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Dense Water Formation in the North-Western Mediterranean area during HyMeX-SOP2 in 1/36° ocean simulations: Ocean-atmosphere coupling impact

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Key Points.

- The AROME-NEMO WMED coupled model was run over HyMeX SOP2 and compared to a uncoupled ocean simulation validated against observations
- Air-Sea coupling induces small differences in surface fluxes and in Dense Water Formation (chronology, characteristics and volumes)
- Fine-scale ocean structures around and interacting with the convective patch are the most sensitive to the air-sea coupling

Abstract. The north-western Mediterranean Sea is a key location for the thermohaline circulation of the basin. The area is characterized by intense air-sea exchanges favoured by the succession of strong northerly and north-westerly wind situations (mistral and tramontane) in autumn and winter. Such meteorological conditions lead to significant evaporation and ocean heat loss that are well known as the main triggering factor for the Dense Water Formation (DWF) and winter deep convection episodes.

During the HyMeX second field campaign (SOP2, 1 February to 15 March 2013), several platforms were deployed in the area in order to document the DWF and the ocean deep convection, as the air-sea interface conditions.

This study investigates the role of the ocean-atmosphere coupling on DWF during winter 2012-2013. The coupled system, based on the NEMO-WMED36 ocean model ($1/36^\circ$ resolution) and the AROME-WMED atmospheric model (2.5 km-resolution), was run during two months covering the SOP2 and is compared to an ocean-only simulation forced by AROME-WMED real-time forecasts and to observations collected in the north-western Mediterranean

area during the HyMeX SOP2. The comparison shows small differences in terms of net heat, water and momentum fluxes. On average, DWF is slightly sensitive to air-sea coupling. However fine-scale ocean processes, such as shelf DWF and export or eddies and fronts at the rim of the convective patch are significantly modified. The wind-current interactions constitute an efficient coupled process at fine scale, acting as a turbulence propagating vectors, producing large mixing and convection at the rim of the convective patch.

1. Introduction

1 The north-western Mediterranean Sea is a key location for the thermohaline circulation
2 of the basin. In the Gulf of Lion (GoL), the general circulation in the area is character-
3 ized by a cyclonic gyre [*Millot, 1999*] with three distinct layers, despite a relatively weak
4 stratification: Atlantic Water (AW) in the upper layer, above Levantine Intermediate
5 Water (LIW), itself above Western Mediterranean Deep Water (WMDW). The succession
6 of strong-wind situations in winter is well known as the major triggering factor for the
7 Dense Water Formation (DWF) in the western Mediterranean [*Schott et al., 1996; Mar-*
8 *shall and Schott, 1999*]. The DWF interannual variability is strongly controlled by the
9 interannual variability of the winter-integrated buoyancy loss, which is connected to the
10 heat loss variability during the winter [*Somot et al., 2016*]. A strong buoyancy loss was
11 notably responsible for the exceptional DWF that occurred in the area in winter 2005, in
12 terms of extension and volume of newly formed WMDW [*Herrmann et al., 2010*]. Indeed,
13 in the north-western Mediterranean Sea region, air-sea fluxes present a large variability
14 in space and time. Intense air-sea exchanges (strong momentum flux, evaporation and
15 heat loss) notably occur when the mistral and tramontane (northerly and north-westerly
16 wind, respectively) affect the area in autumn and winter. They induce extreme cooling
17 and salting of the surface layer. If the surface water is enough dense, a violent mixing
18 occurs, sometimes reaching the seafloor (2500m-depth). This process is known as deep
19 ocean convection.

20 The estimation and representation of DWF in ocean model is still challenging. Large
21 uncertainties are notably due to the calculation of the exchanges (heat, freshwater, mo-

22 momentum and kinetic energy) at the air-sea interface [*Caniaux et al.*, 2017], which strongly
23 control DWF [*Herrmann and Somot*, 2008; *Herrmann et al.*, 2010; *Carniel et al.*, 2016;
24 *Estournel et al.*, 2016a; *Somot et al.*, 2016]. From the ocean modelling point of view, the
25 surface forcing can be of two kinds. The first forcing method ("bulk" method) consists
26 in using the atmospheric fields (wind, humidity and air temperature, etc.) produced by
27 an atmospheric model simulation. The air-sea fluxes are then computed in the ocean
28 model using its explicit Sea Surface Temperature (SST) and currents. The second way
29 is done by directly using the surface fluxes from an atmospheric model. This method is
30 called "flux forcing". These two methods however lead to inconsistency. In the "bulk"
31 method, there are differences in fluxes seen by the ocean and atmospheric models. These
32 differences can be enlarged when different bulk formulations are used in the two com-
33 ponent models, especially during strong wind events as the bulk parameterizations show
34 the largest discrepancies in such meteorological conditions [*e.g. Lebeaupin Brossier et al.*,
35 2008; *Olabarrieta et al.*, 2012; *Brodeau et al.*, 2017]. In the second case, the inconsis-
36 tency arises because of differences in SST. Besides, the ocean feedbacks are generally not
37 taken into account in the fluxes calculation and during the atmospheric model integration.
38 Indeed, a constant initial SST field throughout the simulation is generally used in high-
39 resolution short-range Numerical Weather Prediction models. This was proved to lead to
40 significant errors in the representation of air-sea fluxes during intense events [*Rainaud et*
41 *al.*, 2016; *Ricchi et al.*, 2016]. Ocean-atmosphere coupled system permits the calculation
42 of the surface fluxes consistently in the ocean and the atmosphere, taking jointly their
43 dynamics into account.

44 Furthermore, intense fluxes at the air-sea interface are associated with fine-scale in-
45 termittent processes in and above [below] the two boundary layers. Such processes are
46 frequent in the north-western Mediterranean sub-basin: mesoscale atmospheric systems,
47 storms, wind jets, surface temperature variations, diurnal cycle or gradients linked to ed-
48 dies, filaments or upwelling [downwelling], low-salinity lenses. To better understand and
49 represent such fine-scale and short-term intense exchanges, the development of ocean-
50 atmosphere coupled system at high-resolution is needed. Such system permits at the
51 same time to accurately solve the mesoscale systems in the two compartment models
52 and to interactively update the near-surface solutions and the exchanges between them.
53 During intense weather events, ocean-atmosphere coupling generally tends to improve the
54 air-sea fluxes and to finally moderate the corresponding atmospheric or oceanic responses.
55 For example, in the studies of *Lebeaupin Brossier et al.* [2009]; *Small et al.* [2011, 2012]
56 over the Gulf of Lion and Ligurian Sea, coupling induces in the ocean component less
57 cooling and less mixing compared to an uncoupled run. But, these two studies only focus
58 on short strong wind events in summer or autumn when the north-western Mediterranean
59 stratification is high. *Carniel et al.* [2016] investigated the coupling (including atmo-
60 sphere, ocean and waves) impact on a DWF event in the northern Adriatic Sea using the
61 COAWST system [*Warner et al.*, 2010] at high-resolution (7 km for the atmosphere and 1
62 km for the ocean [and waves]). They notably showed that the ocean-atmosphere coupling
63 improves the results in particular the total heat flux, by taking into account the dynamic
64 SST prediction in the system. Overall, they concluded that coupling ocean and atmo-
65 sphere even in a sub-region of the model domain, may significantly change the circulation
66 and water mass characteristics even in a wider area and can strongly affect the volume

67 of water involved in the densification and its contribution in the deep sea ventilation.
68 Several studies in the Mediterranean already highlight that taking waves into account
69 significantly modify the representation of the atmosphere stability, the wind, the ocean
70 cooling and mixing [*Renault et al.*, 2012; *Ricchi et al.*, 2016; *Carniel et al.*, 2016]. Indeed,
71 waves play a significant role on the surface roughness length and on the turbulent flux
72 estimation [*Janssen*, 2004]. The momentum flux parameterization is a key parameter for
73 the three components, as it intervenes in the air-sea, air-waves and waves-sea exchanges.
74 Moreover, waves strongly modify the upper-ocean turbulence [*Craig and Banner*, 1994;
75 *Ardhuin and Jenkins*, 2006], and thus can interplay with convection and DWF.

76 To validate ocean-atmosphere coupled models, simultaneous and co-localized observa-
77 tions of the two boundary layers are also needed. The HyMeX project (*Hydrological cycle*
78 *in the Mediterranean Experiment*) [*Drobinski et al.*, 2014] investigates the hydrological
79 cycle in the Mediterranean region. The second Special Observations Period (SOP2) over
80 the north-western Mediterranean area in February-March 2013 [*Estournel et al.*, 2016b]
81 was dedicated to the documentation of the DWF. One objective of the field campaign
82 was to better understand the fine scale processes involved in the DWF and ocean deep
83 convection, in particular the intense air-sea interactions role and feedbacks. Several atmo-
84 spheric and ocean platforms were deployed in the north-western Mediterranean Sea during
85 SOP2: aircraft with turbulent measurements, pressurized boundary layer balloons, radio-
86 soundings, drifting buoys, profiling floats, gliders, XBTs and CTDs from several ships in
87 the area, etc. This observation dataset represents a challenging opportunity to identify
88 the coupled processes and small scale ingredients leading to DWF.

89 The objective of this study is to evaluate the role of coupled processes in terms of air-sea
90 exchanges and of DWF rates, characteristics and extent, taking benefit of the dense data
91 collection obtained during the SOP2. To do so, two numerical experiments are compared:
92 the AROME-NEMO WMED coupled run and an ocean-only (uncoupled) simulation run
93 with NEMO-WMED36 alone. This latter run was forced by air-sea fluxes extracted from
94 the AROME-WMED real-time forecasts [*Fourrié et al.*, 2015], where the ocean is only
95 seen in the form of a SST analysis updated daily to compute the surface fluxes. This
96 can be seen as a classical flux-forced approach. The reference ocean-only simulation was
97 chosen after a large comparison and validation against HyMeX SOP2 observations done
98 in *Léger et al.* [2016], where it was shown as the most realistic run in terms of dense
99 water mass characteristics and formation chronology from a sensitivity study to initial
100 conditions, despite a low initial stratification inducing a wide convective patch.

101 The numerical coupled system and the two experiments are presented in details in
102 section 2. Section 3 analyzes the air-sea interactions at fine scale. The sensitivity of DWF
103 to the coupling is then evaluated in section 4, before focusing on mesoscale ocean features
104 and coupled processes role in section 5. Finally, a summary and concluding remarks are
105 given in section 6.

2. Numerical experiments

2.1. The coupled system: AROME-NEMO WMED

106 The AROME-NEMO WMED coupled system combines the non-hydrostatic convective-
107 scale Numerical Weather Prediction system of Météo-France, AROME [*Seity et al.*, 2011]
108 and the ocean model NEMO [*Madec et al.*, 2008].

2.1.1. The atmospheric model

110 The atmospheric model configuration is AROME-WMED [Fourrié *et al.*, 2015].
111 AROME-WMED has a 2.5 km-horizontal resolution and cover the whole Western Mediter-
112 ranean Sea (Fig. 1a). It has 60 vertical η -levels ranging from 10 m above the ground
113 to 1 hPa. AROME-WMED uses a 1-moment microphysical parameterization [Pinty and
114 Jabouille, 1998; Caniaux *et al.*, 1994], which takes into account five classes of hydrometeors
115 (cloud liquid water, cloud ice, rain, snow and graupel). The vertical turbulent transport
116 in the boundary layer is represented by two schemes: an eddy diffusivity part based on
117 a prognostic turbulent kinetic energy parametrization following Cuxart *et al.* [2000] and
118 a dry thermal and shallow convection mass flux scheme following Pergaud *et al.* [2009].
119 Thanks to its resolution, the deep convection is explicitly resolved in AROME-WMED.
120 The radiative schemes are: the six spectral bands scheme from Fouquart and Bonnel [1980]
121 for short-wave radiation (SW) and the Rapid Radiative Transfer Model (RRTM) [Mlawer
122 *et al.*, 1997] for long-wave radiation (LW). The surface scheme in AROME-WMED is
123 SURFEX [Masson *et al.*, 2013]. Each grid mesh is split into four tiles: land, towns, sea,
124 and inland waters (lakes and rivers). Output fluxes are weight averaged inside each grid
125 box according to the fraction occupied by each respective tile, before being provided to the
126 atmospheric model. The Interactions between Soil, Biosphere, and Atmosphere (ISBA)
127 parameterization [Noilhan and Planton, 1989] with two vertical layers inside the ground
128 is activated over land tile. The Town Energy Budget (TEB) scheme used for urban tiles
129 [Masson, 2000] simulates urban microclimate features, such as urban heat islands. Con-
130 cerning inland waters, the Charnock [1955]’s formulation is used. The sea surface fluxes
131 parameterization used by AROME-WMED/SURFEX is described in section 2.1.3.

132 The atmospheric lateral boundary conditions come from the 10 km-resolution ARPEGE
133 [*Courtier et al.*, 1991] global operational forecasts with a hourly frequency.

134 **2.1.2. The ocean model**

135 The ocean compartment model is NEMO-WMED36 [*Lebeaupin Brossier et al.*, 2014]
136 with a horizontal resolution of $1/36^\circ$ over an ORCA grid (Fig. 1a). In the vertical, 50
137 stretched z-levels are used. The vertical level thickness is 1 m in surface and around 400 m
138 for the last levels (*i.e.* at 4000 m-depth). The model has two radiative open boundaries:
139 one west boundary at $\sim 4.8^\circ\text{W}$ (60 km east of the Strait of Gibraltar), one south boundary
140 across the Sicily Channel ($\sim 37^\circ\text{N}$). The Strait of Messina between Sicily and continental
141 Italy is closed. The horizontal eddy viscosity coefficient is fixed to $-1 \times 10^9 \text{ m}^2 \cdot \text{s}^{-1}$ for the
142 dynamics (velocity) with the use of a bi-Laplacian operator. The TVD scheme is used
143 for tracer advection in order to conserve energy and enstrophy [*Barnier et al.*, 2006]. The
144 vertical diffusion is performed by the standard turbulent kinetic energy model of NEMO
145 [*Blanke and Delecluse*, 1993], and in case of instabilities, a higher diffusivity coefficient
146 of $10 \text{ m}^2 \cdot \text{s}^{-1}$ [*Lazar et al.*, 1999] is used to parameterize convection (see more details in
147 Appendix A). The filtered free surface of *Roullet and Madec* [2000] is used to keep the sea
148 volume constant. A no-slip lateral boundary condition is applied and the bottom friction
149 is parameterized by a quadratic function with a coefficient depending on the 2D mean
150 tidal energy [*Lyard et al.*, 2006; *Bewier et al.*, 2012]. The runoffs are prescribed from a
151 climatology [*Bewier et al.*, 2010] and applied in surface.

152 **2.1.3. The coupling interface and air-sea exchanges**

153 The coupling interface is the SURFEX-OASIS interface [*Voldoire et al.*, 2017, sub] which
154 involves SURFEX and the OASIS3-MCT coupler [*Valcke et al.*, 2013]. This interface man-

155 ages the exchanges of heat, water and momentum between the ocean and the atmosphere
 156 (Fig. 2). The corresponding fluxes at the air-sea interface - the solar heat flux Q_{sol} , the
 157 non-solar heat flux Q_{ns} , the freshwater flux F_{wat} and the momentum flux (or wind stress)
 158 $\vec{\tau}$ - are computed only once within SURFEX taking into account near-surface atmospheric
 159 and oceanic parameters and their evolutions, following the radiative schemes and the bulk
 160 parameterization, and are used consistently in AROME-WMED and NEMO-WMED36:

$$161 \quad Q_{sol} = (1 - \alpha)SW_{down} \quad (1)$$

$$162 \quad Q_{ns} = LW_{down} - \epsilon\sigma T_s^4 - H - LE \quad (2)$$

164 where SW_{down} and LW_{down} are the incoming short-wave (solar) and long-wave (infrared)
 165 radiative heat fluxes, respectively. H and LE are the sensible and latent heat fluxes,
 166 respectively, calculated by the ECUME sea surface turbulent flux bulk parameterization
 167 [Belamari, 2005; Belamari and Pirani, 2007]. They depend on the wind speed and air-sea
 168 gradients of temperature and humidity, respectively. α is albedo, ϵ is emissivity and σ is
 169 the Stefan-Boltzman constant. T_s is the Sea Surface Temperature (SST).

$$170 \quad F_{wat} = E - P_l - P_s \quad (3)$$

171 where E is evaporation, corresponding to LE/\mathcal{L} , \mathcal{L} is the vaporization heat constant. P_l
 172 and P_s are liquid and solid precipitation in surface, respectively (directly coming from
 173 AROME to SURFEX).

$$174 \quad \vec{\tau} = (\tau_u, \tau_v) = \rho_a C_D (U_a - U_s)(\vec{U}_a - \vec{U}_s) \quad (4)$$

175 where \vec{U}_a is the wind at the lowest atmospheric level (~ 10 m). C_D is the drag coefficient
 176 calculated by the ECUME sea surface turbulent fluxes parameterization. ρ_a is the air
 177 density.

178 \vec{U}_s is the ocean near-surface horizontal current, and with T_s , they are here the only
 179 oceanic parameters needed to compute the air-sea exchanges, and thus transferred to
 180 SURFEX (Fig. 2). In return, SURFEX transfers the sea surface fluxes values to OASIS
 181 for NEMO.

182 The coupling only applies on the western Mediterranean Sea: The Atlantic Ocean, the
 183 Adriatic Sea and the western Ionian Sea are uncoupled. In these areas (grey marine
 184 zones in Fig. 1a), SST comes from the SURFEX (AROME-WMED) initial state (*i.e.* the
 185 surface analysis at 00UT each day) and remains constant during 24 hours, and, horizontal
 186 current is considered as null.

2.2. Sensitivity experiments

187 The coupled run (CPL, Tab. 1) is compared to an ocean-only simulation (NEMO-
 188 WMED36 in the forced mode) named IMAP and validated in *Léger et al.* [2016].

189 IMAP begins on 1 September 2012 and runs till 15 March 2013 (Tab. 1). The boundary
 190 conditions come from the PSY2V4R4 daily analyses of Mercator-Océan averaged monthly.
 191 The PSY2 operational system [*Lellouche et al.*, 2013] has a $1/12^\circ$ horizontal resolution
 192 and covers the North-East Atlantic Ocean, the North and Baltic Seas and the Mediter-
 193 ranean Sea. The initial conditions were build with the PSY2V4R4 analyse of 1 August
 194 2012 combined with the analysed fields of the MOOSE campaign over the north-western
 195 Mediterranean Sea. The MOOSE campaign took place from 18 July to 5 August 2012 on
 196 board of the R/V *Le Suroit*. The analysed fields, built in the frame of the ASICS-Med

197 project, cover a domain between 40°N and 12°E. They are obtained with an optimal inter-
198 polation of observations from CTD profiles in addition to profiling floats (ARGO type),
199 gliders and also SST from satellite radiometers. A numerical sensitivity study on ini-
200 tial conditions using NEMO-WMED36, proves that the ASICS-MOOSE initial conditions
201 are the most accurate to well represent DWF and thermohaline characteristics during
202 HyMeX-SOP2 [Léger *et al.*, 2016]. IMAP is driven at the air-sea interface by the net
203 heat ($Q = Q_{sol} + Q_{ns}$), freshwater (F_{wat}) and momentum fluxes ($\vec{\tau}$) taken each day from
204 the AROME-WMED hourly real-time forecasts, for ranges +1 to +24 h. This means
205 that the SST used to calculate the IMAP surface forcing is the AROME-WMED SST
206 analyses over the whole domain (see the next section and *Rainaud et al.* [2016] for the
207 complete description of the AROME-WMED SST analyses) and that the momentum flux
208 computation takes no horizontal current into account. In IMAP, the Sea Surface Salinity
209 (SSS) is relaxed towards the monthly PSY2V4R4 SSS analyses.

210 The coupled run, named CPL, starts on 15 January 2013, from the same ocean state
211 than obtained in IMAP for that day. The ocean open-boundary conditions and runoffs
212 are the same as in IMAP, *i.e.* the monthly-averaged PSY2V4R4 analyses provided by
213 Mercator Océan and the *Beuvier et al.* [2010]’s climatology, respectively. The SSS re-
214 laxation is turned off in CPL. From the ocean point of view, CPL is a continuous run
215 (NEMO-WMED36 restarts each day from the ocean state of the previous day), whereas
216 the atmospheric component (AROME-WMED) is rerun each day at 00UTC, from ini-
217 tial atmospheric conditions coming from the AROME-WMED analyses (Fig. 3). The
218 coupling frequency is 1 hour and the interpolation method used by OASIS is bilinear.

219 The two experiments run without any assimilation, neither in the ocean model nor in
220 the atmospheric model for what concerns CPL.

3. Air-sea interface

221 In this part, we compare the air-sea exchanges computed in CPL and in AROME-
222 WMED real-time forecast (which later drives the IMAP simulation). As the main dif-
223 ferences come from the SST used to compute the turbulent fluxes, we evaluate in the
224 following section the SST fields used by AROME-WMED forecast and those simulated
225 by the CPL experiment.

3.1. AROME-WMED forecast versus CPL SST

226 In AROME-WMED real-time forecasts, the SST used is the 00UT analysis obtained in
227 two steps. First, a 2D optimal interpolation (CANARI scheme, *Taillefer* [2002]) of in-situ
228 data is done using the previous 3-hourly analysis as the first guess and using a correlation
229 length of 200 km. Every 3h, about 20-25 buoy and ship observations are assimilated
230 over the AROME-WMED Mediterranean domain [*Rainaud et al.*, 2016]. This analysed
231 SST field (SST_a) is secondly blended with the daily OSTIA product (SST_o , *Donlon et*
232 *al.* [2012]) to obtain a final analysis [$SST_f = (1 - \alpha)SST_a + \alpha SST_o$ with $\alpha = 0.05$]. The
233 OSTIA SST is provided each day at 06UT with a global coverage on $1/20^\circ$ -resolution grid
234 and integrates various satellite data using an observation window of 36h centred at 12UT
235 on the previous day. Finally, the effective resolution of the SST analysis is ~ 50 -100 km
236 and, for the day D, the analysis integrates satellite-based observations since 18UT of D-3,
237 with blending. In addition, there is no SST evolution during the forecast, meaning the
238 SST is kept constant to the 00UTC analysis.

239 In CPL, the SST is prognostic (solved by NEMO-WMED36) and evolves interactively
240 according to the surface fluxes with a 1h-frequency.

241 Figure 4 shows an illustration of these SST fields, with a comparison to the MyOcean
242 L3S SST satellite product [*Buongiorno et al.*, 2012] for one day well observed (2 March
243 2013) of the SOP2. This figure shows that:

- 244 • The Northern Current path is very well simulated in CPL but brings too warm
245 AW. The AW path is also visible in the AROME-WMED SST analysis thanks to data
246 assimilation in the eastern part, but not well seen in the western part. This is probably
247 due to the large variability of the current in this area [*Conan and Millot*, 1995; *Millot*,
248 1999], with eddies and meanders which detach or enter in the shelf area, and make the
249 current path difficult to capture considering the low effective resolution (around 50-100
250 km) and despite the data assimilation.

- 251 • In CPL, the Balearic Front is thin and warm eddies - as described in *Millot and*
252 *Taupier-Letage* [2005] - are simulated in the southern part. In the AROME-WMED
253 analysis the Balearic front is smooth and no eddy can be seen.

- 254 • The cold (and fresh) shelf waters [*Estournel et al.*, 2003] are well visible in the two
255 SST fields, but the offshore convective patch is only clearly seen in CPL.

256 Due to the limitation of the direct satellite observation in winter, the comparison done
257 here can only be qualitative. Nevertheless, these difference patterns can generally be found
258 when considering the SOP2 (Fig. 5a). The Northern Current is explicitly reproduced in
259 CPL but too warm (+0.5°C), whereas it is not well captured in the AROME-WMED
260 analysis especially in the western part. The smooth Balearic front and the lack of the

261 cold offshore convective patch in AROME-WMED lead to a too high SST in the southern
262 part of the north-western Mediterranean area.

263 In addition, the interactive evolution of the SST in CPL allows to take into account the
264 diurnal variation (in case of calm situation) or rapid surface cooling (response to mistral),
265 whereas it is not the case in the real-time AROME-WMED forecasts (*see Figure S1 in*
266 *the supplementary material document*).

267 To conclude, the coupling permits to take into account the SST small-scale patterns and
268 rapid variations for the heat fluxes (and evaporation) computation. More important is
269 that there is a balance between SST and fluxes in CPL. The fluxes computed in AROME-
270 WMED real forecast (and driving IMAP) are indeed unbalanced with the ocean and have
271 also a relatively low resolution.

3.2. Sea surface fluxes

272 The time-series of the net heat flux, freshwater flux and wind stress during SOP2 are
273 shown in Figure 6. They are almost similar between CPL and the AROME-WMED
274 forecast.

275 The largest differences in net heat flux are found during strong wind events when a
276 slightly lower net heat loss is produced in CPL. At the same periods, the wind stress
277 is lower in CPL, whereas the freshwater flux (dominated by evaporation) is the same
278 between CPL and AROME-WMED (IMAP). The total differences after two months of
279 integration are finally of 660 W.m^{-2} for the net heat flux (corresponding to -6.9% of the
280 AROME-WMED [IMAP] total heat loss during SOP2), of $7 \times 10^{-4} \text{ kg.m}^{-2}.\text{s}^{-1}$ for the
281 freshwater flux but reductions of both E (-2.5%) and precipitation $P_l + P_s$ (-0.5%) in CPL
282 (not shown), and of -0.5 N.m^{-2} (-3.8%) for the stress (Fig. 6).

283 The mean flux fields during SOP2 in the two experiments as their mean differences
284 are presented in Figure 7. It shows that even if the differences on average over the
285 north-western Mediterranean Sea are small (Fig. 6), the local differences can be large.
286 The two experiments evidence the large heat loss in the area induced by strong mistral
287 and tramontane. The patterns are almost similar, however, the differences in net heat
288 flux show two areas responding differently to coupling. In CPL, compared to AROME-
289 WMED, less heat is lost in the southern offshore area, whereas the heat loss is larger
290 along the coasts. These two areas correspond well to the differences found in the SST
291 fields (Fig. 5a), with the CPL SST higher along the Northern Current and over the shelf
292 area linked to the AW (warm surface water) circulation. It produces larger turbulent heat
293 fluxes and thus a larger net heat loss (lower net heat flux). On the other hand, CPL
294 SST is lower offshore near the Balearic Islands and thus induces a lower net heat loss.
295 Even if the freshwater flux fields are more noisy, as precipitation occurs very locally, the
296 difference patterns show similitudes with the net heat flux differences (Fig. 7c). Indeed,
297 evaporation is generally reduced in the open-sea convective area whereas it is increased in
298 the coastal area, in particular over the shelf. These patterns are related to the differences
299 in the SST field between AROME-WMED forecasts and CPL (Fig. 5a). Wind stress is
300 slightly changed but differences show a reduction of the momentum flux in the center and
301 southern part of the north-western Mediterranean area (Fig. 7c). On the contrary, an
302 increase is found close to the Italian coasts and over the GoL shelf. These differences do
303 not correspond to the differences in low-level wind shown in Figure 5b. They seem to be
304 linked to differences in SST (Fig. 5a), with a small increase in the wind stress where the
305 SST is largely higher in CPL. Elsewhere, the stress is reduced because of the reduction

306 due to the surface current (see Eq. 4, as U_s is null in AROME-WMED forecasts), and
307 also offshore because of a colder surface (Fig. 5a), probably linked to a stabilization of
308 the atmospheric boundary layer and thus to a reduction of the near-surface wind (Fig.
309 5b) [*Pullen et al.*, 2006].

4. Dense water formation sensitivity to coupling

4.1. Mixed layer depth

310 Figure 8 presents the mean and maximum Mixed Layer Depths (MLDs) from a density
311 criteria (MLD is defined as the depth with a density gradient of 0.01 kg.m^{-3} with the
312 surface) during SOP2 for the two experiments. It shows that they have a quite similar
313 convective patch, from the GoL to the Ligurian Sea. Although some deep mixed profiles
314 were observed in the Ligurian Sea during SOP2, the convection in IMAP is overestimated
315 in this area due to a low initial stratification [*Léger et al.*, 2016]. The mean MLD is
316 generally lower in CPL than in IMAP (by 300 to 500 m, corresponding to ~ -15 to -40%),
317 except over the shelf area where it is larger by ~ 50 m (~ 2 times larger than IMAP). The
318 same difference patterns are found when considering the maximum MLD. The two distinct
319 responses for the shelf and the offshore regions correspond directly to the differences in
320 surface fluxes shown previously: in CPL, the mixing is lower in the GoL because of a lower
321 net heat loss (and evaporation and stress), whereas it is larger over the shelf due to a larger
322 net heat loss. The largest differences between CPL and IMAP MLDs (up to -2000 to -2400
323 m) are found at the rim of the deep convective patch area. In fact, they correspond to
324 some grid meshes where deep convection does not occur at all in CPL. The comparison to
325 observations is done using floats (ARGO) and CTD profiles and the spatio-temporally co-
326 localized simulated profiles: 213 profiles, located offshore, are considered and the observed

327 MLD is obtained with the same density criteria. The MLD distribution (Fig. 9) confirms
328 that the number of very deep-mixed simulated profiles ($\text{MLD} > 1750$ m) is lower and closer
329 to observations in CPL than IMAP (*see also Figure S2 in the supplementary material*
330 *document*). However, for the other MLD classes, the number of profiles is closer to the
331 observations in IMAP. Figure 9 also highlights that during SOP2 either the water column
332 is stratified with MLD shallower than 250 m, or, the whole column is mixed and the MLD
333 is deeper than 1750 m.

334 To further evaluate the differences in the MLD fields, we computed skill scores as
335 classically done to qualify mesoscale prediction of severe events (*see Ducrocq et al.* [2002]
336 and Appendix B) using the 213 "observed" MLDs as verification (*see Figure S2 in the*
337 *supplementary material document*). Indeed, these skill scores measure the ability of the
338 high-resolution models to reproduce the deep [extreme] convection event with a good
339 intensity, size and location and allow to evaluate more finely the reliability of the two
340 simulations for the deep ocean convection. Done for several MLD thresholds (Fig. 10),
341 CPL shows an improvement of the deep convective patch representation: deeper the
342 threshold is, better CPL is compared to IMAP. For the threshold of 1750 m-depth, the
343 HSS shows a good representation of the deep mixing event for the two experiments, better
344 than a random prediction. The HSS is 0.49 for CPL and 0.41 for IMAP proving that the
345 localization of the convective patch is a little better in the coupled simulation. The FBias
346 is 1.41 for IMAP against 1.21 for CPL, which shows the overestimation of the mixed patch
347 in both simulations, but more significant in IMAP. The strong ability to create more events
348 above the threshold, leads to a higher and better POD (0.72) but a higher and worse FAR

349 (0.49) in IMAP than in CPL (respectively 0.71 and 0.41). On the contrary, for the smaller
350 thresholds the skill scores present better results for IMAP.

4.2. Water mass characteristics

351 The θ/S characteristics over the north-western Mediterranean area strongly change
352 during SOP2 (Fig. 11). After two months, both simulations show a cooling and an increase
353 in salinity for the ocean upper layers (0-350 m). The LIW are less pronounced in mid-
354 March than in mid-January, with a decrease in salinity and temperature at 350 m- to 600
355 m-depth, corresponding to LIW mixing with the upper-layer water. The WMDW shows
356 an increase in salinity (+0.002 psu) and a small increase in temperature below 1500 m-
357 depth (+0.005 °C), corresponding to the newly formed dense water. CPL and IMAP have
358 similar θ/S characteristics for WMDW, which are only a very little warmer (+0.015°C)
359 and saltier (+0.002 psu) than observed (at 1950 m-depth). Considering that the "observed
360 θ/S diagram" is an unweighted average over the North-Western Mediterranean Sea and
361 over the whole SOP2, *i.e.* it is built from an inhomogeneous dataset in space and time,
362 such differences can be considered as not significant. This result is confirmed by the mean
363 vertical biases against the observed profiles from floats obtained using a co-localization in
364 space and time (*see Figure S3 in the supplementary material document*) and the biases
365 and standard deviations computed for three layers and considering the whole SOP2 (Tab.
366 2). These scores show that the two experiments are very close to each other. The largest
367 differences are found for the upper layers (0-150 m). The mean differences for the whole
368 north-western Mediterranean area between CPL and IMAP is of +0.025 °C and +0.03
369 psu. When only considering the simulated profiles co-localized with (Argo type) floats
370 (unevenly distributed over the area), the differences between CPL and IMAP are of +0.039

371 °C and -0.005 psu in the 0-150 m layer (Tab. 2). The differences are +0.007 °C and -
372 0.002 psu when only the profiles co-localized with CTDs are considered (Tab. 2). For the
373 three layers (0-150 m, 150-600 m, 600 m to bottom), the two experiments are close to
374 observations with very small biases and standard deviations (Tab. 2). The only significant
375 modification is finally found for temperature in the upper layer (0-150 m), where the
376 coupling shows an improvement.

377 The two simulations are finally compared to the data collected at Lion (4.7°E-
378 42.1°N, Fig.1b) by the MOOSE mooring line and the surface buoy (*SST doi:*
379 *10.6096/HyMeX.LionBuoy.Thermosalinograh.20100308 and SSS doi:10.6096/MISTRALS-*
380 *HyMex-MOOSE.1025*) in Figure 12. The observed time-series of temperature and salinity
381 between 1 February and 15 March 2013 show three phases (Fig. 12): first a "mixing"
382 phase progressively reaching the seafloor and characterized by salinity and temperature
383 increases at 1500 m-depth (3 February) and at 2000 m-depth (8-9 February). The LIW
384 appears already mixed at the beginning of February. Then a "mixed" phase is visible with
385 small changes in S and θ , ended by a convective event marked by a new increase in θ/S
386 (27-28 February). Finally, a restratification period is seen with a high temporal variability
387 in the observations for all levels and marked in surface by θ diurnal cycles and short de-
388 creases in SSS. This restratification period ended by a new convection event from IOP28
389 on 15 March. The simulations show first a lower variability of the θ/S time-series. In
390 surface and at 300 m-depth the simulated values are close to observations. Despite initial
391 biases, the two simulations well reproduce the rapid θ/S increases at 1500 m-depth (but
392 in advance of one day) and at 2000 m-depth. During the "mixed" phase, IMAP and CPL
393 simulate increases in θ/S at 1500 and 2000 m-depth which are not observed. The largest

394 differences between IMAP and CPL clearly appears from IOP24, during the restratifica-
395 tion period, for all the levels considered, but in particular with different behaviour in the
396 very upper layer (see section 5.2).

4.3. Dense water volumes and formation rates

397 The time-series of the dense water volumes in the north-western Mediterranean Sea are
398 presented in Figure 13 for the two simulations. Almost the same evolutions of dense water
399 volume are found, with similar chronologies. However, progressively along SOP2, CPL
400 produces less water denser than 29.11 and 29.12 kg.m^{-3} than IMAP. On 15 March 2013,
401 compared to IMAP, the volume of water denser than 29.11 kg.m^{-3} is decreased in CPL by
402 4%, and of water denser than 29.12 kg.m^{-3} by 49%. The 29.11 kg.m^{-3} production rate,
403 computed by only considering the volume increasing phases during the period, is 2.59 Sv
404 in IMAP and 2.38 Sv (-8%) in CPL, and, the 29.12 kg.m^{-3} production rate is 0.77 Sv in
405 IMAP and 0.56 Sv (-27%) in CPL. On the other hand, the volume of water denser than
406 29.13 kg.m^{-3} is larger in CPL than in IMAP, but stays low (up to 320 km^3 on 3 March
407 against 50 km^3 for IMAP, Fig. 13c). This dense water is in fact a signature of the dense
408 water production in the shelf area, where the surface fluxes are larger in CPL (see section
409 5). *Waldman et al.* [2016] estimated the integral formation rate for the whole North-
410 Western Mediterranean Sea using an *Observing System Simulation Experiment* method
411 to be 2.3 ± 0.5 Sv for winter 2012-2013. They also obtained a volume of water with density
412 $\rho > 29.11 \text{ kg/m}^3$ of $17.7 \pm 0.9 \times 10^4 \text{ km}^3$ on April 2013. The coupled run with lower volumes
413 (Fig. 13) and formation rates is thus slightly in better agreement with the estimation of
414 *Waldman et al.* [2016] than IMAP.

415 In summary, at the scale of the north-western Mediterranean area, the DWF directly
416 responds to the surface flux modifications due to coupling, *i.e.* a change in the SST field
417 seen by AROME-WMED. In the offshore area, a lower heat loss, evaporation and wind
418 stress lead to a decrease in DWF. On the contrary over the shelf area, the increase of the
419 turbulent fluxes induces a larger production of dense water.

5. Mesoscale features

420 The general circulation at the basin scale is very similar between IMAP and CPL (*see*
421 *Figure S4 in the supplementary material document*). But, as preliminary indicated by
422 the modification of the convective patch perimeter (Fig. 8) and by a larger shelf DWF in
423 CPL (Fig. 13), the fine-scale ocean circulation and structures seem to be very sensitive
424 to the air-sea coupled processes. The objective of this section is to illustrate some fine-
425 scale structures response to coupling and to preliminary examine in CPL some coupled
426 processes acting at the rim of the convective zone.

5.1. Shelf DWF and export

427 Figure 14 shows an instant view (2 March 2013) of the DWF in the two experiments.
428 It highlights that, at that time, new dense water is formed over the shelf and offshore in
429 both simulations. In the offshore zone, the 29.12 kg.m^{-3} isopycnal has almost the same
430 patterns and homogeneous characteristics ($\theta=12.9^\circ\text{C}$, $S=38.45\text{-}38.5 \text{ psu}$), but it is less
431 deep and covers a wider area in IMAP than in CPL, indicating that a more intense deep
432 convection occurred in IMAP. The deep eddies at $4.4^\circ\text{E}\text{-}40.9^\circ\text{N}$ and at $4.7^\circ\text{E}\text{-}\sim 41.5^\circ\text{N}$,
433 containing and propagating deepwards and southwards the new dense water, are the most
434 significantly changed. Over the shelf, the new dense water is constrained along the coast

435 in IMAP at upper level. Its temperature is below 11 °C and its salinity is below 38 psu.
436 In CPL, the shelf dense water is warmer (~ 11.6 °C) and saltier (38.1-38.2 psu) than in
437 IMAP and flows between the surface and 75 m-depth along meanders. At the Cap Creus
438 canyon (3.5°E - 42.3°N), it overflows. When it leaves the shelf, this dense water volume is
439 rapidly integrated to the WMDW within the offshore mixed patch and diffused.

440 So, it appears that the local modifications of the surface fluxes due to coupling strongly
441 constrain the circulation over the shelf. As a result, the dense shelf water volume is
442 strongly increased (6 times larger for water denser than $29.13 \text{ kg}\cdot\text{m}^{-3}$ in CPL) and over-
443 flows in canyon. The dense shelf water salinity is significantly increased (+0.2 to +0.3
444 psu) because of the larger evaporation and of a larger mixing (related to the larger wind
445 stress) in the area (Fig. 7) and the temperature is higher in CPL than in IMAP ($+0.6^{\circ}\text{C}$)
446 despite a larger net heat loss locally ($+20 \text{ W}\cdot\text{m}^{-2}$, Fig. 7), but related to a larger mixing
447 and (warm) AW intrusion (*see* Figs. 4 [for the same date] and 15).

5.2. Offshore eddy

448 The comparison to the Lion surface buoy and mooring dataset previously showed that
449 the two simulations are very similar in terms of chronology and close to the in-situ ob-
450 servations in surface, except at the end of SOP2 (5-13 March 2013) when restratification
451 occurs. CPL shows negative biases in temperature and salinity, maximum on 6 March.
452 As highlighted by the profile time-series (Fig. 12), these biases are due to too cold and
453 fresh water in the 0-50 m layer coming at Lion. Almost the same cold bias is found in
454 IMAP, but with a delay of ~ 4 days. Indeed, Figure 15 presents the SST and SSS maps for
455 6 March 2013. It shows that the fresh and cold water intrusion is due to a very fine eddy
456 reaching the Lion buoy in CPL (Fig. 15b), whereas the cold and fresh eddy is located 10

457 km south of the moored buoy in IMAP (Fig. 15a). Elsewhere, the CPL SST is higher.
 458 Compared to IMAP, the CPL SSS is larger inside the GoL gyre and over the shelf, but
 459 lower in the Balearic front. Below 50 m-depth, the water column stays well mixed. From
 460 9 March, a diurnal warming occurs in the very thin (0-5m) near-surface layer due to a
 461 radiative heating larger than the turbulent heat loss (Fig. 12). Then, between 13 and
 462 15 March, as a new mixing event occurs, both simulations have surface temperature and
 463 salinity in agreement with observed values (Fig. 12).

5.3. Wind Energy Flux

464 In the following, the fine-scale coupled dynamical processes related to DWF are prelim-
 465 inarily evaluated. For that purpose, rather than the buoyancy flux largely controlled by
 466 the atmospheric fields (not shown), the surface Wind Energy Flux (WEF) which quan-
 467 tifies the kinetic energy flux injected in to the ocean by the wind stress at the air-sea
 468 interface [*Giordani et al.*, 2013] is computed. Indeed, the WEF is the dot product of the
 469 wind stress $\vec{\tau} = (\tau_u, \tau_v)$ with the surface horizontal ocean velocity $\vec{U}_s = (u_s, v_s)$:

$$470 \quad WEF = \vec{\tau} \cdot \vec{U}_s = \tau_u u_s + \tau_v v_s \quad (5)$$

471 When the WEF is positive, the wind stress and the surface current have the same direction
 472 and thus the atmosphere can increase the ocean mean kinetic energy; and conversely when
 473 the WEF is negative [*Giordani et al.*, 2006]. The WEF is examined in the CPL experiment
 474 as the relationship between wind/stress/currents/mixing is explicit thank to coupling.

475 Figure 16 presents the daily-mean surface fluxes and circulation, the mixed layer depth
 476 from the turbulence (where $K_z \geq 5 \text{ cm}^2 \cdot \text{s}^{-1}$) and density criteria, the daily-mean WEF
 477 and the vertical velocity for 7 February 2013 corresponding to the mistral/tramontane

478 event of IOP21c and just before the convection reach the seafloor. It shows that the
479 WEF maxima are located at the rim of the mixed patch. These maxima correspond to
480 the locations where the Northern Current and the cyclonic circulation are in the same
481 direction as the north/north-westerly wind (stress) and thus are the places where the
482 surface wind energy is efficiently injected into the currents inside the mixed layer. The
483 vertical response to the WEF forcing is a production of vertical velocity. The largest
484 intensities of the vertical velocity (Fig. 16d,e,f, up to 800 m.day^{-1} in absolute value) are
485 indeed found to be close to the locations with high WEF (for example in the western
486 $[3.7^{\circ}\text{E}-42^{\circ}\text{N}]$ and southern $[4.5^{\circ}\text{E}-41.5^{\circ}\text{N}]$ parts of the mixed (dense) patch, Fig. 16c). It
487 also shows the permutation of downward motion with upward motion, with a characteristic
488 size of $\sim 10 \text{ km}$. The injected kinetic energy participates to the destabilization of the front
489 and is a key parameter for the turbulent mixing [*Giordani et al.*, 2013]. It adjusts the
490 "mixing" layer with here a rapid and larger increase of the MLD from a turbulent criteria,
491 in particular at the western and southern boundaries (Fig. 16b) of the convective zone.

492 This indicates a conversion of the kinetic energy into turbulence and vertical motion in
493 the frontal zone and thus illustrates the major role of the wind/stress/current interactions
494 at the rim of the convective patch on turbulent mixing. Nevertheless, additional analyses
495 must be conducted to further investigate the mechanical coupled processes acting on
496 convection and DWF, as suggested by *Giordani et al.* [2017].

6. Summary and Conclusion

497 This study evaluates the mesoscale air-sea coupling impacts on DWF. For that, the
498 coupling between the NEMO-WMED36 ocean model and the AROME-WMED numerical
499 weather prediction (atmospheric) model was developed and run over two months

500 covering the HyMeX SOP2. The AROME-NEMO WMED coupled simulation (CPL)
501 was compared to an ocean-only simulation (IMAP) forced by AROME-WMED real-time
502 forecasts. A comparison to observations collected during the field campaign was also
503 done and constitutes a first validation of the high-resolution air-sea coupled system for
504 ocean purposes. This validation shows that the two simulations represent in a realistic
505 way the winter 2013 convection (MLD and chronology) and DWF event (volume and
506 characteristics) that was sampled by the field campaign.

507 The results, summarized in Figure 17, show that, first, the air-sea fluxes are slightly
508 decreased on average in the coupled simulation. The fluxes are in fact modified in rela-
509 tionship with the change in the SST field seen by SURFEX and AROME. In CPL, the
510 heat loss and evaporation are increased over the shelf and in the coastal area, whereas a
511 decrease is found elsewhere, notably over the GoL. The modifications of the wind stress
512 are small.

513 As a consequence, the offshore DWF is reduced in CPL and the deep convective patch
514 is slightly smaller corresponding to an improvement when compared to the MLD deduced
515 from in-situ profiles, but the thermohaline characteristics are not significantly changed.
516 From the categorical scores computed considering MLD thresholds, it appears that the two
517 simulations are almost similar in term of deep convective (mixed) patch. But, considering
518 the dense network of observations obtained during the field campaign, there is a high
519 potential of such skill scores when comparing ocean model abilities in representing the
520 deep convection intensity, size and location that could be useful, notably in a context
521 of inter-comparison. Over the shelf, the coupled simulation shows a high sensitivity of
522 the mixing to coupling and a larger (but limited) production of dense water ($\rho \geq 29.13$

523 $\text{kg}\cdot\text{m}^{-3}$). Despite the ocean model limitations due to the horizontal resolution of $1/36^\circ$,
524 the z-coordinate levels and the hydrostatic assumption, CPL produces an overflow of the
525 shelf DWF in the Cap Creus Canyon whose occurrence (referred as "cascading") is also
526 suggested by some observations [*Estournel et al.*, 2016b; *Testor et al.*, 2017, rev].

527 The main differences between the coupled and forced simulations are found in the frontal
528 zones, more specifically at the rim of the cyclonic gyre. The fine-scale ocean structures
529 around the mixed patch, like coastal currents, eddies, fronts and meanders, seem to be
530 very sensitive to the air-sea coupled processes. Precisely, these ocean mesoscale features
531 are in strong interaction with the convective zone, so they can control the 3D transport
532 of AW and LIW increasing locally the stratification, or, on the contrary, the transport
533 of well-mixed (dense) water columns. In addition, the configuration of the north-western
534 Mediterranean region, with characteristic strong northerly winds with fine jets and a
535 mesoscale ocean circulation marked by numerous fine-scale ocean structures, often leads
536 to optimal wind-current interactions. It results in significant vertical motion at the rim of
537 the convective patch, triggered by the kinetic energy injection from the atmosphere to the
538 mixing layer and by the front destabilization. This coupled mechanism acts efficiently and
539 at fine-scale as a turbulence propagating vector, producing large mixing and convection.

540 Even if this result must be further investigated, for example for other case studies and
541 with other coupled models, it already gives the first insights of how coupled processes like
542 mesoscale ocean structures/strong wind interactions could significantly affect the verti-
543 cal motion and convection associated with DWF and the thermohaline circulation. The
544 perspective of this work will be to use a potential vorticity approach in order to further
545 analyse the coupled processes between the surface wind and the rim of the cyclonic gyre,

546 because of their strong impacts on the ocean dynamics and dense water formation [*Gior-*
547 *dani et al.*, 2017]. Then, a vertical scheme considering the mass flux as in *Pergaud et al.*
548 [2009], which is under development for ocean, will also be used in order to improve the
549 ocean convection representation in the coupled system. Finally, using a sea state forcing
550 or introducing a wave model in the coupled system will also be considered.

Appendix A: Turbulent mixing scheme in NEMO-WMED36

551 The vertical eddy viscosity A^{vm} and diffusivity A^{vT} coefficients are computed from a
 552 TKE turbulent closure model based on a prognostic equation for the turbulent kinetic
 553 energy \bar{e} and a closure assumption for the turbulent length scales. This turbulent closure
 554 model has been developed by *Bougeault and Lacarrère* [1989] in the atmospheric case,
 555 adapted by *Gaspar et al.* [1990] for the oceanic case, and implemented in OPA by *Blanke*
 556 *and Delecluse* [1993] then by *Madec et al.* [1998] in NEMO.

557 The time evolution of \bar{e} is the result of the production of \bar{e} through vertical shear, its
 558 destruction through stratification, its vertical diffusion, and its dissipation of *Kolmogorov*
 559 [1942] type, which can be numerically written as (k is the vertical coordinate):

$$560 \quad \frac{\partial \bar{e}}{\partial t} = \frac{A^{vm}}{e_3} \left[\left(\frac{\partial u}{\partial k} \right)^2 + \left(\frac{\partial v}{\partial k} \right)^2 \right] - A^{vT} N^2 + \frac{1}{e_3} \frac{\partial}{\partial k} \left[\frac{A^{vm}}{e_3} \frac{\partial \bar{e}}{\partial k} \right] - C_\epsilon \frac{\bar{e}^{-3/2}}{l_\epsilon} \quad (\text{A1})$$

$$561 \quad A^{vm} = C_k l_k \sqrt{\bar{e}} \quad (\text{A2})$$

$$562 \quad A^{vT} = A^{vm} / P_{rt} \quad (\text{A3})$$

565 where e_3 is the level thickness, u and v are the horizontal components of the velocity, N
 566 is the local Brunt-Vaisl frequency, l_ϵ and l_k are the dissipation and mixing length scales,
 567 P_{rt} is the *Prantl* number which is a function of the Richardson number [see *Blanke and*
 568 *Delecluse*, 1993]. The constants C_k and C_ϵ are set to 0.7 and 0.1, respectively, to deal
 569 with vertical mixing at any depth.

570 The mixing length are obtained by $l_\epsilon = l_k = \sqrt{2\bar{e}}/N$ with an extra assumption con-
 571 cerning their vertical gradient: $\frac{1}{e_3} \left| \frac{\partial l}{\partial k} \right| \leq 1$ with $l = l_k = l_\epsilon$. For that two additional length
 572 scales are introduced: $l_{up}^{(k)} = \min \left(l^{(k)}, l_{up}^{(k+1)} + e_3^{(k)} \right)$ from $k = 1$ to jpj (*i.e.* the bottom
 573 level) and $l_{down}^{(k)} = \min \left(l^{(k)}, l_{down}^{(k-1)} + e_3^{(k-1)} \right)$ from $k = jpj$ to 1 with $l^{(k)} = \sqrt{2\bar{e}^{(k)}/N^{(k)}}$.

574 Then $l_k = \sqrt{l_{up}l_{down}}$ and $l_\epsilon = \min(l_{up}, l_{down})$.

575 At the surface, $\bar{\epsilon} = 60||\vec{\tau}||$ with a minimum value of $10^{-4}\text{m}^2\text{s}^{-2}$. At the bottom, $\bar{\epsilon}$
576 is assumed to be equal to the value of the level just above. Cut-offs are applied on $\bar{\epsilon}$,
577 A^{vm} and A^{vT} with minimum value of $10^{-6} \text{ m}^2\text{s}^{-2}$, $10^{-4} \text{ m}^2\text{s}^{-1}$ and $10^{-5} \text{ m}^2\text{s}^{-1}$ respective
578 minimum values.

579 The reader is referred to *Bougeault and Lacarrère* [1989]; *Gaspar et al.* [1990]; *Blanke*
580 *and Delecluse* [1993] for a complete description of the TKE vertical mixing scheme and
581 to *Madec et al.* [1998, 2008] for the implementation.

582 Furthermore, as the NEMO model is hydrostatic, convection is not explicitly solve in
583 case of static instabilities (when a profile has a low density under a high density). For that
584 purpose, the Enhanced Vertical Diffusion parameterization is used to represent convection.
585 So, in case of unstable conditions, a constant $A^{vEVD} = 10 \text{ m}^2\text{s}^{-1}$ is added on the vertical
586 eddy coefficient A^{vT} [*Lazar et al.*, 1999].

Appendix B: Skill scores for Mixed Layer Depth evaluation

587 In a similar manner than *Ducrocq et al.* [2002], we use a 2×2 contingency table (Tab.

588 A) considering different thresholds of MLD to compute:

- 589 • the frequency bias $FBIAS = (b + d)/(c + d)$;
- 590 • the probability of detection $POD = d/(c + d)$;
- 591 • the false alarm rate $FAR = b/(b + d)$;
- 592 • the Heidke skill score $HSS = (a + d - T)/(N - T)$;

593 with $N = a + b + c + d$ the total number of observations (density profiles from floats

594 (ARGO type) and R/V *Le Suroit* CTDs), $T = [(a + c)(a + b) + (b + d)(c + d)]/N$ referring

595 to the expected number of all the correct simulated values with a random simulation. The

596 FBIAS measures the ability of the model to predict the occurrence of the event "over the

597 threshold". The POD describes the ability in representing the size of the event and should

598 be pondered with the FAR, which considers the rate of false detection of the intense event.

599 It does not take into account localization errors. The HSS score measures the ability to

600 predict the event relatively to the accuracy of random simulation.

601 A perfect prediction has FAR equal to 0 and FBIAS, POD and HSS equal to 1. A

602 random prediction has HSS equal to 0.

Notations*Models and parameterizations*

AROME	Application of Research to Operations at MEscale
AROME-WMED	Western Mediterranean configuration of AROME
ARPEGE	Action de Recherche Petite Echelle Grande Echelle
ECUME	Exchange Coefficients from Unified Multi-campaign Estimates
ISBA	Interactions between Soil, Biosphere and Atmosphere
NEMO	Nucleus for European Modelling of the Ocean
(NEMO-)WMED36	Western Mediterranean basin configuration of NEMO (1/36°-resolution)
OASIS	Ocean Atmosphere Sea Ice Soil
OASIS3-MCT	version of OASIS
PSY2(V4R4)	Regional operational NEMO configuration from Mercator Océan (1/12°-resolution)
RRTM	Rapid Radiative Transfer Model
SURFEX	Surface Externalized
TEB	Town Energy Budget
TVD	Total Variance Dissipation scheme

Simulations

CPL	AROME-NEMO WMED coupled simulation
IMAP	NEMO-WMED36 simulation, initialization with the MOOSE-ASICS analysis and PSY2

Fields and constants

α	Albedo
C_D	Drag coefficient
ϵ	Emissivity
E	Evaporation
F_{wat}	Freshwater flux
H	Sensible heat flux
\mathcal{L}	Latent heat of vaporization
LE	Latent heat flux
LW	Long-wave radiative flux
LW_{down}	Downward long-wave radiative flux
MLD	Mixed Layer Depth
P_l	Liquid precipitation
P_s	Solid precipitation
Q	Net heat flux
Q_{ns}	Non-solar heat flux
Q_{sol}	Solar heat flux
ρ	Ocean density
ρ_a	Air density
σ	Stefan-Boltzman constant
S	Salinity
SSS	Sea Surface Salinity
SST or T_s	Sea Surface Temperature

SW	Short-wave radiative flux
SW _{down}	Downward short-wave radiative flux
θ	Potential ocean temperature
τ, τ_u, τ_v	Wind stress and components
U_a, u_a, v_a	Near-surface wind and components
U_s, u_s, v_s	Surface ocean velocity and components
w	Ocean vertical velocity
WEF	Wind Energy Flux
 <i>Skill scores</i>	
FAR	False alarm rate
FBIAS	Frequency bias
HSS	Heidke skill score
POD	Probability of detection
 <i>Observations</i>	
CTD	Conductivity Temperature Depth
IOP	Intensive Observations Period (The reader is referred to <i>Léger et al.</i> [2016] for IOP numbers)
SOP	Special Observations Period
R/V	Research-Vessel
XBT	eXpendable BathyThermograph
 <i>Water masses, processes and locations</i>	
AW	Atlantic Water
DWF	Dense Water Formation
GoL	Gulf of Lion
LIW	Levantine Intermediate Water
WMDW	Western Mediterranean Dense Water
 <i>Projects</i>	
ASICS-Med	Air-Sea Interaction and Coupling with Submesoscale structures in the Mediterranean
HyMeX	Hydrological cycle in the Mediterranean Experiment
MISTRALS	Mediterranean Integrated STudies at Regional And Local Scales
MOOSE	Mediterranean Ocean Observing System for the Environment
SiMed	Simulation of the Mediterranean Sea

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References

- 626 Ardhuin, F., A. D. Jenkins (2006), On the interaction of surface waves and upper ocean
 627 turbulence, *J. Phys. Ocean.*, *36* (3), 551-557, doi:10.1175/JPO2862.1.
- 628 Barnier, B., G., Madec, T., Penduff, J.-M., Molines, A.-M., Tréguier, J., Le Sommer, A.,
 629 Beckmann, A., Biastoch, C., Böning, J., Dengg, C., Derval, E., Durand, S., Gulev, E.,
 630 Rémy, C., Talandier, S., Theetten, M.E., Maltrud, J., McClean, and B., De Cuevas
 631 (2006), Impact of partial steps and momentum advection schemes in a global ocean
 632 circulation model at eddy-permitting resolution, *Ocean Dyn.*, *56* (5-6), 543-567, doi:
 633 10.1007/s10236-006-0082-1.
- 634 Belamari S. (2005), Report on uncertainty estimates of an optimal bulk formulation for
 635 turbulent fluxes. *MERSEA IP Deliverable, D.4.1.2*, 31pp
- 636 Belamari, S. and A. Pirani (2007), Validation of the optimal heat and momentum fluxes
 637 using the ORCA-LIM global ocean-ice model. *MERSEA IP Deliverable, D.4.1.3*, 88pp.
- 638 Beuvier, J., F. Sevault, M. Herrmann, H. Kontoyiannis, W. Ludwig, M. Rixen, E. Stanev,
 639 K. Béranger, S. Somot (2010), Modelling the Mediterranean Sea interannual variability
 640 during 1961-2000: Focus on the Eastern Mediterranean Transient (EMT), *J. Geophys.*
 641 *Res.*, *115* C08517, doi:10.1029/2009JC005950.
- 642 Beuvier, J., K. Béranger, C. Lebeaupin Brossier, S. Somot, F. Sevault, Y. Drillet, R.
 643 Bourdallé-Badie, N. Ferry, B. Levier, and F. Lyard (2012), Spreading of the Western
 644 Mediterranean Deep Water after winter 2005, time-scales and deep cyclone transport.
 645 *J. Geophys. Res.*, *117*, C07022, doi:10.1029/2011JC007679.
- 646 Blanke, B., and P. Delecluse (1993), Variability of the tropical Atlantic ocean simulated by
 647 a general circulation model with two different mixed layer physics, *J. Phys. Oceanogr.*,

- 648 23, 1363-1388.
- 649 Brodeau, L., B. Barnier, S. K. Gulev, C. Woods (2017), Climatologically significant effects
650 of some approximations in the bulk parameterizations of turbulent air-sea fluxes. *J.*
651 *Phys. Ocean.*, 47 (1), 5-28, doi:10.1175/JPO-D-16-0169.1.
- 652 Bougeault, P. and P. Lacarrère (1989), Parameterization of orography-induced turbulence
653 in a meso-beta scale model. *Mon. Wea. Rev.*, 117, 1872-1890.
- 654 Buongiorno Nardelli, B., C. Tronconi, A. Pisano, R. Santoreli (2012), High and Ultra-high
655 resolution processing of satellite Sea Surface Temperature data over Southern European
656 Seas in the framework of MyOcean project. *Remote Sensing Environment*, 129, 1-16,
657 doi:10.1016/j.rse.2012.10.012.
- 658 Caniaux, G., Redelsperger, J-L., and Lafore, J-P. (1994), A numerical study of the strati-
659 form region of a fast-moving squall line, Part I: General description and water and heat
660 budgets. *J. Atmos. Sci.*, 51, 2046-2074.
- 661 Caniaux, G., Prieur, L., Giordani, H. Redelsperger, J-L. (2017), An inverse method to de-
662 rive surface fluxes from the closure of oceanic heat and water budgets: Application to the
663 north-western Mediterranean Sea. *J. Geophys. Res. Ocean*, doi:10.1002/2016JC012167.
- 664 Carniel, S., A. Benetazzo, D. Bonaldo, F. M. Falcieri, M. M. Miglietta, A. Ricchi, M.
665 Sclavo (2016), Scratching beneath the surface while coupling atmosphere, ocean and
666 waves: Analysis of a dense water formation event. *Ocean Modelling*, 101, 101-112, doi:
667 10.1016/j.ocemod.2016.03.007.
- 668 Charnock, H. (1955), Wind stress over a water surface. *Quart. J. Roy. Meteor. Soc.*, 81,
669 639-640.

- 670 Craig, P. D., M. L. Banner (1994), Modeling wave-enhanced turbulence in the ocean
671 surface layer. *J. Phys. Ocean.*, *24*, 2546-2559.
- 672 Conan, P. and C. Millot (1995), Variability of the northern current off Marseilles, western
673 Mediterranean Sea, from February to June 1992. *Oceanologica Acta*, *18* (2), 193-205.
- 674 Courtier, P., C. Freydier, J-F. Geleyn, F. Rabier, and M. Rochas (1991), The ARPEGE
675 project at Météo-France. *ECMWF workshop on numerical methods in atmospheric mod-*
676 *eling*, *2*, 193-231.
- 677 Cuxart, J., P. Bougeault, and J-L. Redelsberger (2000), A turbulence scheme allowing for
678 mesoscale and large-eddy simulations. *Quart. J. Roy. Meteor. Soc.*, *126*, 1-30.
- 679 Donlon, C. J., M. Martin, J. Stark, J. Roberts-Jones, E. Fiedler, W. (2012), The oper-
680 ational sea surface temperature and sea ice analysis (OSTIA) system. *Remote Sensing*
681 *of Environment*, *116*, 140158.
- 682 Drobinski, P., V. Ducrocq, P. Alpert, E. Anagnostou, K. Béranger, M. Borga, I. Braud, A.
683 Chanzy, S. Davolio, G. Delrieu, C. Estournel, N. Filali Boubrahmi, J. Font, V. Grubisic,
684 S. Gualdi, V. Homar, B. Ivancan-Picek, C. Kottmeier, V. Kotroni, K. Lagouvardos, P.
685 Lionello, M. C. Llasat, W. Ludwig, C. Lutoff, A. Mariotti, E. Richard, R. Romera, R.
686 Rotunno, O. Roussot, I. Ruin, S. Somot, I. Taupier-Letage, J. Tintore, R. Uijlenhoet,
687 H. Wernli (2014), HyMeX, a 10-year multidisciplinary program on the Mediterranean
688 water cycle, *BAMS*, *95*, 1063-1082, doi:10.1175/BAMS-D-12-00242.1.
- 689 Ducrocq, V., D. Ricard, J-P. Lafore, F. Orain (2002), Storm-scale numerical rainfall
690 prediction for five precipitating events over France: On the importance of the initial
691 humidity field, *Wea. and Forecast.*, *17*, 1236-1256.

- 692 Estournel, C., X. Durrieu de Madron, P. Marsaleix, F. Auclair, C. Julliand, R. Vehil
693 (2003), Observation and modeling of the winter coastal oceanic circulation in the Gulf
694 of Lion under wind conditions influenced by the continental orography (FETCH exper-
695 iment). *J. Geophys. Res.*, *108* (C3), 8059, doi:10.1029/2001JC000825.
- 696 Estournel, C., P. Testor, P. Damien, F. D’Ortenzio, P. Marsaleix, P. Conan, F. Kessouri,
697 X. Durrieu de Madron, L. Coppola, J-M. Lellouch, S. Belamari, L. Mortier, C. Ulses,
698 M-N. Bouin, L. Prieur (2016a), High resolution modeling of dense water formation in
699 the north-western Mediterranean during winter 2012-2013: Processes and budget. *J.*
700 *Geophys. Res. Oceans*, *121* (7), 5367-5392, doi:10.1002/2016JC011935.
- 701 Estournel, C., P. Testor, I. Taupier-Letage, M-N. Bouin, L. Coppola, P. Durand, P. Conan,
702 A. Bosse, P-E. Brilouet, L. Beguery, S. Belamari, K. Branger, J. Beuvier, D. Bourras,
703 G. Canut, A. Doerenbecher, X. Durrieu de Madron, F. D’Ortenzio, P. Drobinski, V.
704 Ducrocq, N. Fourrié, H. Giordani, L. Houpert, L. Labatut, C. Lebeaupin Brossier,
705 M. Nuret, L. Prieur, O. Roussot, L. Seyfried, S. Somot, (2016b), HyMeX-SOP2: The
706 field campaign dedicated to dense water formation in the Northwestern Mediterranean.
707 *Oceanography*, *29* (4), 196-206, doi:10.5670/oceanog.2016.94.
- 708 Fouquart, Y. and B. Bonnel (1980), Computations of solar heating of the earth’s atmo-
709 sphere: A new parameterization. *Beitr. Phys. Atmos.*, *53*, 35-62.
- 710 Fourrié, N., E. Bresson, M. Nuret, C. Jany, P. Brousseau, A. Doerenbecher, M. Kreitz, O.
711 Nuissier, E. Sevault, H. Bénichou, M. Amodei, and F. Pouponneau (2015), AROME-
712 WMED, a real-time mesoscale model designed for HyMeX Special Observation Periods.
713 *Geo. Model. Dev.*, *8*, 1919-1941, doi:10.5194/gmd-8-1919-2015.

- 714 Gaspar, P., Y. Grégoris, and J.-M. Lefevre (1990), A simple Eddy Kinetic Energy model
715 for simulations of the oceanic vertical mixing: Tests at station Papa and Long-Term
716 Upper Ocean Study site. *J. Geophys. Res.*, *95*, 16179-16193.
- 717 Giordani, H., L. Prieur, and G. Caniaux (2006), Advanced insights into sources of vertical
718 velocity in the ocean. *Ocean Dyn.*, *56* (5-6), doi:10.1007/s10236-005-0050-1.
- 719 Giordani, H., G. Caniaux, A. Voldoire (2013), Intraseasonal mixed-layer heat budget in
720 the equatorial Atlantic during the cold tongue development in 2006. *J. Geophys. Res.*
721 *Oceans*, *118*, 650-671, doi:10.1029/2012JC008280.
- 722 Giordani, H., C. Lebeaupin Brossier, F. Léger, G. Caniaux (2017), A PV-Approach for
723 Dense Water Formation along Fronts: Application to the Northwestern Mediterranean.
724 *J. Geophys. Res. Oceans*, *122* (2), 995-1015, doi:10.1002/2016JC012019.
- 725 Herrmann, M., S. Somot (2008), Relevance of ERA40 dynamical downscaling for mod-
726 eling deep convection in the Mediterranean Sea. *Geophys. Res. Lett.*, *35*, L04607, doi:
727 10.1029/2007GL032442.
- 728 Herrmann, M., F. Sevault, J. Beuvier, S. Somot (2010), What induced the exceptional
729 2005 convection event in the northwestern Mediterranean basin? Answers from a mod-
730 eling study. *J. Geophys. Res.*, *115*, C12051, doi:10.1029/2010JC006162.
- 731 Janssen, P. (2004), The Interaction of Ocean Waves and Wind. Ed: *Cambridge University*
732 *Press*, Cambridge, UK.
- 733 Kolmogorov, A. N. (1942), The equation of turbulent motion in an incompressible fluid.
734 *Izv. Akad. Nauk SSSR, Ser. Fiz.*, *6*, 5658.
- 735 Lazar, A., G. Madec, P. Delecluse (1999), The deep interior downwelling in the Veronis ef-
736 fect and mesoscale tracer transport parameterizations in an OGCM. *J. Phys. Oceanogr.*,

- 737 29, 2945-2961.
- 738 Lebeau-pin Brossier, C., V. Ducrocq, H. Giordani (2008), Sensitivity of three
739 Mediterranean heavy rain events to two different sea surface fluxes parameteriza-
740 tions in high-resolution numerical modeling. *J. Geophys. Res.*, 113, D21109, doi:
741 10.1029/2007JD009613.
- 742 Lebeau-pin Brossier, C., V. Ducrocq, H. Giordani (2009), Two-way one-dimensional high-
743 resolution air-sea coupled modelling applied to Mediterranean heavy rain events. *Q. J.*
744 *Roy. Meteorol. Soc.*, 135, 187-204, doi:10.1002/qj.338.
- 745 Lebeau-pin Brossier, C., T. Arsouze, K. Béranger, M-N. Bouin, E. Bresson, V. Ducrocq, H.
746 Giordani, M. Nuret, R. Rainaud, I. Taupier-Letage, 2014 : Ocean mixed layer responses
747 to intense meteorological events during HyMeX-SOP1 from a high-resolution ocean
748 simulation. *Ocean Modelling*, 84, 84-103, doi:10.1016/j.ocemod.2014.09.009.
- 749 Léger, F., C. Lebeau-pin Brossier, H. Giordani, T. Arsouze, J. Beuvier, M-N. Bouin,
750 E. Bresson, V. Ducrocq, N. Fourrié, M., Nuret, (2016), Dense Water Formation in
751 the North-Western Mediterranean area during HyMeX-SOP2 in 1/36° ocean simula-
752 tions: Sensitivity to initial conditions. *J. Geophys. Res. Oceans*, 121 (8), 5549-5569,
753 doi:10.1002/2015JC011542.
- 754 Lellouche, J-M., O. Le Galloudec, M. Drévil-lon, C. Régnier, E. Greiner, G. Garric, N.
755 Ferry, C. Desportes, C-E. Testut, C. Bricaud, R. Bourdallé-Badie, B. Tranchant, M.
756 Benkiran, Y. Drillet, A. Daudin, C. De Nicola (2013), Evaluation of global monitoring
757 and forecasting systems at Mercator Océan. *Ocean Sci.*, 9, doi:10.5194/os-9-57-2013.
- 758 Lyard, F., F. Lefevre, T. Letellier, and O. Francis (2006), Modelling the global ocean
759 tides: modern insights from FES2004, *Ocean Dynamics*, 56 (5-6), 394-415.

- 760 Madec, G., P. Delecluse, M. Imbard, and C. Levy (1998), OPA8 ocean general circulation
761 model - reference manual. *Tech. rep., LODYC/IPSL Note 11*.
- 762 Madec, G. and the NEMO team (2008), NEMO ocean engine, *Note du Pole de*
763 *modélisation, Institut Pierre-Simon Laplace (IPSL), France*, No 27 ISSN No 1288-1619.
- 764 Marshall, J., and F. Schott (1999), Open-ocean convection: Observations, theory, and
765 models, *Reviews of Geophysics*, *37* (1), 1-64.
- 766 Masson, V. (2000), A physically-based scheme for the urban energy budget in atmospheric
767 models. *Bound.-Layer Meteor.*, *94*, 357-397.
- 768 Masson, V., P. Le Moigne, E. Martin, S. Faroux, A. Alias, R. Alkama, S. Belamari,
769 A. Barbu, A. Boone, F. Bouyssel, P. Brousseau, E. Brun, J.-C. Calvet, D. Carrer, B.
770 Decharme, C. Delire, S. Donier, K. Essaouini, A.-L. Gibelin, H. Giordani, F. Habets, M.
771 Jidane, G. Kerdraon, E. Kourzeneva, M. Lafaysse, S. Lafont, C. Lebeaupin Brossier,
772 A. Lemonsu, J.-F. Mahfouf, P. Marguinaud, M. Mokhtari, S. Morin, G. Pigeon, R.
773 Salgado, Y. Seity, F. Taillefer, G. Tanguy, P. Tulet, B. Vincendon, V. Vionnet, and A.
774 Voltaire (2013), The SURFEXv7.2 land and ocean surface platform for coupled and
775 offline simulation of earth surface variables and fluxes. *Geosci. Model Dev.*, *6*, 929-960,
776 doi:10.5194/gmd-6-929-2013.
- 777 Millot, C., 1999: Circulation in the Western Mediterranean Sea. *J. Mar. Syst.*, *20* (14),
778 423-442.
- 779 Millot, C. and I. Taupier-Letage, (2005), Circulation in the Mediterranean Sea. In
780 *The Mediterranean Sea*, Saliot (ed.). *Handbook of Environmental Chemistry*, *5*, 29-66.
781 Springer. Berlin and Heidelberg, Germany, doi:10.1007/b107143.

- 782 Mlawer, E.J., S.J. Taubman, P.D. Brown, M.J. Iacono and S.A. Clough, (1997), Radiative
783 transfer for inhomogeneous atmospheres: RRTM, a validated correlated-k model for the
784 longwave. *J. Geophys. Res.*, *102*, 16663-16682.
- 785 Noilhan, J., and S. Planton (1989), simple parameterization of land surface processes for
786 meteorological models. *Mon. Wea. Rev.*, *117*, 536-549.
- 787 Olabarrieta, M., J. C. Warner, B. Armstrong, J. B. Zambon, R. He (2012), Ocean-
788 atmosphere dynamics during hurricane Ida and nor'Ida: an application of the cou-
789 pled ocean-atmosphere-wave-sediment transport (COAWST) modeling system. *Ocean*
790 *Model.*, *43-44*, 112-137. doi:10.1016/j.ocemod.2011.12.008.
- 791 Pergaud, J., V. Masson, S. Malardel, and F. Couvreur (2009), A parameterization of dry
792 thermals and shallow cumuli for mesoscale numerical weather prediction. *Bound.-Layer*
793 *Meteor.*, *132*, 83-106
- 794 Pinty, J.-P., and P. Jabouille (1998), A mixed-phased cloud parameterization for use
795 in a mesoscale non-hydrostatic model: Simulations of a squall line and of orographic
796 precipitation. Preprints, *Conf. on Cloud Physics*, Everett, WA, Amer. Meteor. Soc.,
797 217-220.
- 798 Pullen, J., J. D. Doyle, and R. P. Signell (2006), Two-way air-sea coupling: A study of
799 the Adriatic, *Mon. Wea. Rev.*, *134*, 1465-1483.
- 800 Rainaud, R., C. Lebeaupin Brossier, V. Ducrocq, H. Giordani, M. Nuret, N. Fourrié, M-N.
801 Bouin, I. Taupier-Letage, and D. Legain (2016), Characterization of air-sea exchanges
802 over the Western Mediterranean Sea during HyMeX SOP1 using the AROME-WMED
803 model. *Quart. J. Roy. Meteorol. Soc.*, *142* (S1), 173-187, doi:10.1002/qj.2480.

- 804 Renault, L., J. Chiggiato, J. C. Warner, M. Gomez, G. Vizoso, J. Tintoré (2012), Coupled
805 atmosphere-ocean-wave simulations of a storm event over the Gulf of Lion and Balearic
806 Sea. *J. Geophys. Res.*, *117*, C09019, doi:10.1029/2012JC007924.
- 807 Ricchi, A., M. M. Miglietta, P. P. Falco, A. Benetazzo, D. Bonaldo, A. Bergamasco,
808 M. Sclavo, S. Carnier (2016), On the use of a coupled ocean-atmosphere-wave model
809 during an extreme cold air outbreak over the Adriatic Sea. *Atmos. Res.*, *172-173*, 48-65,
810 doi:10.1016/j.atomsres.2015.12.023.
- 811 Roulet G. and G. Madec (2000), Salt conservation, free surface and varying levels: A
812 new formulation for ocean general circulation models, *J. Geophys. Res.*, *105* (C10),
813 23927-23942, doi:10.1029/2000JC900089.
- 814 Schott, F., M. Visbeck, U. Send, J. Fischer, L. Stramma, Y. Desaubies (1996), Observations
815 of deep convection in the Gulf of Lion, Northern Mediterranean, during the winter of
816 1991/1992. *J. Phys. Oceanogr.*, *26*, 505-524.
- 817 Seity, Y., P. Brousseau, S. Malardel, G. Hello, P. Bénard, F. Bouttier, C. Lac, V. Masson
818 (2011), The AROME-France convective scale operational model, *Mon. Wea. Rev.*, *139*,
819 976-991
- 820 Small, R. J., T. Campbell, J. Teixeira, S. Carniel, T. A. Smith, J. Dykes, S. Chen, R.
821 Allard (2011), Air-Sea interaction in the Ligurian Sea: Assessment of a coupled ocean-
822 atmosphere model using in-situ data from LASIE07. *Mon. Wea. Rev.*, *139*, 1785-1808,
823 doi:10.1175/2010MWR3431.1.
- 824 Small, R. J., S. Carniel, T. Campbell, J. Teixeira, R. Allard (2012), The response of the
825 Ligurian and Tyrrhenian Seas to a summer mistral event: a coupled atmosphere-ocean
826 approach. *Ocean Modelling*, *48*, 30-44.

- 827 Somot, S., L. Houpert, F. Sevault, P. Testor, A. Bosse, I. Taupier-Letage, M-N. Bouin,
828 R. Waldman, C. Cassou, E. Sanchez-Gomez, X. Durrieu de Madron, F. Adloff, P.
829 Nabat, M. Herrmann (2016), Characterizing, modelling and understanding the climate
830 variability of deep water formation in the North-Western Mediterranean Sea. *Clim.*
831 *Dyn.*, doi:10.1007/s00382-016-3295-0.
- 832 Taillefer, F. (2002), "CANARI (Code for the Analysis Necessary for
833 ARPEGE, for its Rejects and its Initialization): Technical documenta-
834 tion". Technical report, CNRM/GMAP, Météo-France, Toulouse, France.
835 <http://cnrm.meteo.fr/gmapdoc/spip.php?article3>.
- 836 Testor, P., A. Bosse, L. Houpert, F. Margirier, L. Mortier, H. Legoff, D. Dausse, M.
837 Labaste, J. Karstensen, D. Hayes, A. Olita, E. Heslop, F. D'Ortenzio, N. Mayot, H. Lav-
838 igne, O. de Fommervault, L. Coppola, L. Prieur, V. Taillandier, X. Durrieu de Madron,
839 F. Bourrin, G. Many, P. Damien, C. Estournel, P. Marsaleix, I. Taupier-Letage, P.
840 Raimbault, R. Waldman, M-N. Bouin, H. Giordani, G. Caniaux, S. Somot, V. Ducrocq,
841 P. Conan (2017, rev), Multi-scale observations of deep convection in the northwest-
842 ern Mediterranean Sea during winter 2012-2013 from a multi-platform approach. *J.*
843 *Geophys. Res. Oceans*.
- 844 Valcke, S., T. Craig, L. Coquart (2013), OASIS3-MCT user guide. *Technical note*,
845 *TR/CMGC/13/17*, CERFACS, Toulouse, France.
- 846 Voldoire, A., B. Decharme, J. Pianezze, C. Lebeaupin Brossier, F. Sevault, L. Seyfried,
847 V. Garnier, S. Bielli, S. Valcke, A. Alias, M. Accensi, F. Ardhuin, M-N. Bouin, V.
848 Ducrocq, S. Faroux, H. Giordani, F. Léger, P. Marsaleix, R. Rainaud, J-L. Redelsperger,
849 E. Richard, S. Riette (2017, sub), The seamless and multi-model coupling between

850 atmosphere, land, hydrology, ocean, waves and sea-ice models based on SURFEX v8
851 surface model using OASIS3-MCT. *Geosci. Model Dev. Disc.*

852 Waldman, R., S. Somot, M. Herrmann, P. Testor, C. Estournel, F. Sevault, L. Prieur, L.
853 Mortier, L. Coppola, V. Taillandier, P. Conan, D. Dausse (2016), Estimating dense wa-
854 ter volume and its evolution for the year 2012-2013 in the Northwestern Mediterranean
855 Sea: An observing system simulation experiment approach. *J. Geophys. Res. Oceans*,
856 *121* (9), 6696-6716, doi:10.1002/2016JC011694.

857 Warner, J. C., B. Armstrong, R. He, J. B. Zambon (2010), Development of a Coupled
858 Ocean-Atmosphere-Wave-Sediment Transport (COAWST) Modeling System. *Ocean*
859 *Modelling*, *35*, 230-244.

Table 1. Description of the numerical experiments. IC stands for initial conditions, BC for boundary conditions, and SURF for surface conditions.

Name	Begin	End	IC.	BC	SURF
IMAP	1 September 2012	15 March 2013	MOOSE-ASICS + PSY2 <i>Summer 2012</i>	PSY2 (monthly)	forced mode AROME-WMED forecast (fluxes, 2.5km, 1h)
CPL	15 January 2013	15 March 2013	restart from IMAP <i>15 Jan. 2013</i>	PSY2 (monthly)	coupled with AROME-WMED (2.5km) 1h frequency

Table 2. Biases and standard deviations against Argo floats and CTD profiles for potential temperature (θ), salinity (S) and density (ρ).

Argo floats	θ ($^{\circ}\text{C}$)			S (psu)			ρ ($\text{kg}\cdot\text{m}^{-3}$)		
	0-150m	150-600m	600-2000m total	0-150m	150-600m	600-2000m total	0-150m	150-600m	600-2000m total
IMAP bias	-0.120	-0.003	0.013	-0.028	-0.13×10^{-3}	2.98×10^{-3}	3.18×10^{-3}	0.51×10^{-3}	-0.30×10^{-3}
std dev	0.268	0.124	0.042	0.110	0.043	0.010	0.101	0.030	0.005
CPL bias	-0.081	0.012	0.014	-0.033	2.90×10^{-3}	2.56×10^{-3}	-8.57×10^{-3}	-0.20×10^{-3}	-0.99×10^{-3}
std dev	0.237	0.132	0.047	0.112	0.041	0.005	0.100	0.027	0.005
CTD profiles	θ ($^{\circ}\text{C}$)			S (psu)			ρ ($\text{kg}\cdot\text{m}^{-3}$)		
	0-150m	150-600m	600m-bottom total	0-150m	150-600m	600m-bottom total	0-150m	150-600m	600m-bottom total
IMAP bias	-0.048	-0.059	0.017	-0.039	-0.026	-0.31×10^{-3}	-0.020	-7.88×10^{-3}	-3.78×10^{-3}
std dev	0.255	0.165	0.073	0.113	0.056	0.020	0.110	0.043	0.004
CPL bias	-0.041	-0.054	0.022	-0.041	-0.027	0.33×10^{-3}	-0.024	-9.56×10^{-3}	-4.20×10^{-3}
std dev	0.225	0.159	0.075	0.105	0.056	0.021	0.101	0.042	0.004

Table A. Schematic 2×2 contingency table for the definition of scores, given a threshold thr for the MLD.

	simulation $< thr$	simulation $\geq thr$
observation $< thr$	a	b
observation $\geq thr$	c	d

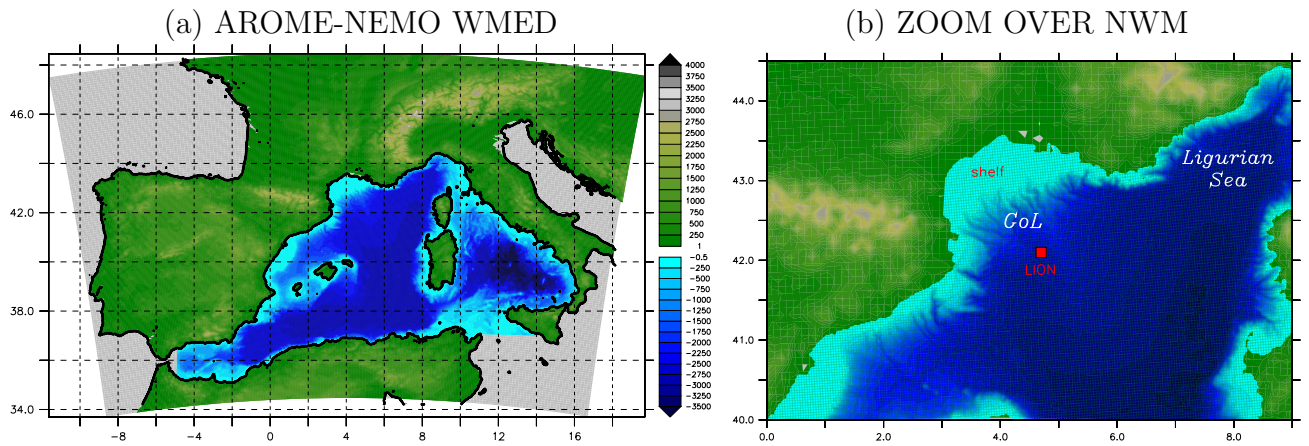


Figure 1. (a) AROME-NEMO WMED domain: AROME-WMED topography (in green) and NEMO-WMED36 bathymetry (in blue). The grey areas are the uncoupled marine zones. (b) Details of the north-western Mediterranean area. The red square indicates the Lion surface buoy and mooring line location.

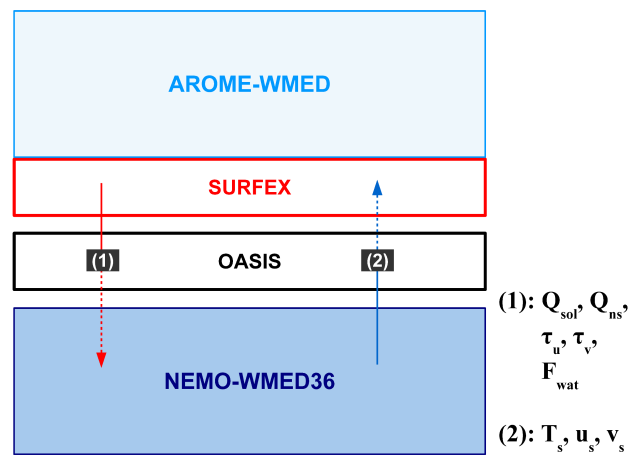


Figure 2. Description of the exchanges between the different components of the AROME-NEMO WMED coupled system.

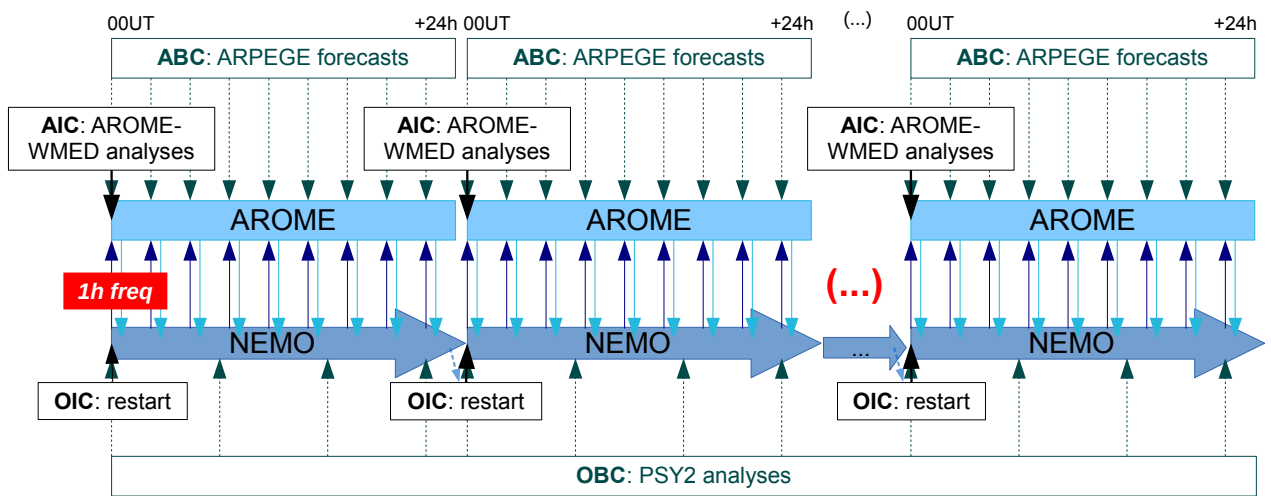


Figure 3. Numerical setup for the CPL experiment. ABC [OBC] stands for Atmospheric [Ocean] Boundary Conditions and AIC [OIC] for Atmospheric [Ocean] Initial Conditions.

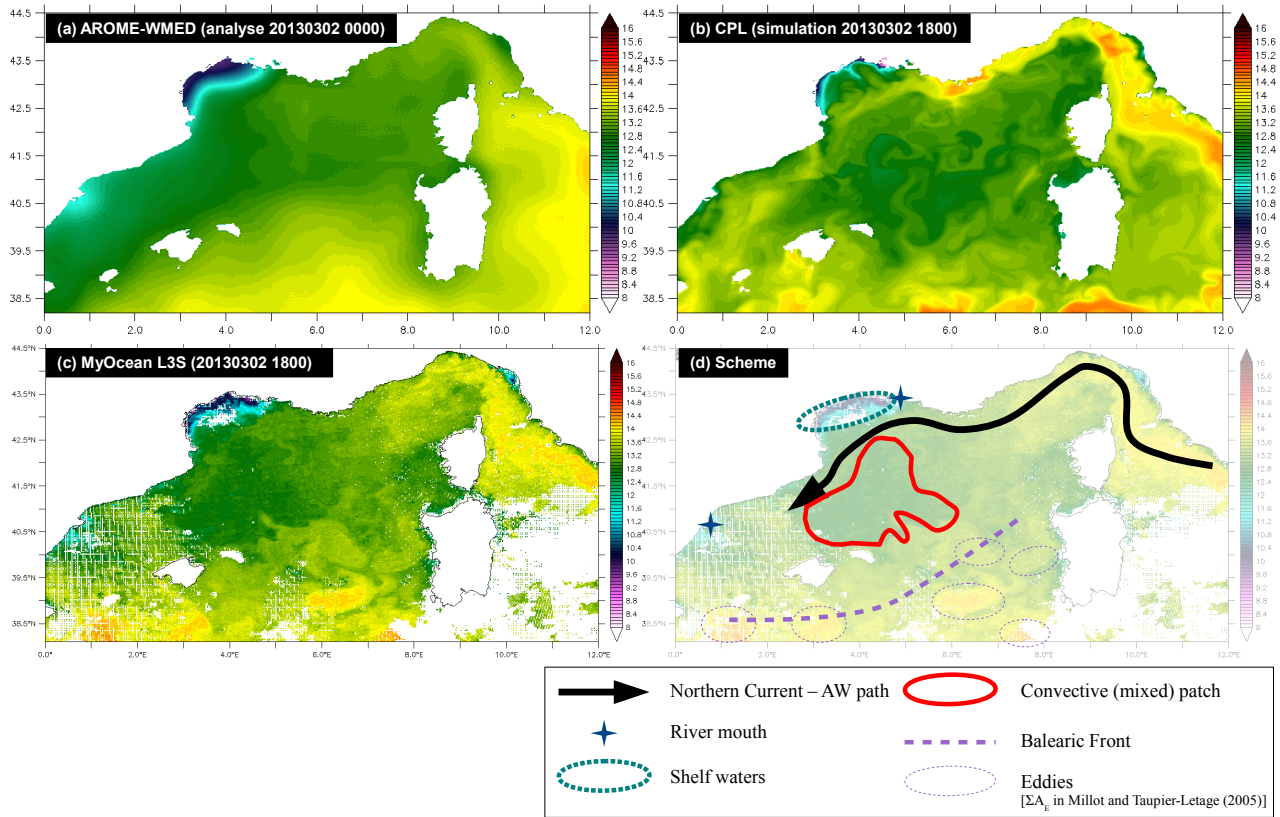


Figure 4. SST fields on 2 March 2013 18UT. (a) AROME-WMED forecast corresponding to the analysis at 00UT, (b) CPL simulation, (c) MyOcean L3S supercollated product (resolution: 0.01°; source: <http://hoc.sedoo.fr> - restricted access) and (d) schematic view of the SST patterns and related processes according to the L3S SST field in (c).

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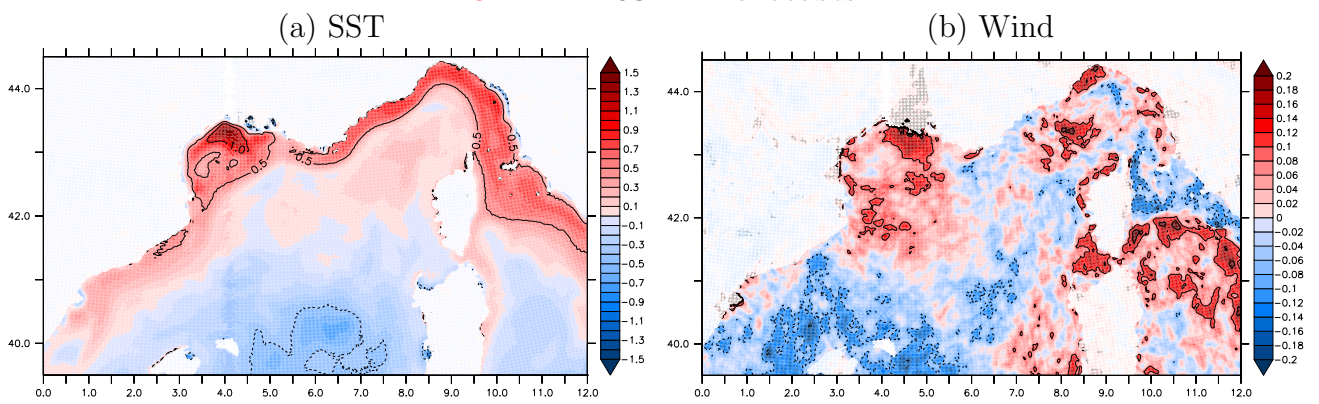


Figure 5. Mean differences during SOP2 in (a) SST (K, contours every 0.5 K) and (b) wind speed (m.s^{-1} , contours every 0.1 m.s^{-1}) at the first atmospheric level ($\sim 10 \text{ m}$), between CPL and the AROME-WMED operational forecasts.

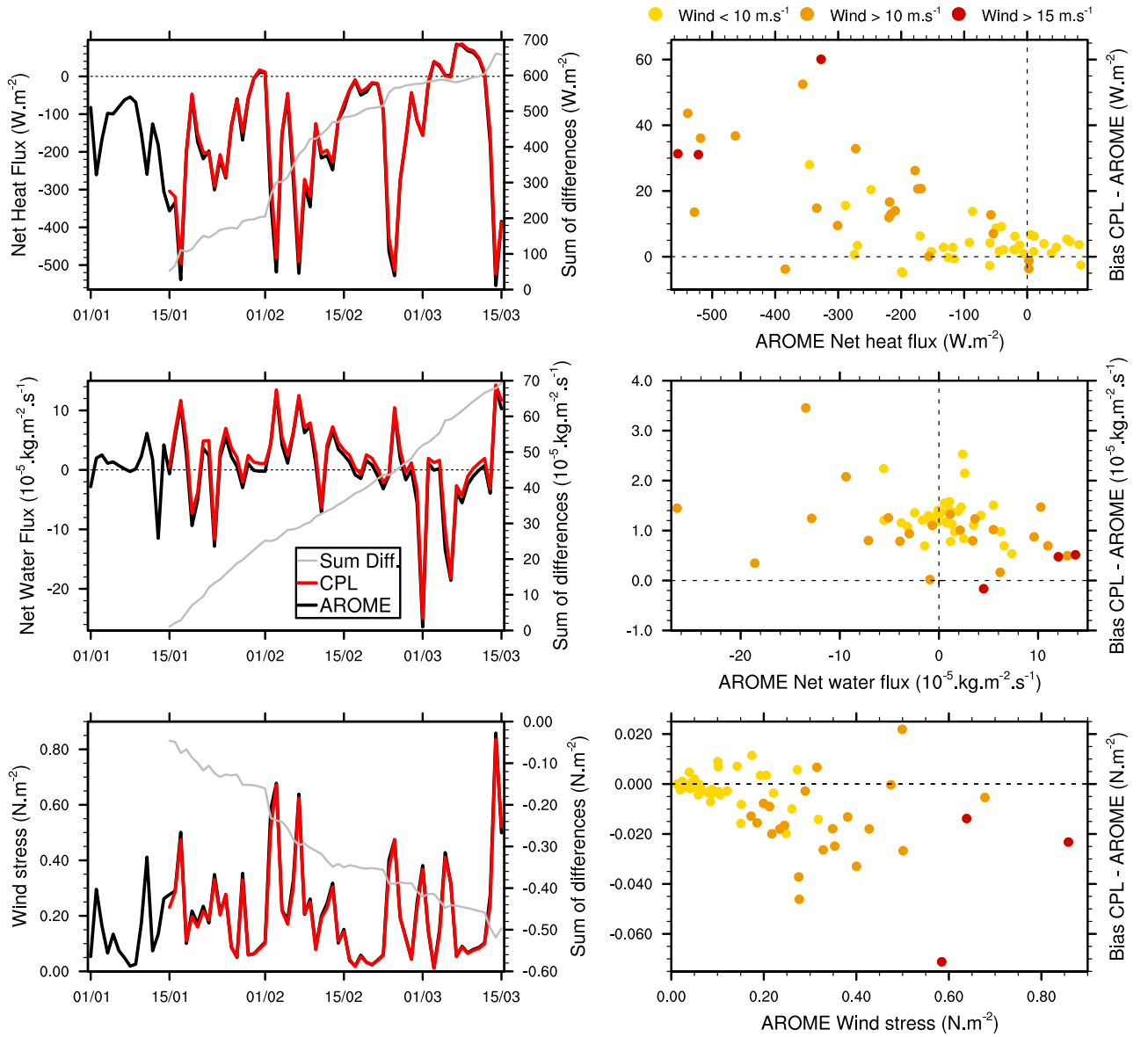


Figure 6. *Left panels:* Daily time-series of the net heat flux (W.m^{-2}), of the freshwater flux ($\text{kg.m}^{-2}.\text{s}^{-1}$) and of the momentum flux (N.m^{-2}) intensity over the north-western Mediterranean Sea in AROME-WMED forecasts (used to compute the surface forcing for IMAP) and in CPL. The grey lines are the sums since 15 January of the differences between CPL and AROME-WMED (scales on the right). *Right panels:* Daily differences in the net heat flux, the freshwater flux and the momentum flux between CPL and AROME-WMED as a function of the daily flux values in AROME-WMED. The color indicates the range of the corresponding daily mean wind speed in AROME-WMED forecasts over the north-western Mediterranean Sea.

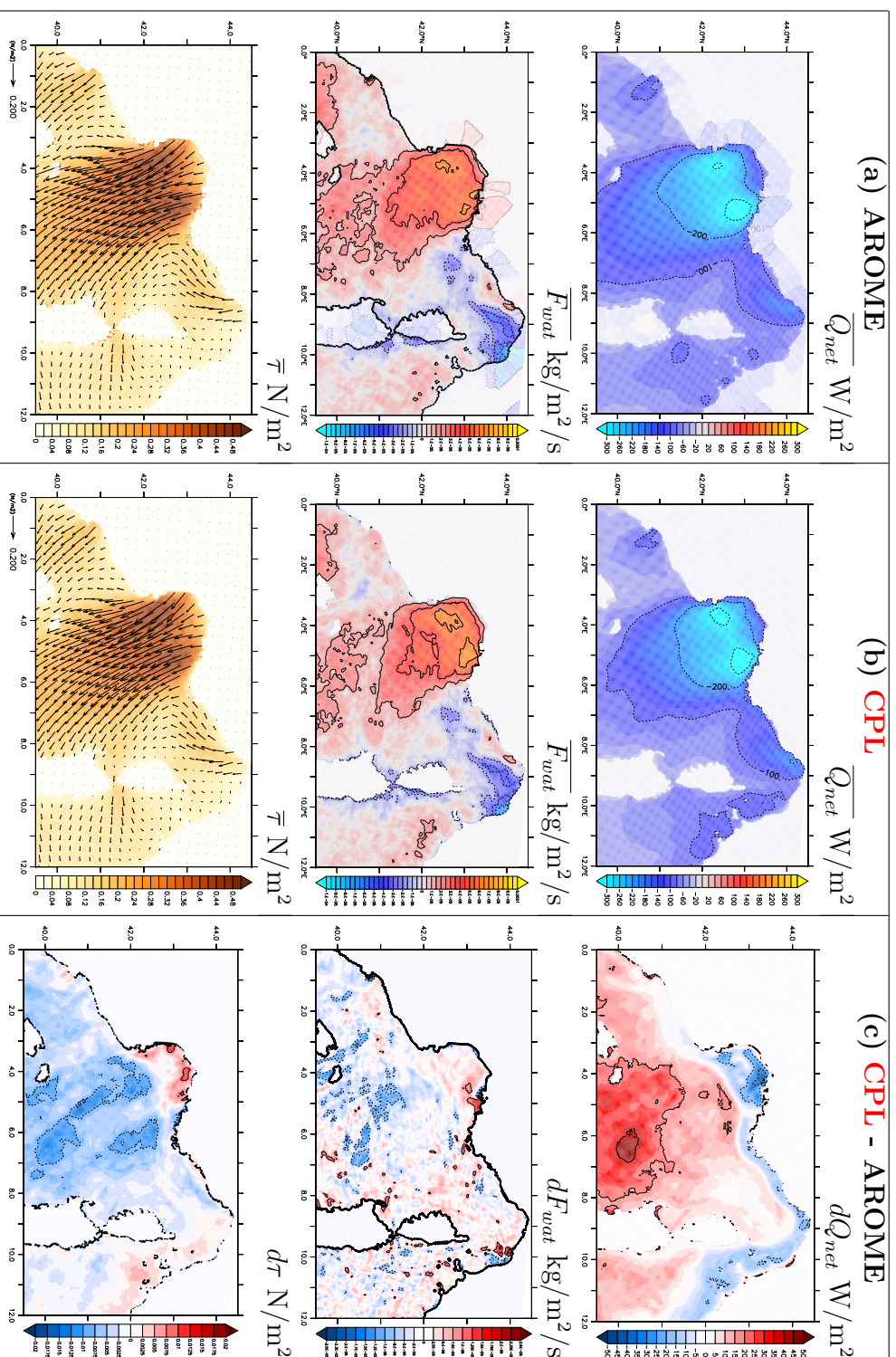


Figure 7. Mean net heat flux ($W \cdot m^{-2}$), freshwater flux ($kg \cdot m^{-2} \cdot s^{-1}$) and momentum flux ($N \cdot m^{-2}$) during SOP2 in (a) AROME-WMED forecasts and (b) in CPL, and (c) mean differences between CPL and IMAP.

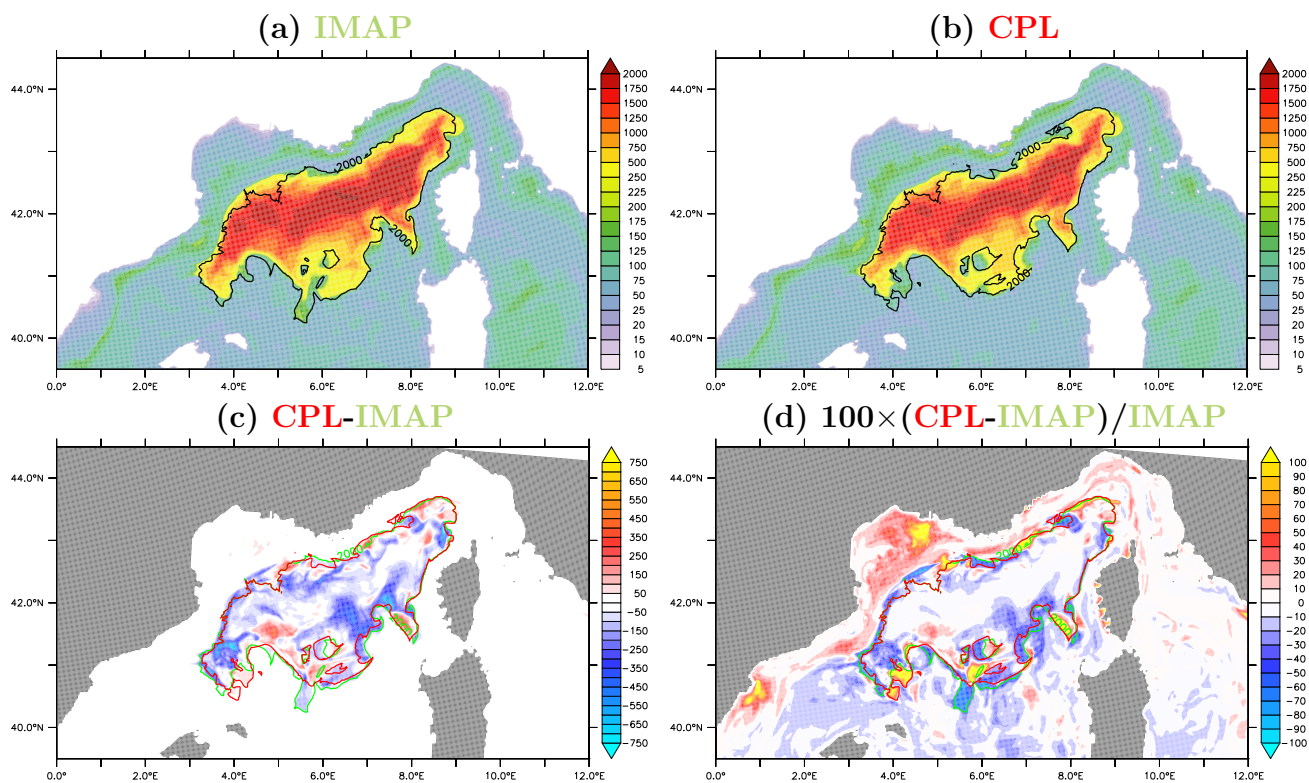


Figure 8. Mean Mixed Layer Depth (colors, meters) from a density criteria in (a) IMAP and (b) CPL. (c) Absolute (in meters) and (d) relative (in %) differences in the mean MLD between CPL and IMAP. The contours indicate the area where the maximum MLD simulated during SOP2 is larger than 2000m-depth (green for IMAP and red for CPL in c and d).

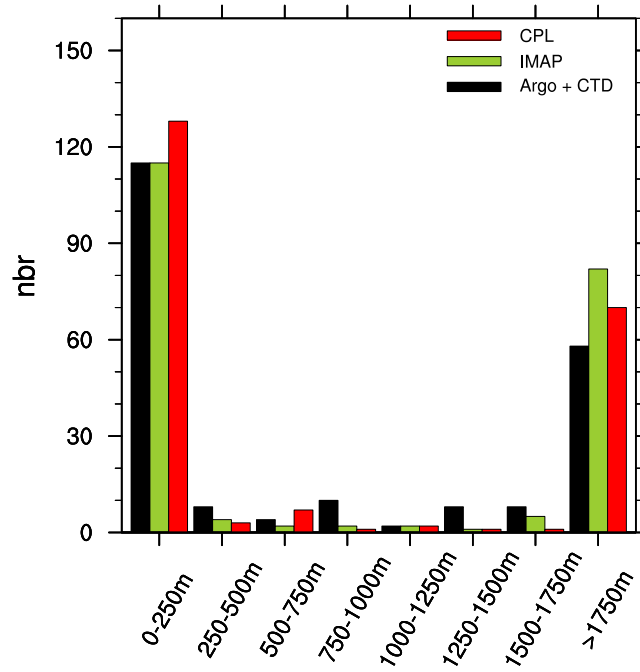


Figure 9. Distribution (number of profiles) of the MLD (meters) from density in-situ profiles (floats [ARGO type] and R/V *Le Suroit* CTDs) during SOP2 in the north-western Mediterranean and spatio-temporally colocalized in the two simulations IMAP and CPL.

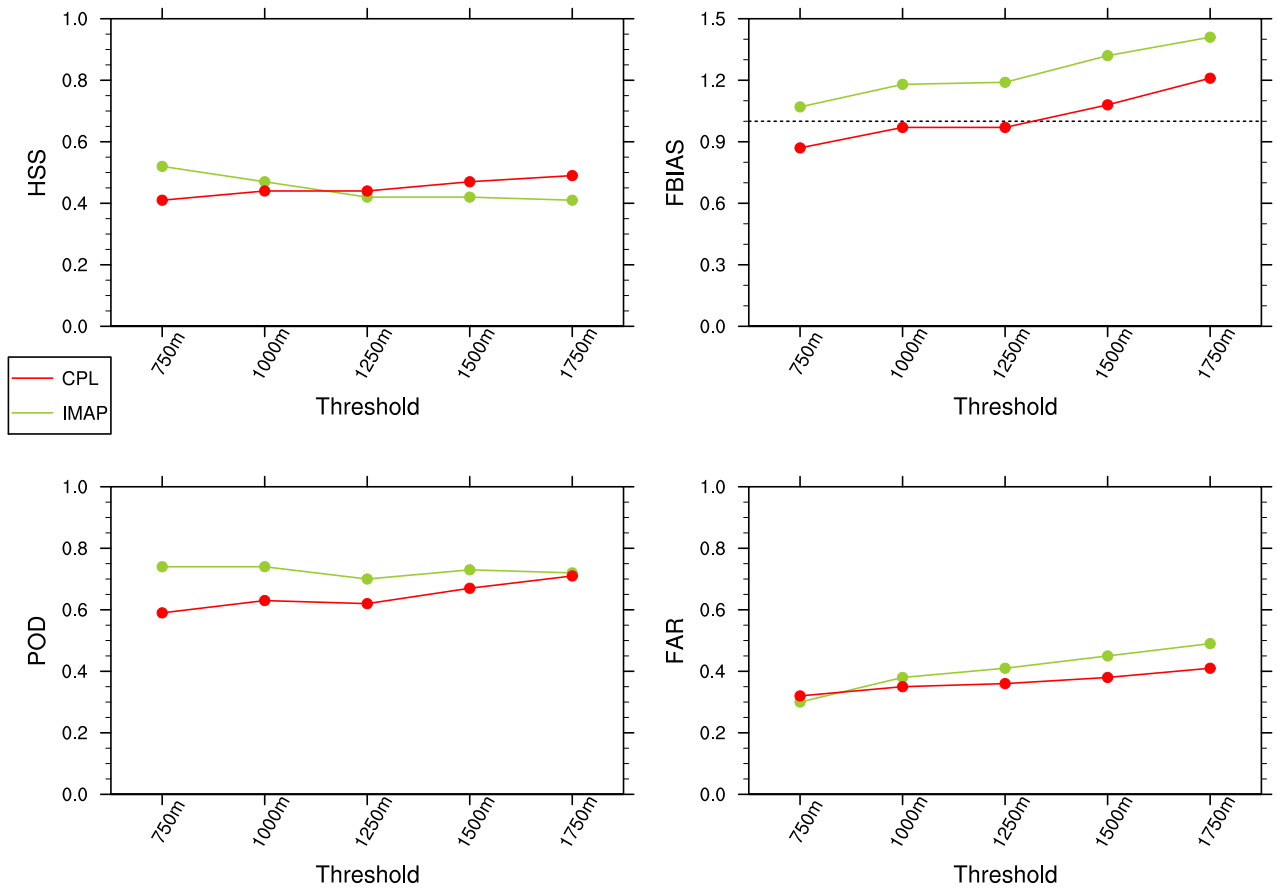


Figure 10. Skill scores (HSS, FBIAS, POD and FAR, see Annexe B) for IMAP and CPL obtained when compared to observed MLD in density in-situ profiles (floats [ARGO type] and R/V *Le Suroit* CTDs) and considering various MLD thresholds.

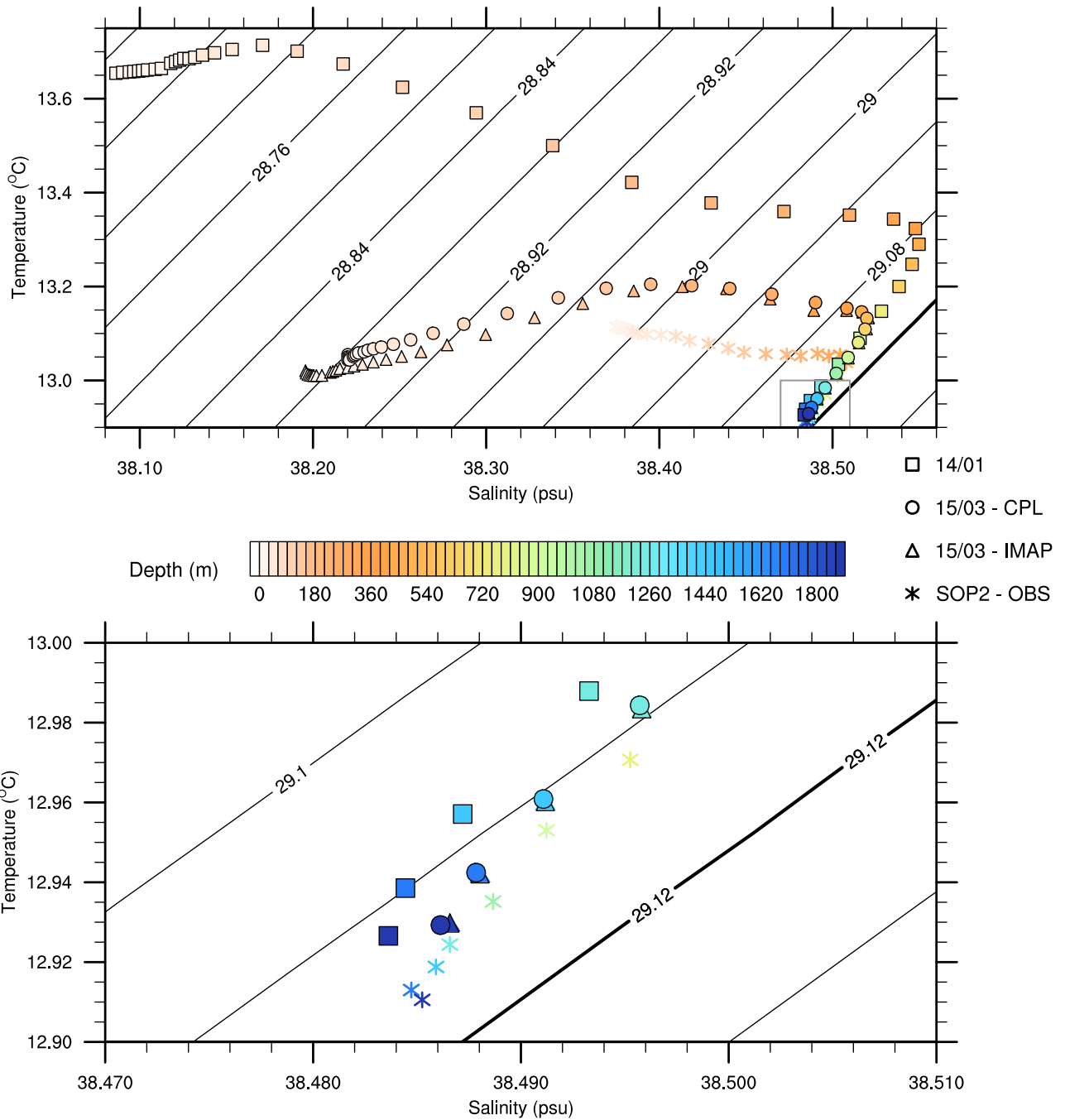


Figure 11. *Top panel:* θ/S diagram averaged in the north-western Mediterranean area before the convection (14 January 2013, squares) and at the end of SOP2 (15 March 2013) for the two experiments (triangles for IMAP and circles for CPL). *Bottom panel:* zoom for the WMDW (dashed rectangle in the top panel). The mean θ/S diagram from in-situ floats [ARGO type] averaged over SOP2 is indicated with stars.

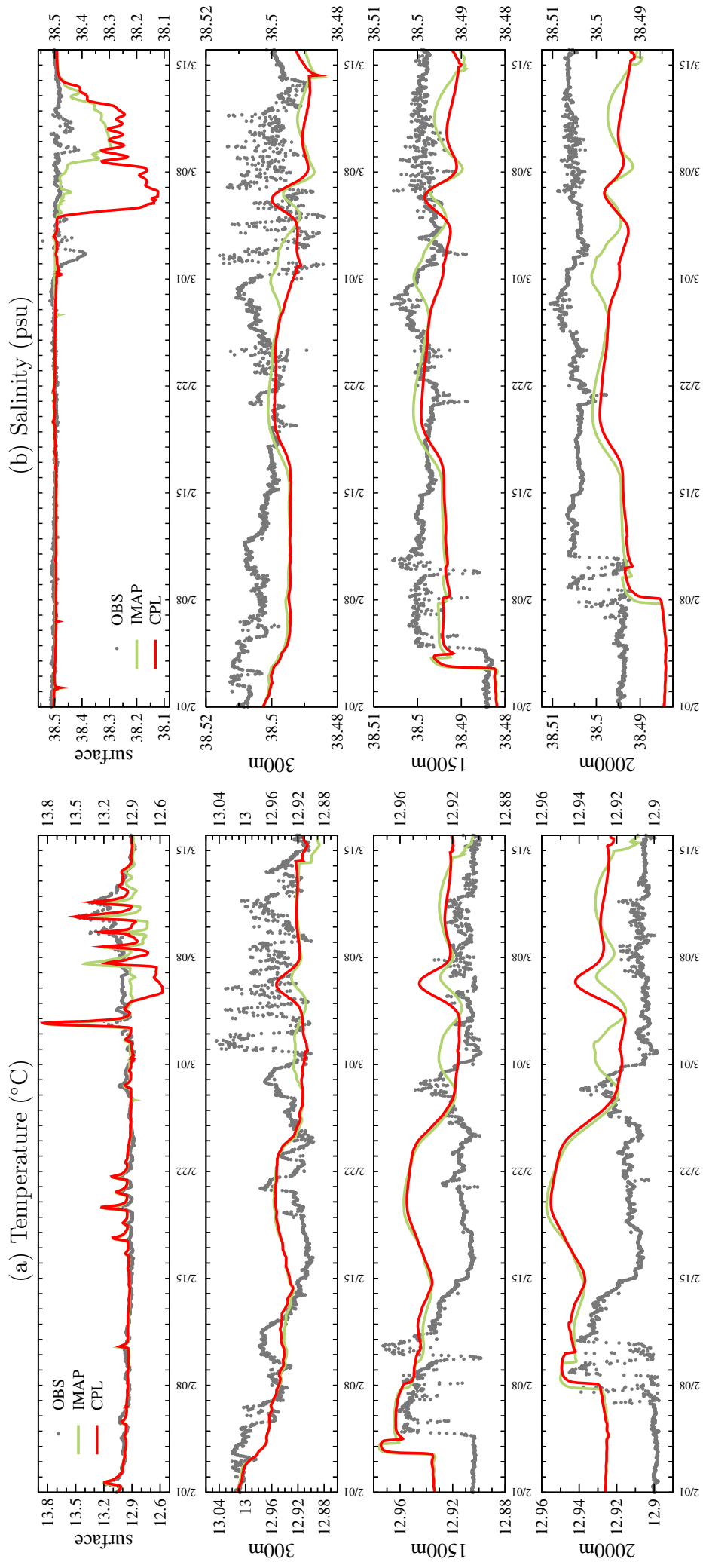


Figure 12. Lion buoy and mooring [4.7°E-42.1°N]: Simulated and observed Temperature (°C) and Salinity (psu) in surface, at 300 m-depth, at 1500 m-depth and at 2000 m-depth.

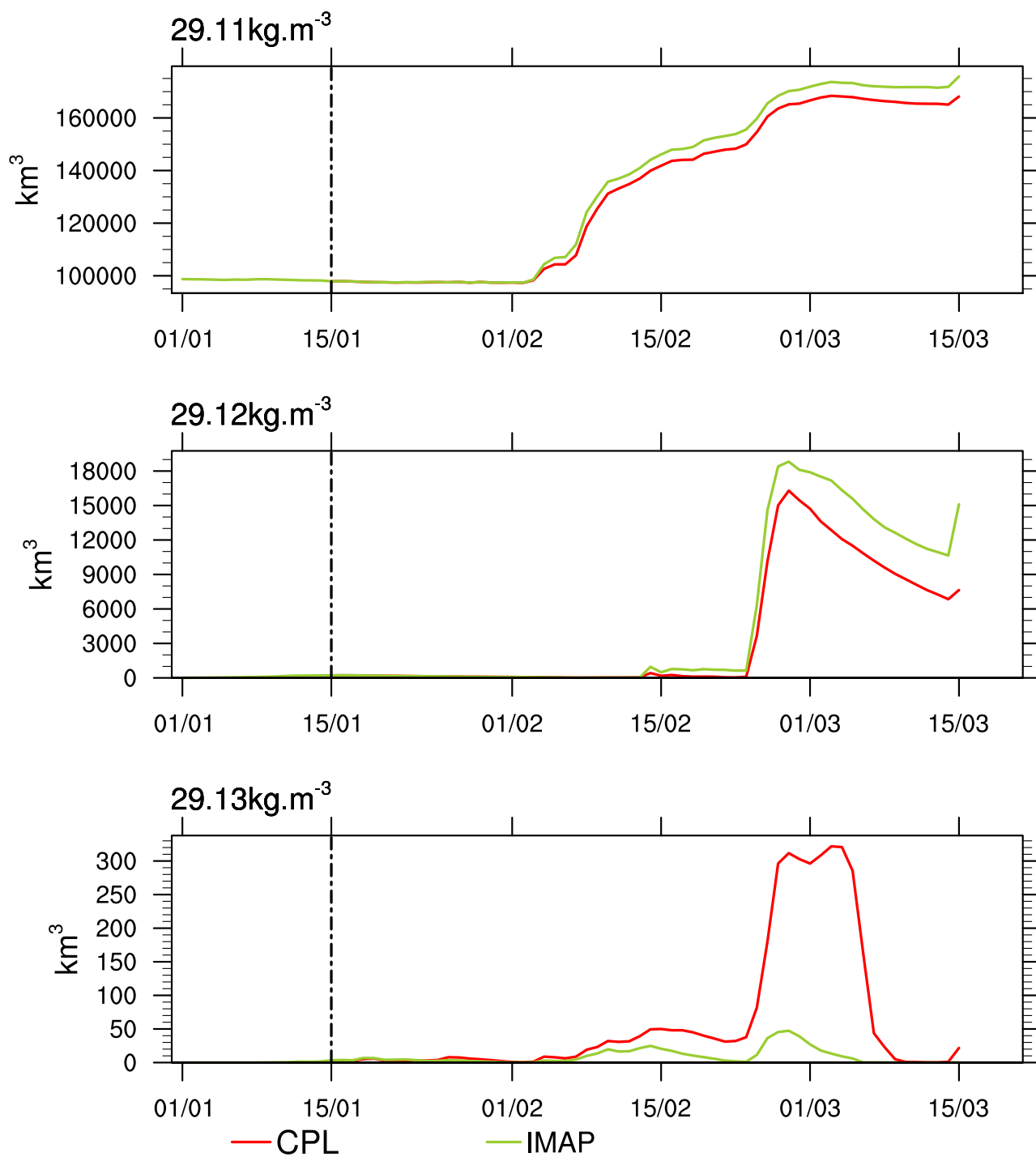


Figure 13. Time-series of the dense water volumes (km³): water denser than (a) 29.11 kg.m⁻³, (b) 29.12 kg.m⁻³ and (c) 29.13 kg.m⁻³.

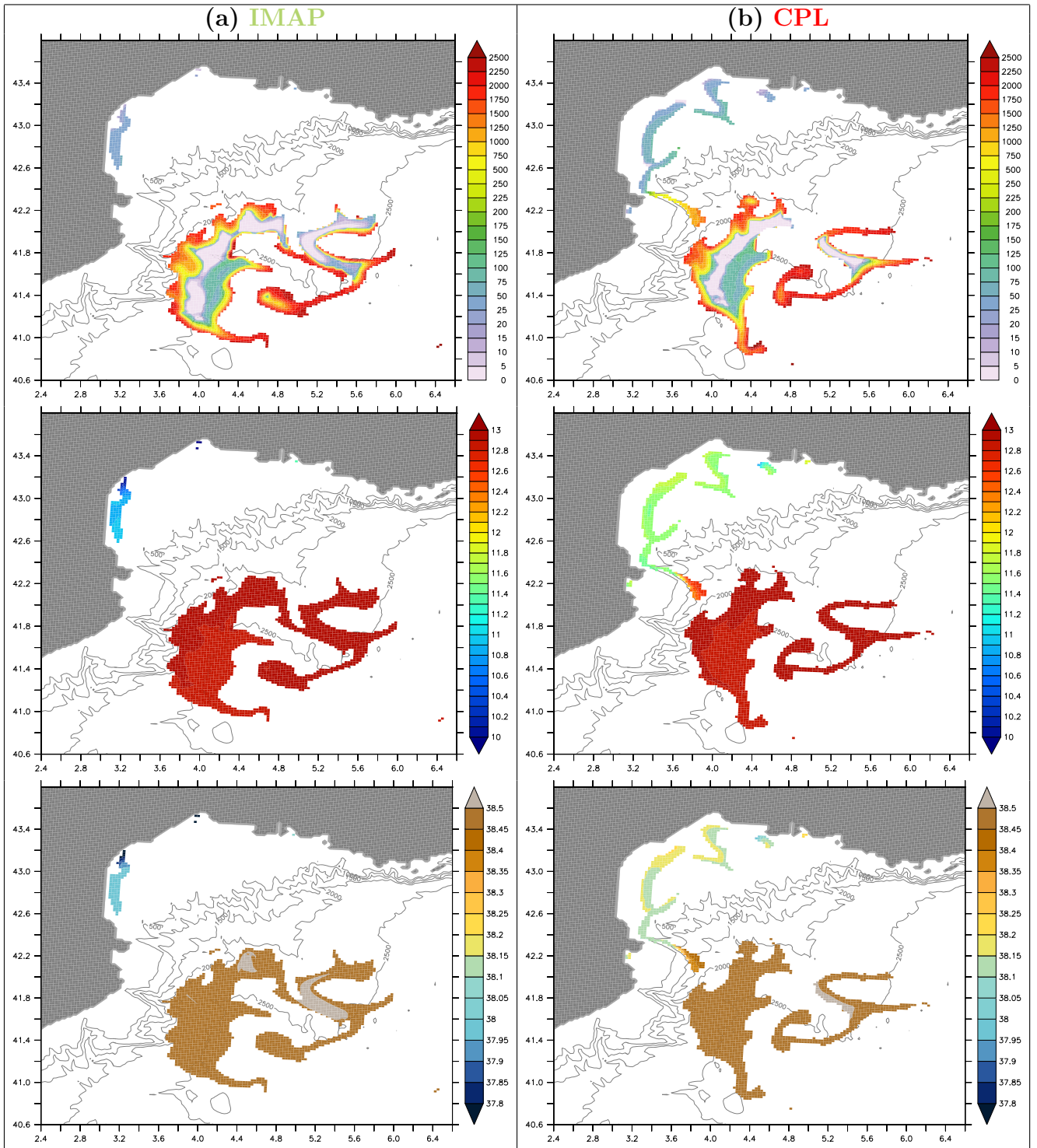


Figure 14. 29.12 kg.m⁻³ isopycnal surface depth (meters, top panels), temperature (°C, middle panels) and salinity (psu, bottom panels) simulated on 2 March 2013 12UT by (a) IMAP and (b) CPL.

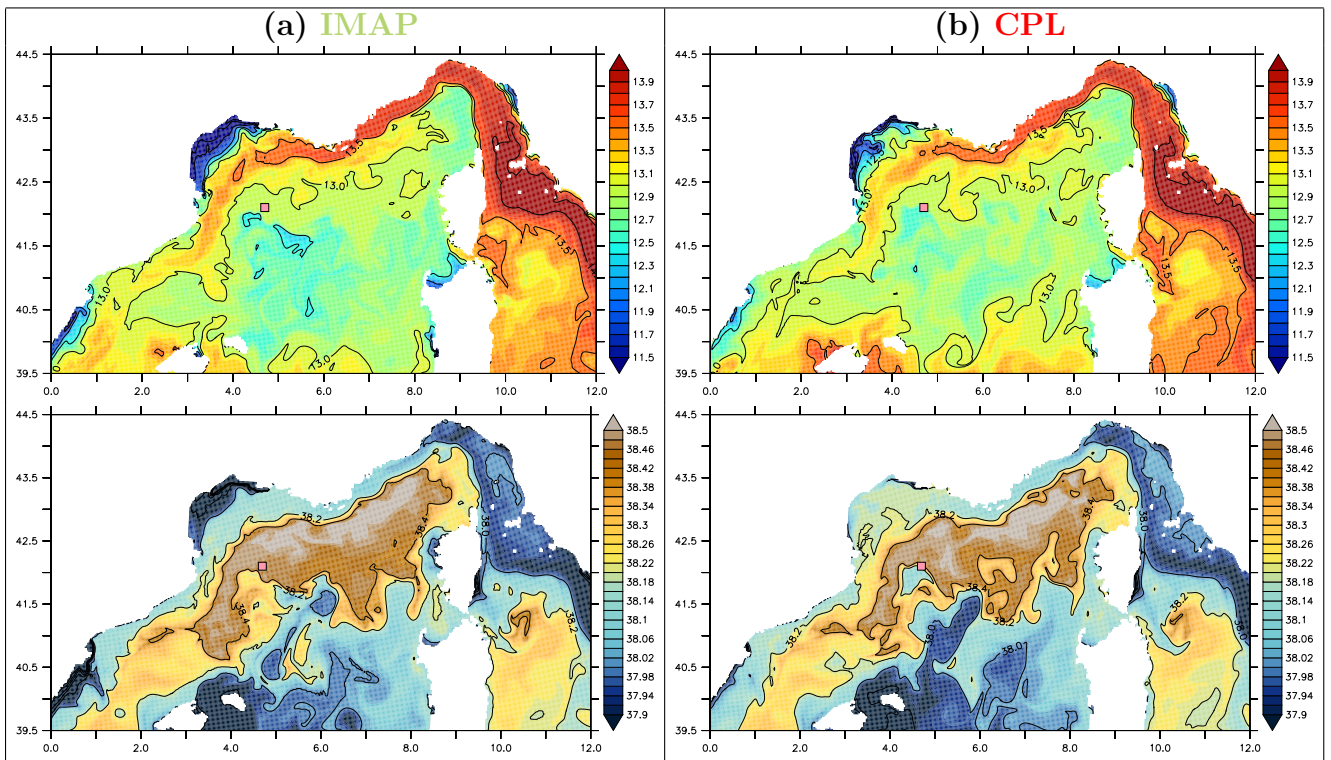


Figure 15. Sea Surface Temperature ($^{\circ}\text{C}$, top panels) and Salinity (psu, bottom panels) simulated on 6 March 2013 00UT in (a) IMAP and (b) CPL. The pink square indicates the Lion buoy location.

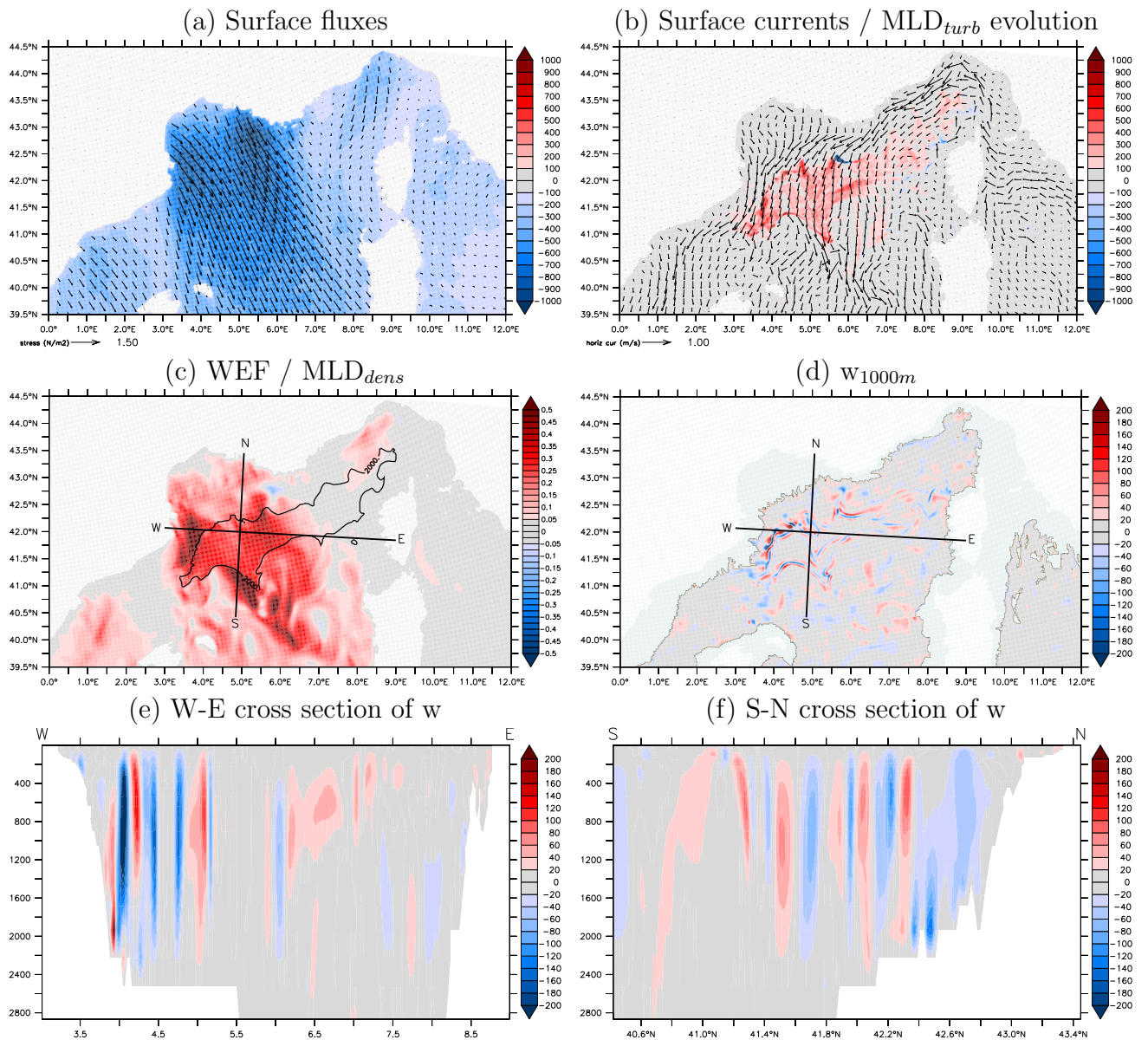


Figure 16. 7 February 2013 (IOP21c) in CPL: (a) Daily-mean net heat flux (colors, $\text{W}\cdot\text{m}^{-2}$) and wind stress (arrows, $\text{N}\cdot\text{m}^{-2}$). (b) Surface current (arrows, $\text{m}\cdot\text{s}^{-1}$) and daily evolution of the MLD from a turbulence criteria (colors, in meters [per day]). (c) Daily-mean WEF (colors, $\text{N}\cdot\text{m}^{-1}\cdot\text{s}^{-2}$). The black contour indicates where the daily-maximum MLD from a density criteria reaches 2000 m. (d) Daily-mean vertical velocity (w , in meters per day) at 1000m-depth. (e,f) Vertical cross sections (thick solid black lines in c,d) of the daily-mean vertical velocity (in meters per day).

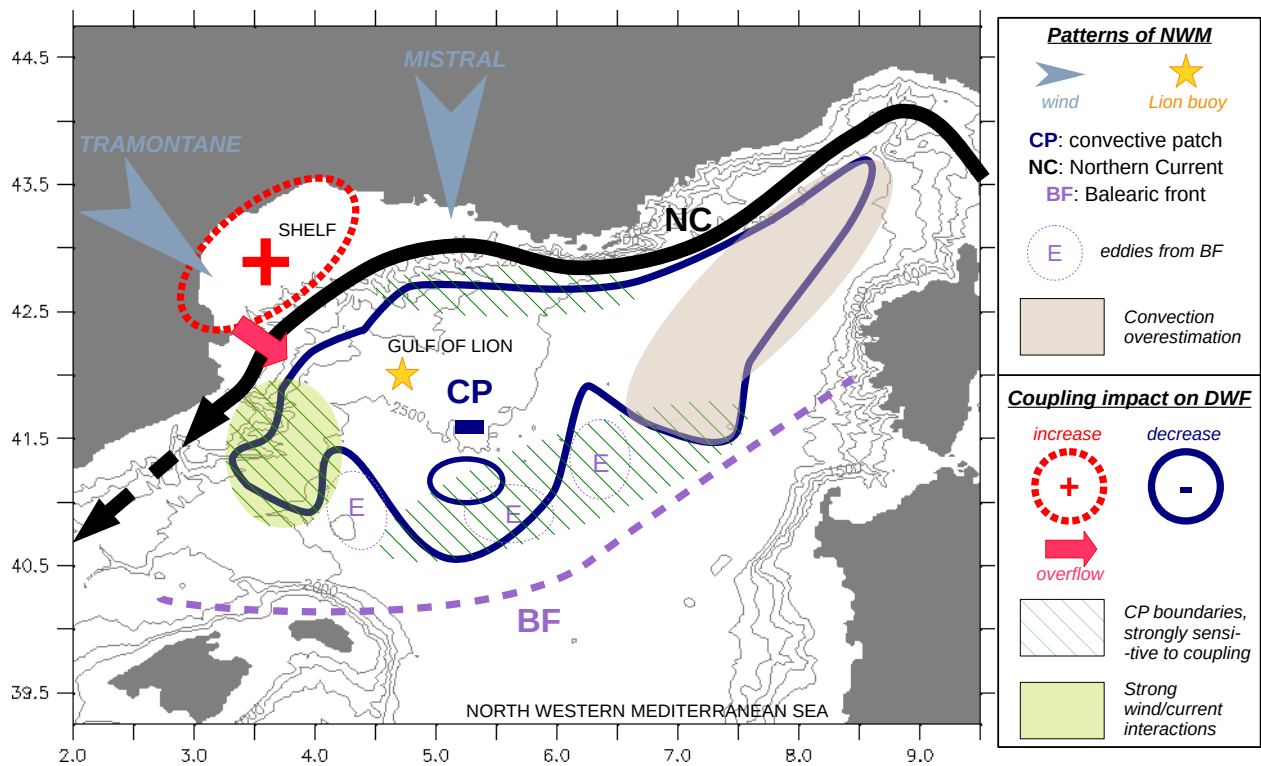


Figure 17. Schematic summary of the ocean-atmosphere coupling impacts on DWF in the North-Western Mediterranean Sea during HyMeX SOP2, deduced from AROME-NEMO WMED simulations.