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Influence of the sea state on Mediterranean heavy precipitation: a case-study from HyMeX SOP1

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Abstract :

Sea state can influence the turbulent air–sea exchanges, especially the momentum flux, by modifying the sea-surface roughness. The high-resolution non-hydrostatic convection-permitting model MESO-NH is used here to investigate the impact of a more realistic representation of the waves on heavy precipitation during the Intense Observation Period (IOP) 16a of the first HyMeX Special Observation Period (SOP1). Several quasi-stationary mesoscale convective systems developed over the western Mediterranean region, two of them over the sea, and resulted in heavy precipitation on the French and Italian coasts on 26 October 2012. Three different bulk parametrizations are tested in this study: a reference case (NOWAV) without any wave effect, a parametrization taking into account theoretical wave effects (WAV) and a last one with realistic wave characteristics from the MFWAM analyses (WAM). Using a realistic wave representation in WAM significantly increases the roughness length and the friction velocity with respect to NOWAV and WAV. The three MESO-NH sensitivity experiments of the IOP16a show that this surface-roughness increase in WAM generates higher momentum fluxes and directly impacts the low-level dynamics of the atmosphere, with a slowdown of the 10 m wind, when and where the wind speed exceeds 10 m s⁻¹ and the sea state differs from the idealized one. The turbulent heat fluxes are not significantly influenced by the waves, these fluxes being controlled by the moisture content rather than by the wind speed in the simulations. Although the convective activity is globally well reproduced by all the simulations, the difference in the low-level dynamics of the atmosphere influences the localization of the simulated heavy precipitation. Objective evaluation of the daily rainfall amount and of the 10 m wind speed against the observations confirms the positive impact of the realistic wave representation on this simulation of heavy precipitation.

Keywords : air–sea exchanges, Mediterranean Sea, HyMeX, MESO-NH, sea state, roughness length, turbulent fluxes, heavy precipitation

1. Introduction

Regularly during the autumn, Heavy Precipitating Events (HPEs) occur over the western Mediterranean basin and more particularly the mountainous coastal regions of Spain, France and Italy. These events generate high rainfall amount in a very short time on localized areas, often leading to flash flood events with dramatic consequences on goods and people (Llasat *et al.* 2013). Numerical Weather Prediction (NWP) of HPEs is still challenging. Because they represent an important source of societal damages, a better understanding of the underlying mechanisms as well as an improvement of their representation by NWP models is therefore a key step towards mitigating their impact. This is one of the objectives of the 10-year programme called HyMeX (Hydrological cycle in the Mediterranean Experiment, Drobinski *et al.* 2014) launched in 2010 which aims at improving our understanding of the Mediterranean water cycle with a specific attention on intense weather events. In particular, a large field campaign dedicated to heavy precipitation and flash flooding (called Special Observation Period - SOP1) took place during autumn 2012 in the western Mediterranean region (Ducrocq *et al.* 2014).

Two different meteorological situations can generate HPEs in the Mediterranean basin. On one hand large rainfall amount accumulate for several days at the same location with the slowdown of a frontal disturbance. On the other hand, heavy precipitation can be observed within a few hours over a small area where a MCS (Mesoscale Convective System) remains quasi-stationary (Nuissier *et al.* 2008). The second case is more favourable to flash flood events (Ducrocq *et al.* 2003, 2004). A combination of conducive factors is necessary to the generation of a quasi-stationary MCS at the origin of HPEs. First, a slow-evolving synoptic situation induces marine low-level jets advecting warm and moist air from the Mediterranean Sea to the coasts (Homar *et al.* 1999; Nuissier *et al.* 2011; Ricard *et al.* 2012). Conditional instability is then released if the low-level flow is forced to lift when encountering the coastal mountains. Triggering of deep convection can also occur upwind

the mountains due to low-level convergence over the sea or due to a cold pool beneath the convective systems (Ducrocq *et al.* 2008).

Several studies examined the influence of the Mediterranean Sea in the generation of HPEs. Duffourg and Ducrocq (2011) who focused on the origin of the moisture feeding the precipitating systems showed that the evaporation over the Mediterranean Sea represents a major source of humidity (between 40 and 60%) transported by the low-level jet towards the MCS. The other sources of humidity come from the Atlantic Ocean and Africa. Strong exchanges of moisture and heat between the ocean and the atmosphere are specifically achieved through the latent heat flux. Lebeaupin *et al.* (2006) witnessed also a strong influence of the SST (Sea Surface Temperature) variations on the atmospheric low-level dynamics for convective-scale numerical simulations of three HPEs, with an increase of latent and sensible heat fluxes for warmer SST. This directly produces an increase of the convective activity during the event with consequently more rainfall amount. The influence of the bulk parametrization of the turbulent fluxes used within the model on HPE simulations has also been shown by Lebeaupin-Brossier *et al.* (2008) with a comparison between the formulation of Louis (1979) and the COARE (Coupled Ocean-Atmosphere Response Experiment) bulk algorithm from Fairall *et al.* (2003). Strong differences were obtained between both parametrizations, especially regarding the momentum and latent heat fluxes with lower values of both fluxes in strong-wind conditions by the COARE parametrization. These reduced air-sea exchanges led to a decrease in the moisture feeding of the convective system with lower simulated rainfall amount.

Past studies based on the analysis of *in situ* data have highlighted the wave influence on sea-surface exchanges through the dependence of the roughness length to the wave age (Smith *et al.* 1992; Drennan *et al.* 2003). This relationship has a direct impact on the wind stress and therefore on the near surface winds and on the low-level dynamics. A dependence between the roughness length and the wave age is included by several bulk parametrizations, among which the COARE 3.0 parametrization (Fairall *et al.* 2003). For operational medium-range forecasts, the European Center for Medium-range Weather Forecasts (ECMWF) has been running since 1998 a coupled system between the atmospheric and the wave modelling parts using the wind input

term of the wave model to estimate the Charnock parameter which, in turn, determines the surface roughness (Janssen 2004). Oppositely, the approach used in the present study uses the wave parameters from the wave model as an input of the COARE 3.0 parametrization of turbulent fluxes in the atmospheric model.

This study focuses on the influence of the waves in the simulation of HPEs and more particularly in the intensity and the localization of the precipitation. It investigates the well-documented HPEs which occurred during HyMeX SOP1 on 26 October 2012. During this event, most of the MCSs affecting the area initiated and developed at sea. This case is thus well adapted for studying the impact of the waves on the low-level jet feeding the MCS and on the precipitation forecast through the air-sea fluxes parametrization. The meteorological environment and sea state encountered during this HPE are presented in details in section 2. Then, a description of how the waves are taken into account by the bulk algorithm COARE 3.0 (Fairall *et al.* 2003) as well as an evaluation of the sensitivity of the COARE parametrization to these wave representations are given in section 3. The numerical experiments using the convection-permitting MESO-NH model (Lafore *et al.* 1997) are described in section 4 and their results are discussed in section 5 before concluding remarks in section 6.

2. Case study: IOP16a

2.1. Synoptic situation

This case study focuses on the HPEs that occurred on 26 October 2012 over the northwestern Mediterranean region, which corresponds to the Intense Observation Period (IOP) 16a of the HyMeX SOP1 (Ducrocq *et al.* 2014).

The upper-level synoptic meteorological situation is shown in Figure 1. It is characterized by a cut-off low centered over Portugal on 25 October, associated with a southwesterly and diffluent upper-level flow over the northwestern Mediterranean where deep convection triggered. The pressure low progressed eastward while deepening and evolved in a thalweg extended from southeastern France to Morocco on 27 October, 00 UTC.

The AROME-WMED analysis (Fourrié *et al.* 2015) at 2.5-km horizontal resolution provides a description of the low-level atmospheric circulation over the western Mediterranean

(Fig. 2). On 25 October at 12 UTC (Fig. 2a), the low-level circulation over southern Spain is associated with a low pressure off Portugal. Low-level flow over sea is weak, except over the western side with southwesterly to southerly flow facing the Spanish coastal mountains. As the low-level pressure decreases over the northwestern Mediterranean on 26 October, the low-level southerly flow associated with moist and warm air reinforces over the western Mediterranean. A convergence line develops on the morning of 26 October between the southerly flow and southwesterly colder winds (Fig. 2b). A southerly moist and warm flow over the Tyrrhenian Sea from Tunisia to Gulf of Genoa establishes during the morning of 26 October (Fig. 2b and 2c).

2.2. Chronology of the convective systems

Deep convection triggers in several places during the night from 25 to 26 October and the following day as evidenced by the infrared temperature (Fig. 3). A first MCS (called hereafter MCS0) forms over the sea between the Spanish coast and the Balearic Islands around 22 UTC on 25 October. This quasi-stationary V-shape MCS begins to decay around 04 UTC on 26 October. Northward, new convective cells triggered near 05 UTC forming a MCS (called MCS1) over the Gulf of Lion. MCS1 splits in two MCSs (MCS1a and MCS1b). MCS1a progresses northward, with convective rainfall reaching the southwest French coast around 10 UTC. This MCS progressively decays after reaching the coast, with however orographic precipitation remaining till late afternoon over the Cévennes. In the same time, MCS1b maintains and strengthens over the Mediterranean Sea while moving northeastward to the French coasts on the morning of 26 October. The mature system remains quasi-stationary over and offshore the southeast French coasts until 17 UTC. High hourly surface rainfall totals up to 50 mm are observed by the rain gauges over land. This MCS evacuates eastward and decreases after 17 UTC. Local flash flooding made two casualties in Toulon (southeast France). Meanwhile, a fourth quasi-stationary MCS (MCS2) develops on the Italian coast. It initiates near 06 UTC on 26 October and remains quasi-stationary all the morning (Fig. 3c). Both MCS1b and MCS2 lead to heavy precipitation. It must be noticed that as a large part of the MCS development occurs over the sea, larger precipitation amounts may occur over the sea as

156 well. This is however not possible to confirm due to the lack of
157 direct measurements at sea.

158 2.3. Evolution of the sea state

159 The mean sea state can be described using two main
160 characteristics of the waves: the significant wave height is the
161 average height (trough to crest) of the highest one third of the
162 waves, and the peak period of the waves is the period at which
163 the waves reach their maximum of energy, given by the wave
164 energy spectrum. During the second half of 25 October 2012,
165 the sea state of the northwestern part of the Mediterranean is
166 characterized by a smooth surface with significant wave height
167 inferior to 0.5 m and a peak period inferior or equal to 5 s,
168 except south of the Balearic Islands where the significant wave
169 height ranged between 1 and 1.3 m with an associated peak
170 period between 5 and 6 s. On 26 October 2012, the sea state
171 is globally rougher with a significant wave height around 1.5 m
172 (Fig. 4c, d) except locally where it is superior to 2.5 m (over
173 the Gulf of Lion and west of Sardinia). The peak period ranges
174 between 6 and 7 s the whole day (Fig. 4a, b). These values are
175 issued from the 3-hourly, 10-km resolution analysis of the regional
176 wave forecasting model *MFWAM* and witness a typical mixed
177 wind sea. *MFWAM* is a third generation ocean wave prediction
178 model (The WAMDI Group 1988) used operationally by Météo-
179 France and forced every 6 hours by the 10-m wind of the global
180 *ARPEGE* forecasting model of Météo-France at 10-km resolution.

181 The regional *MFWAM* analyses used in this study do not use any
182 data assimilation, but are forced as a boundary condition by the
183 *MFWAM* global model which assimilated, at the time period of
184 the experiment, satellite altimetry data from Jason-1 and Jason-2.

185 Two moored buoys are deployed in the northwestern
186 Mediterranean Sea, one in the Gulf of Lion (Lion buoy, 42.06°N
187 4.64°E) and one off the southeast French coast (Azur buoy,
188 43.38°N 7.83°E) recording hourly atmospheric and oceanic
189 parameters (see locations in Fig. 4). The significant wave height
190 and peak period recorded by these buoys are not assimilated in the
191 *MFWAM* analyses and can thus be used to independently assess
192 the quality of the *MFWAM* products at these two locations. The
193 comparison on the IOP16a time period (from the 25 October at
194 12 UTC to the 27 October at 00 UTC, Fig. 5a,b) highlights an

195 underestimation of the significant wave height by the *MFWAM*
196 analysis (negative bias of -0.18 m for Azur and -0.42 m for
197 Lion) but no significant bias for the peak period. The temporal
198 variations of both parameters are well reproduced by the wave
199 model as witnessed by the correlation coefficient superior to 85%
200 for each parameter. Ultimately, the scatter index gives a relative
201 uncertainty between 35 and 50% for the wave height and 15% for
202 the peak period so that this latter parameter is better reproduced
203 by the model.

3. Influence of the sea state on the turbulent fluxes

3.1. Parametrization of the turbulent fluxes at the air-sea interface

204 Different parametrizations can be used to determine the sea-
205 surface turbulent fluxes (*i.e.* momentum τ , sensible H_s and latent
206 H_l heat fluxes). The *COARE* parametrization is a commonly-used
207 bulk parametrization for the computation of turbulent fluxes in
208 numerical models and has already been used for the study of HPEs
209 (Lebeau-pin-Brossier *et al.* 2008). The reader is referred to Fairall
210 *et al.* (1996, 2003) for a comprehensive description of the *COARE*
211 algorithm which is summarized in the Appendix. 212 213 214

215 The version 3.0 of the *COARE* algorithm allows to take into
216 account the waves in the computation of the turbulent fluxes. The
217 waves, characterized by the dimensionless wave age χ , modulate
218 the roughness length z_0 defined in Eq. (A-10) and which is a
219 key parameter in the determination of the turbulent fluxes by
220 the *COARE* algorithm (cf. Appendix). The modulation of the
221 roughness length by the waves is accounted for through the
222 Charnock parameter α_{ch} which can be expressed as a function
223 of the wave age according to the formulation of Oost *et al.* (2002)
224 (Eq. (1)).

$$\alpha_{ch} = 50\chi^{-2.5} \quad (1)$$

$$\chi = \frac{gT_p}{2\pi u_*} \quad (2)$$

225 The dimensionless wave age χ depends on the friction velocity u_*
226 and on the peak period of the waves T_p only (Eq. (2)). In this study,
227 the impact of two different representations of T_p on the turbulent

228 fluxes is evaluated and compared to the default situation where the
 229 wave age is not taken into account. In the default situation without
 230 any wave impact, the formulation of the Charnock coefficient from
 231 Hare *et al.* (1999) is used. α_{ch} is set to 0.011 for wind speed
 232 below 10 m s^{-1} , then increases linearly up to 0.018 at 18 m s^{-1} ,
 233 and remains constant for larger wind speed values. Otherwise, the
 234 formulation of Oost *et al.* (2002) is used with T_p either computed
 235 empirically or obtained from an output of a wave model. In the
 236 first case, T_p is linearly dependent of the 10-m wind speed U with
 237 $T_p = 0.729U$. In the second case, T_p is given by the *MFWAM*
 238 analysis.

239 3.2. Influence of the sea state on z_0 and u_*

240 Using the *COARE* 3.0 parametrization, a first test of the influence
 241 of the waves on z_0 during IOP16a is run through 3 experiments
 242 (NOWAV, WAV and WAM) with an improved representation of
 243 the sea state from one to another. To be consistent with the use
 244 of *MFWAM* analyses, all these experiments are forced by the
 245 wind field of the atmospheric model *ARPEGE* used to drive the
 246 wave model *MFWAM*. For the first experiment called NOWAV,
 247 the wave age is not taken into account for the roughness length:
 248 it corresponds to the default situation presented above. The two
 249 other experiments (WAV and WAM) consider the formulation of
 250 Oost *et al.* (2002) (Eq. (1) and (2)). For WAV, T_p is given by the
 251 10-m wind speed of *ARPEGE*. For WAM, T_p is directly given by
 252 the analysis of the regional wave forecasting model *MFWAM* with
 253 a resolution of 0.1° , updated every 3 hours.

254 The roughness length values obtained from the *COARE*
 255 parametrization over the northwestern part of the Mediterranean
 256 Sea at 12 UTC on 26 October 2012 are displayed in Figure 6.
 257 Almost the same patterns of z_0 are displayed by the three
 258 experiments, although the roughness length is globally lower for
 259 WAV compared to NOWAV. On the contrary, the roughness length
 260 maxima, especially over the Gulf of Lion, can reach values 10
 261 times higher in WAM compared to NOWAV (from 2.10^{-4} to
 262 1.10^{-3} m). This difference includes also more variability in z_0
 263 coming from the variability in the *MFWAM* peak period. These
 264 first tests show that a more realistic description of the wave field
 265 has a stronger impact on the surface roughness than the use of
 266 an empirical formulation based on surface winds to determine the

267 wave characteristics. This strong albeit not systematic change of
 268 the roughness length in the WAM experiment has a direct impact
 269 on the associated friction velocity. Friction velocity differences
 270 between WAV and NOWAV and between WAM and NOWAV
 271 are displayed at 09 and 12 UTC on 26 October 2012 (Fig. 7).
 272 Positive values up to 0.08 m s^{-1} in the friction velocity differences
 273 between WAM and NOWAV are obtained along the southeast
 274 French coasts, the Gulf of Lion and off the Spanish coasts at
 275 both time steps. The use of the *MFWAM* waves induces higher
 276 values of the roughness length over most of the northwestern
 277 Mediterranean and consequently a stronger friction velocity. A
 278 comparison of the values of the drag coefficient C_d obtained
 279 using the three parametrizations in offline mode with observed
 280 atmospheric parameters at the Lion and Azur buoys is shown
 281 Fig. 8. The drag including the realistic wave effects (WAM, red
 282 dots) is significantly larger than the ones obtained with the two
 283 other simulations for wind speed superior to 6 m s^{-1} . It is thus
 284 expected a corresponding slowdown of the near surface winds.
 285 The atmospheric simulations discussed in the following have been
 286 designed to verify this assumption and to examine the impact on
 287 the intensity and location of the convective systems.

288 4. Atmospheric numerical experiments

289 4.1. The MESO-NH model

290 The non-hydrostatic atmospheric French research model *MESO-*
 291 *NH* (Lafore *et al.* 1997) is used for studying the effect of the sea
 292 state on the simulation of the convective precipitating systems of
 293 IOP16a. The simulation domain covers an area of $750 \times 1250 \text{ km}$
 294 including a large part of the western Mediterranean Sea region
 295 (Fig. 9). The marine domain here represents thus more than half
 296 of the full domain, in order to cover a large part of the upstream
 297 zone and to evaluate the impact of an improved representation of
 298 the waves on the sea-surface fluxes.

299 The model resolution and associated physical parametrization
 300 package are the same as those used in previous studies of HPEs
 301 using *MESO-NH* (e.g. Nuissier *et al.* 2008). The horizontal
 302 grid has a 2.5-km horizontal resolution. The vertical grid has
 303 55 stretched levels from about 19 m to 21 km (Gal-Chen and
 304 Somerville 1975).

305 The prognostic variables of the model are the three components
 306 of the wind, the dry potential temperature, the turbulent kinetic
 307 energy and the mixing ratios of the water vapor and of five
 308 different classes of hydrometeors (cloud water, rain water, primary
 309 ice, snow aggregates, and graupel). The evolution of the water
 310 species are governed by a bulk microphysical scheme (Caniaux
 311 *et al.* 1994; Pinty and Jabouille 1998). The parametrization of the
 312 turbulence is based on a 1.5-order closure (Cuxart *et al.* 2000).
 313 Thanks to its high horizontal resolution, the atmospheric deep
 314 convection is explicitly solved by the model.

315 The surface conditions and the air-surface exchanges are
 316 governed by the SURFEX surface model (Masson *et al.* 2013).
 317 The sea-surface turbulent fluxes parametrization is the COARE 3.0
 318 parametrization (see section 3.2).

319 4.2. Sensitivity simulations

320 Three sensitivity experiments using three different configurations
 321 of the COARE parametrization are performed, using the same
 322 experimental design as in section 3.2 (Table 1).

323 The three companion MESO-NH simulations all start on 25
 324 October 2012 at 12 UTC and last 36 hours. They are initialised
 325 and driven at their lateral boundaries every 3 hours by the high-
 326 resolution AROME-WMED analysis (Fourrié *et al.* 2015). The
 327 SST field comes from the initial AROME-WMED analysis, which
 328 is built with the 2D Optimal Interpolation of *in situ* measurements
 329 (CANARI, Taillefer (2002)) blended with the Operational Sea
 330 Surface Temperature Ice Analysis (OSTIA, Donlon *et al.* (2012)).
 331 The SST remains constant during the 36-h integration.

332 5. Results

333 5.1. The reference experiment NOWAV

334 The ability of the NOWAV simulation in representing the IOP16a
 335 is evaluated here.

336 5.1.1. Convective systems

337 Convection over the northeastern Spain and offshore is simulated
 338 in NOWAV in the afternoon and late evening of 25 October.
 339 After 2130 UTC, it is organized in a MCS corresponding to
 340 the observed MCS0 over the sea between the Balearic Islands

and northeastern Spain, and along the coast. MCS1 forms in the 341
 simulation at the tip of Catalonia around 03 UTC on 26 October 342
 2012. The system then moves eastward over the Gulf of Lion 343
 (Fig. 10c), with a location that corresponds quite well to the 344
 observed one (Fig. 10a). The splitting in two MCSs is however not 345
 represented in the simulation. MCS1 moves northeastward like 346
 the observed MCS1b. When reaching the Var coast, the simulated 347
 MCS1 is less organized and intense than previously over the 348
 sea (Fig. 10d). It then moves northeastward over the French and 349
 then Italian coasts. When compared with the observed MCS1b 350
 (Fig. 10b), the simulated convective system progresses eastward 351
 too rapidly during the afternoon and progressively loses its MCS 352
 organization. A convective system, corresponding to the observed 353
 MCS2, forms in the simulation around 02 UTC on 26 October 354
 over the northwestern Italian coast between Genoa and La Spezia 355
 and stays at the same location till about 13 UTC. This system 356
 is simulated only a few tens of kilometers north of the observed 357
 MCS2 (not shown). 358

To sum up, the deep convection over the sea and coastal regions 359
 is globally reproduced in NOWAV. The chronology and location 360
 of MCS0 and MCS2 are well represented. The development of 361
 MCS1 over the sea is also well simulated, even though MCS1a 362
 does not appear in the simulation. Moreover, it is in advance 363
 and less organized than in the observations, where it reaches the 364
 French coast during the afternoon. 365

366 5.1.2. 10-m wind speed

The simulated 10-m wind speed over the Mediterranean Sea 367
 provides a broad picture of the low-level dynamics of the 368
 atmosphere (Fig. 11). In the early morning of 26 October, the 369
 Mediterranean Sea is affected by a southeasterly flow coming 370
 from Sardinia and a southwesterly flow over the Balearic Islands 371
 region, resulting in a convergence line (Fig. 11a). The simulated 372
 MCS1 is located over the northern end of the convergence area. 373
 North of the convergence line, an easterly flow resulting from the 374
 deflection of the low-level flow by the Alps is simulated along 375
 the French coast. At that time, 10-m wind speeds are close to 376
 10 m s^{-1} . During the morning of 26 October (Fig. 11b), the 377
 intensity of the southeasterly flow increases, with peak values 378
 superior to 15 m s^{-1} . The convergence line is thus reinforced. 379

Figure 12a and 12b compares the 10-m wind speed from French coastal weather stations, moored buoys and ships and from the NOWAV simulation at 09 UTC (54 stations). The speed of the easterly to southeasterly flow along the coast in the Gulf of Lion is overestimated by 1.0 to 1.5 m s^{-1} in the simulation, probably linked to the absence of MCS1a in the simulation whereas the speed of the southeasterly flow over the Gulf of Lion is underestimated by 1.4 to 3.4 m s^{-1} . The southwesterly and the southeasterly flows are progressively shifted eastward, the extent of the southeasterly flow diminishing as it is pushed against Corsica and Sardinia (Fig. 11c). In the afternoon of 26 October, as the eastward shift continues, the southeasterly flow is limited to a narrow region from Corsica to the Var region with weaker wind speeds ($< 14 \text{m s}^{-1}$) compared to the situation in the morning (Fig. 11d). It vanishes progressively and disappears after 17 UTC. Between the Italian coast and Corsica, a southerly flow prevails in the morning and the afternoon of 26 October and shifts to a southeasterly flow in the evening (after 1730 UTC). It remains below 8m s^{-1} in the morning except over the Ligurian Sea where it feeds MCS2 with values around 10m s^{-1} (Fig. 11b). It then strengthens for the rest of the day with 10-m wind speed between 10 and 15m s^{-1} (Fig. 11d).

5.1.3. Turbulent fluxes

Latent heat flux is quite low over the western Mediterranean Sea (Fig. 13a). The largest values of latent heat flux, higher than 200W m^{-2} , are associated with the low-level southwesterly winds, located in the morning of 26 October between Spain and the Balearic Islands (Fig. 11). This area of moderate winds progresses eastward during the day and affects, at the end of the day, the whole western Mediterranean basin between Spain and Sardinia, south of Gulf of Lion. Dry air (relative humidity $< 70\%$) is associated with this southwesterly wind area (Fig. 13b) and produces the large evaporation. The strong low-level southeasterly winds (from Sardinia to Var at 12 UTC, see Fig. 11c) do not produce high latent heat fluxes as this low-level flow is nearly saturated (Fig. 13b). The high values of relative humidity prevent large evaporation to occur.

Sensible heat flux remains below 30W m^{-2} (not shown) the whole time except beneath the MCSs with localized peak

values of sensible heat flux between 100W m^{-2} and 150W m^{-2} , corresponding to strong low-level cooling induced by evaporation of the falling precipitation. The low values of sensible heat fluxes can be partly explained by the weak differences between the SST and the 2-m air temperature over the domain (1 to 3°C locally).

Finally, the momentum flux remains lower than 0.2N m^{-2} during the simulation, except under the main south-southeasterly flow directed towards the French coasts where the momentum flux is stronger than 0.4N m^{-2} with peak values close to 1N m^{-2} between 07 and 12 UTC on 26 October (Fig. 13c).

5.2. Sensitivity to sea state

In the following paragraphs, we evaluate the sensitivity of the sea-surface fluxes, of the atmospheric low-level conditions, and of the convective systems at the origin of heavy precipitation to the sea-state representation.

5.2.1. Sea surface turbulent fluxes

The momentum flux simulated by the WAM simulation is significantly larger than the NOWAV and WAV momentum fluxes between 07 and 12 UTC on 26 October, when and where the momentum flux is the strongest for the three simulations. The differences between the friction velocity simulated by WAM and NOWAV reach values close to 0.1m s^{-1} in the south-southeasterly flow area (Fig. 14a). As shown in section 3.2, the friction velocity is strongly linked to the roughness length which is clearly larger for WAM, and more particularly over the Gulf of Lion (Fig. 6). The larger roughness length in WAM influences thus directly the momentum flux. It has however no impact on the turbulent heat fluxes. In addition to the fact that the turbulent heat fluxes are globally low during the simulations, the area of the Mediterranean Sea with a strong impact of the waves on the roughness length is indeed associated with a low-level nearly-saturated air flow (Fig. 13) which limits latent heat flux variations. Also, the sensible heat fluxes are low in all the simulations (see section 5.1.3) as the SST and the 2-m temperature are very close to each other.

454 5.2.2. *Low-level winds*

455 In response to larger momentum fluxes, the 10-m wind speed
 456 simulated with the WAM parametrization is 1 to 3 m s^{-1} lower
 457 than in the two other simulations (Fig. 14b). It results in a strong
 458 slowdown of the low-level moist south-southeasterly flow that
 459 feeds the convective systems. Ten-meter wind with speed higher
 460 than 10 m s^{-1} are particularly affected by this decrease, as already
 461 shown in section 3.2 (Fig. 8, see the drag coefficient C_d with
 462 respect to the neutral 10-m wind speed at the Lion and Azur
 463 buoys). The largest differences occur for wind speed above 7
 464 m s^{-1} , with C_d values significantly higher when realistic wave
 465 characteristics are taken into account. On the other hand, using
 466 a theoretical peak period dependent on the 10-m wind speed
 467 only as in the WAV parametrization shows almost no difference
 468 with NOWAV whatever the wind speed. Accounting for the wave
 469 effects in the momentum flux using a realistic wave field as the
 470 one provided by the *MFWAM* analysis is of interest especially in
 471 areas with moderate to strong winds.

472 Table 2 lists the mean bias and the standard deviation of the
 473 difference (SDD) for NOWAV and WAM simulated 10-m wind
 474 speed against the observations taken every hour between 00 and 11
 475 UTC on 26 October. The observed 10-m wind speed is obtained
 476 using a logarithmic profile from the wind speed at the height of
 477 the measurement (4 m for the moored buoys), and the simulated
 478 wind speed is extracted at the closest grid point of the model. The
 479 mean bias for both simulations is larger between 07 and 10 UTC,
 480 when the area is concerned by the stronger low-level south-
 481 southeasterly winds. The mean bias for WAM is however reduced
 482 compared to NOWAV for this period, and more broadly between
 483 03 and 10 UTC even though the reduction is not significant due
 484 to large uncertainties. The slowdown of the 10-m winds by taking
 485 into account the wave effects thus improves the simulated winds
 486 with respect to the observations. No significant improvement is
 487 observed for the SDD between both sets of data, meaning that
 488 there is no modification of the spatial variations of the 10-m wind
 489 speed in WAM compared to NOWAV.

490 5.2.3. *Precipitation*

491 Figure 15 displays the 24-h accumulated precipitation from
 492 the three simulations. Rain gauges show three areas of intense

precipitation (more than 100 mm in 24 h) from west to east: the 493
 first one over the Cévennes mountains, the second one along the 494
 French Var coast and the last one along the northwestern Italian 495
 coast, associated with MCS1a, MCS1b and MCS2, respectively. 496
 The main difference between the simulations concerns the 497
 precipitation associated with MCS1b. The maximum simulated 498
 rainfall amounts match the observations (142 mm/24h) with 499
 maximum daily precipitation of 123 mm/24h (NOWAV), 107 500
 mm/24h (WAV) and 120 mm/24h (WAM). However these maxima 501
 are located inland in NOWAV and WAV, whereas they are located 502
 closer to the coast in WAM and in the observations. Over Italy, 503
 the maximum daily rainfall amounts are shifted northwestward 504
 compared to the observations in the three experiments. WAM 505
 performs a little better in extending the heavy precipitation area 506
 southward. 507

The maxima of 24-h rainfall totals simulated by NOWAV, 508
 WAV and WAM are close to the observed one, above 200 mm. 509
 Large precipitation amounts are simulated over the Cévennes 510
 region by the three simulations. WAM produces more intense 511
 precipitation upwind the mountain range whereas NOWAV places 512
 it over the northern part of the region. This behaviour is 513
 consistent with a weaker south-to-southeasterly flow feeding 514
 MCS1a and MCS1b during the morning. Indeed, based on 515
 idealized numerical simulations, [Bresson *et al.* \(2012\)](#) showed that 516
 convective precipitation upwind [respectively over] the Cévennes 517
 mountain range is favoured when the impinging feeding flow is 518
 weaker [resp. stronger]. 519

To assess more precisely the skill of the simulations, scores 520
 against rain gauge observations over the whole simulation domain 521
 have been computed. The simulated daily rainfall amounts are 522
 extracted at the closest grid point to the 2144 rain gauge stations. 523
 The mean bias, the SDD, and the correlation coefficient (r) have 524
 been computed, as well as two categorical scores: the Equitable 525
 Threat Score (ETS; Schaefer, 1990) and the Hanssen and Kuipers 526
 discriminant (HK; Hanssen and Kuipers, 1965). A perfect forecast 527
 would give ETS and HK equal to 1, and null ETS and HK 528
 indicate no skill. The mean bias is reduced in WAV and WAM 529
 compared to NOWAV, and the correlation coefficient is increased 530
 (Table 3). Once again, due to large standard deviations of the 531
 differences, the reduction of bias is not statistically significant, 532

533 whereas the increase of the correlation coefficient is significant
 534 at 95% with a two-sided hypothesis (Fisher test). NOWAV and
 535 WAM give almost the same SDD values whereas it is larger
 536 for WAV. The categorical scores for the 5 mm, 10 mm and 25
 537 mm thresholds indicate that WAM performs better than NOWAV
 538 for all thresholds and scores. On the opposite, WAV performs
 539 slightly worse than NOWAV. The improvement in both ETS and
 540 HK scores is significant with a 95% probability. This objective
 541 evaluation confirms the better skill of the WAM experiment with
 542 respect to NOWAV in representing the precipitation.

543 To sum up, the main impact of the slowdown of the low-
 544 level flow in the WAM simulation, in better agreement with
 545 the observed 10-m winds, concerns the location of the intense
 546 precipitation. No significant impact on the amplitude of the
 547 maximum of precipitation is evidenced. This better match in
 548 location with the observed precipitation leads to globally better
 549 scores.

550 6. Conclusions

551 This study examines the impact of a better representation of
 552 the wave effect on the turbulent fluxes in the convection-
 553 permitting simulation of coastal heavy precipitation. During
 554 HyMeX IOP16a, three MCSs produced heavy precipitation over
 555 the Cévennes mountains, on the southeast French coast and on
 556 the northwestern Italian coast with two of these systems forming
 557 over the Mediterranean Sea prior to reaching the coasts. A more
 558 realistic representation of the wave effect on the turbulent fluxes
 559 has been used in the *COARE* 3.0 parametrization (Fairall *et al.*
 560 2003) with wave characteristics, namely peak period converted
 561 into wave age, coming from the 3-h *MFWAM* analyses at 10-km
 562 resolution. First, the study highlights the theoretical impact of the
 563 waves on the roughness length and on the wind friction velocity
 564 using the turbulent fluxes parametrization alone. Comparison of
 565 the roughness length z_0 and of the friction velocity u_* show a
 566 strong increase of both parameters when realistic wave parameters
 567 and the formulation of Oost *et al.* (2002) are used.

568 In a second set of experiments, three numerical simulations
 569 of IOP16a using the non-hydrostatic atmospheric model *MESO-*
 570 *NH* at 2.5-km resolution were performed using the same
 571 discrimination in the turbulent fluxes parametrization. The

increase in the surface roughness highlighted in WAM directly
 impacts the low-level dynamics of the atmosphere when and
 where the wind speed is higher than 10 m s^{-1} and the sea surface
 state is significantly different from the idealized one. It results in
 a significant slowdown of the 10-m wind in WAM compared to
 the two other simulations, due to higher momentum flux. Before
 and during the event, the highest latent heat exchanges at the
 air-sea interface correspond to areas where the low-level flow is
 not saturated in humidity, and not to the low-level wind maxima.
 As such, they are almost not sensitive to the wave representation
 in the fluxes parametrization. Although the convective activity is
 globally well reproduced by the three simulations, the difference
 in the low-level dynamics influences the localization of the
 simulated daily precipitation. The objective evaluation against
 the observations over the entire simulated domain for the daily
 rainfall amounts and along the French coast for the 10-m wind
 speed confirms a better representation of both parameters by the
 WAM simulation. The case study of IOP16a is thus sensitive
 to a more realistic representation of the waves with a better
 representation of the simulated precipitation especially due to a
 better representation of the low-level moist jet feeding the French
 coastal precipitating systems.

This study shows that, even in a moderate-wind context, the
 sea-surface roughness due to waves can significantly influence
 the low-level flow and the marine atmospheric boundary layer
 dynamics. As location and intensity of heavy precipitation have
 been shown to be very sensitive to the characteristics of the
 moist low-level inflow, a more realistic representation of the
 wave influence on the turbulent fluxes modifies the simulated
 precipitation. It demonstrates that wind-wave interaction is also
 important in convection-permitting NWP models. Large impacts
 may notably concern strong-wind events like mistral, midlatitude
 storms, tropical storms and cyclones. Therefore, the use of a
 simplistic wind-waves transfer function such as in WAV is in some
 cases not sufficient to represent the variability of the sea-surface
 roughness and momentum flux due to the sea state.

The perspectives of this study include different steps. First,
 we plan the study of a second HPE (HyMeX IOP13) during
 which heat transfers between the ocean and the atmosphere
 are more intense prior to the event, and where the sensitivity

612 of the simulated heat fluxes to the waves and sea-roughness
 613 representation should be higher. The surface wind speeds are
 614 higher and one can expect a stronger impact on the atmospheric
 615 surface layer and on the HPE chronology. Then, a two-way
 616 coupling between the atmospheric and wave models could be
 617 considered in a second step. The effect of a full coupling between
 618 atmospheric and wave models should be reduced with respect
 619 to the effect of a forcing as used in the present study. Indeed,
 620 the slowdown of the low-level wind obtained with the WAM
 621 configuration should partly damp the sea state and reduce the
 622 surface roughness, with a possible negative feedback on the
 623 atmosphere. Finally, it could be of interest to distinguish the total
 624 sea-state effect as taken into account in this study (the peak period
 625 may correspond in some cases to swell) from the pure wind-
 626 sea effect. The parametrization of Oost *et al.* (2002) used here
 627 corresponds to the instantaneous wind-wave equilibrium and is
 628 supposed to be constrained by the characteristics of the wind sea
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 641 model.

642 Appendix - Bulk parametrization of turbulent fluxes

643 Bulk parametrizations of turbulent fluxes relate the latter to the
 644 vertical gradients between atmospheric and oceanic parameters
 645 close to the surface, using linear transfer coefficients C_d , C_h ,
 646 C_q for τ , H_s , H_l respectively. According to the Monin-Obukhov
 647 (MO) similarity theory, the turbulent fluxes can also be defined
 648 thanks to the scale parameters u_* , θ_* , q_* of wind, potential
 649 temperature and humidity, respectively:

$$\tau = \rho C_d (\Delta U)^2 = \rho u_*^2 \quad (\text{A-3})$$

$$H_s = \rho c_p C_h \Delta U \Delta \theta = \rho c_p u_* \theta_* \quad (\text{A-4})$$

$$H_l = \rho L_v C_q \Delta U \Delta q = \rho L_v u_* q_* \quad (\text{A-5})$$

where ΔU , $\Delta \theta$, Δq are the air-sea gradients of velocity, 650
 potential temperature and specific humidity close to the interface. 651
 C_p is the air heat capacity, L_v is the vaporization heat constant 652
 and ρ is the air density. 653

In the COARE parametrization, the transfer coefficients are 654
 determined after iterations over the MO scale parameters, the 655
 roughness length z_0 , and the MO length L using Eq. (A-6) to 656
 (A-10). The stability functions ψ_u , ψ_θ , and ψ_q used in Eq. (A-6) 657
 to (A-8) correspond to the generalization of atmospheric profiles 658
 in neutral conditions to non-neutral conditions and depend only 659
 on the stability parameter $\zeta = z/L$, z being the reference height. 660

$$u_* = \frac{k \Delta U}{\ln\left(\frac{z}{z_0}\right) - \psi_u(\zeta)} \quad (\text{A-6})$$

$$\theta_* = \frac{k \Delta \theta}{\ln\left(\frac{z}{z_0}\right) - \psi_\theta(\zeta)} \quad (\text{A-7})$$

$$q_* = \frac{k \Delta q}{\ln\left(\frac{z}{z_0}\right) - \psi_q(\zeta)} \quad (\text{A-8})$$

$$L = \frac{T u_*^2 (1 + aq)}{\theta_* (1 + aq) + a q_* T} \text{ is the MO scale height} \quad (\text{A-9})$$

with $a \approx 0.61$ and $k = 0.4$ is the constant of von Karman.

$$z_0 = \alpha_{ch} \frac{u_*^2}{g} + 0.11 \frac{\nu}{u_*} \quad (\text{A-10})$$

α_{ch} is the Charnock parameter (see below) and ν is the 661
 kinematic viscosity of dry air. 662

Starting with a first guess of u_* , θ_* and q_* , the roughness length 663

664 z_0 is obtained from Eq. (A-10) and the stability parameter $\zeta =$
 665 z/L from Eq. (A-9). Both parameters are then used to reassess
 666 u_* , θ_* , q_* following Eq. (A-6) to (A-8). The whole process is
 667 reiterated up to three times if the stability parameter $\zeta \leq 50$, else
 668 the computation is ended and the scale parameters are no more
 669 modified. At the end, the transfer coefficients and the turbulent
 670 fluxes are determined using the last iterated values of u_* , θ_* , q_*
 671 according to Eq. (A-3), (A-4), (A-5).

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<i>MESO-NH</i>		
Initial time: 25 October 2012, 12 UTC		
Duration: 36 h		
Initial and boundary conditions: <i>AROME-WMED</i> analysis		
Sea surface turb. flux param.: <i>COARE 3.0</i>		
Name	α_{ch}	Wave (T_p)
NOWAV	<i>Hare et al. (1999)</i> $0.011 \leq \alpha_{ch} \leq 0.018$	None
WAV	<i>Oost et al. (2002)</i>	$T_p = 0.729U$
WAM	<i>Oost et al. (2002)</i>	<i>MFWAM</i> 3-hourly analysis

Table 1. Description of the three *MESO-NH* experiments.

UTC	NOWAV-Obs		WAM-Obs	
	Mean bias	SDD	Mean bias	SDD
00	-0.06	1.78	-0.08	1.98
01	-0.03	1.96	0.06	1.97
02	0.19	2.27	0.20	2.18
03	-0.13	2.45	-0.10	2.74
04	0.82	2.30	0.60	2.04
05	0.27	2.23	0	1.98
06	0.67	2.08	0.66	2.12
07	1.01	2.32	0.81	2.54
08	1.47	2.32	1.28	2.40
09	1.26	2.73	0.84	2.87
10	1.39	3.27	1.32	3.02
11	0.84	3.12	1.0	3.15

Table 2. Ten-meter wind speed statistical analysis (m s^{-1}); SDD = standard deviation of the difference.

		NOWAV	WAV	WAM
	Mean bias	2.96	2.77	2.24
	SDD	22.21	24.47	22.30
	Correlation	0.577	0.591	0.589
5 mm	ETS	0.296	0.286	0.321
	HK	0.496	0.473	0.522
10 mm	ETS	0.245	0.240	0.281
	HK	0.441	0.441	0.491
25 mm	ETS	0.210	0.191	0.229
	HK	0.338	0.319	0.369

Table 3. Scores of the *MESO-NH* experiments against 24-h cumulated rain gauge observations on 26 October 2012 (mean bias and SDD in mm).

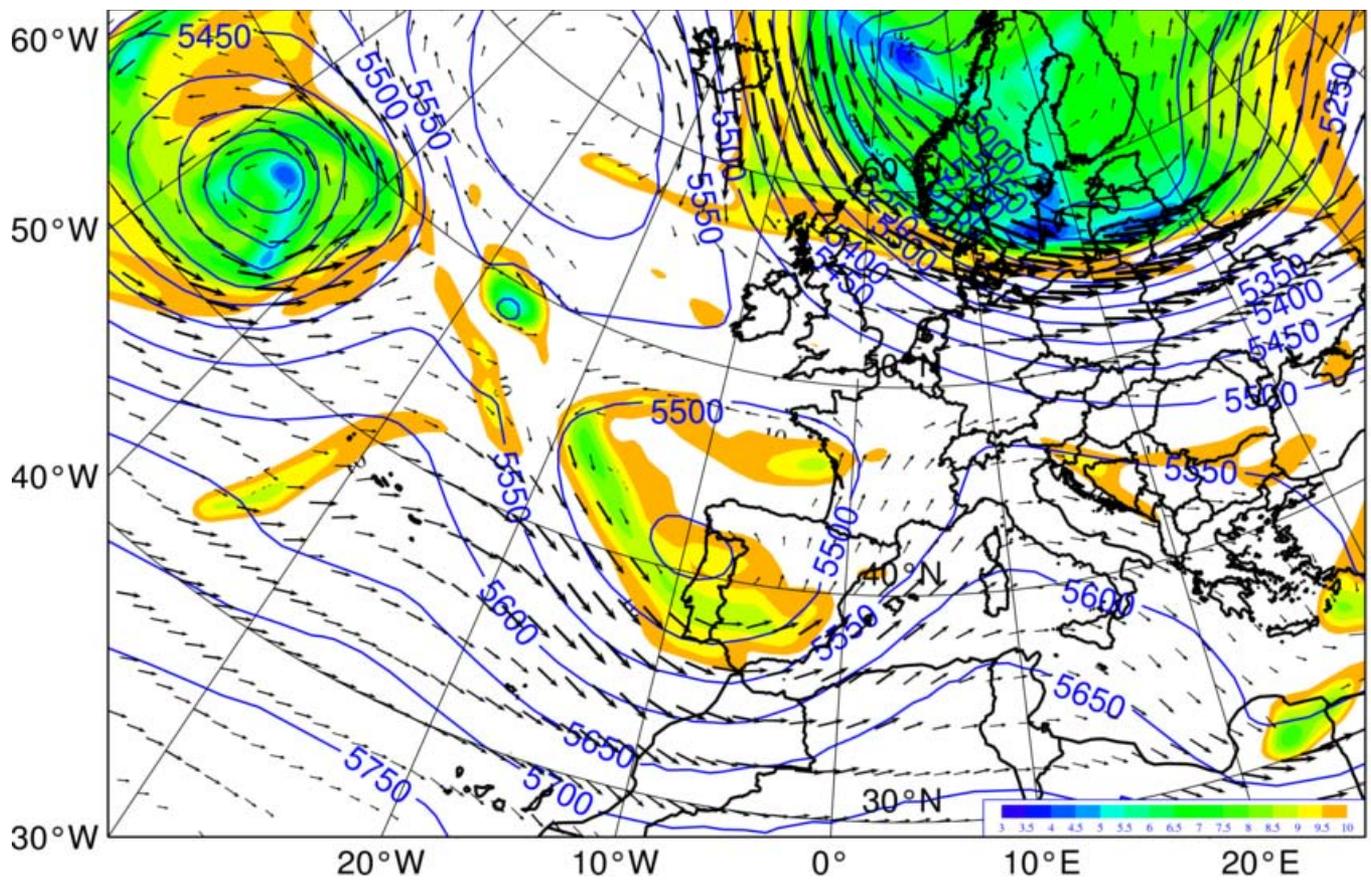


Figure 1. ARPEGE analysis at 00UTC, 26 October 2012: geopotential at 500 hPa (isolines, m), height of the 2 PVU iso surface (colour) and wind vectors at 300 hPa (above 10 m s^{-1}).

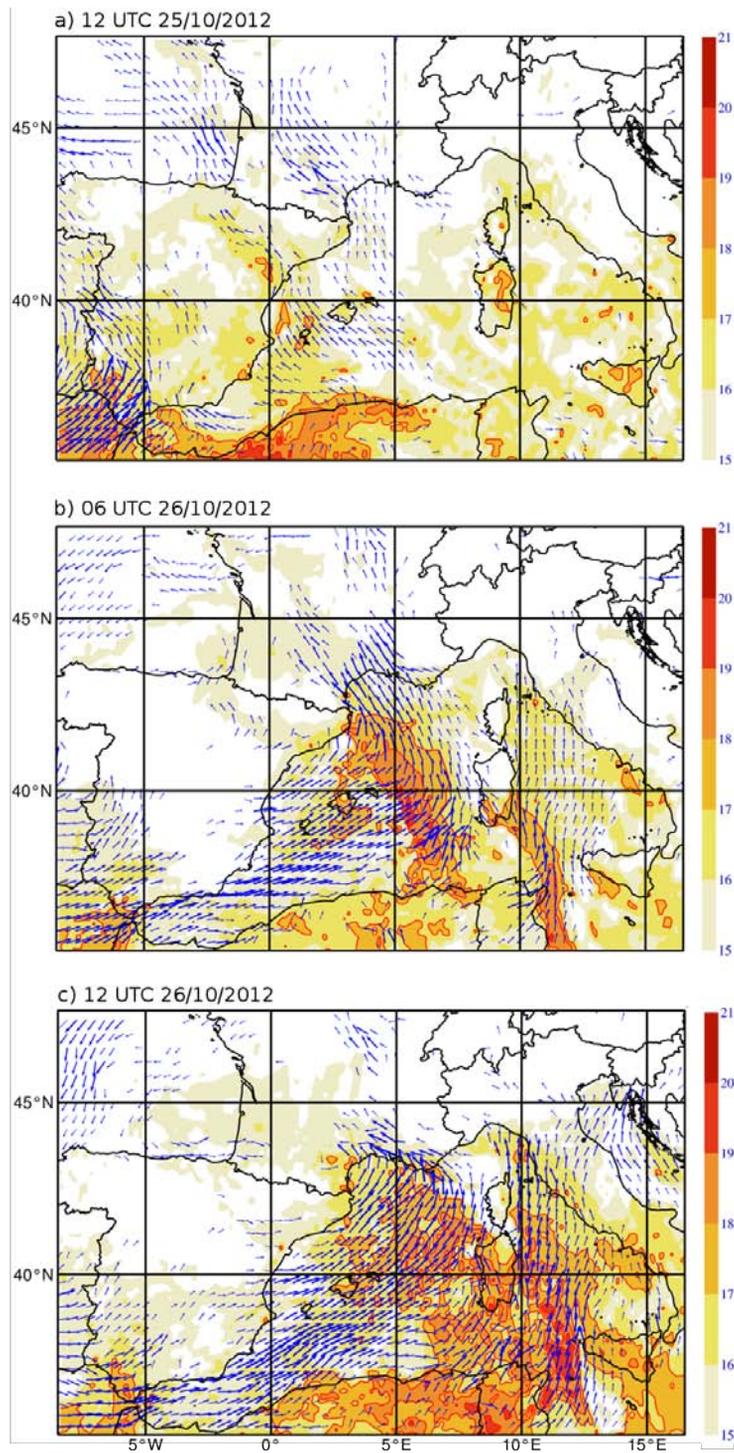


Figure 2. Wind vectors and wet bulb potential temperature at 925 hPa at 12 UTC on 25 October 2012 (a) and at 06 and 12 UTC on 26 October 2012 (b and c) from AROME-WMED analysis (Fourrié *et al.* 2015).

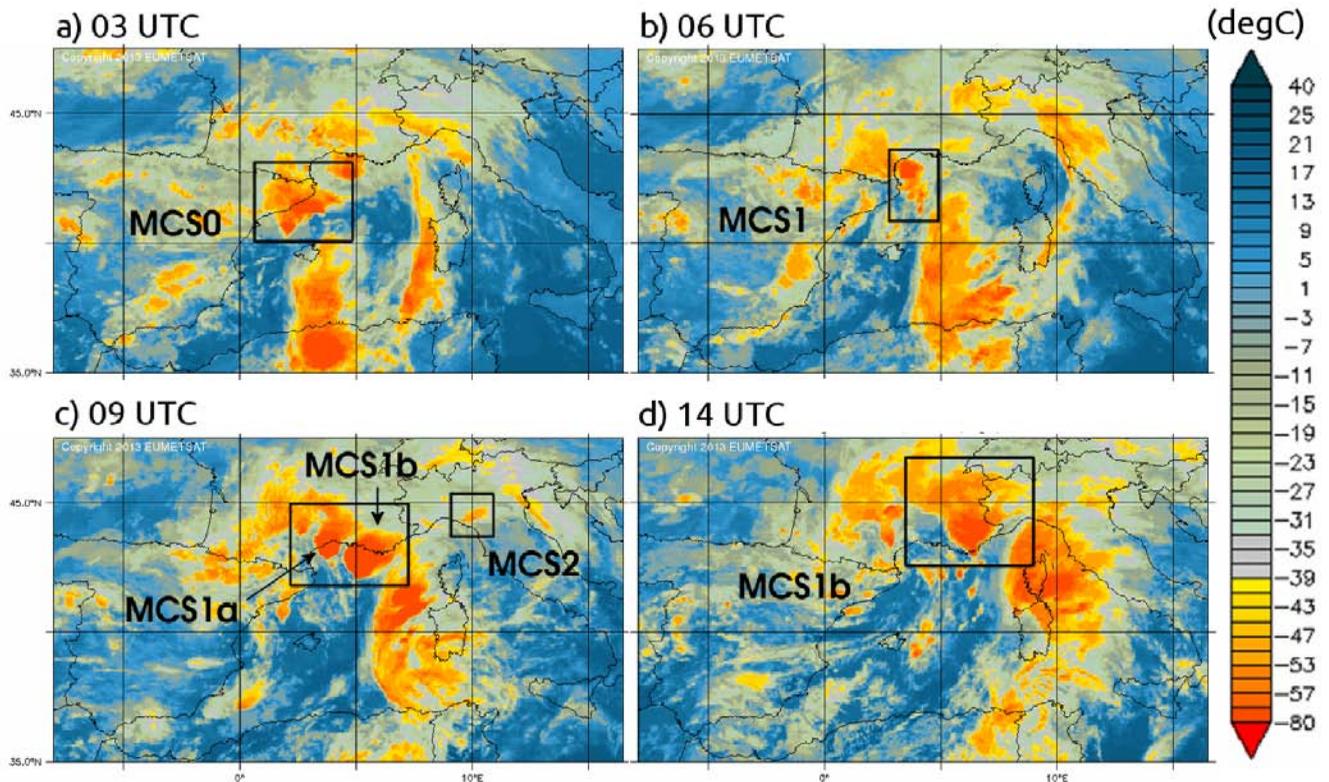


Figure 3. Météosat Second Generation infrared brightness temperature ($^{\circ}\text{C}$) on 26 October 2012 at 03 (a), 06 (b), 09 (c), 14 (d) UTC.

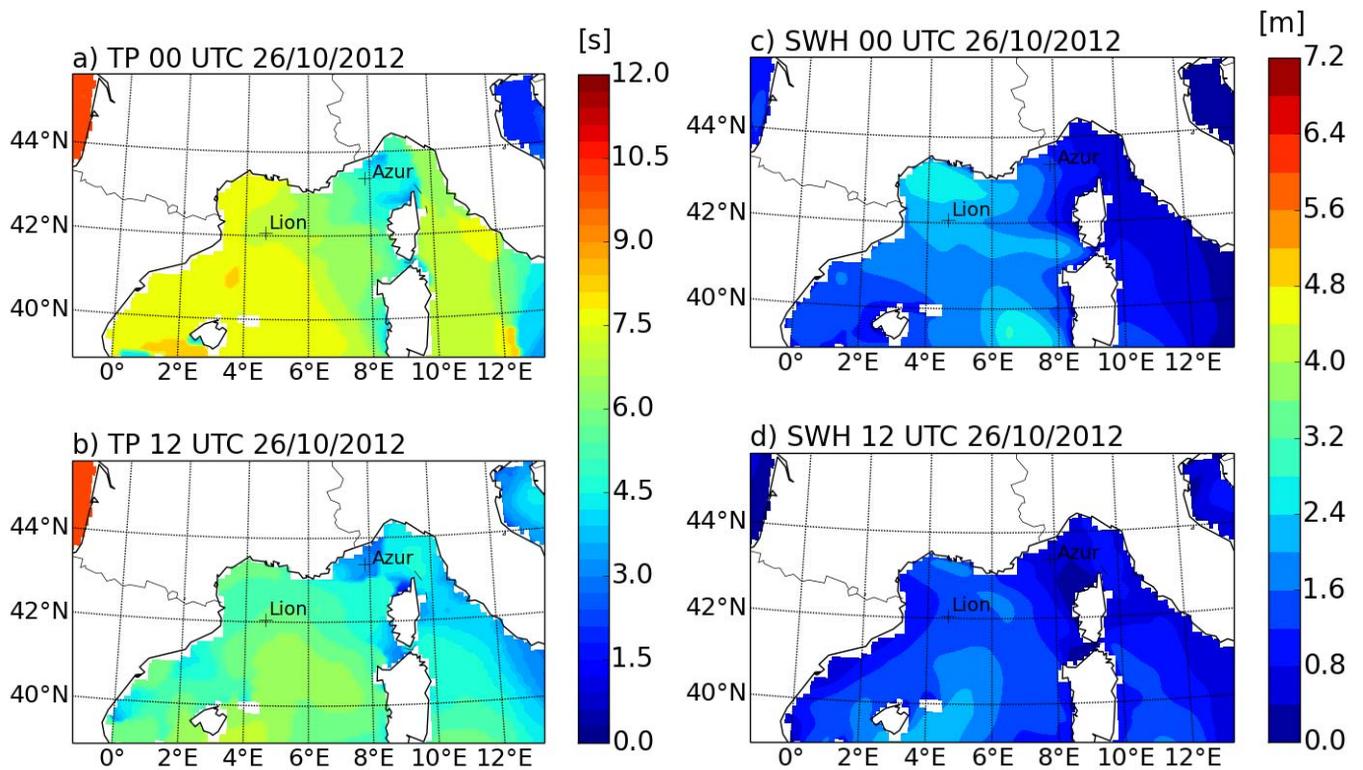


Figure 4. MFWAM analysis of the peak period (TP, left) and the significant wave height (SWH, right) on the 26 October 2012 at 00 UTC (a and c) and 12 UTC (b and d).

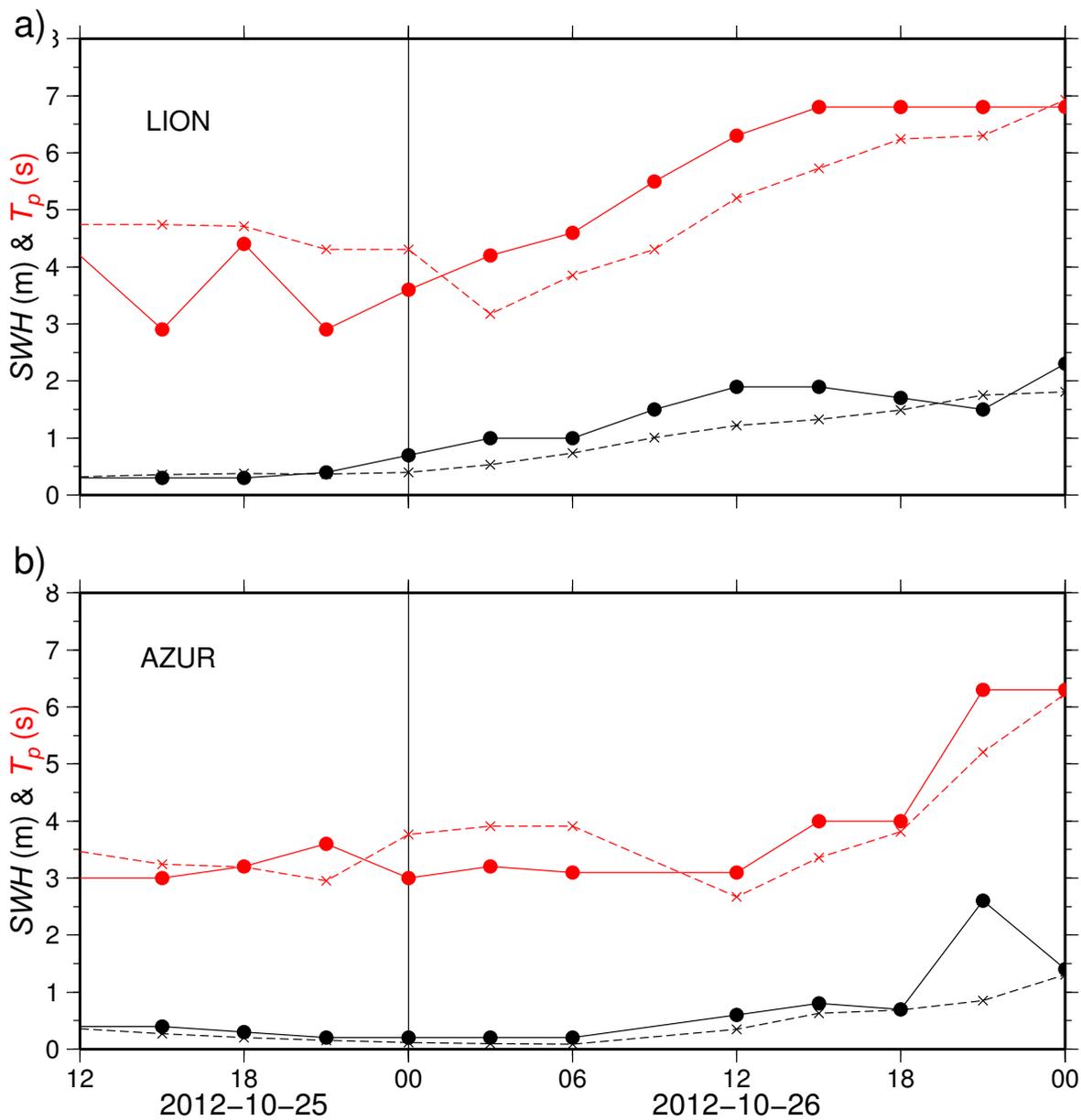


Figure 5. Time series of the wave significant height (SWH , m, black) and of the peak period (T_p , s, red) observed (solid line and dots) and modeled by *MFWAM* (dashed line, crosses) at (a) the Lion buoy, and (b) the Azur buoy, for the time of the simulation.

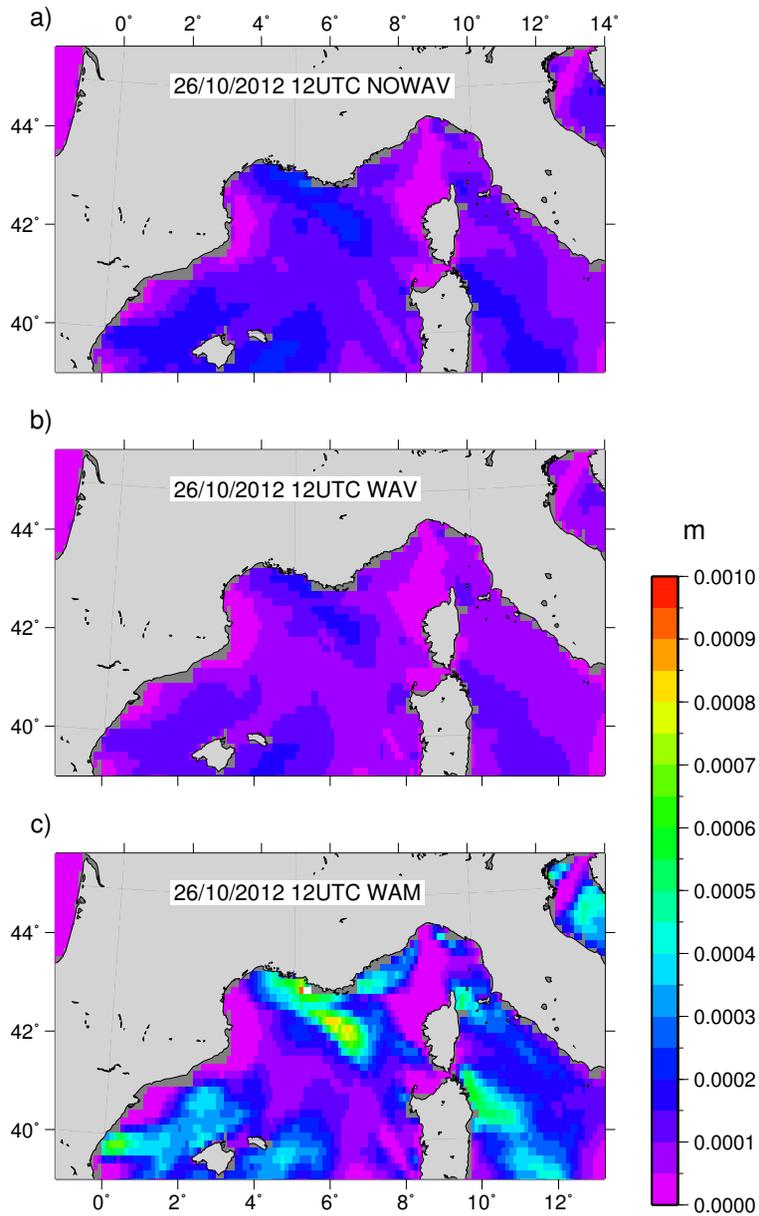


Figure 6. Roughness length (z_0 , m) for (a) NOWAV, (b) WAV and (c) WAM at 12 UTC on 26 October 2012.

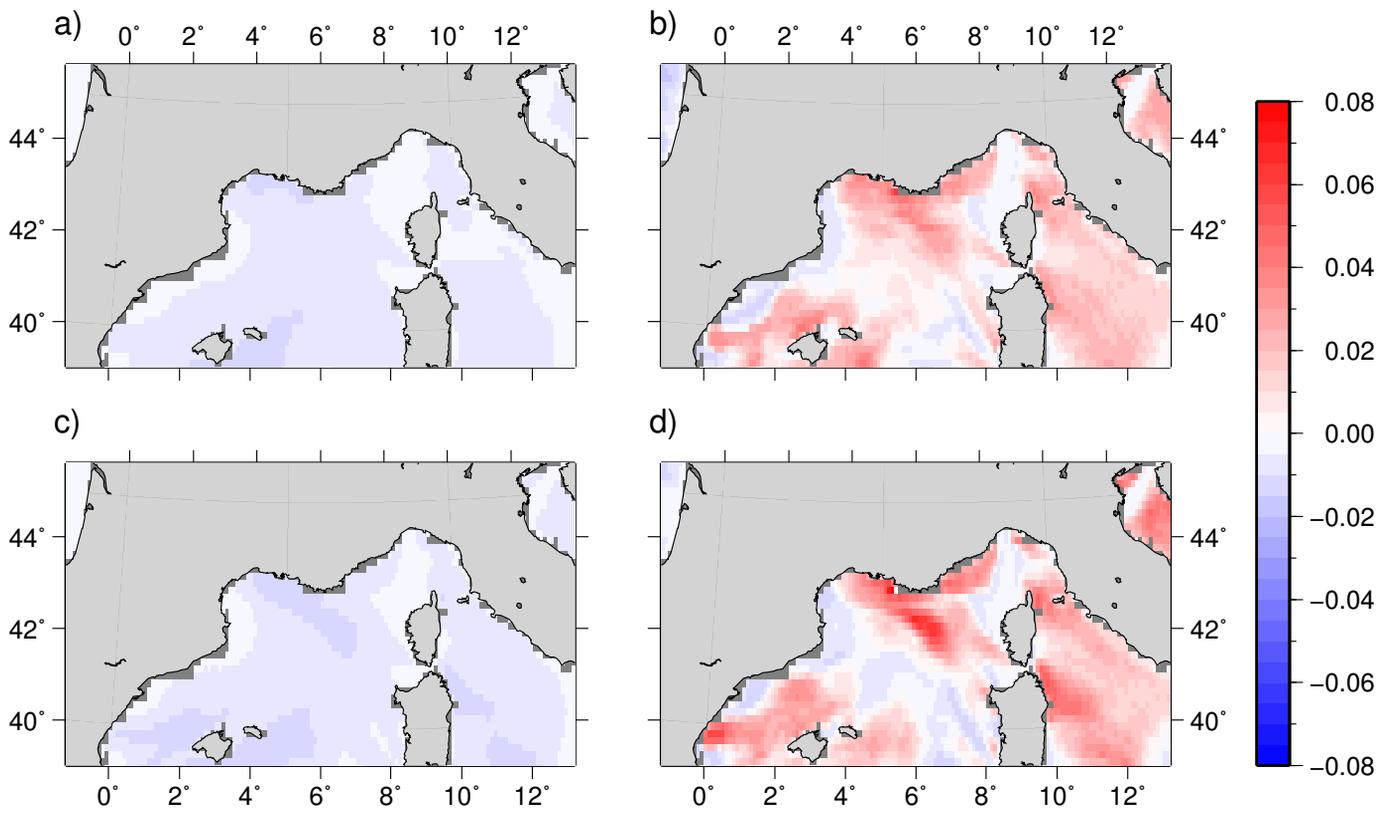


Figure 7. Friction velocity (u_* , m s^{-1}) differences between WAV and NOWAV at 09 (a) and 12 UTC (c) on 26 October 2012, WAM and NOWAV at 09 (b) and 12 UTC (d).

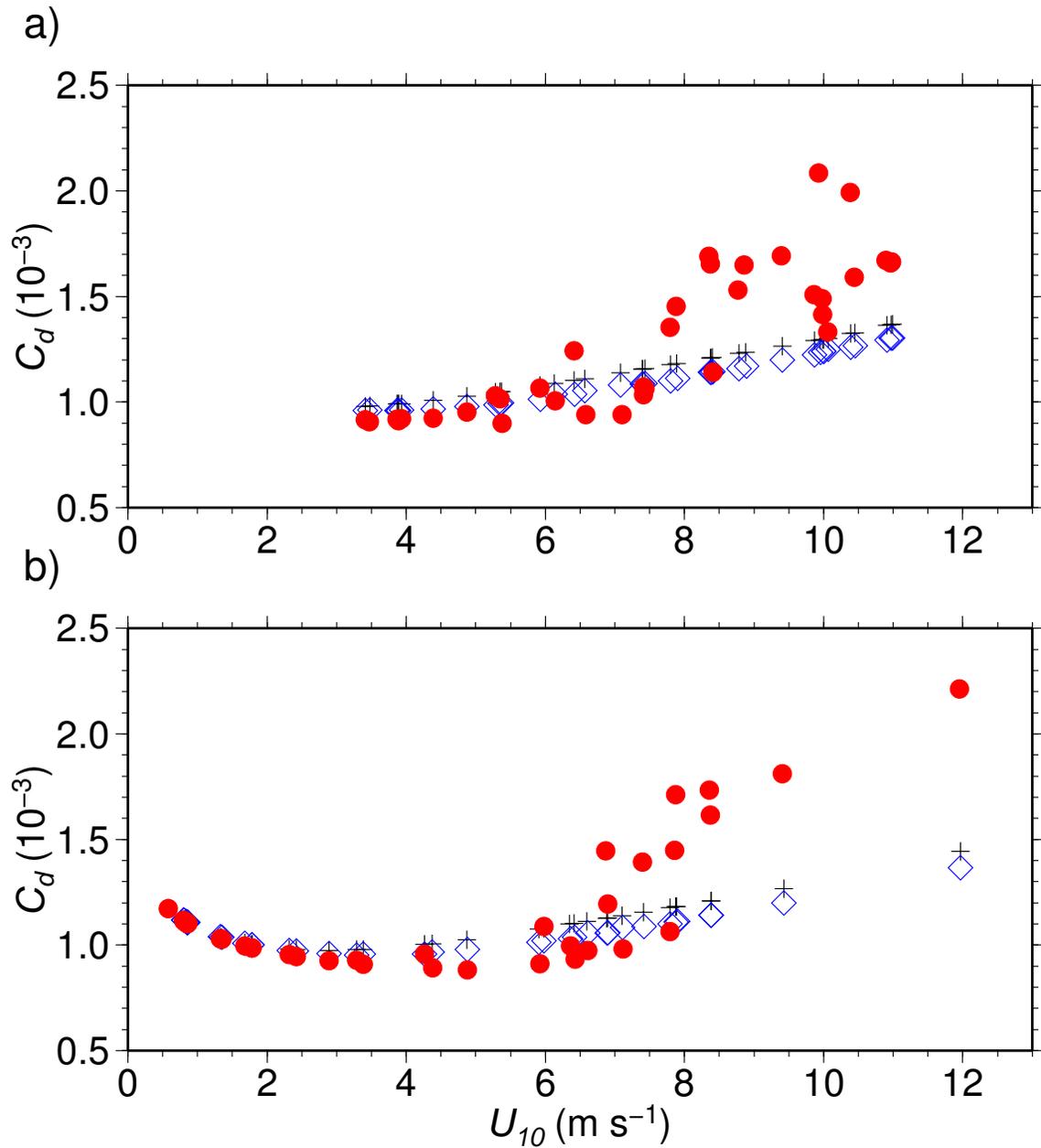


Figure 8. Evolution of the transfer coefficient C_d associated to the momentum flux with the 10-m neutral wind speed at (a) the Lion and (b) Azur buoys for the three simulations: NOWAV (black plusses), WAV (blue diamonds) and WAM (red dots).

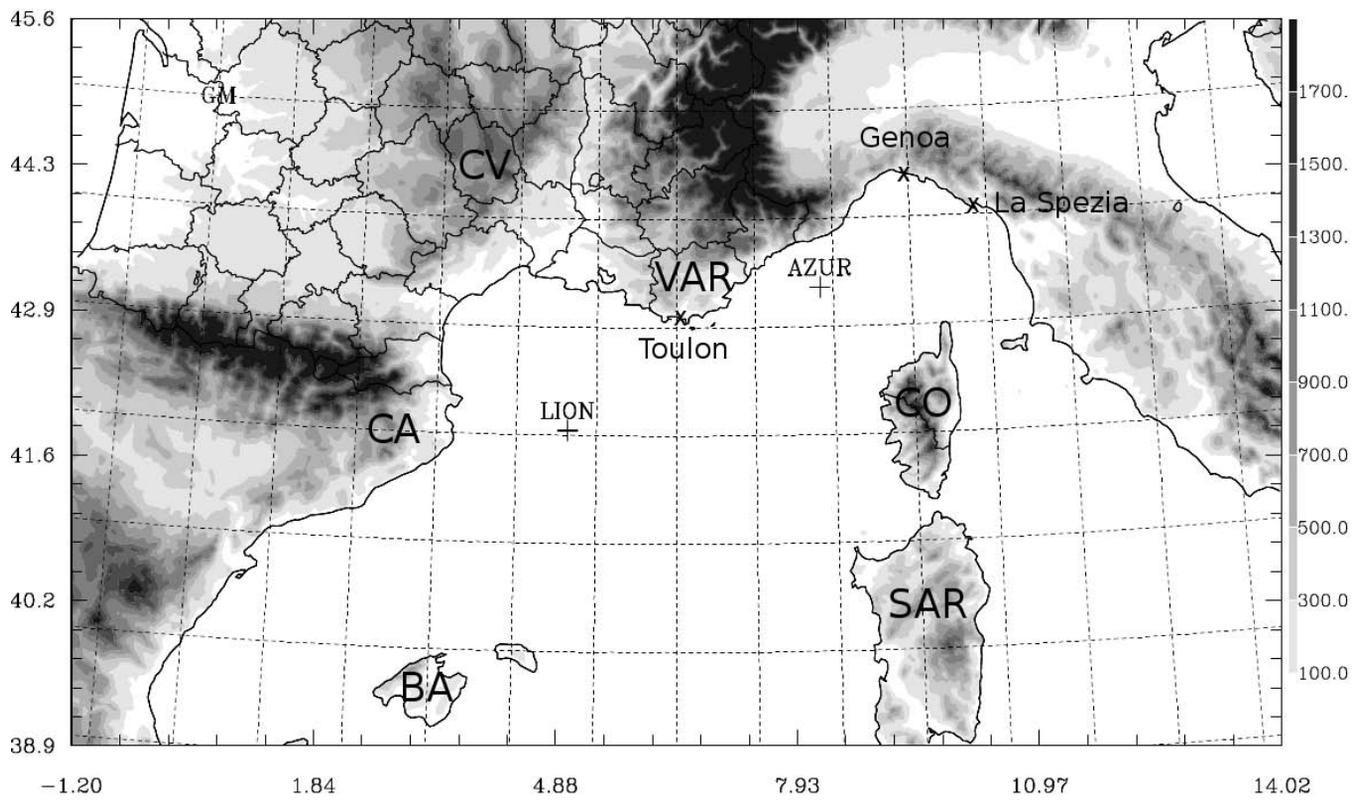


Figure 9. Domain used in the three *MESO-NH* simulations with some specific areas: CA, Catalonia; CV: Cévennes; CO, Corsica; SAR, Sardinia; BA, Balearic Islands; and VAR for the French Var department.

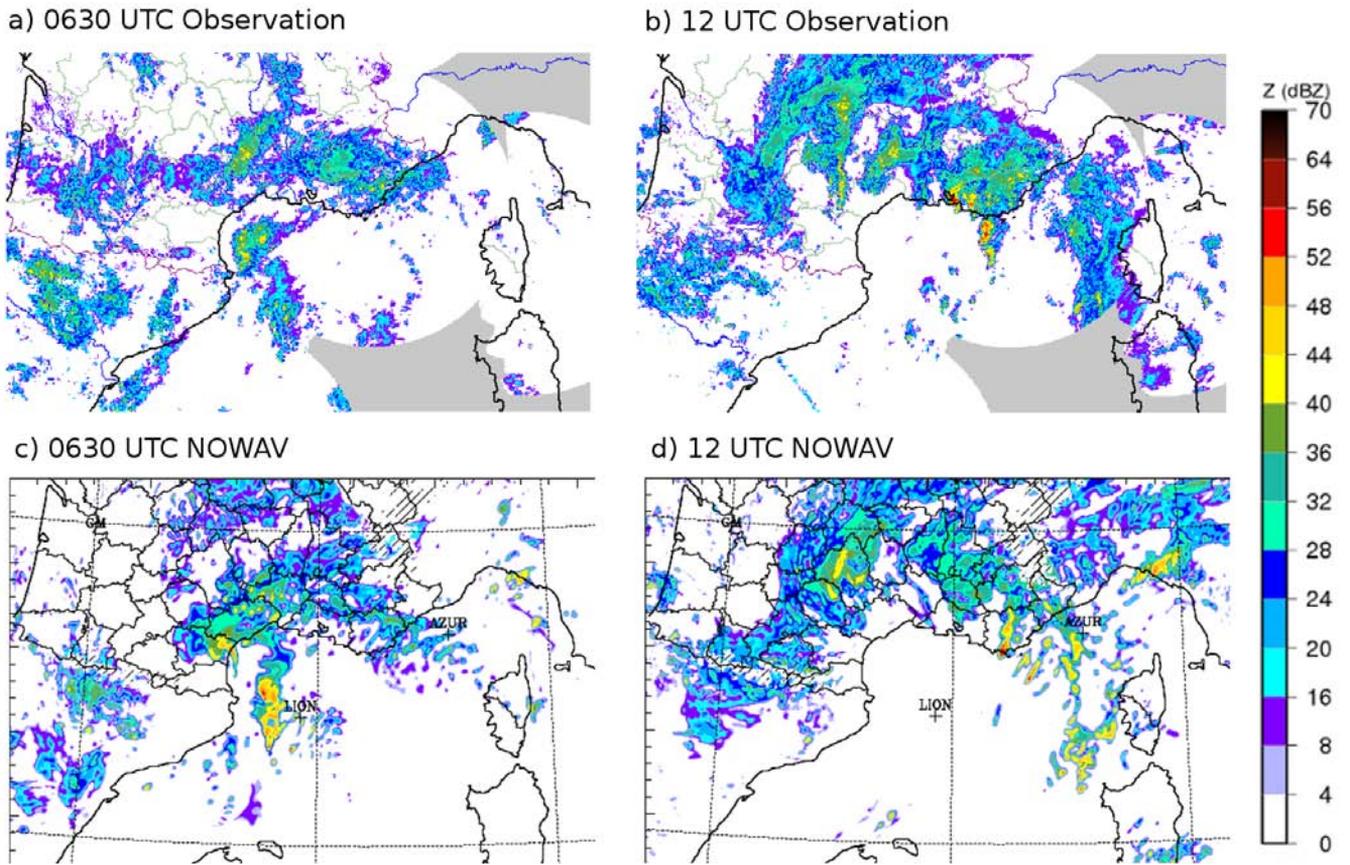


Figure 10. Radar reflectivities at 2000 m (dBZ): observed (top) versus simulated by NOWAV (bottom) at a) 0630 UTC and b) 12 UTC on 26 October 2012.

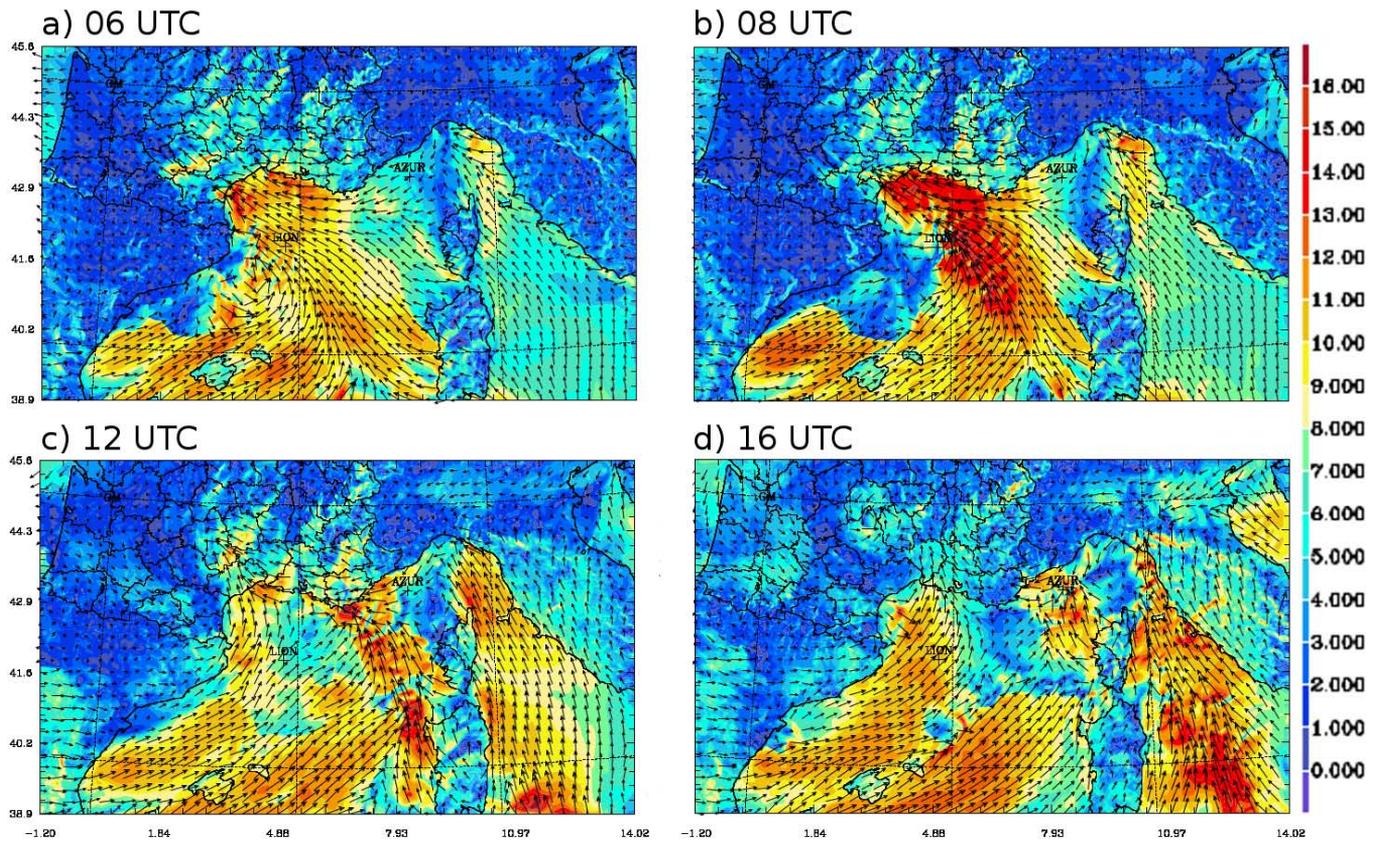


Figure 11. 10-m wind speed (m s^{-1}) and direction simulated by NOWAV at a) 06 UTC, b) 08 UTC, c) 12 UTC and d) 16 UTC on 26 October 2012.

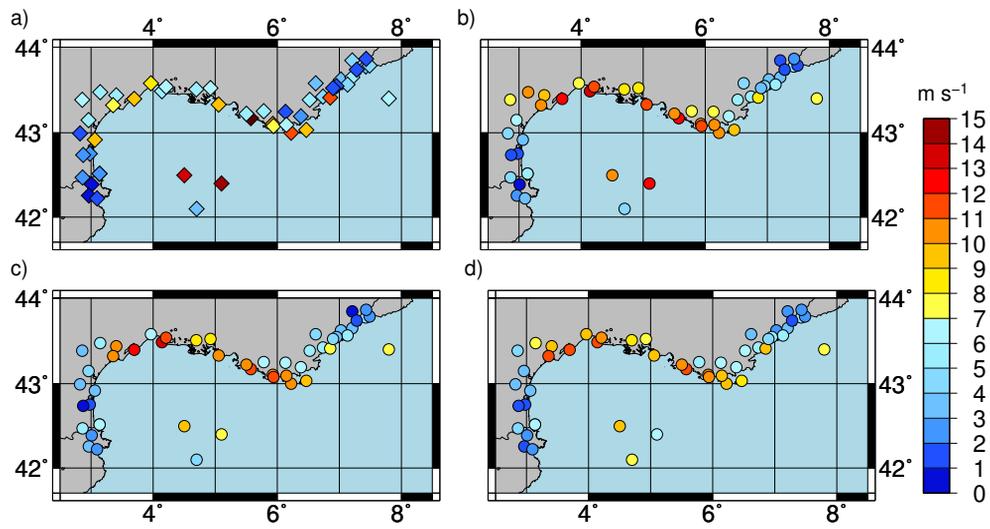


Figure 12. 10-m wind speed observed (a) and simulated by the NOWAV (b), WAV (c) and WAM (d) configurations at 09 UTC along the French coast.

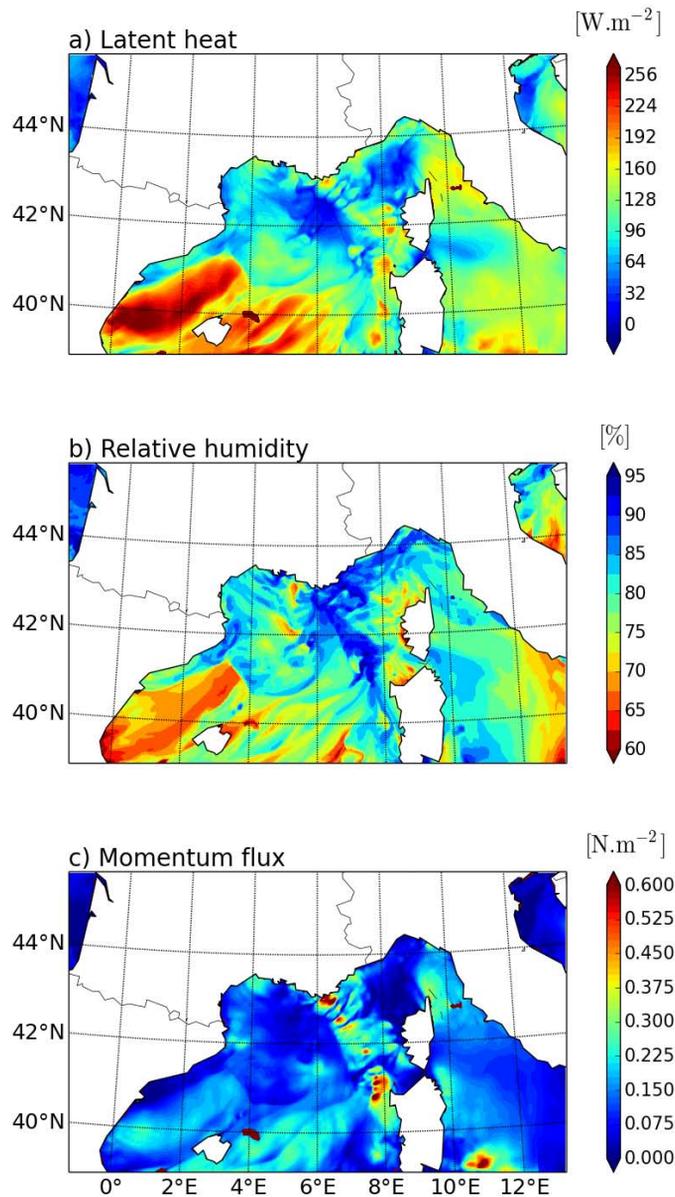


Figure 13. Latent heat flux (a), relative humidity (b) and momentum flux (c) of the NOWAV simulation at 12 UTC on 26 October 2012.

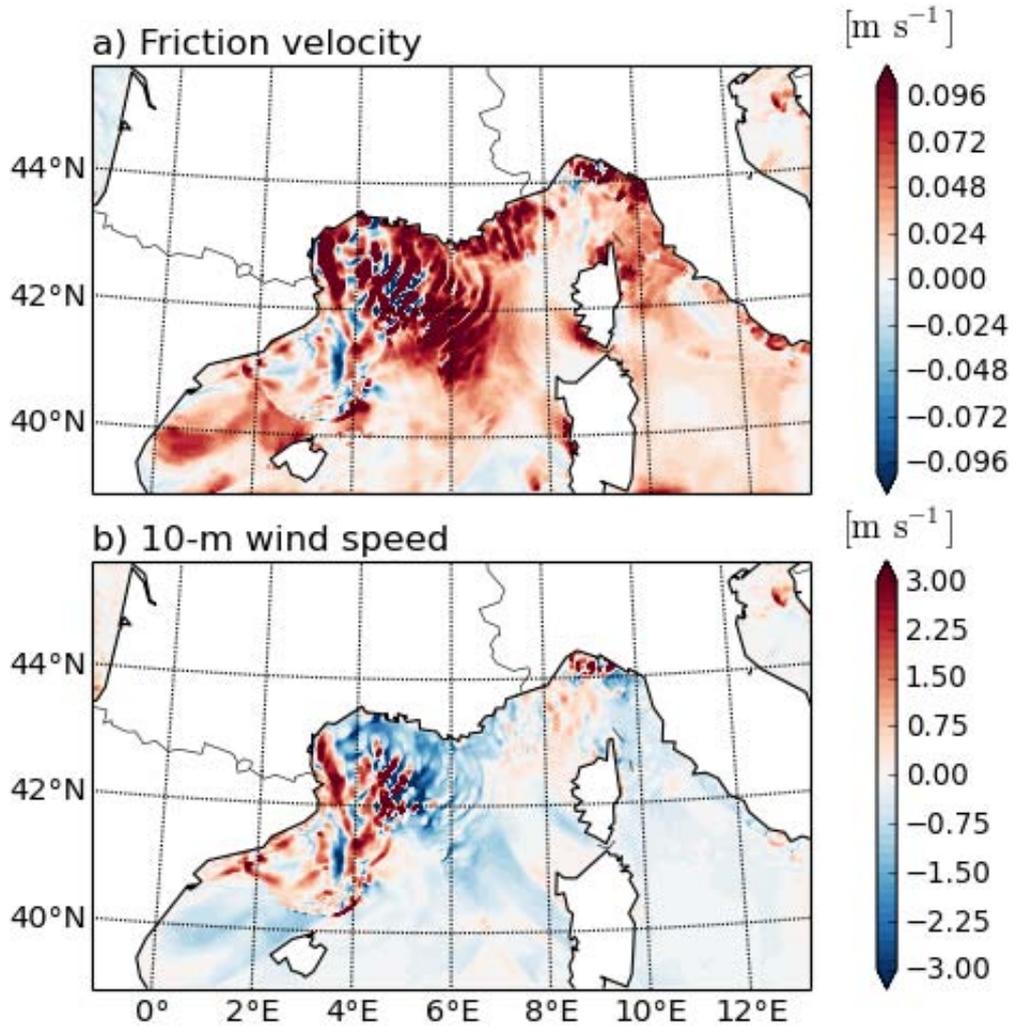


Figure 14. Difference between WAM and NOWAV at 08 UTC for (a) the friction velocity (m s^{-1}) and (b) the 10-m wind speed (m s^{-1}).

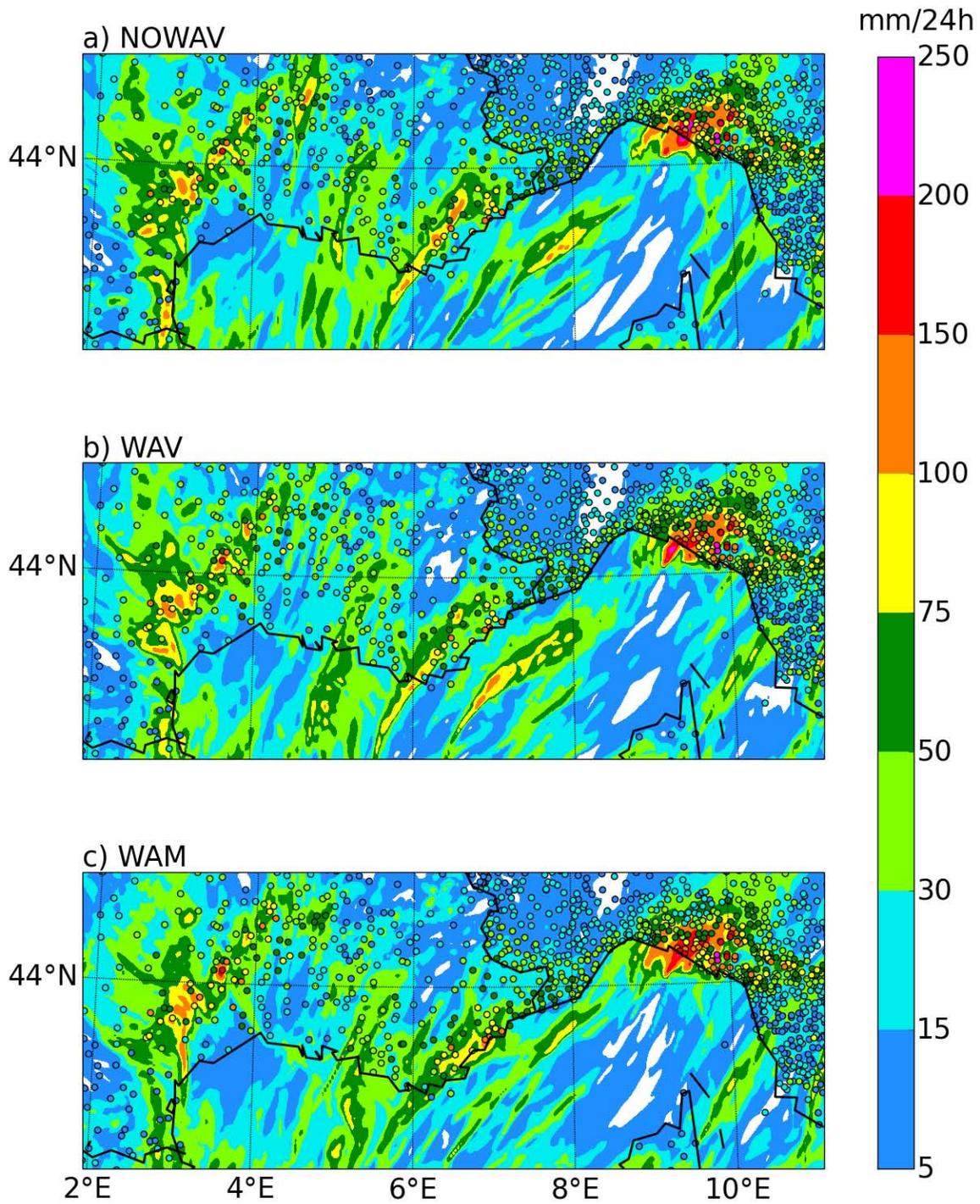


Figure 15. 24-h accumulated rainfall (mm) on 27 October 2012 at 00 UTC from (a) NOWAV, (b) WAV and (c) WAM. Coloured bullets are for the 24-h cumulative rainfall from rain gauge observations.