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Marie-Noëlle Bouin, Jean-Luc Redelsperger, Cindy Lebeaupin Brossier

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Processes leading to deep convection and sensitivity to sea-state representation during HyMeX IOP8 heavy precipitation event

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Review

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4 **Processes leading to deep convection and sensitivity to sea-state**
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7 **representation during HyMeX IOP8 heavy precipitation event**
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10 M.-N. Bouin^{1,2}, J.-L. Redelsperger², C. Lebeaupin Brossier¹

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12
13 ¹CNRM, UMR 3589, Météo-France & CNRS, Toulouse, France
14
15

16 ²CNRS, Ifremer, IRD, UBO /Laboratoire d'Océanographie Physique et Spatiale (LOPS), UMR
17
18 6523, IUEM, Plouzané, France.
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ABSTRACT

9 During the first observation period of the HyMeX programme, the Mediterranean coasts of Spain
10 were impacted by several heavy precipitating events (HPEs). The most damaging one occurred
11 during IOP 8 resulting in cumulative rainfall amount over 180 mm in the area of Murcia-Valencia.
12 Numerical simulations using a high-resolution atmospheric model provide a very realistic
13 representation of the mesoscale convective systems (MCSs) at the origin of this HPE and of the
14 associated low-level conditions, consisting in two cold sectors surrounding a warm sector. This
15 study provides a detailed analysis of the mechanisms of deep convection initiation and maintenance
16 between 1200 UTC on 28 September and 0000 UTC on 29 September 2012. On the coastal
17 mountainous area, the conditionally unstable inflow feeding the MCS is uplifted by the relief
18 whereas at sea, a strong low-level convergence plays the same role. At the coast, cold pools are
19 generated and strengthened by a strong low-level jet (LLJ) carrying cold dense air parcels from the
20 Gulf of Lion and by evaporation and cooling under the precipitating systems. These cold pools play
21 a key role in triggering the deep convection, either by directly uplifting the air masses or by
22 deflecting horizontally the inflow. They largely control the localisation and distribution of the heavy

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4 23 precipitation at sea near Valencia. A weak barrier wind over the cold pools and a secondary cyclonic
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6 24 circulation result in a bending of the convergence line at sea, in agreement with radar observations.
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9 25 A sensitivity study to a more realistic representation of the sea state in the air-sea exchanges shows
10
11 26 that the LLJ is decelerated by the increased sea-surface roughness, resulting in cold pools extending
12
13 27 further at sea and shifting the precipitation patterns 50 km offshore.
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16 28 **Keywords:** *heavy precipitation event; HyMeX; mesoscale convective system; sea state;*
17
18 29 *Mediterranean Sea; air-sea exchanges; cold pool; deep convection*
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21 30 **1. Introduction**

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25 31 Heavy precipitation events (HPEs) affect regularly the western Mediterranean coastal regions in
26
27 32 autumn and result often in flash flooding and landslides susceptible to produce thousands of
28
29 33 millions euros damages and even casualties (e.g. Ricard *et al.*, 2012; Llasat *et al.*, 2013). Daily
30
31 34 rainfall cumulative amounts higher than 150 mm are frequently observed as generated by multi-cell
32
33 35 quasi-stationary mesoscale convective systems (MCSs; e.g. Ducrocq *et al.*, 2008; Buzzi *et al.*,
34
35 36 2014). These MCSs generally develop on the eastern edge of an upper-level trough extending from
36
37 37 the British Isles to off the Iberian Peninsula associated with potential vorticity anomaly likely to
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39 38 generate low-level instability (Nuissier *et al.*, 2011; Duffourg *et al.*, 2016). A major ingredient
40
41 39 common to all HPEs is the conditionally unstable moist marine flow extracting energy from the
42
43 40 western Mediterranean sea surface, which is at its warmest in autumn and acts as a heat and
44
45 41 moisture reservoir. This flow is directed towards the Spanish, Italian or French mountainous coastal
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47 42 regions where it undergoes continuously renewed lifting: the same local mechanisms enable
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49 43 uplifting at the same place. This lifting marks the onset of deep convection and precipitation on the
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51 44 coastal areas, and is a key ingredient of quasi stationarity. The complexity of the mountainous
52
53 45 terrain and coastal topography has been demonstrated to be a factor influencing the orographic
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4 46 triggering of the convection, along with the characteristics of the low-level inflow (Miglietta and
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6 47 Rotunno, 2010; 2012; Bresson *et al.*, 2012). The precise role of the relief in determining the
7
8 48 location and distribution of the heavy precipitation has been recently demonstrated on the
9
10 49 Mediterranean islands (Corsica and Sardinia; Barthlott and Kirshbaum, 2013) and on the alpine
11
12 50 foothills on the northern Adriatic coastal area (Di Muzio, 2014; Davolio *et al.*, 2016). Low-level
13
14 51 convergence upstream of the relief or over the sea is another mechanism triggering deep convection
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16 52 (Ducrocq *et al.*, 2008; Duffourg *et al.*, 2016).

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20 53 Besides the direct orographic forcing by the elevated terrain behind the coasts and the low-
21
22 54 level moisture convergence at sea, the formation of a low-level cold pool is another lifting
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24 55 mechanism susceptible to favour deep convection. Cold pools are usually generated by rainfall
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26 56 evaporation in subsaturated air masses around 1.5 km above sea level (asl - Ducrocq *et al.*, 2008;
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28 57 Bresson *et al.*, 2012). The resulting cold air parcels propagate downwards and spread to form cold
29
30 58 pools. This cold pool, which is denser than the incoming low-level flow, acts like a relief by
31
32 59 blocking and uplifting the moist and warm flow, resulting in a shift of the deep convection
33
34 60 upstream. This cooling mechanism and the resulting cold pool has been demonstrated by a
35
36 61 sensitivity study to be the main mesoscale ingredient for the quasi-stationarity of a MCS that
37
38 62 occurred in September 2002 over the Gard plains rather than over the Massif Central foothills
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40 63 (Ducrocq *et al.*, 2008). Cold pools may interact or compete with orographic lifting to determine the
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42 64 positioning of heavy precipitation depending upon the characteristics of the incoming conditionally
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44 65 unstable low-level flow. Air lifting and triggering of deep convection by cold pools are especially
45
46 66 efficient when the incoming flow is relatively dry or weak (relative humidity around 85 % and low-
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48 67 level wind around 15 m s^{-1} , Bresson *et al.*, 2012). Besides direct lifting of the air masses, cold pools
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50 68 may also deflect horizontally the incoming flow and favour or reinforce the low-level convergence
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52 69 (Duffourg *et al.*, 2016).

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4 70 Better understanding these competing or interacting processes at the origin of deep
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6 71 convection and heavy precipitation on the Mediterranean coasts in autumn is part of the objectives
7
8 72 of the HyMeX (HYdrological cycle in the Mediterranean EXperiment) programme. HyMeX is an
9
10 73 international programme dedicated to the study of the hydrological cycle and related processes in
11
12 74 the Mediterranean (Drobinski *et al.*, 2014). The first special observation period (SOP1) in autumn
13
14 75 2012 aimed at monitoring and studying these heavy precipitating events over the Mediterranean
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16 76 coastal regions (Ducrocq *et al.*, 2014). Several Intensive Operation Periods (IOPs) focused on the
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18 77 observation of the MCSs at the origin of these high-impact events and on investigating the
19
20 78 associated processes.
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25 79 IOP8 on 28 September 2012 led to heavy precipitation on the Spanish Mediterranean coasts
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27 80 (more than 200 mm (24h)⁻¹ in Andalusia, 190 mm (24h)⁻¹ in the Murcia region and 140 mm (24h)⁻¹
28
29 81 in Valencia) and was the most damaging event observed during the SOP1 (Jansà *et al.*, 2014;
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31 82 Ducrocq *et al.*, 2016). The chronology of the event and an overview of the associated mechanisms
32
33 83 have been the subject of previous studies. Using high-resolution numerical modelling, Röhner *et al.*
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35 84 (2016) identified orographic forcing as the main mechanism for the triggering of deep convection
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37 85 over the coastal mountainous area, and strong low-level convergence as the mechanism at the origin
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39 86 of deep convection at sea. Using backwards trajectories helped them to identify heat and moisture
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41 87 local source for the conditionally unstable air masses in the Western Mediterranean basin for the last
42
43 88 and more damaging part of the event. Khodayar *et al.* (2016) evaluated the adequacy of
44
45 89 observational networks to provide information on the HPE. Describing the associated atmospheric
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47 90 circulation and stratification at broad scale, they identified three different convergence mechanisms
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49 91 leading to the development of deep convection over land, at the coast or at sea. These studies
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51 92 showed that the representation of IOP8 using high-resolution numerical modelling is good, and that
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53 93 its low-level dynamics is reasonably represented by the observations, making IOP8 a suitable case
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4 94 for process study and sensitivity tests. However, they provide a rather general view of the events. In
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6 95 particular, none of these studies investigate in details the processes likely responsible for triggering
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8 96 deep convection along the coasts or on the mountainous foothills and their interplay. Especially, the
9
10 97 presence of cold pools and their possible role in triggering deep convection are not included in the
11
12 98 previous IOP8 studies.

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15 99 Consequently, the first objective of this study is to provide a detailed analysis of the
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17 100 processes at the origin of the heavy precipitation that impacted the Murcia and Valencia region
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19 101 between 1200 UTC on 28 September 2012 and 0000 UTC on 29 September 2012. The study area
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21 102 and associated timing were chosen because they are susceptible to involve most of the uplifting
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23 103 processes described above. It constitutes a good test bed for investigating the interaction of these
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25 104 mechanisms initiating and maintaining deep convection on the mountainous range, over sea and on
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27 105 the coast. The effect of the waves generated by the wind (hereafter wind sea) on the low-level flow
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29 106 is usually implicitly represented in atmospheric models through the Charnock parameter and
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31 107 roughness length (see Eq.1 in section 3.1). As shown by a previous study on the HyMeX SOP1
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33 108 IOP16a (Thévenot *et al.* 2016), a more realistic representation of the wind sea may have a
34
35 109 significant impact on the turbulent air-sea exchanges and in the low-level flow. These authors
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37 110 obtained a slowing down of the low-level flow feeding the MCSs, resulting in a shift of the
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39 111 precipitation pattern towards sea, in better agreement with the observations. The second objective of
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41 112 the present study is thus to assess the impact of a better representation of the wind sea on the sea-
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43 113 surface roughness, momentum and heat transfer between the ocean and the atmosphere, low-level
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45 114 wind field and precipitation.

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52 115 This paper is organized as follows. Section 2 provides an overview of the IOP8 with the
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54 116 synoptic context leading to the development of the HPE, the chronology of the events and a
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56 117 summary of the atmospheric conditions. In section 3, the numerical simulations performed on this
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4 118 case study are presented together with their evaluation using satellite and field campaign
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6 119 observations. The outputs of the reference simulation are analysed in section 4 to provide a detailed
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8 120 description of the processes at the origin of deep convection and heavy precipitation on the Murcia-
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10 121 Valencia region. The impact of the representation of the sea state in the parameterization of the air-
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12 122 sea exchanges on these processes and on the precipitation is presented in section 5. Section 6
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14 123 concludes with a summary of the results.

18 124 **2. IOP8 overview**

21 125 *2.1. Synoptic context*

24 126 The synoptic situation of the HyMeX IOP8 is characterized by an upper-level cut-off low over
25
26 127 southern Portugal at 0000 UTC on 28 September 2012, progressing eastwards and reaching eastern
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28 128 Spain at 0000 UTC on 29 September 2012 (see Fig. 1a for the ARPEGE - Action de Recherche
29
30 129 Petite Echelle Grande Echelle - analysis at 0000 UTC on 28 September, Fig. 1b for the synoptic
31
32 130 analysis at 1800 UTC). The associated warm and cold surface fronts progressed northwards over
33
34 131 the southwestern Mediterranean basin between 0600 UTC on 28 September 2012 and 0000 UTC on
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36 132 29 September 2012.

40 133 *2.2 Chronology of the precipitating events*

43 134 The Spanish and Catalanian networks of rain gauges (MeteoCat) provide hourly observations of the
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45 135 precipitation amounts with a good geographical coverage (Fig. 2; see also Fig. 3b for the names of
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47 136 major cities and geographical locations). Heavy precipitation affected Andalusia in the morning of
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49 137 the 28 September between 0600 and 1200 UTC, with cumulated amount above 200 mm (Fig. 2a,
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51 138 2b), then progressed eastwards to the Murcia region between 1200 and 1800 UTC (Fig. 2c, 2d) with
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53 139 cumulated amount of 190 mm. The Valencia region was also affected by heavy rain between 1800
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55 140 and 2400 UTC on the same day (Fig. 2e, 2f).

141 *2.3 Mesoscale convective systems and low level conditions*

142 The multi-cell MCS responsible for the heavy rainfall on the Murcia and Valencia region originally
143 developed around 1100 UTC over Almeria and progressed northeastwards before splitting into
144 several systems around 1600 UTC. The near-surface wind field, as estimated by the ASCAT
145 (Advanced SCATterometer, EUMETSAT) scatterometer onboard the Metop-A satellite between
146 2100 UTC on 28 September and 0000 UTC on 29 September shows three different low-level flows
147 (Fig. 3a). North of the Mallorca island, a flow initiates in the Gulf of Lion with moderate-to-fairly-
148 strong northerly-to-northeasterly winds converging with a slightly stronger northeasterly flow north
149 of the Balearic Islands. This moderate low-level flow corresponds to the cold sector ahead of the
150 warm front (hereafter ahead cold sector; Fig. 1b) and is blowing from the right side to the left side
151 of the front. It reinforces locally to strong wind and is to some extent channeled between the
152 Balearic Islands and the coast of Catalonia, before reaching the coasts of the Valencia region. South
153 of Ibiza and between North Africa and the coasts of southeastern Spain, a southwesterly moderate
154 flow characterizes the cold sector behind the cold front (hereafter rear cold sector). Between these
155 two regions, the warm sector is characterized by a light-to-moderate easterly-to-northeasterly low-
156 level flow. This low-level flow distribution results in a strong surface convergence line located
157 along the warm front between the warm and moist easterly flow on the warm sector and the rapid
158 southwesterly flow between southern Spain and North Africa. In the following, we focus on the
159 Murcia-Valencia region and we investigate the processes at the origin of deep convection
160 (especially the possible role of cold pools) using the modelling tools that are presented in the next
161 part.

162 3. Meso-NH simulations

163 3.1. Meso-NH model and set-up

164 The numerical simulations were performed with the non hydrostatic numerical research model
165 Meso-NH (Lafore *et al.*, 1998). The Meso-NH model was run over a $1125 \times 1125 \text{ km}^2$ domain
166 covering the western Mediterranean basin from Gibraltar to the west of Sardinia (Fig. 3b). It
167 encompasses the precipitating systems during their whole life-cycle as well as the different marine
168 low-level moisture-supplying flows. The horizontal grid has a 2.5-km resolution, and the vertical
169 grid is defined with 55 stretched vertical levels (Gal-Chen and Somerville, 1975) up to 20 km, with
170 18 levels in the lowest 1500 m above the surface. The model resolution and associated physical
171 parameterization package are the same as those used in previous studies of HPEs using Meso-NH
172 (e.g. Nuissier *et al.*, 2008, Thévenot *et al.*, 2016, Duffourg *et al.*, 2016).

173 The prognostic variables of the model are the three components of the wind, the dry potential
174 temperature, the turbulent kinetic energy and the mixing ratios of the water vapour and of five
175 different classes of hydrometeors (cloudwater, rainwater, primary ice, snow aggregates, and
176 graupel). The evolution of the water species are governed by a bulk microphysical scheme (Caniaux
177 *et al.*, 1994; Pinty and Jabouille, 1998). The parameterization of the turbulence is based on a 1.5-
178 order closure (Cuxart *et al.*, 2000) with the diagnostic mixing length option following Bougeault
179 and Lacarrère (1989). Thanks to its high horizontal resolution, the atmospheric deep convection is
180 explicitly solved by the model. The surface conditions and the air–surface exchanges are governed
181 by the SURFEX surface model (Masson *et al.*, 2013). The sea-surface turbulent fluxes are
182 parameterized using the COARE 3.0 formula (Fairall *et al.*, 2003). The sea-surface roughness is
183 related to the friction velocity u_* using the Charnock coefficient α_{ch} :

$$z_0 = \alpha_{ch} \frac{u_*^2}{g} + 0.11 \frac{\nu}{u_*} \quad (1),$$

where g is the gravitational acceleration and ν the kinematic viscosity.

Simulations start on 28 September 2012 at 0000 UTC and last 24 h. They are initialised and driven at their lateral boundaries every 3 h from the high-resolution AROME-WMED analyses (Application of Research to Operations at Mesoscale – Western Mediterranean; Fourrié *et al.*, 2015). The sea-surface temperature (SST) field comes from the initial AROME-WMED analysis, which is built with the 2D Optimal Interpolation of *in situ* measurements CANARI (Code d'Analyse Nécessaire à Arpege pour ses Rejets et son Initialisation; Taillefer, 2002) blended with the Operational Sea Surface Temperature Ice Analysis (OSTIA; Donlon *et al.*, 2012). This SST field remains constant during the 24-h integration.

3.2 Sensitivity to sea state representation

In order to test the sensitivity of the boundary-layer processes and of the HPEs to the sea state, two simulations were performed with and without explicitly representing the wave impact into the surface turbulent fluxes parameterization. The methodology used here follows closely the one used by Thévenot *et al.* (2016) with the use of the sea state analyses taken from the run of the Wavewatch III® wave model (Ardhuin *et al.*, 2010; Tolman, 2014) in the Previmer project framework (<http://www.previmer.org>) as a surface forcing of the Meso-NH/SURFEX model.

In the experiment with no wave impact (NOWAV hereafter, used throughout this study as a reference simulation), the COARE 3.0 algorithm is used with the formulation of the Charnock coefficient from Hare *et al.* (1999), *i.e.* α_{ch} is set to 0.011 for 10-m wind speed below 10 m s^{-1} , then increases linearly up to 0.018 at 18 m s^{-1} , and remains constant for larger wind speeds. In the experiment with explicit wave representation (WW3_F hereafter), the formulation of Oost *et al.*

206 (2002) is used with the wave peak period provided by the analysis every 3 h:

$$207 \quad \alpha_{ch} = 50 (c_p / u_*)^{-2.5} \quad (2),$$

208 where c_p is the wave phase velocity. The Wavewatch III® wave parameters have been compared
209 with two Météo-France buoy observations in the Gulf of Lion and Azur sites. Mean biases are of
210 0.16 and 0.36 m for the significant wave height (root mean square 0.25 m) and of 1.0 and 0.9 s for
211 the peak period (root mean square 0.6 and 0.5 s), respectively. This level of agreement is
212 comparable to what has been obtained in a previous study (Thévenot *et al.*, 2016) and makes us
213 confident in using these analysed parameters as a forcing representative of the actual sea state over
214 the western Mediterranean basin.

215 **4 Validation of the reference simulation**

216 The results of the NOWAV reference simulation were compared to observations in order to assess
217 the skill of the simulation in representing the processes and the chronology of the events. All the
218 observations used in this work were collected and have been made available in the HyMeX project
219 framework.

220 **4.1 Precipitation**

221 The observations used in this part are taken from the Spanish rain gauge network including the
222 dense network of METEOCAT. The amount and location of the precipitation simulated in NOWAV
223 display a good match with the rain gauge 3-h cumulative precipitation from 0600 UTC on 28
224 September 2012 to 0000 UTC on 29 September 2012 (Fig. 2a to 2f). The maximum cumulative
225 amount observed by the rain gauges available for this study is 204 mm (24h)⁻¹ over the whole
226 domain, that compares favourably with the maximum predicted by the NOWAV simulation of
227 212 mm (24h)⁻¹. To assess more precisely the skill of our two simulations, scores against rain gauge

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4 228 observations for the 24-h period starting at 0000 UTC on 28 September over the whole simulation
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6 229 domain have been computed. The simulated daily rainfall amounts are extracted at the closest grid
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8 230 point to the 1233 rain gauge stations. The mean bias, the standard deviation of the difference
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10 231 (SDD), and the correlation coefficient (r) are given in Table 1, as well as two categorical scores
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12 232 widely used for assessing skills of precipitation forecast: the Equitable Threat Score (ETS; Schaefer,
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14 233 1990) and the Hanssen and Kuipers discriminant (HK; Hanssen and Kuipers, 1965). Both scores
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16 234 assess the ability of the model to detect rainfall amounts above a given threshold (see Table 1) and
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18 235 give values between -1 and 1. The thresholds used here correspond roughly to the 4-, 5- and 6-
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20 236 quantiles in the observations. ETS excludes the probability of detecting an event by chance, values
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22 237 equal to (resp. below) zero indicates that model skills are equal to (resp. below) a random forecast.
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24 238 HK measures the accuracy of the forecast both for events and non events, HK=0 corresponds to a
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26 239 constant forecast and HK=-1 to zero hits both for events and non-events forecasts.
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31 240 The scores obtained here are high, showing that the representation of this event by our two
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33 241 simulations is much better than random or constant forecast and performs very well in predicting
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35 242 both events and non events.
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39 243 ***4.2 Low-level flow and stability***

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41 244 The wind field at the first vertical level of the model over sea is compared with the wind field
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43 245 obtained from the ASCAT scatterometer onboard Metop-A (Fig. 3) and shows a good agreement in
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45 246 intensity and direction. Especially, the near-surface convergence line corresponding to the front
46
47 247 between Ibiza and Mallorca is well represented in the simulation, as well as the lower intensity of
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49 248 the low-level wind in the warm sector area. Here and in the rest of the study the cold and warm
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51 249 fronts are determined using the 2-m temperature gradients. The occluded front is determined using
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53 250 the low-level wind convergence.
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4 251 The radiosoundings performed in Palma de Mallorca and Barcelona on 28 September at
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6 252 1800 UTC and on 29 September at 0000 UTC are used to assess the skills of the NOWAV
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8 253 simulation in reproducing the wind field at different levels and the air masses stability profiles from
9
10 254 the equivalent potential temperature (θ_e , Fig. 4). These radiosoundings were assimilated in the
11
12 255 AROME-WMED analyses used as large-scale forcing of our simulations (Fourrié *et al.*, 2015).
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14 256 Their impact on the NOWAV simulation is probably weak, since they occur late in the simulation
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16 257 and far from the domain boundaries. Observed and simulated radiosounding profiles generally
17
18 258 agree well on the θ_e and wind values. In Barcelona, the simulated wind speed is systematically
19
20 259 lower than in observations especially at upper level, though the agreement on the direction is very
21
22 260 good. In particular, the veering of the wind at 1800 UTC from east-north-east at the surface to south
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24 261 at 5000 m asl and south-west at upper level that can be observed on the two stations is well
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26 262 reproduced by the simulation. The tropopause height is close to 12 km at both locations and times.
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31 263 At 1800 UTC, the Palma sounding is still located in the cold sector ahead of the incoming
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33 264 warm front (Fig. 4a). The ground equivalent potential temperature is 50-55°C in the simulation,
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35 265 increasing up to 60°C at 1000 m asl, then decreasing to 50°C at 3000 m asl. The low-level wind is
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37 266 northeasterly, close to 12 m s⁻¹, weakening to 5 m s⁻¹ at 2000 m asl. At 0000 UTC on 29 September
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39 267 (Fig. 4b) Palma is now in the warm sector as shown by the increase of the surface θ_e up to 55-65 °C
40
41 268 in the simulation. The simulated potential temperature at low level is generally higher than in
42
43 269 observations though their profiles up to 3000 m asl are very similar. This warm and moist low-level
44
45 270 layer goes up to 2000 m asl with some instability remaining up to 4000 m asl in the simulated
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47 271 profile. The low-level wind is weak (up to 4000 m asl) in the warm sector, as already noticed
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49 272 (section 2.3 and Fig. 3).
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54 273 The two Barcelona profiles exhibit a similar evolution between 1800 UTC and 0000 UTC
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56 274 (Fig. 4c and d), with equivalent potential temperatures close to 40°C at the surface. Barcelona
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4 275 profiles are in the ahead cold sector at both time periods, with the occluded front approaching from
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6 276 the south on 29 September at 0000 UTC. At both times the wind is weak from east close to the
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8 277 ground, about 15 m s^{-1} at 1000 m asl, and from south to southeast at upper level. At 1800 UTC, the
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10 278 low- θ_e air masses close to the ground transported from the sea by the northeasterly low-level flow
11
12 279 are topped by a thick layer of warmer, stable air (θ_e close to 55°C) from 700 to 3000 m asl. At
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14 280 0000 UTC on 29 September, the low- θ_e layer below 700 m asl is still present with a sharper vertical
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16 281 gradient (likely maintained by the low-level cold and moist easterly flow) but the warmer air layer
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18 282 located just above is now thinner with θ_e close to 50°C up from 2000 m asl. This can be explained
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20 283 by a strong and homogeneous flow observed in the simulation outputs between 2000 and 4800 m
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22 284 asl transporting drier air with θ_e below 48°C from the south of the western Mediterranean basin
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26 285 (not shown).

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29 286 In summary, the reference simulation succeeds in reproducing the stability (θ_e) variations
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31 287 observed in the radiosoundings. The profiles of wind direction and wind speed at first order is also
32
33 288 well reproduced by the model (except in Barcelona for the wind speed). We consider in the
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35 289 following that this simulation can be confidently used to investigate the fine-scale processes leading
36
37 290 to deep convection and heavy precipitation in the Murcia-Valencia region.

40 291 **5. Process study**

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44 292 In this part, the reference simulation outputs are used to study the different processes at play on the
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46 293 eastern Spanish coast between Murcia and Valencia between the 28 September 2012, 1600 UTC
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48 294 (corresponding to the start of convective precipitation in the Murcia region) and the 29 September
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50 295 2012, 0000 UTC.

51 52 53 296 **5.1 Conditional instability, moisture convergence and low-level jet**

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56 297 As seen in the previous part, the low-level flow over the western Mediterranean on the afternoon of
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4 298 the 28 September is organized in three main areas: the ahead cold sector on the Balearic Sea and the
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6 299 Gulf of Lion; the warm sector from the Balearic Islands to Murcia and the coasts of North Africa;
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8 300 and the rear cold sector on the Alboran Sea up to Murcia. This organization is well reproduced by
9
10 301 the NOWAV simulation with a strong low-level convergence line (Fig. 5a) along the surface cold
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12 302 and occluded fronts. The equivalent potential temperature at the first level of the model (Fig. 5b)
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14 303 shows value above 64°C in the warm sector with maximum values above 70°C on a large area north
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16 304 of the Algerian coasts. This warm sector is limited by very sharp θ_e gradients: to the south by a low-
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18 305 level rather uniform southerly flow bringing warm and much drier air from the African continent
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20 306 and to the southwest by the cold front with the rear cold sector with equivalent potential
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22 307 temperature below 58°C. On its north edge, the θ_e gradient corresponding to the warm front is much
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24 308 smoother. However, θ_e decrease to 54°C on a large area of the Gulf of Lion and the northern
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26 309 Balearic Sea. Over Spain θ_e is lower, confirming that there is no local feeding or development of
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28 310 convective systems on land away from the coasts (Röhner *et al.*, 2016). The Convective Available
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30 311 Potential Energy (CAPE) at the same time (Fig. 5c) shows values above 1000 J kg⁻¹ and up to more
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32 312 than 2000 J kg⁻¹ on a large area within the warm sector, ahead of the low-level flow close to the
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34 313 cold front. This high-CAPE area is limited to the north by the ahead cold sector and to the south by
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36 314 the low-level still warm but much drier southerly flow from North Africa. Its southwestern
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38 315 boundary corresponds to the cold front off Murcia at that time, where strong moisture convergence
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40 316 occurs. The Convective Inhibition (CIN, Fig. 5d) is relatively high in the main part of the warm
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42 317 sector, especially north of the North African coast due to the warm and dry southerly flow. By
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44 318 contrast, it shows low values in the westernmost part of the warm sector in the lee of the warm and
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46 319 moist flow. This area at the southwestern edge of the warm sector – with high CAPE and low CIN
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48 320 values – is thus the area the most prone to feed deep convection processes.
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56 321 The simulated radar reflectivities show values above 40 dBz corresponding to convective
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4 322 precipitation at two different places (Fig. 5b to d, green contour): along the occluded front the low-
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6 323 level moisture convergence above $3 \cdot 10^{-5} \text{ kg m}^{-3} \text{ s}^{-1}$ (Fig. 5a, green contour); and on land close to
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8 324 Murcia, probably as a result of stationarity of a previous convective system. At sea, the mechanism
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10 325 responsible for the convective precipitation is clearly the strong low-level convergence that remains
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12 326 present along the occluded front and on the triple point between 1500 UTC on 28 September and
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15 327 0000 UTC on 29 September.

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18 328 In the ahead cold sector, the 10-m northeasterly wind of 12 to 15 m s^{-1} reinforces to 16 to
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20 329 20 m s^{-1} at 500 m asl (Fig. 5a, 6a). At 1500 m asl , by contrast, the NOWAV simulated wind field is
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22 330 a more uniform southeasterly flow around 12 m s^{-1} on the ahead cold sector and part of the warm
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24 331 sector (Fig. 6b). The ahead cold sector flow corresponds therefore to a strong low-level jet (LLJ) .
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26 332 Note that this LLJ is clearly visible in the observed and simulated radiosounding profiles in Palma
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28 333 de Mallorca at 1800 UTC and in Barcelona at 1800 UTC and at 0000 UTC (Fig. 4). Oppositely to
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30 334 what has been frequently observed in HPE studies (e.g. Ducrocq et al., 2008), this LLJ is not
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32 335 located in the warm sector corresponding to the conditionally unstable feeding of the convective
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34 336 systems but in a cold sector with θ_e at low level between 48 and 60°C . It corresponds likely to a
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36 337 low-level jet associated with the low-level pressure gradient ahead of the warm front (e.g. Shapiro
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38 338 and Keyser, 1990).

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43 339 This low-level flow is not saturated in humidity as shown by the surface latent heat flux
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45 340 values up the 500 W m^{-2} on a large area at 1600 UTC on 28 September (Fig. 7a - corresponding
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47 341 relative humidity values between 70 and 80%). The ahead cold sector is the only place in the
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49 342 western basin where such values are observed at that time (except very locally under the convective
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51 343 systems, probably in link with wind gusts). This is due to the conjunction of this strong LLJ with
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53 344 dry air masses over the Gulf of Lion and with SST above 23°C all over the Balearic Sea as seen in
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55 345 the OSTIA analysis (not shown). By contrast, the SST in the Gulf of Lion is much colder (18 to
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4 346 20°C) due to an episode of Mistral-Tramontane in the previous days; this results in weak
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6 347 evaporation. As a result of this very strong evaporation on the Balearic Sea, the mixing ratio in the
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8 348 LLJ ranges between 8 and 10 g kg⁻¹ over the main evaporative patch (Fig. 7b). It increases
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10 349 gradually when approaching the eastern Spanish coasts, up to 13 g kg⁻¹ (see also Fig. S1 and S2 in
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12 350 the Supporting Information). This rising of humidity from 8 g kg⁻¹ to more than 12 g kg⁻¹ was
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14 351 already seen along backward trajectories in the study of Röhner *et al.* (2016). These air masses
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16 352 coming from the Gulf of Lion have potential temperatures between 18 and 20°C. They warm up
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18 353 slightly over the northern Balearic Sea thanks to the strong turbulent heat transfer and approach the
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20 354 Spanish coasts with potential temperature close to 23°C (see Fig. S1 and S2).

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25 355 Cold and relatively dry air masses originating from the Gulf of Lion are transported by the
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27 356 LLJ to the Spanish coasts close to Valencia. They moisten and get slightly warmer due to turbulent
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29 357 exchanges with a warmer sea over the Balearic Sea.

30 31 32 358 **5.2 Initiation and maintenance processes**

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35 359 We now investigate in more details the processes responsible for the initiation and maintenance of
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37 360 the deep convection in the Murcia-Valencia region, between 1600 and 2300 UTC on 28 September.
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39 361 At this time period, various processes leading to heavy precipitation interact and compete with each
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41 362 other.

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44 363 The virtual potential temperature (θ_v) is directly related to the density of air masses and
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46 364 commonly used as a proxy for cold pools (e. g. Ducrocq *et al.*, 2008; Bresson *et al.*, 2012). In the
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48 365 following, we use θ_v values under 23°C to identify cold pools, in addition to θ_e characterizing the
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50 366 heat and moisture content of the inflow.

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54 367 At 1600 UTC, low values of θ_v are associated with the LLJ in the western part of the Balearic
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56 368 Sea. They correspond to the cold and moist air masses transported from the Gulf of Lion by the
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4 369 LLJ, as previously described. These air masses are sufficiently cold to make them dense in spite of
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6 370 their moisture and they accumulate on the foothills of the relief at the latitude of Valencia (Fig. 8a).
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8 371 The role of this LLJ in initiation of the cold pool on the coast is confirmed on the vertical cross
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10 372 section along this LLJ (Fig. 8c). Dense air masses with low θ_v are present up to 600 m asl and
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12 373 transported by this 15-to-18 m s^{-1} flow towards the coast. Convective precipitation with
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14 374 instantaneous rates above 20 mm h^{-1} are present southwest of Xabia on land and offshore (Fig. 8b).
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16 375 They correspond to strong vertical velocities at 500 m above ground level (agl) (Fig. 8a, white
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18 376 contour) and moisture convergence rates above $3 \cdot 10^{-5} \text{ kg m}^{-3} \text{ s}^{-1}$ at the surface (Fig. 8b, blue
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20 377 contour). Strong convergence at sea is thus the mechanism initiating the convection at that time,
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22 378 with both the LLJ in the ahead cold sector and the weaker flow in the warm sector contributing to
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24 379 the moisture feeding. South of Valencia, along the 2D profile (Fig. 8c) strong convective
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26 380 precipitation correspond to strong vertical velocities (white and black contours) and are located
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28 381 close to the maximum of the topography. The horizontal convergence on the relief slope (vertical
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30 382 blue arrows, Fig. 8c) is the mechanism responsible for the uplift and the corresponding deep
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32 383 convection at that place and time.
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38 384 At 1830 UTC (Fig. 9), the warm and cold fronts bounding the conditionally unstable air mass
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40 385 have moved northeastwards. On the southernmost part of the area, the limit of the convective
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42 386 precipitation with instantaneous rates above 20 mm h^{-1} matches exactly the northern edge of the
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44 387 cold pool (north of Murcia) which is also accurately collocated with strong uplift (Fig. 9a, white
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46 388 contour) and strong moisture convergence rates (Fig. 9b, blue contour). At sea, convective
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48 389 precipitation patches are located on the cold and occluded fronts, and corresponds also to strong
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50 390 convergence rates (Fig. 9b, blue contour). Along the coasts south of Valencia, very high
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52 391 instantaneous rain rates (50 mm h^{-1}) are obtained over the sea not associated with high convergence
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54 392 rates (Fig. 9b, 9c). However, their contour (see the instantaneous rates above 20 mm h^{-1} on Fig. 9b)
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4 393 matches the cold pool present along the coast at the latitude of Valencia, which is the result of both
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6 394 moisture feeding by the northeast LLJ and rain evaporation. Vertical cross sections exhibit two
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8 395 different mechanisms contributing to initiating the deep convection by uplift of the air masses. On
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10 396 the relief as precedently, strong uplift with vertical velocities above 5 m s^{-1} is obtained, resulting in
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12 397 instantaneous rain rate of 50 mm h^{-1} . Over the sea, close to the edge of the cold pool which has
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14 398 developed since 1600 UTC, vertical velocities above 1 m s^{-1} above 1500 m agl (white contour, Fig.
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16 399 9c) result in deep convection. The horizontal convergence (blue vertical arrows, Fig. 9c) is strong at
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18 400 the edge of the cold pool. As no convective cells are present at that place yet, the cold pool is very
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20 401 likely at the origin of the uplift here. On the cold pool itself, strong uplift (white contour) and strong
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22 402 convergence (blue arrows) are collocated with downdraughts with θ_e below 52°C and subsidence
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24 403 below -1 m s^{-1} (black contour), and strong precipitation. At this place, the active convective cells
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26 404 and strong precipitation seem to maintain (and enhance) the cold pool.
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31 405 At 2000 UTC, the main convergence zone along the occluded front is now located south of
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33 406 Valencia (Fig. 10a). The two areas of strong convective precipitation are still present on the relief,
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35 407 northwest of Xabia and south of Valencia. They are collocated with strong convergence rates (Fig.
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37 408 10a, white contour; Fig. 10b, blue contour) and high evaporation at low level on the relief and on
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39 409 the cold pool (not shown). Conversely, at sea off Valencia, strong precipitation rates are not
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41 410 collocated with strong moisture convergence (Fig. 10b), they rather correspond to the coldest patch
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43 411 of the cold pool at sea (Fig. 10a). The surface covered by the cold pool over the sea has slightly
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45 412 increased and the minimum θ_v on land has decreased down to 17°C . Strong uplifts (white contour,
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47 413 Fig. 10c) collocated with strong horizontal convergence (blue arrows, Fig. 10c) and strong
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49 414 precipitation (Fig. 10b) are located at the edge of the cold pool. Convergence and uplift at the edge
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51 415 of the cold pool are the processes resulting in deep convection here. The already active convection
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53 416 on the relief generates downdraughts (black contour, Fig. 10c) that in turn contribute to the
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4 417 maintenance of this cold pool on the relief foothills and at sea. On these cold pools at the latitude of
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6 418 Valencia, a deflection of the low-level wind from east-north-east to north-north-east corresponds to
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8 419 the initiation of a barrier wind (Fig. 10a, insert). Barrier winds have been observed and modelled in
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10 420 previous studies, mostly around the Adriatic Sea (Di Muzio, 2014, Davolio *et al.*, 2016). The
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12 421 altitude of the mountainous area around Valencia makes it comparable to what is obtained by Di
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14 422 Muzio (2014) with a medium-height orography.

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18 423 At 2130 UTC, the surface covered by the cold pools has further extended at sea, in particular
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20 424 off Xabia (Fig. 11a) and the main moisture convergence zone has reached the latitude of Valencia
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22 425 (Fig. 11b, blue contour, Fig. 11c). The very strong convergence zone offshore is still present with
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24 426 high precipitation rates (see Fig. 11c, blue arrows) in collocation with modelled uplift and
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26 427 downdraughts. On the relief at the latitude of Valencia, high precipitation rates are also collocated
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28 428 with the strong convergence line that progresses northwards with the precipitation patterns.
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30 429 Between these two areas, large patches of very strong precipitation (instantaneous rain rate up to
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32 430 130 mm h^{-1} , Fig. 11c) are obtained along the coast, at the latitude of Valencia, in very good
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34 431 collocation with the cold pool extending at sea. Clearly, low-level convergence (blue arrows Fig.
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36 432 11c) is the mechanism responsible for the uplift triggering the deep convection here. It is however
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38 433 difficult to attribute this convergence to the cold pools effect only, as convective cells are already
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40 434 present here. A secondary cyclonic circulation has formed at low level (Fig. 11a, insert), due to the
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42 435 deflection of the low-level wind by the cold-pool edge. It results in shifting the eastern edge of the
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44 436 cold pool southeastwards, and the patches of strong moisture convergence northwards (Fig. 11b).
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46 437 This low-level secondary circulation has already been observed on HPEs (Berthou *et al.*, 2016;
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48 438 Duffourg *et al.*, 2016) as a result of a barrier wind formation. It corresponds here to the minimum of
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50 439 sea-level pressure (see Fig. 12a) and results in a slight bending of the convergence zone at sea. In
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52 440 order to check whether these instantaneous high rain-rates at sea are also present in the
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4 441 observations, the radar reflectivities obtained in the NOWAV simulation at 2100 UTC are compared
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6 442 with the corresponding observations from the OPERA radar reflectivity Odyssey composite (Fig.
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8 443 13; the Odyssey product is not available at 2130 UTC; <http://www.knmi.nl/opera/>). The rain
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10 444 amounts are comparable between observations and simulation outputs, and the bending of the
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12 445 convection patterns due to the cold pool and associated to the secondary cyclonic circulation in the
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14 446 NOWAV simulation is also present in the observations.

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18 447 The mechanisms at the origin of heavy precipitation in the Murcia-Valencia region involve
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20 448 orographic forcing on the coastal mountainous range, very strong moisture low-level convergence
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22 449 at sea along the cold front and lifting of the moist LLJ by cold pools that developed on the relief
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24 450 foothills and propagated offshore. These cold pools induce a northerly-to-northwesterly weak
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26 451 barrier wind along the coast, and the setup of a low-level secondary cyclonic circulation resulting in
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28 452 the bending of the convective patterns. These processes play a key role in the location and
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30 453 distribution of heavy precipitation (rainfall cumulative amount above 140 mm in 6 hours) in the
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32 454 Valencia region. In particular, they explain the rainfall patterns at sea as observed in the radar
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34 455 reflectivity images (Fig. 13).

35 36 37 38 39 456 **6. Impact of wave representation**

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42 457 In this part, we discuss the impact of representing explicitly the waves in the WW3_F simulation
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44 458 through the forcing of the Meso-NH model at the surface using the 3-h Wavewatch III® analyses.
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46 459 The objective here is to test the sensitivity of the atmospheric response to the representation of the
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48 460 surface roughness. As shown by Thévenot *et al.* (2016) on another HPE case study, representing the
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50 461 surface roughness variability due to the wind sea can result in a slowing down of the low-level wind
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52 462 and in a shift of the precipitation place. As shown by the scores of the WW3_F simulation, which
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54 463 are not significantly different from those of NOWAV (Table 1), the overall representation of the
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4 464 high precipitation is comparable in both simulations.
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7 465 ***6.1 Instantaneous effect on the low-level flow***
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10 466 We first examine the wave parameters as simulated by the Wavewatch III® model at 1500 UTC on
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12 467 28 September, and their instantaneous impact on the oceanic and atmospheric surface parameters.
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14 468 The simulated significant wave heights are the largest (more than 2.5 m) on the Balearic Sea and
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16 469 south of the Balearic Islands along the coasts of Spain, between Murcia and Xabia (Fig. 14a). The
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18 470 resulting surface roughness difference as parameterized in COARE 3.0 depends mainly on the 10-m
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20 471 wind speed at sea and on the state of development of the wind sea (represented by the significant
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22 472 wave height). This difference is then the biggest in the ahead cold sector, where both strong winds
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24 473 at 10 m (Fig. 5a) and waves in equilibrium with the wind (Fig. 14a) are present. Roughness length
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26 474 differences are above $5 \cdot 10^{-4}$ m over a large area (Fig. 14b), which result in a difference on the drag
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28 475 coefficient C_d of $0.2 \cdot 10^{-3}$ over the whole ahead cold sector (Fig. 14c). Local strong increase of the
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30 476 roughness length (up to $8 \cdot 10^{-3}$ m) and of the drag coefficient is also observed more locally under the
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32 477 convective systems and in the rear cold sector. These general increases over the ahead cold sector
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34 478 result in a slowing down of the 10-m wind of more than 0.5 m s^{-1} over a very large area of the
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36 479 ahead cold sector (Fig. 14d). The part of the rear cold sector close to the front shows slowing down
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38 480 of the same magnitude and local patches of wind slowing down by more than 5 m s^{-1} are also
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40 481 observed under the convective systems, collocated with the surface roughness increase. These
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42 482 patches concern very limited areas and are collocated with strong wind and gusts under the
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44 483 convective clouds.
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50 484 This instantaneous impact of the waves on the low-level flow is different from what has
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52 485 been obtained on another case study (IOP16a) using the same methodology (Thévenot *et al.*, 2016)
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54 486 by two aspects. Firstly, the wind slowing down is weaker than what was obtained by these authors
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4 487 in average (0.5 m s^{-1} to be compared to 3 m s^{-1}), but concerns a larger area. Secondly, this slowing
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6 488 down concerns the two cold sectors, especially the ahead cold sector that contribute to the cold pool
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8 489 initiation and maintenance, but does not impact the warm sector and the feeding of the convective
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10 490 systems by conditionally unstable air masses.

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13 491 In atmospheric and oceanic surface conditions corresponding to HPEs, the turbulent heat
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15 492 transfer over the sea is generally dominated by the latent heat flux. This is the case here, with strong
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17 493 evaporative fluxes in a large area of the ahead cold sector (see Fig. 7a) between 1500 and
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19 494 1900 UTC, as well as locally in the rear cold sector. The differences on the latent heat flux due to
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21 495 the wave representation in the surface forcing of the model are however very weak: 5 W m^{-2}
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23 496 increase on average with mean values of the latent heat flux close to 200 W m^{-2} in the two cold
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25 497 sectors of this simulation. The effect on the low-level air masses participating in the feeding of the
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27 498 systems (θ_e , mixing ratio) is consequently not significant. As in the study of Thévenot *et al.* (2016),
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29 499 the mechanisms responsible for a possible change in the deep convection and associated heavy
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31 500 precipitation are to be found more likely in dynamical effects of the low-level flow than in the
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33 501 thermal exchanges.

34 35 36 37 38 502 **6.2 Impact on the precipitation field**

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41 503 We now compare the equivalent and virtual potential temperature fields at the first level of the
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43 504 model, at 2130 UTC in the region of Murcia-Valencia, in order to investigate the impact of the wave
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45 505 representation on the convective systems and precipitation. Few differences are observed in the
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47 506 low-level flow feeding the systems itself (Fig. 12; see also Fig. S3 in Supporting Information). The
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49 507 main differences concern the cold pools that extend further east at sea in the WW3_F simulation,
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51 508 especially at the latitude of Valencia. They also show colder θ_v on the coast north of Valencia. As a
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53 509 consequence of this cold pool shift, the precipitation field above 5 mm h^{-1} (instantaneous rain rate;
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4 510 see yellow contour in Fig. S3) is displaced eastwards. In every simulation, the patterns
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6 511 corresponding to 5 mm h^{-1} rain rate closely follow the edge of the cold pools (corresponding to θ_v
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8 512 under 23°C). This is due to the displacement of the moisture convergence line at the surface on the
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10 513 edge of the cold pool (Fig. 12).

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14 514 The secondary cyclonic circulation at the eastern edge of the cold pool, which is observed at
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16 515 2130 UTC in the NOWAV simulation (section 5.2, Fig. 11a and 12a), is weakened and displaced at
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18 516 sea in the WW3_F simulation (Fig. 12b). In both simulation outputs, the 10-m wind field is
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20 517 deflected and bended by the cold pool edge, resulting in the WW3_F simulation in a shift of this
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22 518 cyclonic circulation 30 km eastwards. The sea-level pressure field at 2130 UTC on 28 September
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24 519 2012 in the reference simulation NOWAV shows two minima close to 1000 hPa at the latitude of
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26 520 Valencia, corresponding to the well-organized secondary low-level circulation (Fig. 12a). In the
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28 521 WW3_F simulation, higher sea-level pressures up to 1012 hPa are observed near the coast, and the
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30 522 sea-level pressure minimum of 1001 hPa is displaced 30 km towards Ibiza, out of the cold pool
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32 523 (Fig. 12b). This weakening of the local sea-level pressure minimum has been observed in previous
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34 524 sensitivity studies and can be explained by the slowing down of the low-level wind in WW3_F.
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36 525 Relative higher pressures result, in turn, in an anticyclonic shift of the surface wind and of the
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38 526 displacement of the cyclonic circulation towards sea (Berthou *et al.*, 2016). A weak barrier wind
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40 527 effect is also observed on the cold pools in both simulation.
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45 528 As a result, the precipitation cumulated over 6 h from 1800 UTC on 28 September to
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47 529 0000 UTC on 29 September 2012 show a significant difference in both their amount and their
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49 530 location between WW3_F and NOWAV (Fig. 15). Taking into account the wave effect on the
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51 531 surface roughness results in rain amount more distributed along an east-west profile (Fig. 15b). The
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53 532 peak of precipitation on this 6-h time period over the whole simulation domain reaches 159 mm in
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55 533 the WW3_F simulation with respect to 258 mm in the NOWAV simulation. The total amount on the
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4 534 Valencia area is the same in both simulations with an average of 2.05 mm km^{-2} . As seen in Figure
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6 535 15a, the precipitation pattern in the WW3_F simulation results is displaced of 50 km towards sea.
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8 536 Note however that this displacement concern precipitation located mainly at sea and does not affect
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10 537 the forecast scores (Table 1) since they were computed using raingauges observations.
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13 538 This impact of a weakening of the low-level wind field of less than 1 m s^{-1} due to wave
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15 539 representation on the formation and maintenance of cold pools in the coastal domain can be
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17 540 explained by the results of previous studies. Cold pools are known to develop more easily and over
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19 541 larger areas with slower low-level winds (Bresson *et al.*, 2012; Davolio *et al.*, 2016). Lower wind
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21 542 speeds results in more time for the convective cells to develop and for the cold pools to strengthen
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23 543 themselves from evaporating precipitation (Bresson *et al.*, 2012). This relatively low but uniform
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25 544 weakening of the low-level wind induces a significant change on the cold pool extent and location,
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27 545 resulting in a corresponding displacement of the heavy precipitation patterns.
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32 546 **7. Summary and conclusion**

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35 547 This study is part of the HyMeX programme, the objectives of which include a better understanding
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37 548 of the mechanisms triggering the deep convection responsible for heavy precipitation, and the role
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39 549 of the air-sea exchanges in the case of Mediterranean HPEs. In the framework of this programme,
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41 550 the SOP1 IOP8 has been the subject of several studies, either using high-resolution simulations for
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43 551 describing the main mechanisms responsible for deep convection (Röhner *et al.*, 2016), or using
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45 552 observations for a general description of the event (Khodayar *et al.*, 2016). These authors do not
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47 553 provide a detailed study neither of the different mechanisms responsible for uplifting the air masses
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49 554 in a coastal area, leading to strong precipitation, nor of the role of the sea surface exchanges in the
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51 555 event.
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55 556 Here, we focused on the local processes leading to heavy precipitation in the Murcia-

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4 557 Valencia region between 1600 and 2300 UTC on the 28 September 2012, on their interaction, and
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6 558 on the role of the sea state representation in a model. The use of the non-hydrostatic, convection-
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8 559 permitting model Meso-NH without and with a more realistic representation of the sea surface
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10 560 roughness provides very accurate representation of the low-level flows and atmospheric stability
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12 561 patterns. In particular, the precipitation amounts, distribution and chronology from 1200 UTC on 28
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14 562 September to 0000 UTC on 29 September are well reproduced by both simulations. The
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16 563 precipitation scores of the reference simulation computed with different thresholds give also
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18 564 excellent results. We then consider these reference simulation outputs suitable for investigating the
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20 565 mechanisms at the origin of this HPE with maximum cumulative rain amount above 200 mm in
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22 566 24 h.
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27 567 Several mechanisms represented in the simulation outputs have been described in previous
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29 568 studies on the same HPE (Röhner *et al.*, 2016; Khodayar *et al.*, 2016). For instance, the warm and
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31 569 cold sector associated play a role in the moisture and heat feeding of the system when it moves
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33 570 northwards along the Mediterranean Spanish coast from Gibraltar to Catalonia. Low-level
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35 571 conditionally unstable air masses with very high CAPE and low CIN reach eastern Spain in a
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37 572 easterly weak-to-moderate flow. The two main mechanisms mentioned in previous studies,
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39 573 responsible for the lifting of these air masses and leading to the triggering of the deep convection -
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41 574 namely orographic forcing on land and strong low-level moisture convergence at sea - were
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43 575 assessed in our reference simulation. The present study reveals other mechanisms at the origin of
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45 576 the deep convection and heavy precipitation (see Fig. 16 for a synthetic scheme). The LLJ is
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47 577 observed here in the ahead cold sector rather than in the main warm low-level fuelling the MCS. It
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49 578 originates in the Gulf of Lion and transports very cold air masses, which barely warm up over the
50
51 579 warmer surface waters of the Balearic Sea. These cold thus dense air masses contribute to initiate
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53 580 cold pools on the relief foothills close to the sea. Rain evaporation in the subsaturated mid-level
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4 581 layer results in downdraughts further developing and maintaining these cold pools. They eventually
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6 582 act as triggering mechanisms of the deep convection when the air masses at their leading edge are
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8 583 thick and dense enough to uplift the incoming conditionally unstable air masses. A weak barrier
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10 584 wind develops from 2000 UTC on 28 September at low level over the cold pool at sea, leading to a
11
12 585 cyclonic deflection of the incoming low-level flow (Fig. 11a). A secondary cyclonic circulation
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14 586 results in a bending of the cold pool edge, and in a deflection of the strong low-level convergence
15
16 587 zones at sea and of the corresponding convective precipitation. Cold pools play here a key role in
17
18 588 the localization of the deep convection and associated heavy precipitation at sea off Valencia rather
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20 589 than on the relief foothills.
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24 590 Representing in a more realistic way the sea state using the 3-hourly Wavewatch III®
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26 591 analyses as a forcing has a significant impact on the simulation results. It first increases the surface
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28 592 roughness, resulting in an increase of the wind stress, thus in a slight decrease of the low-level
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30 593 wind. The main conditionally unstable low-level flow feeding the convective systems in warm and
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32 594 moist air is not impacted, but the moderate wind slowing down over a large part of the secondary
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34 595 flow in the ahead cold sector (low-level jet) is sufficient to affect significantly the location of the
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36 596 heavy precipitation at sea in the Valencia region. The cold pools are displaced further at sea,
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38 597 strengthened, and shift accordingly the place of strong convergence and convective precipitation.
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40 598 The sea-level pressure field is changed towards higher pressures and the low-level wind field is
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42 599 modified accordingly by an anticyclonic shift.
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47 600 Conversely, no impact of the wind sea on the heat exchanges at the sea surface has been
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49 601 observed in this study. The mean change of the latent heat flux of 5 W m^{-2} over the Balearic Sea
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51 602 does not change significantly the moisture content or equivalent potential temperature in the LLJ
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53 603 feeding the convective systems. This reinforces the similar results obtained in a previous study on
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55 604 the HyMeX SOP1 IOP16a (Thévenot *et al.*, 2016). Either developed wind sea concerns the area of
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4 605 warm and moist air directly feeding the HPE, as this was the case in IOP16a, and the low-level flow
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6 606 is already saturated in humidity; as a result, the latent heat transfer change due to waves is weak. Or
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8 607 it concerns an area not saturated in humidity yet as this is the case in IOP8, and this flow plays only
9
10 608 a secondary role in the HPE. This shows that the effect of the sea state on the heat exchanges in
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12 609 systems leading to HPEs is likely very small.

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15 610 This study of the sensitivity to wave impact uses an atmospheric model at high resolution
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17 611 forced by realistic modelled sea states. The conclusion of this part of our study would probably be
18
19 612 different if the simulation was carried out using a full coupling between the atmospheric and the
20
21 613 wave model.

22 23 24 25 614 **Acknowledgements**

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	Mean bias	SDD	Correlation r	ETS			HK		
<i>Threshold</i>				<i>5 mm</i>	<i>15 mm</i>	<i>25 mm</i>	<i>5 mm</i>	<i>15 mm</i>	<i>25 mm</i>
NOWAV	0.4	15.5	0.74	0.63	0.78	0.64	0.79	0.90	0.80
WW3_F	0.4	16.0	0.72	0.63	0.74	0.61	0.79	0.89	0.79

Table 1: Mean bias, standard deviation difference, ETS and HK scores (see text) of the NOWAV and WW3_F simulations against 24-h accumulated rain gauge observations on 28 September 2012 (mean bias and SDD in mm).

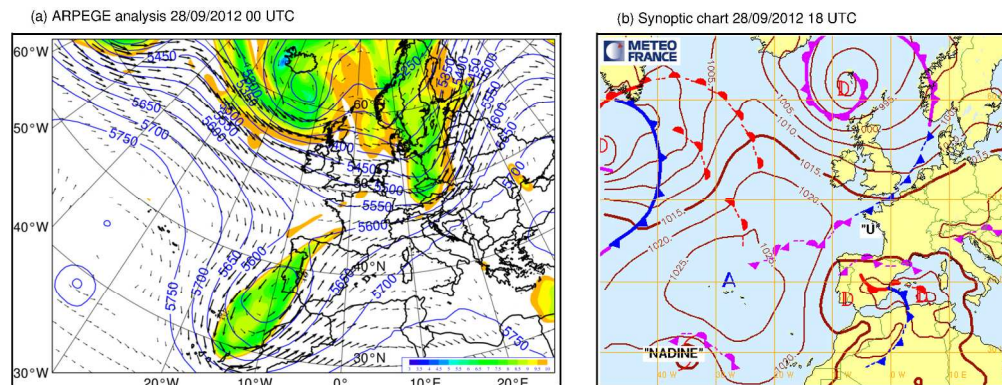


Figure 1: Synoptic conditions of the IOP8 with (a) ARPEGE analysis showing the height of potential vorticity anomaly equal to 2.0 PVU (colour chart, km), the 500 hPa geopotential (isobars, m) and the 300 hPa wind (above 10 m s^{-1}) at 0000 UTC on 28 September 2012 and (b) synoptic Meteo-France analysis with fronts and isobars, at 1800 UTC on 28 September 2012.

273x104mm (300 x 300 DPI)

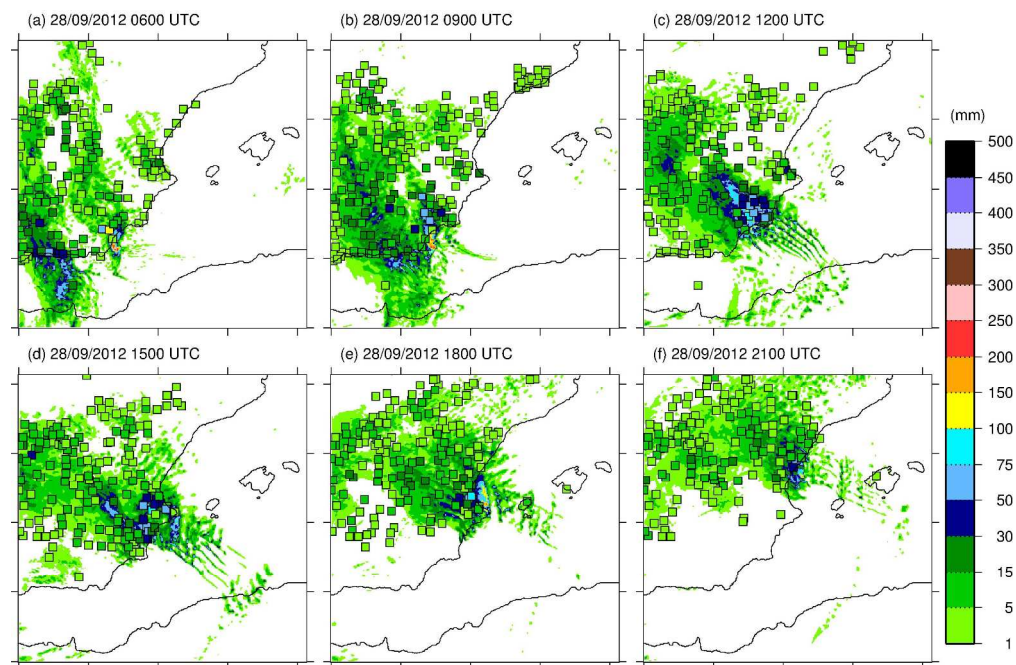


Figure 2: Three-hour cumulated rain amount from the NOWAV simulation (see Table 1 and sections 3.1 and 3.2) and measured by rain gauges (coloured squares) starting at (a) 0600 UTC, (b) 0900 UTC, (c) 1200 UTC, (d) 1500 UTC, (e) 1800 UTC and (f) 2100 UTC on 28 September 2012.

278x183mm (300 x 300 DPI)

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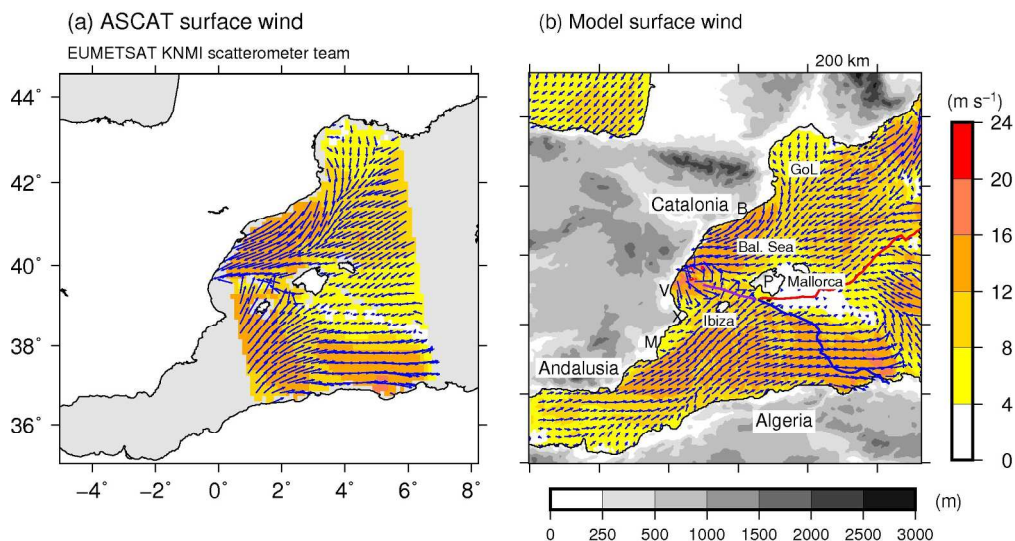


Figure 3: Sea surface wind (m s^{-1}) (a) from ASCAT onboard Metop-A between 2100 UTC on 28 September and 0000 UTC on 29 September and (b) from the NOWAV simulation (see sections 3.1 and 3.2) at 2230 UTC on 28 September 2012, with the surface fronts superimposed. V denotes Valencia, B Barcelona, M Murcia, X Xabia, and P Palma de Mallorca. The topography (grey scale, m) and main geographical areas are also indicated (b).

192x102mm (300 x 300 DPI)

Review

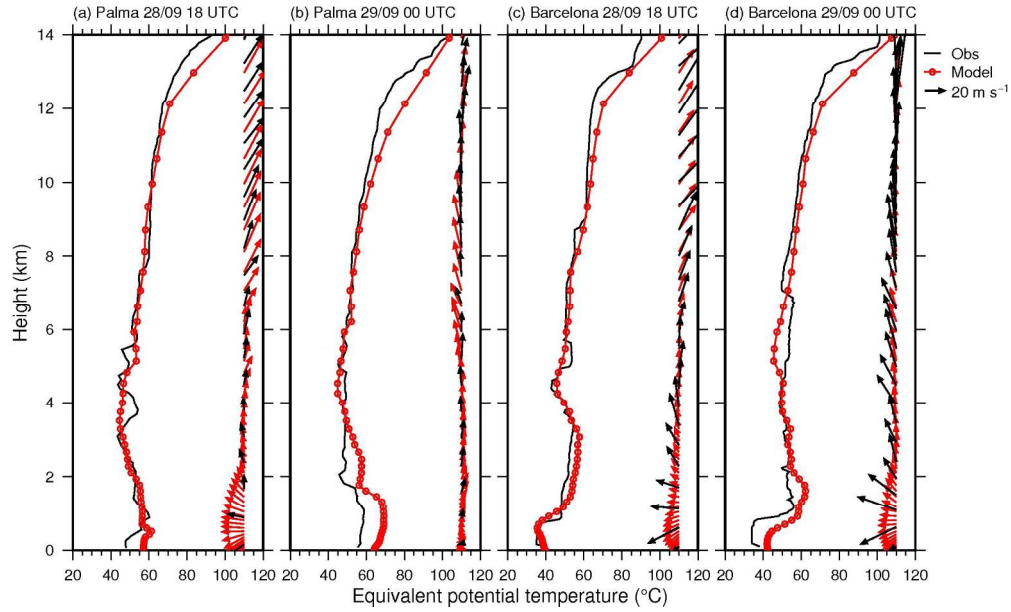


Figure 4: Vertical profiles (height in km) of equivalent potential temperature ($^{\circ}\text{C}$) and horizontal wind (vector) from radio soundings (black) and from NOWAV simulation (red) in Palma de Mallorca (a) at 1800 UTC on 28 September 2012 and (b) at 0000 UTC on 29 September, and in Barcelona (c) at 1800 UTC on 28 September and (d) at 0000 UTC on 29 September 2012.

184x111mm (300 x 300 DPI)

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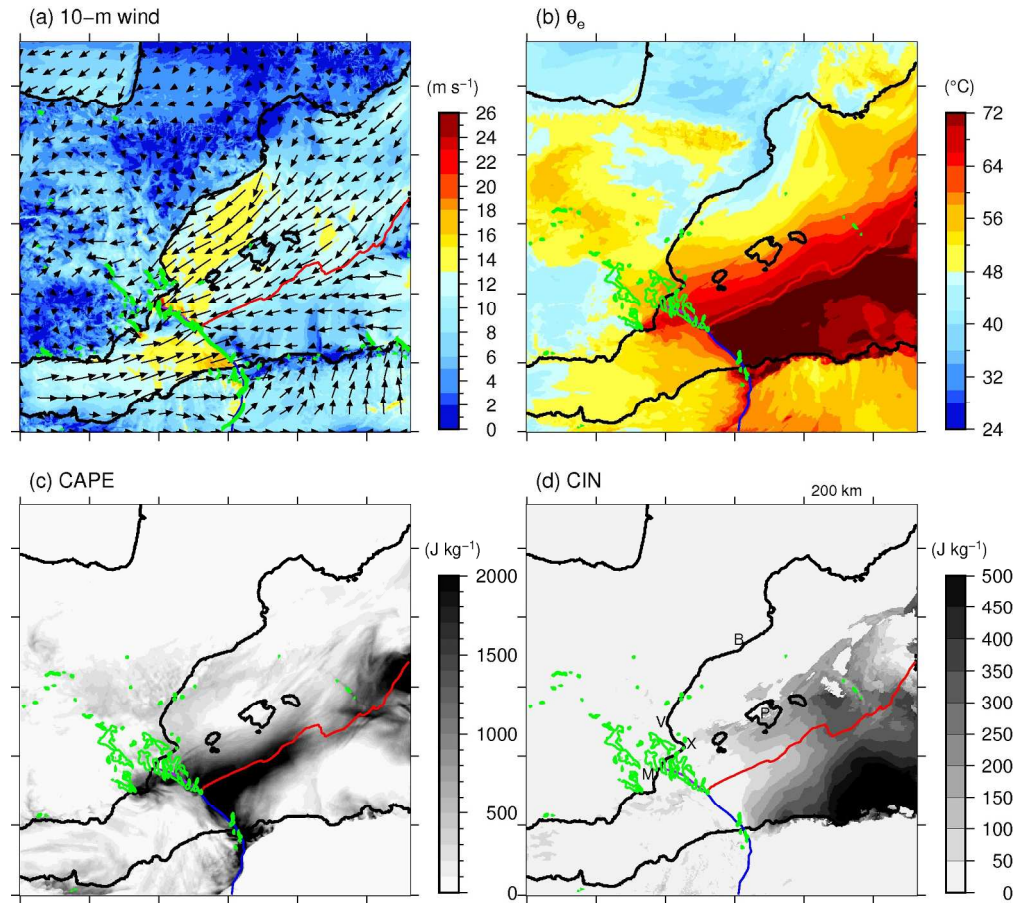


Figure 5: (a) 10-m wind (m s^{-1}), (b) equivalent potential temperature ($^{\circ}\text{C}$), (c) convective available potential energy (J kg^{-1}), and (d) convective inhibition (J kg^{-1}) from the NOWAV simulation, at 1600 UTC on 28 September 2012. The green contour in (a) indicates the moisture convergence rate above $3 \times 10^{-5} \text{ kg m}^{-3} \text{ s}^{-1}$ at the surface. The green contour in (b), (c), (d) indicates the simulated radar reflectivity at 2000 m above 40 dBz. The red (resp. blue) line indicates the warm (rep. cold) front, the purple line the occluded front. V denotes Valencia, B Barcelona, M Murcia, X Xabia, and P Palma de Mallorca.

205x185mm (300 x 300 DPI)

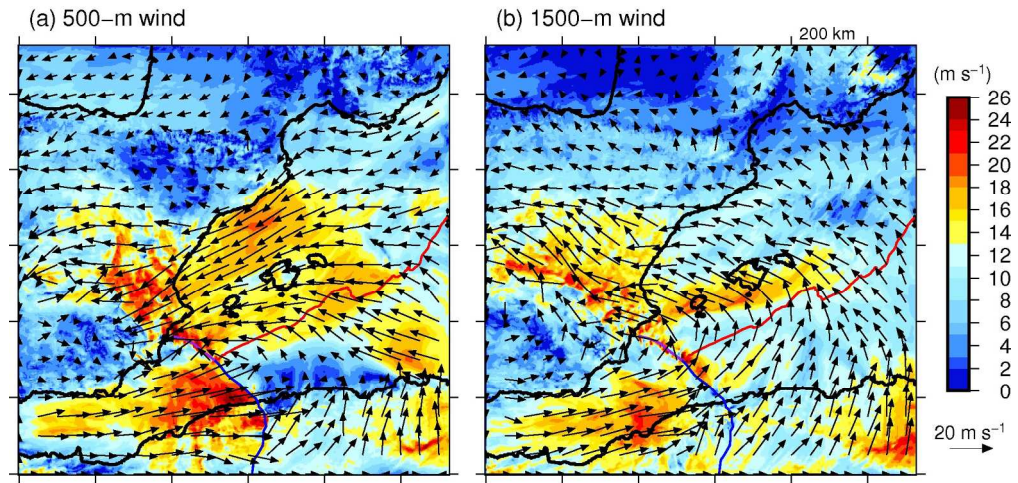


Figure 6: Wind field from the NOWAV simulation (m s^{-1}) at 1600 UTC on 28 September 2012, (a) at 500 m and (b) at 1500 m agl. The red (resp. blue) line indicates the warm (rep. cold) front, the purple line the occluded front.

186x90mm (300 x 300 DPI)

Peer Review

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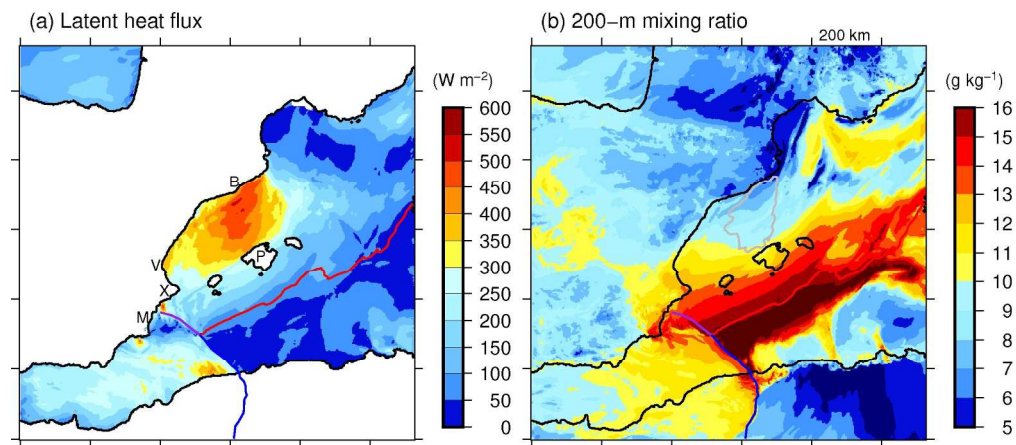


Figure 7: (a) Surface latent heat flux (W m^{-2}) and (b) mixing ratio at 200 m (g kg^{-1}) from the NOWAV simulation at 1600 UTC on 28 September 2012. The grey contour in (b) indicates the area with surface evaporation above 400 W m^{-2} . The red (resp. blue) line indicates the warm (rep. cold) front, the purple line the occluded front. V denotes Valencia, B Barcelona, M Murcia, X Xabia, and P Palma de Mallorca.

203x90mm (300 x 300 DPI)

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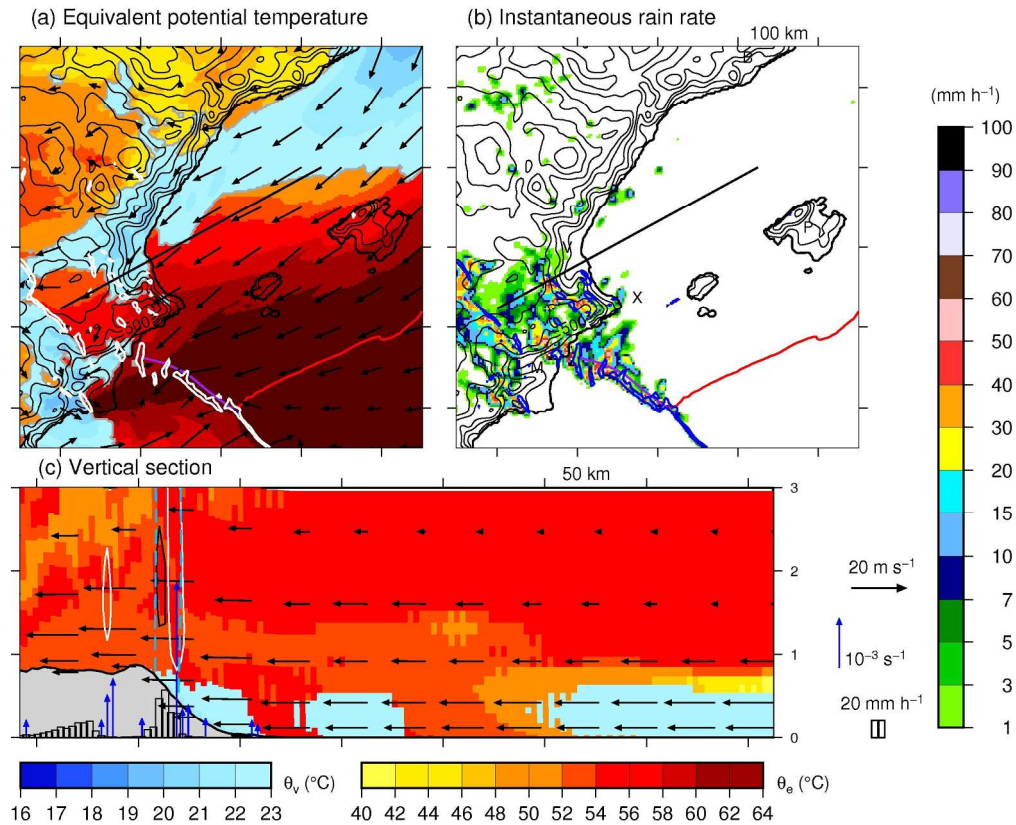


Figure 8: (a) Equivalent potential temperature and virtual potential temperature at the surface ($^{\circ}\text{C}$, colour scale), 10-m wind (vectors), and vertical velocities above 1 m s^{-1} at 500 m agl (white contour); (b) Instantaneous rain rate (mm h^{-1} , colour scale), and moisture convergence rate above $3 \times 10^{-5} \text{ kg m}^{-3} \text{ s}^{-1}$ at the surface (blue contour); (c) vertical profiles of the equivalent potential temperature and virtual potential temperature ($^{\circ}\text{C}$, colour scale ; height scale in km), projection of the wind (black arrows), instantaneous rain rate (vertical bars), vertical velocities above 1 m s^{-1} (white contour) or below -1 m s^{-1} (black contour), cloud precipitating water above 1 g kg^{-1} (light blue, dashed contour), and horizontal convergence at 120 m agl (blue arrows) from the NOWAV simulation at 1600 UTC on 28 September 2012. The black line in (a) indicates the vertical cross section used in (c). The virtual potential temperature in (a) and (c) replaces the equivalent potential temperature when under 23°C to indicate cold pools. The red (resp. blue) line in (a) and (b) indicates the warm (rep. cold) front, the purple line indicates the occluded front. V denotes Valencia, B Barcelona, M Murcia, X Xabia, and P Palma de Mallorca.

199x163mm (300 x 300 DPI)

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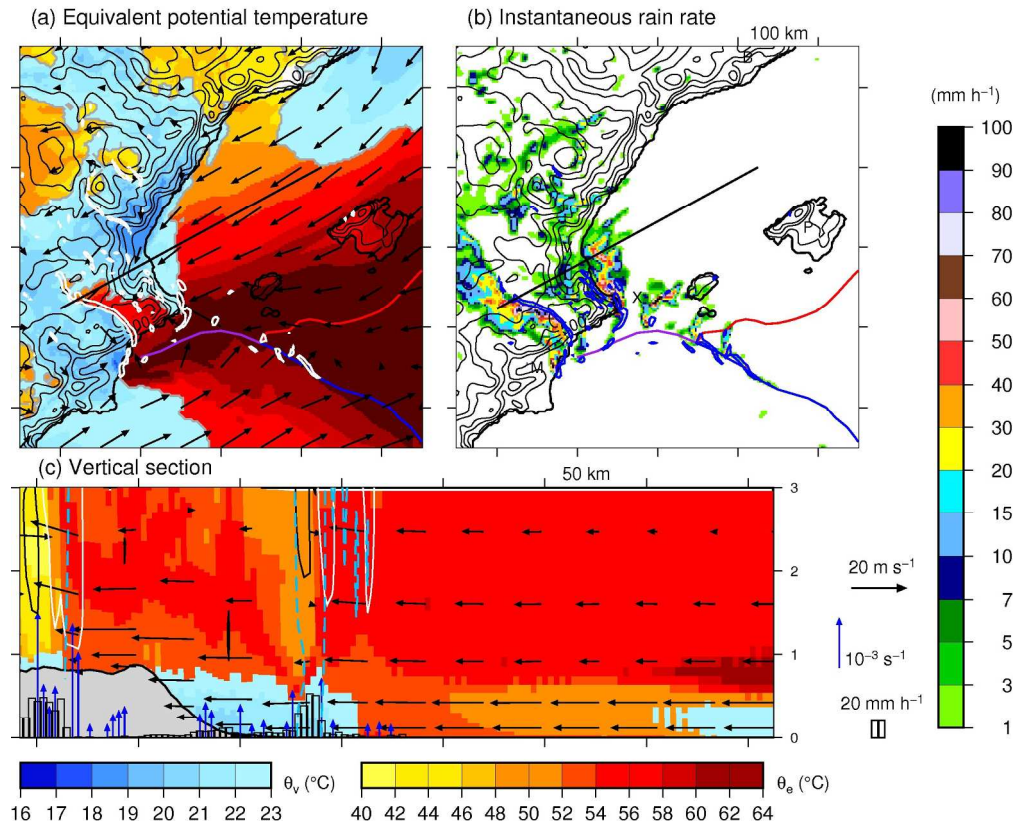


Figure 9: Same as Figure 8, but at 1830 UTC on 28 September 2012.

199x163mm (300 x 300 DPI)

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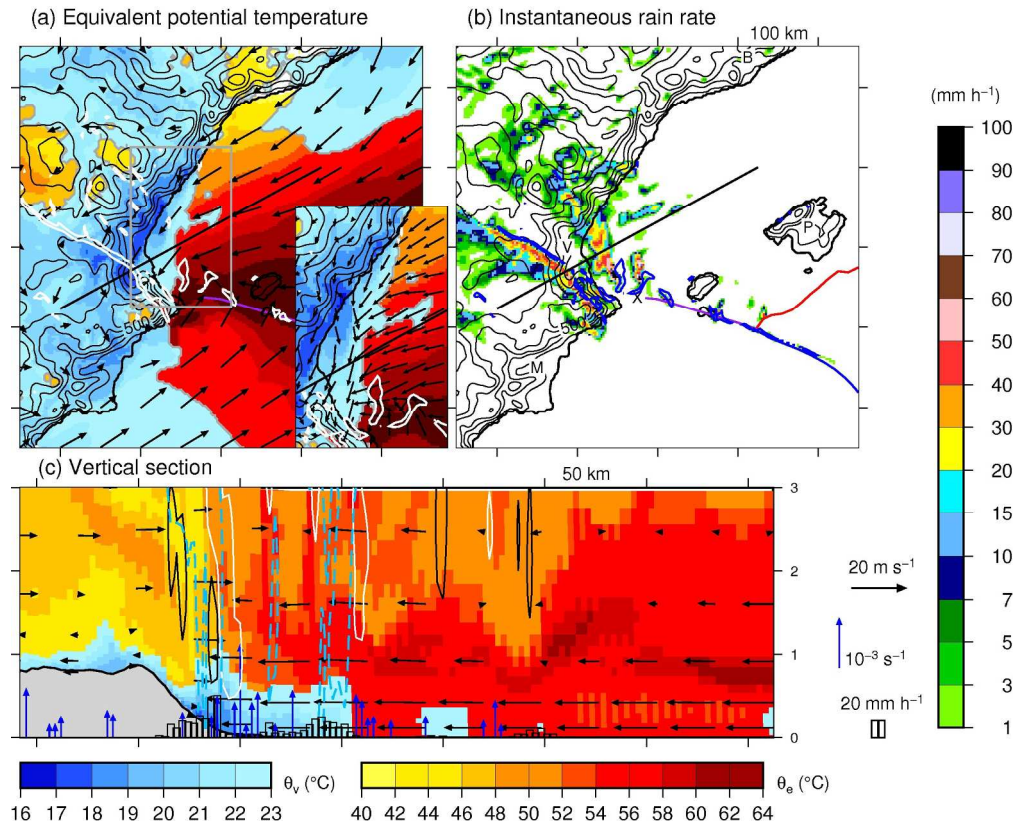


Figure 10: Same as Figure 8, but at 2000 UTC on 28 September 2012. The insert in (a) (grey frame) represents in more details the 10-m wind field on the coast at the level of Valencia.

199x163mm (300 x 300 DPI)

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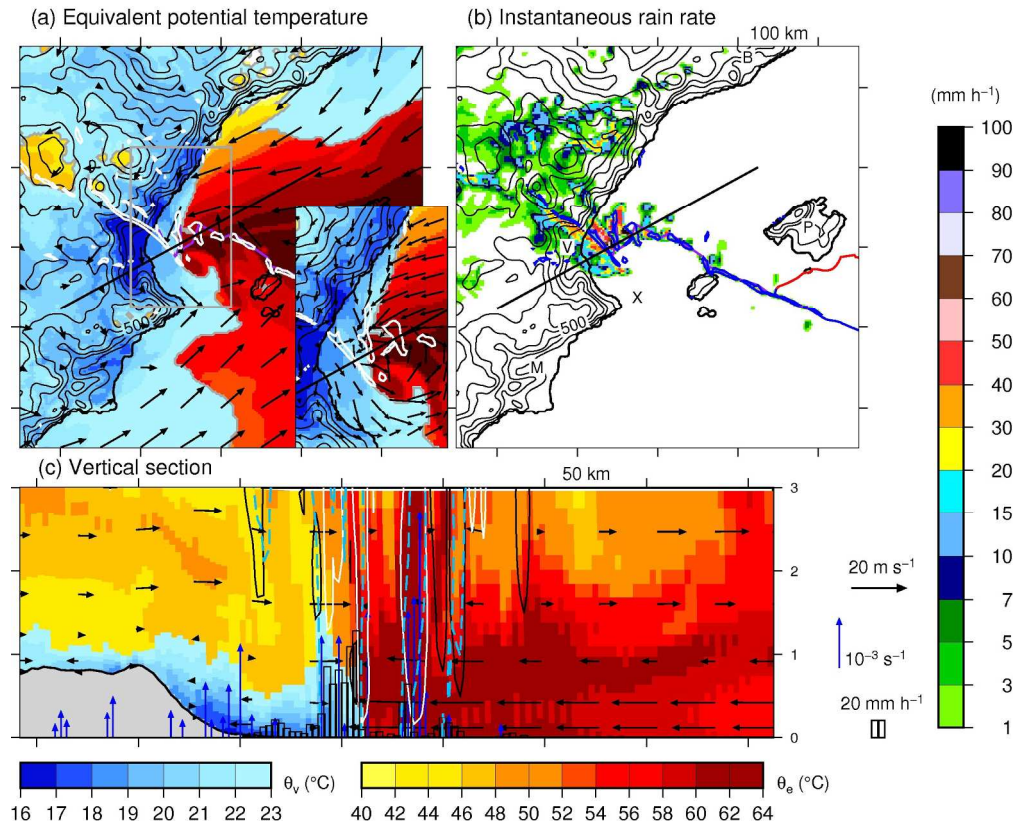


Figure 11: Same as Figure 10, but at 2130 UTC on 28 September 2012.

199x163mm (300 x 300 DPI)

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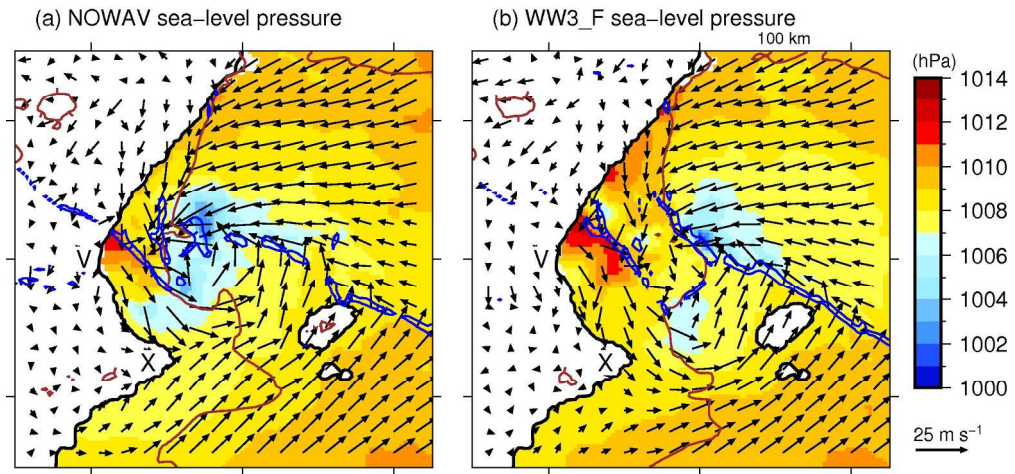
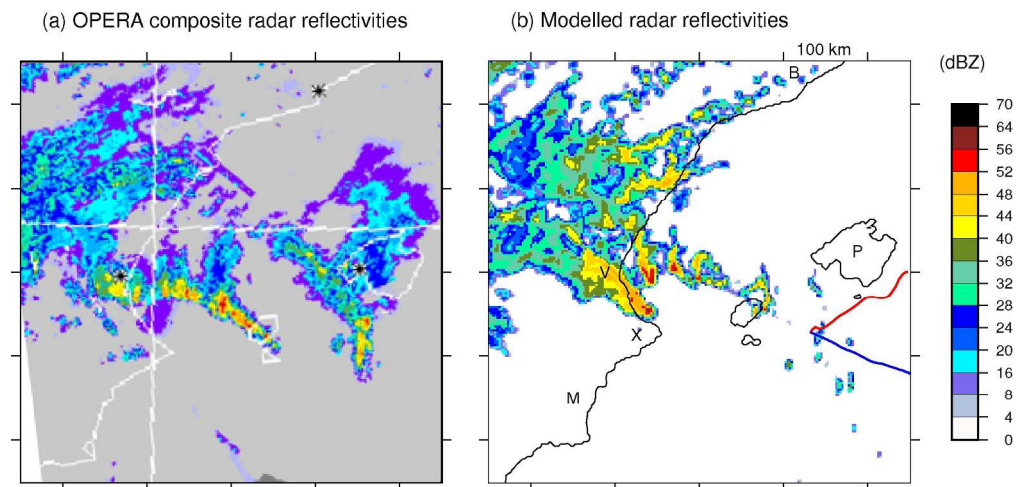


Figure 12: Sea-level pressure (colour scale, hPa), surface wind (black arrows), cold pool edges as identified by $\theta_v \leq 23^\circ\text{C}$ (brown contour), and areas of moisture convergence rate above $3 \times 10^{-5} \text{ kg m}^{-3} \text{ s}^{-1}$ at the surface (blue contour) from the (a) NOWAV and (b) WW3_F simulations at 2130 UTC, on 28 September 2012. V denotes Valencia, and X Xabia.

192x91mm (300 x 300 DPI)

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Figure 13: Radar reflectivities (dBZ) at 2000 m agl in the Valencia region at 2100 UTC on 28 September 2012 (a) from the Odyssey composite product, and (b) from the NOWAV simulation. The red (resp. blue) line indicates the warm (rep. cold) front. V denotes Valencia, B Barcelona, M Murcia, X Xabia, and P Palma de Mallorca.

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190x92mm (300 x 300 DPI)

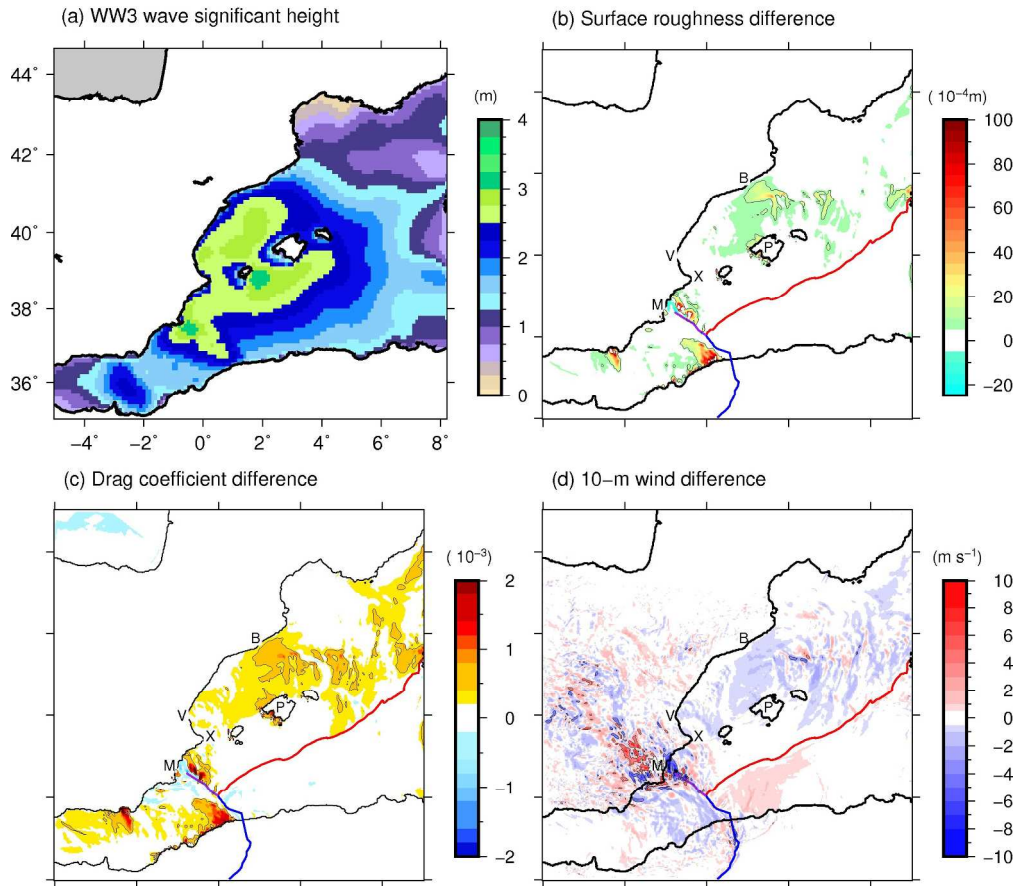


Figure 14: (a) Wave significant height from the Wavewatch III® model outputs (m), (b) surface roughness length difference (10^{-4} m), (c) drag coefficient difference (10^{-3}) and (d) 10-m wind speed difference (m s^{-1}) between the WW3_F and NOWAV simulations at 1500 UTC on 28 September 2012. The red (resp. blue) line indicates the warm (rep. cold) front, the purple line the occluded front. V denotes Valencia, B Barcelona, M Murcia, X Xabia, and P Palma de Mallorca.

217x192mm (300 x 300 DPI)



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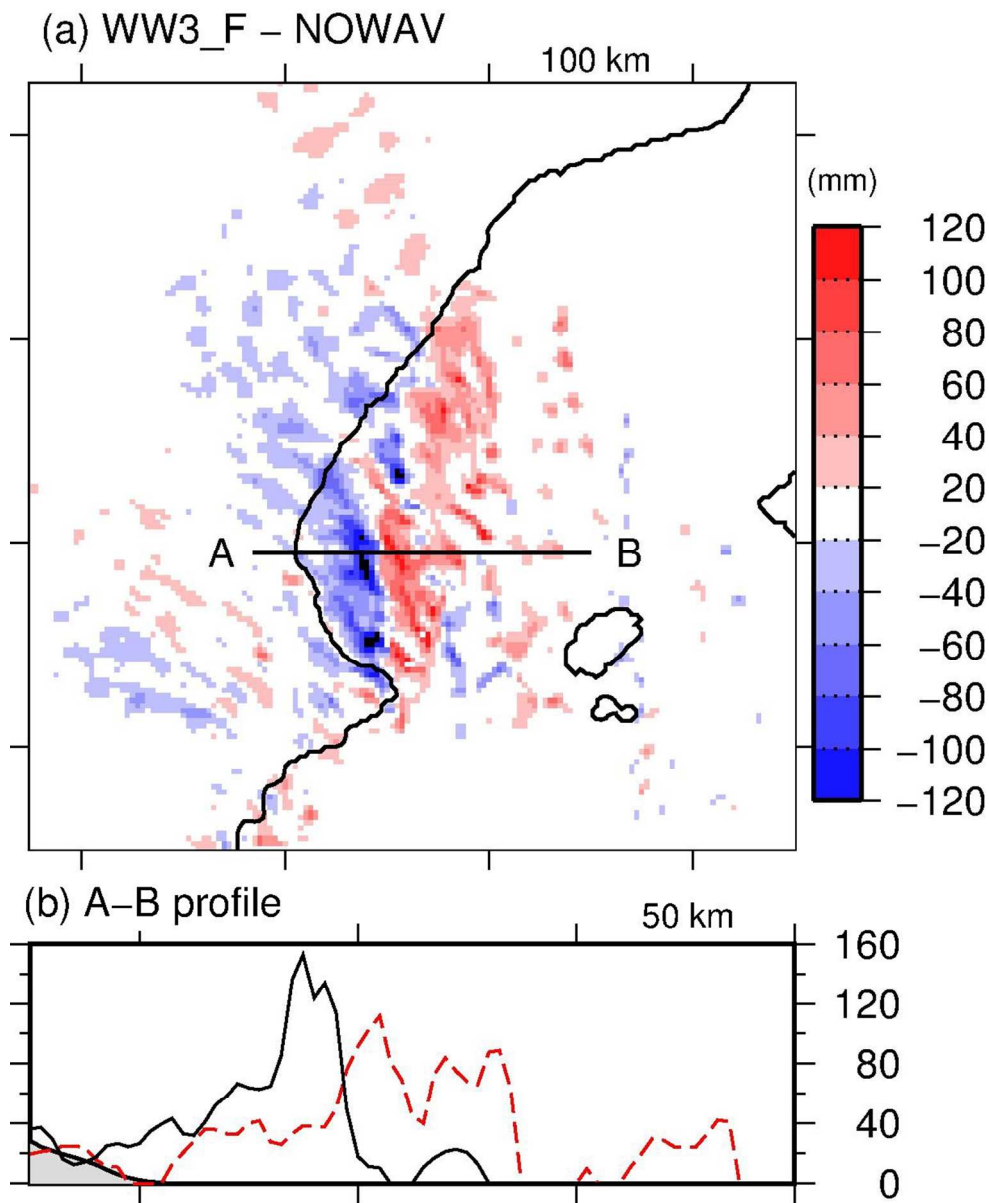


Figure 15: (a) Six-hour cumulated rain difference (mm) between the WW3_F and NOWAV simulations, and (b) six-hour cumulated rain (mm) in the NOWAV (black solid line) and WW3_F (red dashed line) along the A-B profile between 1800 UTC on 28 September 2012 and 0000 UTC on 29 September 2012.

102x125mm (300 x 300 DPI)

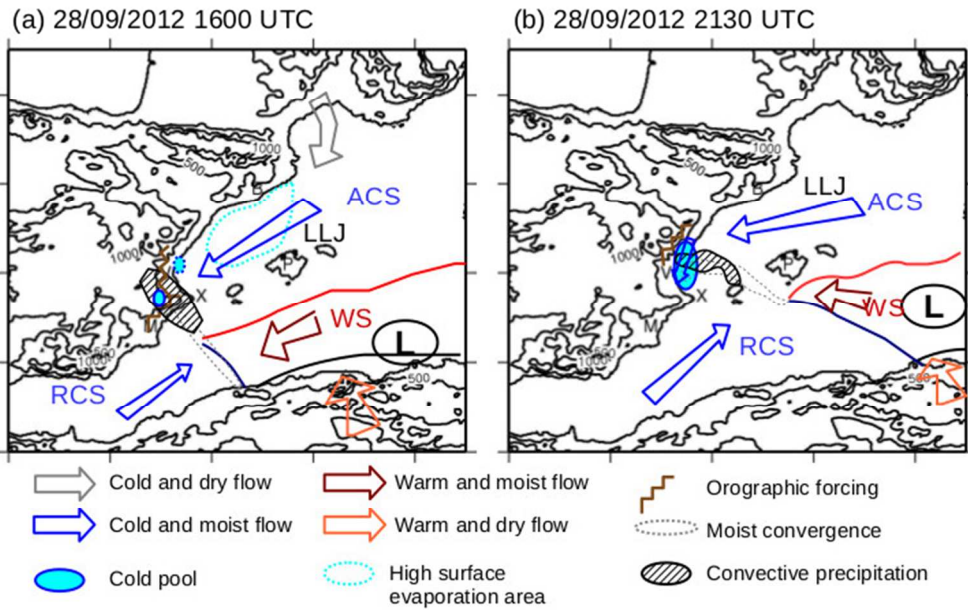


Figure 16: Schemes of the low-level flows and forcing mechanisms resulting in heavy precipitation on the Murcia-Valencia region (a) at 1600 UTC and (b) at 2130 UTC on 28 September 2012. The surface cold (warm) front is indicated with a blue (red) line. LLJ corresponds to low-level jet, ACS to ahead cold sector, WS to warm sector and RCS to rear cold sector.

186x114mm (100 x 100 DPI)

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Processes leading to deep convection and sensitivity to sea-state representation during HyMeX IOP8 heavy precipitation event, M.-N. Bouin*, J.-L. Redelsperger, and C. Lebeaupin Brossier

Using numerical simulations, the processes leading to deep convection and heavy precipitation during the HyMeX-SOP1 IOP8 are investigated. Cold pools and associated inflow deflection play a key role in positioning heavy precipitation, in addition to orographic effects and moisture convergence at sea. Representing in a more realistic way the sea surface roughness results in a displacement of the rain patterns of 50 km offshore.

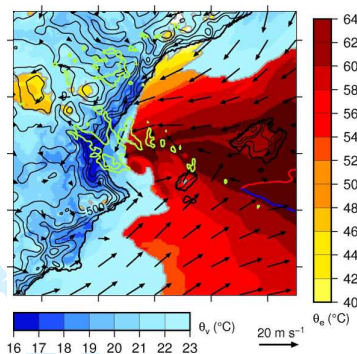


Figure caption: Equivalent potential temperature (°C) and virtual potential temperature (°C) at the surface as a proxy of cold pools, 10-m wind field (arrows) and convective precipitation (green contour) in northeastern Spain at 2130 UTC on the 28 September 2012.