Sensitivity of an intense rain event between atmosphere-only and atmosphere–ocean regional coupled models: 19 September 1996

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The representation of Mediterranean intense rain events in regional coupled models is of great importance for impact studies of climate change. It is investigated through the comparison of an atmosphere-only simulation forced by a coarse-resolution sea-surface temperature (SST), an atmosphere–ocean coupled simulation generating a high-resolution SST and an atmosphere-only simulation forced by a high-resolution monthly smoothed SST. The Cévennes, located in Southern France, is a region of high interest because of its frequent intense rain events and the proximity of the Gulf of Lion, where intense air–sea exchanges happen during mistral and tramontane wind bursts. Focus is given to one of the most intense rain events of the 20 year simulations (1989–2009): 19 September 1996, for which the change in the rain location between the atmosphere–ocean simulation and the atmosphere-only simulation forced by the coarse-resolution SST is large. In this case, the change in the rain location can be attributed mainly to a long-term change in the SST of 1.5 K, with a smaller but significant contribution of submonthly coupled effects such as cooling of the SST in the Gulf of Lion after moderate mistral events that occurred before the precipitation event. The change in the location of precipitation is related to a wind change of 5 m s⁻¹. This wind deviation originated from the combined effects of a surface pressure anomaly, stronger stratification on the western side of the coastal front in the Gulf of Lion and the enhanced blocking effect of the Alps.

Key Words: AORCM; atmosphere–sea coupling; quasi-stationary front; blocking

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1. Introduction

Heavy rain often affects the Mediterranean coast and sometimes causes casualties and damage costing several billions of dollars (Nuissier et al., 2008). Indeed, the configuration of steep-sided valleys in coastal mountains with small-size, high-response watersheds around the Mediterranean sometimes leads to flash floods when intense precipitation occurs in a very short time (typically more than 100 mm in less than 24 h). Since the Mediterranean area is considered as a hotspot for climate change (Giorgi, 2006), it is all the more important to investigate the changes in intensity and frequency of such extreme weather events. A major challenge for climate science is thus to represent them in climate models in order better to assess their possible evolution with climate change and possibly inform decision-making for future urban planning that could withstand the future intensity of such events.

While tools such as global climate models (GCMs) have too coarse a resolution to be able to produce such heavy rain events at present, regional climate models (RCMs) have proved to be better adapted. In fact, Rajczak et al. (2013) and Kysely et al. (2012), among others, investigated the representation of extreme rain events and their projection into the future. Rajczak et al. (2013) showed that extreme rain events are qualitatively well represented by a multiple set of RCMs. They underline the tendency towards an increase in frequency and intensity of extreme and intense short-term (daily or hourly) precipitation events in the Mediterranean in all seasons except summer. However, they stress that considerable uncertainty remains, due to the variability between models. Fowler et al. (2007) explain that this uncertainty comes from both the GCM providing the large-scale features and the RCM exhibiting regional features. Somot et al. (2008) also showed that the coupling between an atmospheric and an oceanic model (Atmosphere–Ocean Regional Climate Model (AORCM)) can be as important as the choice of the RCM and GCM for
evaluating climatological values of atmospheric variables in the future Mediterranean climate. In another direction, the present study addresses the three following questions.

- What processes leading to an intense rain event are represented in a regional climate model?
- How do the long-term and submonthly ocean evolution impact the location and intensity of a heavy precipitation event in an RCM?
- Which processes involved in an intense rain event are sensitive to the numerical coupling with the ocean?

To this aim, one particular intense rain event is in focus. The processes involved in this intense rain event are studied and their changes when an oceanic model is coupled to the atmospheric model are described in order to understand which processes the coupling with an oceanic model can influence.

The studied case occurs in the Cévennes region, located in the south of the Massif Central between the Rhône Valley to the east and the Aude valley to the southwest (Figure 1(a)). This region was chosen for its frequent and well-documented intense rain events. Nuissier et al. (2011) showed that different large-scale conditions could lead to heavy precipitation and floods in the Cévennes. Most of them require a trough located over the French west coast, Ireland or Spain, sometimes associated with a cut-off low. In these conditions, southerly surface winds blow over the sea towards the Cévennes and interactions with the Alps or Pyrenees can further channel the flux on to the southern Massif Central (Bresson et al., 2012; Bousquet and Smull, 2003). Then, the lifting necessary for the triggering of convection is provided either directly by orography or by upstream local convergence enhanced by orography. It can also result from the interaction of the impinging flow with cold air pools originating from a convective system already in place (Ducrocq et al., 2008). The high moisture content of low-level air is provided both by local evaporation over the Mediterranean and by remote sources. The contribution of local sources is greater if cyclonic conditions precede the event, rather than anticyclonic conditions, since the residential time of the impinging air parcels over the sea is larger in the first case (Dufour and Ducrocq, 2011).

When intense rain events occur in the Cévennes, the impinging air flows over the Gulf of Lion before reaching the orography. This marine area is often influenced by northerly and northwesterly winds channelled by the Rhône and Aude valleys, respectively called the mistral (Guénard et al., 2005, 2006; Drobinski et al., 2005) and the tramontane (Drobinski et al., 2001). These winds are particularly intense, dry and cold and they enhance cooling of the sea surface (Lebeaupin Brossier and Drobinski, 2009) as well as the oceanic cyclonic circulation in the Gulf of Lion (Madec et al., 1996). Due to trapping in this cyclonic circulation generating a doming structure, a colder patch of generally 100 km by 100 km is present at the surface in the centre of the Gulf of Lion around 42°N–5°E (Marshall and Schott, 1999). In autumn, this anomaly can persist during successive days under strong wind episodes (Lebeaupin Brossier and Drobinski, 2009) or be removed by oceanic upper-layer restratification when the wind decreases. During winter, due to the repetition of wind episodes, the cooling and salting of the surface water in the gyre can be progressively enhanced and can generate, on interannual scales, formation of western Mediterranean deep water (THETIS Group, 1994; Artale et al., 2002; Herrmann and Somot, 2008; Béranger et al., 2010). This 100 km² small area of cold SST cannot be well represented in the reanalyses (NCEP or ERA-interim) used to force regional climate simulation, since their spatial resolution is too coarse. However, in a coupled model such as the one used in

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this study, finer resolution and better representation of air–sea interactions prove better able to reproduce such a structure of cold SST (Lebeaupin Brossier et al., 2012). This area of cold SST can in turn influence the wind or the moisture content of the air when the flow is southerly and induces precipitation on the coast. In fact, Pullen et al. (2006) showed a reduced mixing in the stabilized boundary layer over the cooler SST induced in the Adriatic Sea by a bora event by comparing a simulation forced by an optimum interpolation of satellite SST or coupled to the COAMPS oceanic model every 6 h.

In fact, the influence of air–sea interactions on the atmosphere has been stressed by several studies. Kirtman et al. (2012) used the National Center for Atmospheric Research (NCAR) Community Climate System Model (CCSM 3.5) to show that the ocean was leading the atmosphere where the small-scale SST variability was strong. Chelton and Xie (2010) showed that the enhancement of the SST resolution improves the intensity of wind variations on 100–1000 km scales. They stress the fact that the influence of SST on the winds at scales smaller than 1000 km depends on the SST resolution, the parametrization of vertical mixing and the vertical grid resolution of models. Lebeaupin et al. (2006) showed that a change in the mean spatial SST of ±3 K had a strong influence on the intensity of rain events through changes in surface heat fluxes in three case studies over the Cévennes area. The change in the heat fluxes has more impact on the mesoscale convective systems than on the quasi-stationary fronts, for which the change results from the interaction between the front dynamics and convection. Regarding the dynamic effects, the influence of mesoscale eddies and fronts in the ocean on surface winds in the midlatitudes was reviewed by Small et al. (2008). Dynamic effects result from changes in the stability of the Marine Atmospheric Boundary Layer (MABL) and in the interactions between the MABL and the free-tropospheric winds, from changes in the pressure field or from changes in the surface friction. The relative importance of each of these effects depends on the state of the atmosphere.

Thus, coupling an atmospheric and an oceanic model can potentially have a strong impact on both the thermodynamic and dynamic fields in the Gulf of Lion, where a cold gyre is persistent, especially in autumn, a season in which intense rain events frequently occur over the Cévennes.

In this line of study, Lebeaupin Brossier et al. (2013) compared a simulation of an RCM forced by ERA-interim SST with an AORCM for an intense precipitation event in the Aude valley during 12–13 November 1999. Seven days before precipitation occurred, a strong mistral event started and lasted 5 days, decreasing the SST by 1 K on average over the Gulf of Lion in both simulations and locally by 2 K in the AORCM. This led to an eastward shift in precipitation, due to the ocean circulation, which had conserved the cold anomaly resulting from the long-lasting mistral event and the long-term difference between both simulations.

In the following study, we used the same simulations as Lebeaupin Brossier et al. (2013) (RCM and AORCM) but added a third simulation free from the submonthly coupling effects. This allowed us to separate the short-term from the long-term effects of coupling an atmospheric to an oceanic model on the precipitation location of one case study. In order to analyze the different processes that determine the sensitivity of intense rain events to the differences in air–sea coupling between these three simulations as precisely as possible, we focused on the rain event of 19 September 1996. This event was chosen amongst the most extreme rainfall events simulated in autumn over the Cévennes in hindcast simulations run from 1989 to 2009. In section 2, the models and simulations are described. Section 3 presents the simulated processes at play before and during the intense rain event of 19 September 1996. In section 4, the precipitation shift is assessed and explained in terms of dynamic shift. Section 4 further explains the origin of the wind change.

### 2. Models and simulations

#### 2.1. Experimental design

The Model Of the Regional Coupled Earth system (MORCE) platform is the framework in which the regional two-way air–sea coupled system used in this study was developed (Drobinski et al., 2012). It is a tool to improve understanding of the role of coupled processes on the regional climate of particularly vulnerable areas. The MORCE system is used in the Hydrological Cycle in the Mediterranean Experiment (HyMex: Drobinski et al., 2014) and the Coordinated Downscaling Experiment (CORDEX) of the World Climate Research Program (WCRP) (Giorgi, 2007).

##### 2.1.1. The atmospheric model

The atmospheric model within the MORCE system is the Weather Research and Forecasting (WRF) model of NCAR (Skamarock et al., 2008). The domain covers the Mediterranean basin with a horizontal resolution of 20 km with 130 grid points in latitude and 240 in longitude. It has 28 vertical levels using sigma coordinates. There are eight levels in the first 1200 m of the atmosphere. Initial and lateral conditions are taken from the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-interim reanalysis (Simmons et al., 2007) provided every 6 h with a 0.75° resolution. Moreover, indiscriminate nudging is used to constrain the fields above the planetary boundary layer with a coefficient of $5 \times 10^{-5}$ s$^{-1}$ for temperature, humidity and velocity components. This reduces the internal variability of the different simulations and allows us to consider that the differences come mostly from the distinct forcings at the surface (Stauffer and Seaman, 1990; Salameh et al., 2010; Omrani et al., 2012). The boundary-layer parametrization is the Yonsei University scheme (YSU), which is a K-profile scheme improved by Noh et al. (2003). The surface layer is the Monin–Obukhov scheme (Stull, 1994). The convective scheme is the Kain–Fritsch scheme (Kain, 2004). The complete set of physical parametrizations can be found in Lebeaupin Brossier et al. (2013).

##### 2.1.2. The ocean model

The ocean model of MORCE is Nucleus for European Modelling of the Ocean (NEMO: Madec, 2008). It is used in a regional eddy-resolving Mediterranean configuration MED12 (Lebeaupin Brossier et al., 2011; Beuvier et al., 2012) with a 1/12° resolution ORCA grid (i.e. 7–8 km resolution). Vertically, MED12 has 50 stretched z levels, with finer levels near the surface. The first level has a thickness of about 1 m. The initial conditions for 3D potential temperature and salinity fields are provided by the MODB4 climatology (Brankart and Brasseur, 1998), except in the Atlantic zone between 11°W and 5.5°W, where the Levitus et al. (2005) climatology is applied. In this area, a 3D relaxation to this monthly climatology is used through the run. River runoff and the Black Sea water input come from a climatology and their freshwater flux is defined at the mouths of the 33 main rivers and at the Dardanelles Strait respectively. Smaller river runoffs are summed and set as a homogeneous coastal runoff around the Mediterranean Sea (Beuvier et al., 2010). Further details on the ocean model parametrization can be found in Lebeaupin Brossier et al. (2011) and Beuvier et al. (2012).

##### 2.1.3. Numerical experiments

The control simulation (CTL) uses WRF on the Mediterranean basin forced on its boundaries by ERA-interim reanalyses from January 1989 to December 2008. It is also nudged with ERA-interim moisture, temperature and wind fields above the planetary boundary layer. Sea-surface temperature (SST) is updated daily
from ERA-interim reanalyses. Outputs are given every 3 h, while the model integration time step is 1 min.

The coupled simulation (named CPL experiment) runs with two-way interactive exchanges between the two compartment models managed by the OASIS coupler version 3 (Valcke, 2006). The exchanged variables are the SST and the heat, water and momentum fluxes. The coupling frequency is 3 h. The coupler uses a bilinear method to interpolate the ocean grid towards the atmospheric grid and vice versa (see also Lebeaupin Brossier et al., 2013).

The smoothed simulation (SMO) is an atmosphere-only simulation with the same characteristics as the CTL simulation except that, instead of the ERA-interim SST, a new SST field has been used for forcing the atmospheric model. This forcing has been designed in order to retain the same climatology and diurnal cycle as the SST of the CPL simulation, but without the submonthly SST variations. To this end, the SST value used to force the SMO atmospheric simulation at each target time step was calculated by performing a central moving average with a 31 day window, retaining only the 31 time steps in the time window that correspond to the same GMT time as the target time step. In this way, the diurnal cycle, as well as its seasonal variations, are preserved, as are all the persistent spatial structures that exist in the CPL run. The high-frequency air–sea coupling effects (submonthly variations), however, are not present in the SMO simulation.

Figure 2 illustrates the differences in SST between the three simulations during the month preceding the 19 September 1996 event. The SST from CTL is represented by the red curve, which is smoother and has no diurnal cycle (ERA-interim SST is updated daily), the SST from CPL by the black curve and the SMO SST by the dashed black curve, which is smoother than the CPL but still has the same diurnal cycle and climatology.

The three simulations run from January 1989 to December 2008. CPL starts with an ocean at rest. In this study, we only focus on the northwestern Mediterranean area in November 1996, i.e. far from any drift.

3. 19 September 1996 in the RCM
3.1. Representation of the intense precipitation event

This case has been chosen among the 20 strongest precipitation events represented by the WRF model for the Cévennes between 1 September and 30 November over the 1989–2009 period. It shows a large difference in the precipitation pattern between the simulation coupled with an ocean model (CPL) and the simulation forced by the ERA-interim SST (CTL).

The rain maximum (∼ 150 mm in 24 h, Figure 1) is located on the first slopes of the Cévennes on the southeastern edge of the Massif Central. The precipitation rate peaks at 0900 UTC (Figure 3 shows the spatial average of rain in the frame presented in Figure 1). The amount of rain mainly comes from explicitly resolved processes (Figure 3).

The synoptic flow is characterized by a trough in the 500 mb geopotential over the west coast of France with a cut-off extension down to Spain. It shows a slow evolution during the day, allowing the synoptic wind to stay southwesterly (see Figure S1). At the surface, a low moves from southwestern France to the Mediterranean and is centred on the western edge of the Gulf of Lion during this day. This, associated with the deflection flow by the Alps, produces at low level a moist southsoutheasterly jet and a converging zone in the Gulf of Lion, which are key ingredients for mesoscale convective systems (see Bresson et al. (2012) for another event in the same area).

In the simulation, the front produces an amount of rain greater than 40 mm day$^{-1}$ over the sea in the Gulf of Lion (Figure 1(b)). The rain intensity is enhanced when the converging zone reaches the Cévennes slope at 0600 UTC. The situation stays the same until 1800 UTC; a total of 150 mm is recorded by the model at the end of the day. The model can be compared with Système d’Analyse Fournissant des Renseignements Atmosphériques à la Neige (SAFRAN) analysis, which provides 8 km gridded rain at the hourly time step using ground data observations (rain-gauges; see Quintana-Seguí et al. (2008) and Figure 1(d)). The maximum daily amount also reaches 150 mm in SAFRAN and the rain-gauges registered a local maximum of 169.5 mm in Saint-Martin-de-Londres (43.79°N, 3.73°E). The model reproduces the intensity of the event with a maximum of 150 mm of precipitation, mostly explicitly resolved (Figures 1(b) and 3). However, the rain location is about 50 km further northeast than in the model. The CPL simulation reproduces the SAFRAN analysis better, though the rain is still located too far to the southwest. The model error comes from many factors (synoptic conditions given by ERA-interim, surface temperature and moisture that lack good representation, parametrization schemes, discrete resolution of dynamical equations, etc.) and the representation of the SST is just one of them. For example, the blocking by the Alps may be misrepresented in the model because of insufficient resolution and uncertainties in boundary-layer parametrization.
3.2. Ocean surface pre-conditioning

Over the month preceding the event, the CPL simulation shows two significant mistral events, characterized by strong northnorthwesterly wind (black solid line in Figure 2(b) and (c)) in the Gulf of Lion at 5°E, 42.4°N. Figure 2(a) shows that the first event cooled the SST by 1.5 K in 5 days on average in the Gulf of Lion, ending on 2 September. Six days before the rainfall event, another mistral event started and lasted from 12 September to 15 September (Figure 2(b) and (c)). It cooled the SST by 1 K. Owing to the representation of the valleys at this resolution (Figure 1), these winds are well represented (Lebeaupin Brossier et al., 2011).

Indeed, the CPL simulation shows similar variations of SST to the high-resolution GOS-SST, which is an optimally interpolated SST with a resolution of 1/16° obtained from the night-time satellite data of the Advanced Very High Resolution Radiometer (Marullo et al., 2007). The variations of the GOS-SST are synchronous with the variations in CPL SST and are of similar amplitude: the representation of the mistral and the response of the ocean are well reproduced for the considered period.

Repeated high-speed wind events over the simulation, such as the ones described here, thus helped to produce a sea-surface cold pool of less than 18°C in the centre of the Gulf of Lion, whereas it is around 20°C in the south and along the French Riviera coast in the CPL simulation (Figure 4(c)). Hence, the structure of SST in the CPL model shows a similar spatial structure to the GOS reanalysis: a colder SST in the centre of the Gulf of Lion with closed contours and a warmer tongue along the French Riviera (Figure 4(a)). In comparison, the SST in the CTL simulation gradually decreases from 21 to 18°C from the south to the north of the area, not showing any cold pool (Figure 4(b)). However, the CPL simulation shows a long-term cold bias of 1.5 K relative to GOS-SST and ERA-interim in the whole area and on average in the Gulf of Lion, as is shown in Figures 4 and 2. As seen before (Figure 2), ERA-interim shows a smoother evolution in time than GOS-SST, but the long-term signal is similar. The CPL simulation shows the same submonthly variations due to the mistral events as GOS-SST, but it has a cold bias compared with reality. Hence, comparing CPL with CTL (ERA-interim) combines the effects of a long-term cold bias of 1.5 K in CPL with those of sharper and more realistic spatial and temporal variations of SST. To isolate the effects of the temporal smoothing of SST in the CTL simulation, the SMO simulation, which has a similar long-term behaviour to the CPL simulation but no submonthly variations, was performed and reproduces similar differences to those between ERA-interim and GOS-SST. Figure 4(d) shows that the submonthly coupled effects in CPL added a cold anomaly of 0.5 K in the north of the Gulf of Lion and a warmer SST by 0.25 K in the south of the Gulf of Lion compared with the smoothed SST of SMO.

4. Northeastward shift of precipitation

Precipitation in CPL and CTL simulations reaches 150 mm day$^{-1}$. However, the rain maximum in the CPL simulation is shifted towards the northeast compared with the CTL simulation (see Figures 1 and 5(a)). The maximum in CPL is enhanced by 3 mm, which is negligible compared with the decrease of 80 mm on the southwestern side of the CTL maximum and the increase of 80 mm on the northeastern side which characterize the precipitation shift. All the simulations have the same large-scale forcing at their boundaries and are nudged with ERA-interim temperature, humidity and velocity fields with a coefficient of $5 \times 10^{-5}$ s$^{-1}$ above the planetary boundary layer. This limits the large-scale variability between simulations (Omran et al., 2012). Thus, the
observed differences essentially result from the different boundary conditions at the ocean surface between the three simulations.

4.1. Attribution of the precipitation shift: long-term versus submonthly SST differences

Figure 5(b) shows the difference in precipitation between CPL and SMO, which only represents the short-term effects of the coupling. The pattern of the difference is similar to the CPL–CTL difference (Figure 5(a)) but the intensity is weaker: −15 mm day$^{-1}$ decrease west of the maximum (19% of CPL–CTL) and 7 mm day$^{-1}$ increase east of it (9% of CPL–CTL). Thus, a smaller but significant part of the change in precipitation is induced by the submonthly air–sea coupling shown by the CPL–SMO difference, while the long-term difference in SST between the CPL and CTL simulations is responsible for most of the change.

4.2. A shift in the wind induces the change in precipitation

The precipitation shift can result from a change in Convective Available Potential Energy (CAPE), in the Level of Free Convection (LFC), in the moisture field or in the wind field induced by the change in SST. There are some changes in the LFC of the incoming air, but in both simulations it stays under 500 m (Figure S2): it is not a factor that can inhibit precipitation on one side and enhance it on the other (the topography is higher than 500 m in this area). The values of CAPE remain above 300 J kg$^{-1}$, varying by 50 J kg$^{-1}$ between both simulations (see Figure S2). However, its spatial structure remains the same (same location of the maximum CAPE between both simulations). This change can thus modulate the intensity of precipitation but cannot explain such a shift either. Table 1 shows the budget of moisture advected into the red and blue boxes presented in Figure 5(a) during the day. This budget is calculated as the daily mean of the inward flux of moisture through the box boundary in the first eight vertical levels ($\lesssim$1200 m) divided by the box area in order to obtain the mean moisture convergence for 19 September 1996. The amount of moisture that enters the box through its lateral boundaries is also precipitated within the box or exits the box upward through its summit: it is thus the moisture available for rainfall generation. The two boxes are respectively centred on the negative and positive differences in rain between the CPL and CTL simulations (Figure 5(a)).

Table 1 examines the relative impacts of the moisture and wind fields on moisture convergence in the two boxes shown in Figure 5. In the first line of Table 1, the convergence was calculated for the CTL simulation. 208 kg m$^{-2}$ day$^{-1}$ entered the western box (blue box in Figure 5(a)) while 40% less entered the eastern box. On the other hand, for the CPL simulation (fourth line) the result is reversed: 220 kg m$^{-2}$ day$^{-1}$ of moisture converged into the eastern box while only 40% less entered the western box. This difference could be due to changes in either the moisture field or the wind field. To determine which of these effects dominates, we performed intermediate calculations. When we used the moisture field from the CPL simulation with the wind field from the CTL simulation (second line), the budget in each box was almost equal to the budget from the CTL simulation: changing the moisture field did not significantly affect the distribution of moisture convergence between the two boxes. Conversely, when the moisture field from the CTL simulation was used with the wind from the CPL simulation, the distribution of moisture convergence between the two boxes was almost the same as in the CPL simulation (third and fourth lines).

Therefore, the change in moisture convergence (and consequently in precipitation) results mainly from a change in the wind field, which is shown in Figure 6(a). On the other hand, the impact of the incoming moisture field is much weaker. The maximum amplitude of the wind change is located in the converging zone next to the coast. It is directed towards the east and its maximal amplitude is 6 m s$^{-1}$ (Figure 6(b)).

The difference between CPL (fourth line) and SMO (last line) is weaker, more pronounced in the western box and also essentially results from a change in the wind field rather than in the moisture field (fifth and sixth lines).

5. Three explanations of the wind shift

The surface winds result from a fine equilibrium between the free tropospheric forcing and the surface interaction through friction, turbulence, orography and local thermal effects. In both simulations, the free tropospheric large-scale forcing remains similar because of the nudging applied above the boundary layer. However, the air–sea interactions can influence boundary-layer processes, interaction of flows with orography (through the stabilization or destabilization of the boundary layer) and surface pressure by local thermal effects. These three mechanisms are studied to understand the differences between the CPL, SMO and CTL simulations and thus to separate the effects of different SST in the long-term and submonthly air–sea coupling.

5.1. A surface pressure anomaly leads to an anticyclonic shift in the wind

Figure 7 displays the surface pressure anomaly between CPL and CTL. Its greatest intensity is 0.6 hPa in the Gulf of Lion. The geostrophic wind associated with this pressure anomaly is northwesterly off the coast and reaches up to 8 m s$^{-1}$ (Figure 7).

Under the hydrostatic approximation, the surface pressure can be expressed as follows:

$$P_S = P_{top} \exp \left( \int_{0}^{z_{top}} \frac{g}{R_{air} T} dz \right). \tag{1}$$

$P_{top}$ is the pressure at the top of the highest level of the model, $g$ the gravitational acceleration and $R_{air}$ the perfect gas constant divided by the dry air molar mass. $T$ is the virtual temperature of the air. The hydrostatic surface pressure anomaly is calculated as the difference between $P_S$ in CPL and $P_S$ in CTL. Even though WRF was run without making the hydrostatic approximation, the pressure anomalies in our case appear to be essentially hydrostatic (Figure 8). The only significant differences between both figures lies around (3.75°E, 43.6°N), where convective rain occurs and consequently the vertical acceleration is not negligible. In other places, and especially above the sea, the pressure anomaly is directly linked to the temperature anomaly between both simulations. The evolution of the temperature anomaly of CPL relative to CTL in the lowest model layers is analyzed, given by

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Table 1: Daily-averaged budget of moisture convergence on 19 September 1996 (kg m$^{-2}$ day$^{-1}$) over the first eight levels for the western box (which contains CTL precipitation maximum) and the eastern box (which contains CPL precipitation maximum); the boxes are shown in Figure 5(a).

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<tr>
<th></th>
<th>Western rain box (blue)</th>
<th>Eastern rain box (red)</th>
<th>Both boxes</th>
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<tbody>
<tr>
<td>$\sum Q_{CTL-CTL}$</td>
<td>208</td>
<td>122</td>
<td>165</td>
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<tr>
<td>$\sum Q_{CTL-CTL}$</td>
<td>208</td>
<td>124</td>
<td>166</td>
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<td>220</td>
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<td>$\sum Q_{SMO-CTL}$</td>
<td>144</td>
<td>224</td>
<td>184</td>
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<tr>
<td>$\sum Q_{SMO-CTL}$</td>
<td>145</td>
<td>224</td>
<td>185</td>
</tr>
</tbody>
</table>
An Intense Rain Event in a RCM and an AORCM

Figure 5. Daily precipitation on 19 September 1996 (mm day$^{-1}$). (a) Colours: difference between CPL and CTL; (b) colours: difference between CPL and SMO. In both panels, thick contours show CPL (50 mm day$^{-1}$) (note the difference in colour scales).

The following equation:

$$\frac{\partial T_{CPL-CTL}}{\partial t} = - \mathbf{u}_{CPL-CTL} \nabla T_{CPL-CTL}$$

Advection of $T$ anomalies

$$\mathbf{u}_{CPL-CTL} \nabla T_{CPL-CTL}$$

Advection of $T$ by wind anomalies

$$\mathbf{u}_{CPL-CTL} \nabla T_{CPL-CTL}$$

Advection of $T$ anomalies by wind anomalies

$$\frac{HF_{CPL-CTL}}{C_P}$$

Turbulent heat flux anomalies

The anomalous field is noted with the ‘CPL–CTL’ index. This equation includes terms linked to the advection of the temperature anomaly by the background wind, advection of the background temperature gradient by the anomalous wind, the cross-product of both anomalies and the heat flux anomaly. The adiabatic compression term is neglected, since this equation is analyzed at the surface layer.

Figures 9 and 10 show a sequence of surface pressure and temperature anomalies both on the day before the event and a few hours before it. In both figures, the last panel (f) presents the anomalous pressure field at 0600 UTC on 19 September 1996, while the first panel (a) presents this field 30 h before the maximum of the rain event. All other panels show the situation at intermediate times.

A day before the event at 0000 UTC, the negative surface heat flux anomaly is collocated with the pressure and temperature anomalies and also with the negative SST anomaly (Figures 4(a) and 9(a)). Thus, the wind blowing over the negative SST anomaly brings less heat to the atmosphere in the CPL simulation than in the CTL simulation, where the SST is warmer and spatially more homogenous. This creates a negative temperature anomaly in the boundary layer and a positive pressure anomaly at the surface (by the hydrostatic relation in Eq. (1)). At this step, the heat flux anomaly (mostly from sensible heat flux) is the dominant term for the temperature evolution (in Eq. (2)). The winds blowing over the Gulf of Lion come from the east and the north: they originate from a depression centred between Catalonia and Corsica in both simulations (Figure 10(b); the green lines represent the 999 hPa isobar of the first layer pressure in CTL).

After 6 h of this situation, the wind convergence is enhanced in the lee of the depression (Figure 10(a) and (b)). The mistral and tramontane converge with easterly winds coming from the Gulf of Genoa. The collocation between the wind convergence and the enhanced surface anomaly shows that the temperature anomalies are advected by the winds and gather vertically in the converging zone. At that moment, both surface heat flux anomalies and advection of temperature anomalies by the prevailing wind are important in Eq. (2). This phenomenon enhances the surface pressure anomaly (Eq. (1)). Figure 10(c) shows that the converging zone progressively disappears after 12 h and the pressure anomaly weakens and is separated into two maxima.
At 0000 UTC on the day of the event (Figure 10(d)), the western maximum is enhanced over the SST anomaly by cooling compared with the CTL simulation (weaker fluxes). The air cools down while it is slowly advected northeastwards over the SST anomaly by weak winds (3–5 m s\(^{-1}\)) (Figure 9(d) and (f)) and the pressure anomaly deepens (Figure 10(d) and (f)). Again, the heat flux anomaly dominates the right-hand side of the equation. The pressure anomaly reaches 0.5 hPa when it arrives in the converging zone at 0600 UTC. This generates an anomalous northeasterly geostrophic wind of about 8 m s\(^{-1}\) (Figure 7). The converging zone seen in Figure 10(d) and (f) does not seem to create any pressure anomaly, since the converging winds come from areas with both positive (along the eastern coast) and negative (in the Gulf of Lion) flux anomalies (Figure 9(d) and (f)).

To summarize, the pressure anomaly in the CPL simulation relative to the CTL simulation was generated consecutively by a convergence of northerly and easterly winds advecting cold temperature anomalies in the CPL simulation and the enhancement of this pressure anomaly by slow advection and simultaneous cooling of air masses.

This pressure effect is not seen in the difference between the CPL and SMO simulations: the pressure difference stays within the intensity of the noise difference between both simulations with no proper pattern. The pressure effect thus essentially results from the long-term SST difference between CPL and CTL.

### 5.2. Enhancement of blocking by the Alps

The surface horizontal wind field evolution is given by the following equation (Laprise, 1992):

\[
\frac{du}{dt} + f \mathbf{k} \times \mathbf{u} = - \nabla_\eta P \rho - \nabla_\eta \phi + \frac{\partial u'w'}{\partial z} \tag{3}
\]

and can be approximated as the result of the balance between the advection (first term in Eq. (3)), Coriolis force (second left term), pressure force (first and second terms on the right-hand side of the equation) and turbulent mixing (last term on the right, horizontal turbulence is considered negligible). Given that the equations are resolved on terrain-following coordinates (\(\eta\)), the third term is composed of two terms: the gradient of pressure along \(\eta\) coordinates added to the gradient of the geopotential along \(\eta\) coordinates. The geopotential gradient corresponds to the gravity acceleration multiplied by the altitude gradient along \(\eta\) coordinates. \(f\) is the Coriolis parameter. \(\mathbf{u}\) is the horizontal projection of the wind vector; \(u'\) and \(w'\) are respectively the high-frequency anomalous horizontal and vertical winds compared with the time average of the wind. \(\rho\) is the density of the fluid and \(\mathbf{k}\) is a vertical unity vector. This equation can be rewritten by decomposing the wind into geostrophic and ageostrophic components:

\[
f \mathbf{k} \times \mathbf{u}_g = - \frac{\nabla_\eta P}{\rho} - \nabla_\eta \phi, \tag{4}
\]

\[
f \mathbf{k} \times \mathbf{u}_{ag} = - \frac{du}{dt} + \frac{\partial u'w'}{\partial z}. \tag{5}
\]

The geostrophic wind \(\mathbf{u}_g\) results from a balance between pressure gradient and Coriolis force (Eq. (4)). The ageostrophic wind \(\mathbf{u}_{ag}\) results from a balance between the remaining terms (momentum transfer, advection, time evolution) and the Coriolis force (Eq. (5)). In practice, \(\mathbf{u}_{ag}\) is calculated as \(\mathbf{u} - \mathbf{u}_g\).

In section 5.1, the geostrophic anomaly between CPL and CTL resulting from the pressure anomaly was presented. Here, the ageostrophic component is shown in Figure 11. An intense
The northerly wind of 20 m s$^{-1}$ can be noticed on the southern edge of the Alps. This wind component offsets the southerly geostrophic wind (see Figure S3): this shows the blocking effect of the Alps. The sum of both components generates an easterly wind along the Alps (Figure 6(a)).

Figure 11(a) and (b) show respectively the ageostrophic wind of the CTL and CPL simulations. Comparing them allows an appreciation the stronger ageostrophic wind intensity in the CPL simulation. In fact, the contour of 20 m s$^{-1}$ right at the southern edge of the Alps goes further south in the CPL simulation and the wind intensity increases by 3 m s$^{-1}$ around (5.25°E, 42.8°N). This could be the result of a change in the drag coefficient, but the Monin–Obukov length does not undergo a great change, nor does $u^*$ or the surface roughness (see Figure S4). Therefore, the change in ageostrophic wind results from a change in the intensity of the blocking effect due to the Alps, in turn affecting the turbulent flux intensity, since the two effects are linked to the wind shear.

To understand how the deflection by the Alps can be stronger in the CPL simulation, the results of Pierrehumbert and Wyman (1985) are used. They consider a 3D zonal flow with a zonal speed $U$ crossing a mountain of height $h_m$ extending meridionally. The fluid has a constant Brunt–Väisälä frequency $N^2 = g/\theta_0 h/\partial z$, meaning a constant stratification ($\theta$ is the potential temperature). Nonlinear Boussinesq hydrostatic equations are used. Adimensionalization is applied, with $L$ being the mountain width. Two parameters arise from it: the Rossby number $Ro = U/L\lambda$ and the Froude number $Fr = Nh_m/U$, which reflects the adimensional height of the mountain. The Rossby number reveals the extent of the advective and time evolution term compared with the geostrophic terms. In the case of the Alps, $U = 20$ m s$^{-1}$, $f = 10^{-4}$ s$^{-1}$ and $L = 200$ km, so $Ro$ is about 0.4. Thus, all the terms are important in the equation. Pierrehumbert and Wyman (1985) explain the initial upstream surge as being the result of the initial $U$ becoming subgeostrophic when the air reaches the mountain, because the air is decelerated. Part of the meridional pressure gradient is no longer balanced by the Coriolis force and $v$, the meridional wind, increases with time: the flow is deflected to the left around the mountain. It is then balanced by the geostrophy in the zonal direction.

Deflection becomes more important as the Froude number increases for a constant Rossby number: the mountain becomes too high for the air to be able to rise over it. Thus, the more stratified the flow, the more deflection it undergoes. For a moist flow, the moist Brunt–Väisälä frequency is used, replacing the potential temperature by the virtual potential temperature to add the change in air density because of water vapour in the air mass. Figure 12 shows the moist Brunt–Väisälä frequency averaged over the first five model layers in (a) CTL and (b) CPL simulations (up to 500 m over the sea). Considering that the incoming air (like the limit conditions in Pierrehumbert and Wyman (1985)) is located at the southern edge of the Alps, out of the deflection zone where $N^2$ is positive (stable zone), Figure 12(b) shows that $N$ is twice as large in the CPL simulation ($8 \times 10^{-3}$ s$^{-1}$) as in the CTL simulation ($4 \times 10^{-3}$ s$^{-1}$), reflecting the weaker sensible heating of the atmosphere by the sea over the colder SST in the Gulf of Lion. Thus, the Froude number varies from 1 to 2 between both simulations. In both cases, the air is deflected to the west because the Rossby number is small enough to show geostrophic effects: the deflection is not symmetrical, as it would be with an infinite Rossby number. When the air then arrives over the Cévennes, the associated Froude number becomes smaller: $U = 18$ m s$^{-1}$, $N = 3 \times 10^{-3}$ s$^{-1}$, $h_m = 1000$ m, $Fr = 0.17$ and the flow goes easily over the Cévennes, as shown by Chen and Lin (2005). Precipitation occurs on the upslope of the Cévennes and the system remains stationary.

However, the ageostrophic wind is not only increased but also deviated towards the north in CPL compared with CTL at (4°E, 43.2°N). This generates a southerly anomaly of the ageostrophic wind between CPL and CTL, which offsets the northerly component of the geostrophic anomaly, which is a northwesterly wind of 8 m s$^{-1}$ (Figure 7). This deviation to the north is further explained in the next section.

5.3. Stabilization of the cold front: northward anomaly

The converging zone associated with precipitation over the sea is a cold front. In fact, the neutral layer on the northeastern side (Figure 12) with larger wind intensity encounters the more stable layer on the southwestern side of the front. This stable area is characterized by colder air near the surface (2 K cooler than in the neutral area). Its Brunt–Väisälä frequency doubles between the CTL and CPL simulations and this zone extends towards the east (Figure 12(a) and (b)). The frontal zone can thus be moved to the east because of the increase of stability of this zone. In fact, Snyder (1998) explains that the scaling factor between the cross-front wind component and the along-front wind component ($\epsilon$) depends on the basic state shear and on the stability of the planetary boundary layer. In simulations of an idealized baroclinic disturbance with constant vertical wind shear ($A$) and constant horizontal gradient and stratification of potential temperature ($N$), the scaling factor can be set as $\epsilon = A/N$ by estimating the ageostrophic flow from the Sawyer–Eliasen equation (Snyder et al., 1993). Thus, the larger the stratification, the smaller the cross-front flow or the larger the along-front flow. In the CPL case, this may explain that the winds on the front edge are deviated towards the along-front direction (Figure 11). Thus, this provides a southerly component to the ageostrophic wind, which offsets the...
Figure 9. Evolution of the temperature anomalies averaged over the first four levels of the atmosphere (380 m) and the surface sensible heat flux anomalies over the day before the rain event (18 September 1996) and the day of the rain event (19 September 1996). Colours: temperature anomaly between CPL and CTL (K); black contours: surface sensible flux anomalies (every 5 W m$^{-2}$, plain line positive, dash–dotted line negative); blue arrows: wind of the CTL simulation (scale indicated on each panel).

northerly component of the geostrophic anomaly. The resulting wind anomaly is thus a westerly wind with 5 m s$^{-1}$ intensity.

The impact of stability seems to be the major explanation for the difference between CPL and SMO, since no significant pressure difference exists between SMO and CPL. In fact, the Brunt–Väisälä frequency increases by $1 \times 10^{-3}$ s$^{-1}$ around 4.25° E, 42.75° N (Figure 12(b) and (c)) in the CPL simulation compared with the SMO simulation. A decrease in the intensity of the blocking by the Alps mixed with a slight eastward move of the converging zone due to the stronger stratification on the western side of the front in the CPL simulation can explain the decrease in precipitation on the western side of the front and the increase on the eastern side.

6. Conclusion

On 19 September 1996, up to 150 mm of precipitation occurred over the day in the Cévennes region. It was reported by the rain-gauge network and reproduced in the regional simulations using either the atmospheric model WRF forced by ERA-interim large-scale fields and SST or the WRF model coupled with the NEMO-MED12 ocean model. Both the atmosphere-only (CTL) and atmosphere–ocean coupled (CPL) simulations captured the
intensity of the event. However, both of them give a location too far to the southwest. All the dynamic ingredients of the heavy rain event (a coastal front due to the blocking of the wind by the Alps and a depression on the northwest of the Gulf of Lion) are represented in the model, leading mostly to explicitly calculated rain, enhanced by the interaction with orography. The mistral and tramontane are also represented by the model a few days before the event, allowing strong air–sea interactions in the coupled simulation. Nevertheless, the model shows a southwestern shift of the location of the rain event compared with SAFRAN reanalyses. The error may come from various factors, including the model resolution and approximations made by parametrization in the model.

Regarding the impact of the long-term and submonthly evolution of the ocean on an intense rain event, the CPL simulation shows a shift of the precipitation location to the northeast, slightly closer to the SAFRAN maximum location compared with the CTL simulation. This event is therefore sensitive to the coupling of an atmospheric model with an oceanic regional model. The differences between the CPL and the CTL simulations result from both the different SST climatology (mean bias of −1.5 K and different spatial structure of NEMO-MED12 in the CPL simulation, compared with ERA-interim in the CTL simulation) and the response of the ocean to rapid events such as the mistral wind burst, which occurred 5 days before the precipitation event only present in the CPL simulation.
Figure 11. Ageostrophic wind intensity (shading) and direction (vectors) for (a) the CTL simulation and (b) the CPL simulation on 19 September 1996 at 0600 UTC.

In practice, ERA-interim SST values have much fewer submonthly variations than NEMO-MED12 SST, which presents similar variations to GOS-SST (high-resolution reanalyses of AVHRR). However, the NEMO-MED12 simulation has a long-term cool bias of SST over the Gulf of Lion. To separate the effects of the submonthly variations from those of the different long-term SST, a third simulation (SMO) has been performed, using the NEMO-MED12 SST from the CPL simulation but smoothing it in time in order to remove the submonthly coupled effects. Therefore the SMO simulation has the same climatological SST as the CPL simulation but without its rapid variations. The shift between SMO and CPL was significant, but smaller than the difference between CTL and CPL. In this case, the location shift can be attributed mainly to the change in long-term SST (about 80% of the signal), with a smaller but significant contribution of the submonthly coupled effects arising from a mistral event, which cooled the sea by about 1 K about 5 days before the precipitation occurred. However, these proportions may change for other events, depending on the chronology between mistral events and the intense precipitation event.

Regarding the sensitivity of the processes involved to numerical coupling with an ocean, the location shift in the CPL simulation relative to the CTL simulation results mainly from an eastward shift of the southerly low-level wind jet, locally producing a difference in intensity up to 5 m s$^{-1}$. This change is caused by a positive pressure anomaly located in the Gulf of Lion, which leads to an 8 m s$^{-1}$ geostrophic northwesterly anomalous wind. The stronger stratification on the western side of the front, in addition to a stronger blocking of the southerly impinging flow by the Alps, counteracts the northerly component of the geostrophic wind by providing a southerly anomalous ageostrophic wind. The resulting wind anomaly is thus eastward. These ingredients are specific to this case, in the sense that the pressure anomaly was generated by the combination of wind convergence gathering the temperature anomalies and enhancement of it by surface fluxes in phase with advection. Moreover, the presence of a front with weak wind on its stable western side and the presence of the blocking effect of the Alps makes stratification a crucial element for the front position and the convergence intensity.

In summary, this study showed that processes such as a pressure anomaly resulting from the accumulation of temperature anomalies led by a wind convergence, enhancement of pressure anomalies in weak wind areas where the anomaly moves with the air parcels, changes in the blocking of winds and changes in the dynamics of a surface front because of changes in stability can result from the use of an AORCM compared with a RCM on a several-year run. This case shows that an intense rain event can be sensitive to an evolving SST, dictated both by the long-term and short-term interactions of the oceanic and atmospheric models.
SST is only one factor responsible for the model error and we cannot conclude with a single case that the represented air–sea coupling improves the model ability to represent heavy rain events. In fact, the present study addressed one case of intense precipitation: 19 September 1996, with one atmospheric model, WRF-ARW model, and one oceanic model, the NEMO-MED12 model. Therefore, in order to strengthen and generalize these results, more studies will be performed along the same lines: intermodel comparison with several ARCM/AORCM pairs of simulations performed within the Mediterranean region COordinated Regional climate Downscaling Experiment (MedCORDEX) programme with different atmospheric and oceanic models and long-term statistical studies to assess the climatological impact of air–sea coupling on the representation of intense rain events using the full duration of the available MedCORDEX simulations (1989–2008, to be continued up to the present).

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Supporting information

The following supporting information is available as part of the online article:

**Figure S1.** Geopotential height and wind direction at 500 hPa on 19 September 1996 at 0600 UTC.

**Figure S2.** CAPE and LFC on 19 September 1996 at 0600 UTC.

**Figure S3.** Geostrophic wind in CPL simulation at 0600 UTC on 19 September 1996.

**Figure S4.** Surface friction velocity of CPL at 0600 UTC on 19 September 1996.

(a) Difference of surface friction velocity on 19 September 1996. (b) Difference of surface friction velocity on 19 September 1996 at 0600 UTC.


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