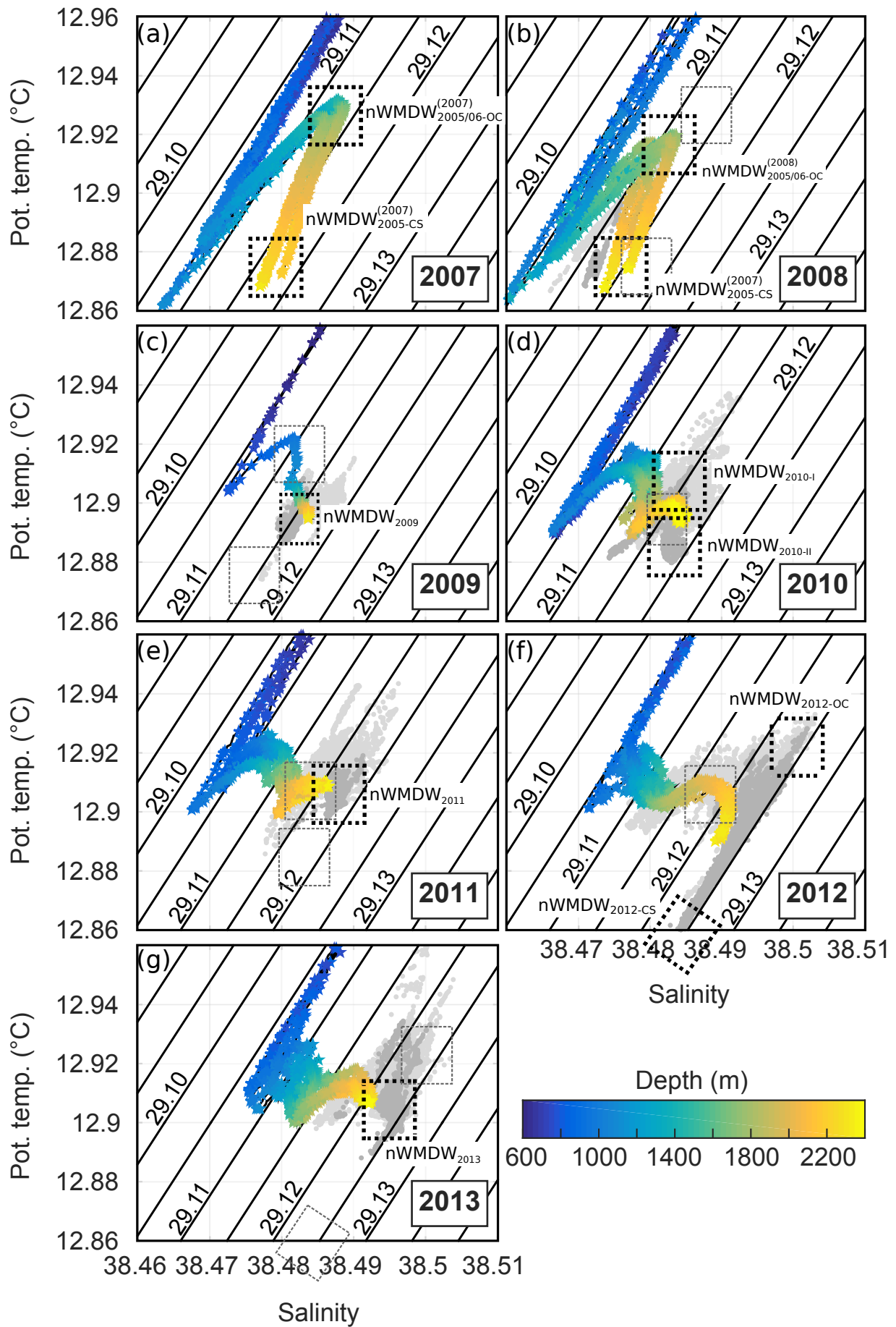


Figure A1. Figure

