Coastally-Trapped Disturbances caused by the Tramontane wind on the North-Western Mediterranean: Numerical study and Sensitivity to Short-wave Radiation

Abstract

The Tramontane-Cierzo wind system is a recurrent feature of the north-western Mediterranean basin in front of Catalan coast (NE Spain). Associated with this feature, northeast wind surges affect occasionally the coast and become a weather hazard for low-level aircraft operations, affecting for example the Barcelona international airport. This paper first reports these surges characterizing them as Coastal-Trapped Disturbances (CTDs). Climatological features are described, showing that CTDs occur frequently during the warm season and between the afternoon and evening. We classified CTDs related to two synoptic patterns related to the location of a mid-level tropospheric geopotential trough and the Iberian Peninsula: pattern A, with the trough crossing eastwards north Spain; and pattern B, with the trough over the Mediterranean, after crossing the Iberian Peninsula. To study the CTDs in detail, numerical simulations were conducted using the non-hydrostatic and convection-permitting NWP model HARMONIE-AROME. Two cases, one for each synoptic pattern, were studied showing that CTDs generate in the discontinuity between cool outflows and warmer air progressing southward as a density current, trapped by the mountain ranges parallel to the coastline. Cool outflows may have two different sources: in Pattern A the origin of the cold air is the Tramontane itself, while in Pattern B convective outflows associated with storm downdrafts play this role. Both cases show similarities with CTDs studied on the California coast, showing an antitriptic and ageostrophic flow behind the CTD. An additional numerical sensitivity experiment was conducted by varying the short-wave radiation to explore the effects of diabatic warming on CTDs. It is demonstrated that a large warming influences on CTDs by enhancing the potential temperature gradient

between the density current and the environment modulating its intensity and speed.

Keywords: coastal-trapped disturbances, low-level circulation, tramontane, sensitivity analysis, numerical simulation, HARMONIE-AROME, weather hazards, aeronautical hazards, western Mediterranean

1. Introduction

Coastal-Trapped Disturbances (CTDs) are a particular case of orographically trapped disturbances, defined as a lower atmosphere perturbation "laterally confined against a suitably large mountain barrier by Coriolis effects and, vertically, by stable stratification" (Reason, 1994). Typical CTD length scales are of the order of 1000 km alongshore and 100 km cross-shore, a life span of 2 to 6 days, and their presence usually implies a wind field reverse and strengthening as well as fog and stratus (Reason and Steyn, 1990). CTDs have been studied in many regions worldwide where dominant atmospheric low-level circulation and geographical features favour their occurrence, such as the Pacific coast of North America, where they are referenced as coastally trapped wind reversals (CTWR) (e.g. Mass and Albright 1987; Nuss et al. 2000; Rahn and Parish 2008; Parish et al. 2015), South Africa (Gill, 1977; Reason and Jury, 1990) and south-eastern Australia (Reason, Tory and Jackson, 1999) among others.

In this paper, we examine mesoscale circulations in the North-Western Mediterranean basin with many similar features to classical CTDs but exhibiting shorter spatial and temporal scales than those previously mentioned. Specifically, Mediterranean CTDs examined have horizontal scales not larger than 200 km alongshore, 50 km cross-shore, the time span is of the order of a few hours instead of a few days, and they do not bring fog or low stratus. However, these events do produce a sudden change in the wind field direction and strength which is hardly captured by operational NWP models, disturbing aeronautical takeoff and landing operations in coastal airports, as has been reported at the Barcelona international airport. Moreover, this study may contribute to a better understanding of low-level circulations of the region which play a crucial role in providing a source of moisture for heavy precipitation events as pointed out in recent case studies examined in the framework of the HyMeX programme - see for example Lee et al. (2017), Röhner et al. (2016) and Bouin et al. (2017), the last two dealing with the an extreme event that holds the all-time record for 4, 5 and 6h rainfall amounts (216 mm, 248 mm and 275 mm) as reported by Gonzalez and Bech (2017).

The main aim of this paper is to provide an overview and description of Coastally-Trapped Disturbances caused by the Tramontane wind on the North-Western Mediterranean to improve our understanding of the mechanisms involved in their formation and evolution. To accomplish these objectives, a background of regional wind-systems in the area of study, the Tramontane-Cierzo winds, is provided in Section 2 and a brief climatology of observed events is presented in Section 3. The HARMONIE-AROME NWP model used and the sensitivity experiments performed are described in Section 4. Section 5 presents and describes the evolution of two different simulated events. Finally, Section 6 compares the sensitivity of the surges to short-wave radiation. A brief summary and conclusions are presented in Section 7.

2. Background and Area of study

Tramontane-Cierzo wind system is the western part of Mistral-Tramontane wind system and takes place during synoptic north and north-western flow events over the North-Western Mediterranean basin, when the incident flow characterized by a low Froude number hits the north side of Pyrenees mountain range (Georgelin and Richard, 1996). This 400 km long west-east oriented range is a major sink of meridional momentum (Bessemoulin *et al.*, 1993) forcing the north and north-west low-level flow to go around both the west and east sides of the mountain range, progressing through two gaps: Ebro valley on the right-hand side of the flow direction (Cierzo wind) and the wide area between Pyrenees and Massif Central on the left-hand side (Tramontane wind) as shown in Figure 1a. These ducts favour an additional acceleration to both flows and canalize them towards the Mediterranean Sea. The interaction of the flow with the mountain barrier produces a perturbation in the surface pressure field at sub-synoptic scale (Bénech *et al.*, 1998) that creates a mesoscale pressure dipole with a high-pressure system windward the Pyrenees mountain range and a low system leeward (Figure 1a).

Tramontane-Cierzo wind system was widely studied during PYREX experiment (Bougeault *et al.*, 1990, 1997; Genovés *et al.*, 1994; Flamant and Pelon, 1996). Bénech et al. (1998) and Koffi et al. (1998) showed a flow asymmetry during northerly flows induced by the Coriolis effect which causes Tramontane to be stronger than Cierzo. Campins et al. (1995) studied the main structure of the Tramontane and found a low-level jet just below an inversion layer located around 1000 m in altitude. They also found that Tramontane was at first accelerated by the pressure gradient force but partially compensated by the frictional effect. When the Tramontane reaches the sea, the frictional effect abruptly drops off, the acceleration is enhanced and wind speed reaches its maximum intensity offshore over the Gulf of Lion (Vazquez, 1995; Georgelin and Richard, 1996). Once the wind acceleration reaches its maximum, the flow becomes inertial and is decelerated by the friction becoming a density current (Campins *et al.*, 1995). Between the outflowing cold dry air of the Tramontane and the previously existing warm and moist Mediterranean air a cold-front is created.

Due to the Coriolis force Tramontane turns anticyclonically (Campins *et al.*, 1995) forming a recurrent convergence line at the Catalonian northeast coast and Balearic Sea when it meets the relatively warm-dry Cierzo wind (Pascual and Callado, 2002) (Figure 1a). This convergence line is not always static, and occasionally moves southward reaching Barcelona as a northeast (NE) surge, where the wind may get reinforced, changing suddenly of direction –from SW to NE– and speed, which can exceed 70 km h^{-1}. This wind surge occurs sometimes with no significant cloudiness and becomes a weather hazard for aircrafts during the take-off and landing operations at the Barcelona airport (Gonzalez and Pascual, 2013). Aircraft pilots' reports describe this phenomenon as a sudden change from tailwind to headwind between 2000 and 3000 ft. when they land at the Barcelona airport from the south-west using the runway 07 (landing in direction 70º in order to land against the wind).

In Catalonia there are two mountain ranges parallel to the coast (Figure 1c). Catalan Coastal Range is the nearest to the sea (less than 1 km away), and has a height scale of about 300 m with the highest peaks between 600 and 750 m. Catalan Pre-Coastal Range is about 20 km away from the sea, and is higher than the latter with a height scale of about 1000 m and the highest peaks over 1700 m. The main gap over both mountain ranges is the relatively narrow Llobregat river valley (about 10 km wide), located a few kilometres south-west of Barcelona.

3. Overview of NE surges at Barcelona

We checked the observational database of AEMET (Spanish Meteorological Service) to characterize climatological aspects of relevant NE surges at Barcelona associated with Tramontane from 2010 to 2016. We selected the cases when the wind gusts at the AEMET surface station 0201D in Barcelona Meteorological Center exceeded 14 m s^{-1} (50 km h^{-1})

and, its direction was comprised between 20º and 100º (roughly NE). At the same time wind over Reus surface station, 100 km south of Barcelona, had to be blowing from directions between 180º and 360º (westerlies associated with the Cierzo) in order to filter synoptic eastern and north-eastern flows. Finally, a visual inspection of observational data (temperature, humidity and wind) at Barcelona station 0201D checking for the presence of a sudden air mass change allowed us to accept or reject each case individually. Notice that we have chosen 14 m s^{-1} gusts in Barcelona as a threshold to filter out weaker surges.

Table 1 shows a list of the events occurred between 1 January 2010 and 1 December 2016. Although Tramontane blows often in the cold season, when the synoptic circulation associated with the jet stream moves to a low latitudes, most episodes are comprised between March and October. This suggests a seasonal behaviour with a maximum centred in the warm season. Notice that NE surges are more common in spring or autumn than in summer, due to the seasonality of the Tramontane. Furthermore, Table 1 shows a preferred time of occurrence of the wind surges, with a maximum between the afternoon and the evening. Both facts suggest that land warming is a key ingredient in these surges as will be discussed in detail in Section 5.

A qualitative analysis based in operational synoptic and mesoscale charts and satellite imagery allows us to classify NE surges into two main different synoptic patterns (shown at Table 1).

- Pattern A is linked to a 500 hPa synoptic trough crossing north Spain with an associated surface cold front. When the front arrives at the Gulf of Lion, the cold surge spreads over the Balearic Sea forming a meso-alpha front that eventually reaches Barcelona. This pattern is often associated with a NE surge overrunning the well-established southwestern sea-breeze.

- Pattern B is associated with a synoptic trough that has already crossed Spain and is located over the Mediterranean Sea. Low level winds behind the trough flow from west or north-west. Tramontane blows directly towards the Mediterranean. We found that all cases in this pattern presented convective clouds with cool pools flowing out to the south forming a mesofront that eventually could reach Barcelona. On this pattern, NE surges often overruns the established western wind (associated with the Cierzo).

Of the 28 events found from 2010 to 2016, there are four events (14%) that could not be classified into patterns A and B (events E01, E05, E09 and E1). Two representative events, one for each pattern, are studied in Section 4.

4. Methodology

4.1 NWP model description

The HARMONIE-AROME (hereafter HARMONIE) meso-scale convection-permitting non-hydrostatic NWP model (Bengtsson *et al.*, 2017) has been used to simulate and investigate Tramontane-Cierzo system and CTDs in north-western Mediterranean coasts (see simulation domain in Figure 1b). HARMONIE is developed by the HIRLAM consortium and it is based on the ALADIN consortium AROME model (Seity *et al.*, 2011) within the framework of the HIRLAM-ALADIN consortiums' joint project.

HARMONIE version 37h1.1 has been integrated on event simulations with 2.5 km horizontal resolution, 65 sigma-pressure hybrid vertical levels and an integration time step of 60 seconds. With this resolution is expected to simulate properly CTDs and their related meso-scale flows due to an accurately enough representation of the Pyrenees mountain massif, the coastal mountain ranges and the coastline. Moreover, the area of the simulation is wide enough to include all mesoscale circulations related to Tramontane-Cierzo system. HARMONIE has been run on case studies up to 24 hours with outputs every hour taking boundary conditions hourly from the global ECMWF-IFS at T1279 (~ 16km).

4.2 Sensitivity experiments

HARMONIE sensitivity experiments of Tramontane-Cierzo system and CTDs due to diabatic heating have been done varying constant solar value (S) on Morcrette short-wave radiation scheme (ECMWF, 2015) called every 15 minutes. Several experiments with distinct S values have been investigated but only two relevant ones are shown in this paper:

- SW06: where S is multiplied by a 0.6 factor decreasing its original value by 40% and,
- SW12: where S is multiplied by a 1.2 factor increasing its original value by 20%.

The results are discussed on Section 5.

4.3 Analysis of the forcing mechanisms

In order to estimate the dynamical force balance and to validate if the forcing mechanisms of the CTD in Barcelona are comparable to the CTWR studied by Rahn and Parish (2008), we computed each term of the horizontal equation of motion:

$$\underbrace{\frac{\partial V}{\partial t}}_{\mathrm{ACC}} = \underbrace{-V \cdot \nabla V}_{ADV} \underbrace{-\nabla \Phi}_{PGF} \underbrace{-fk \times \mathrm{V}}_{COR} + R \qquad (1)$$

simulated by HARMONIE at 1000 hPa. The left-hand side of Equation 1 corresponds to the local acceleration of the horizontal wind (ACC), the first term on the right-hand side is the horizontal advection of the wind (ADV), the second term is the pressure gradient force (PGF), the third term is the Coriolis force (COR), and the last term is the residual (R) that includes surface drag, turbulent mixing and the vertical advection of the wind.

5. CTDs Case Studies

To examine how CTDs evolve in the NW Mediterranean and to explore the forcing that reinforces the wind in the area of Barcelona we have studied in detail two cases, each one representative of each synoptic pattern presented in Section 2. To ensure the reliability of the forcing, we have chosen cases where Tramontane outflow and CTDs were well simulated by the model HARMONIE.

5.1 Pattern A: 28 May 2013

5.1.1 Synoptic setting

Figure 2 shows the synoptic ECMWF analysis on 28 May 2013 at 12:00 UTC. At 500 hPa pressure level (Figure 2a) there is a well-defined trough with a cold core over northern France, with its axis extended along northern Iberian Peninsula. The diffluent downstream section of the trough is located over NW Mediterranean, and a strong temperature gradient surpasses the Pyrenees from northwest to southeast. A temperature gradient along the Pyrenees and the Gulf of Lion is also evident at 850 hPa where a cold front is crossing (Figure 2b). Behind the front, Tramontane-Cierzo system develops. A high sea level pressure (SLP) gradient between the high-pressure area around the Azores and the low-pressure area at the English Channel yields to a long and strong north-western flow over the Pyrenees. The synoptic SLP field is perturbed over the Pyrenees, where the characteristic mesoscale pressure dipole can be identified.

5.1.2 Mesoscale evolution

Figure 3 shows the evolution of the episode using Meteosat Second Generation High Resolution Visible (MSG HRVIS) images and HARMONIE NWP model output. HRVIS (Figure 3a) images show at 12:00 UTC cloudy skies over a large portion of the North-Western Mediterranean, mostly inland, and open cells over the western Pyrenees associated with the cold air mass. Despite there are not many shadows since the sun is close to the zenith, some enhanced clouds with a rope shape can be observed. These enhanced clouds are placed where HARMONIE simulation shows a strong convergence zone (Figure 3b), most of them related to an abrupt change of pseudo-equivalent temperature that exceeds 10 K in few kilometres. Steep changes in pseudo-equivalent temperature (θ_{ep}) are associated with air mass boundaries and rope shape convergence zones where these two air masses meet (Figure 3c).

According to the HARMONIE simulation, three low level air masses collide into a Y shape convergence zone over Catalonia. Figure 4 shows 1000 hPa temperature and moisture content of each air mass at 16:00 UTC. Both variables define θ_{ep} depicted at Figure 3e. Continuous shaded colours (temperature, moisture and θ_{ep} fields in Figure 3 and Figure 4) highlight the abrupt changes at the air mass boundaries. Tramontane air mass (TrAM) is cold and dry with temperatures around 287 K and specific humidity widely below 7 g kg^{-1}. This air mass is related with strong northern winds that make up the Tramontane. Cierzo air mass (CiAM) is very dry (below 4 g kg^{-1}), and warm (293 K), especially downstream.

As explained in Section 2, TrAM and CiAM have the same common source over the Atlantic Ocean and they form when the Atlantic air mass is forced to go around the Pyrenees. TrAM retains almost all its moisture by going across a flat and relatively low altitude zone. Conversely, CiAM loses almost all its moisture when crosses over mountains west of the Pyrenees and is diabatically heated during the daytime hours as it moves along the Ebro valley (see Figure 3). At night, this air mass loses quickly its heat (not shown) following a diurnal cycle behaviour. The remaining air mass located over the Mediterranean Sea (Figure 4) was already described by the pioneering work of Jansà (1959) who called it Mediterranean air mass (MedAM). MedAM is the most representative air mass during the warm season over the Mediterranean Sea when the air becomes relatively stagnated and is characterized by a large amount of low-level moisture with values exceeding 9 g km^{-1} and temperatures slightly higher than TrAM. Considering the buoyancy of the MedAM as zero, TrAM has negative buoyancy and CiAM positive buoyancy. Broadly, changes in moisture are greater than changes in temperature. This can be seen clearly at Figure 3 where TrAM and CiAM have low values of θ_{ep} and, in contrast, MedAM presents large values.

HRVIS in Figure 3d shows the advance of the TrAM boundary corresponding to the Tramontane outflow leading front, against MedAM and CiAM at 16:00 UTC. HARMONIE shows a thermal difference of about 3 °C between TrAM and MedAM in the alongshore cross-sections (Figure 5a) suggesting that flow is driven by the thermal contrast as a density current where TrAM acts as a cold pool spreading over the Mediterranean. From 16:00 UTC at the leading edge of the Tramontane, a strong wind surge next to the coast north of Barcelona can be seen. This surge is enhanced at the coast, and it weakens offshore like a classical CTD. Behind the Tramontane low level front, wind is strong and mainly oriented alongshore heading to southwest (Figure 3f). However, there is a weak on-shore component that can be related as the result of both the curvature effect of the Coriolis force (Campins *et al.*, 1995) and the thermal up-slope flow over land. This small inland component is a key element in the generation of the CTDs.

According to the observational data shown in Figure 6, NE surge reached Barcelona at 15:20 UTC. In few minutes, the wind shifted and intensified with a gust of 18.8 m s^{-1} (67.7

km h^{-1}), and the temperature dropped 3.8 °C in 10 minutes and 6 °C in 30 minutes. However, the outflow simulated by HARMONIE at 16:00 UTC is located north of Barcelona, being the front delayed about 1 hour. Though there are some important differences such as the velocity of the front, HARMONIE reproduces quite well its movement. A comparison between simulated 1000 hPa winds and data from the Advanced Scatterometer (ASCAT) observations of METOP satellite passes at 20:24 and 21:10 UTC (not shown) also supports the consistency of the simulation with observational data. This suggests that HARMONIE is able to simulate the key forcing mechanisms of the event, although some of them might be underestimated.

5.1.3 Cross section analysis

The alongshore cross-section at 16:00 UTC in Figure 5a shows a large stratification over the outflow which has a clear density current structure. The head and the body of the density current reach approximately 850 hPa. The head can be identified in the HRVIS imagery as a rope cloud (Figure 3a,d) as a consequence of the forced lift while the strongly stratified capped environment does not allow strong updrafts over a deep layer. Due to the stratified stability, some Kelvin-Helmholtz waves are developed behind the front enhancing the turbulent mixing between the upper MedAM and the lower TrAM similarly to the numerical simulations described by Xu et al. (1996).

Figures 3h and 3i show the interaction between the three air masses simulated by HARMONIE at 18:00 UTC, when the mesofront along the coast reaches CiAM. The outflow boundary increases its thermal difference to 5 K and the convergence intensifies. Figure 5c shows that CiAM, with lower buoyancy, is forced to overlay the front, increasing the static stability of the air column behind the head front, which leads to a reduction of the boundary layer height and a buffering of the Kelvin-Helmholtz waves (see white arrow). By contrast, head wave deepens beyond 800 hPa, probably due to the environmental vertical shear that points in the direction of outflow motion (Xue, 2000). Notice that the perturbation in the wind field at 18:00 UTC (Figure 3i) extends towards the sea about 50 km, which approximately corresponds to the Rossby radius for this flow (Table 2) which is dynamically coherent with CTDs.

Figure 5b shows the cross-shore A'B' section at 16:00 UTC before the arrival of the density current front, illustrating that MedAM is restricted to the bottom layer of the atmosphere over the sea with weak cross-shore wind. Inland, CiAM flows strongly offshore over the MedAM. At 18:00 UTC after the front crosses the A'B' section (Figure 5d) a MedAM thin layer (higher θ_{ep}) remains between the less buoyant TrAM on the bottom and the more buoyant CiAM on the top. Behind the front, low level on-shore wind is blocked by the mountain, as suggested by the wind deceleration and the MBL step.

In the last steps of the episode, the outflow front advances steadily until 20:00 UTC and then the wind speed suddenly drops. The front becomes stationary at 22:00 UTC around 100 km south of Barcelona. It is worth to remark that θ and θ_{ep} of CiAM, unlike TrAM and MedAM, drop in the afternoon a few degrees (not shown) as a response to the diabatic cooling associated with the diurnal cycle and the low moisture content of the air mass. As a result, there is a decrease in the thermal contrast, and therefore in the density difference of the boundary that weakens and may eventually stop the density current.

5.1.4 Scale analysis

An approximate calculation indicates a high static stability of the outflow with a Brunt-Väisälä frequency value (N) for the boundary layer approximately $1.5 \cdot 10^{-2}$ s^{-1}. A scale analysis indicates the degree of blocking of the hydrodynamic regime by the topographic features: it is almost blocked by the Coastal Range – values of mountain Froude number (F_m) and Burger number (B) of 1.1 and 9, respectively, as seen in Table 2 –, and totally blocked by the Pre-Coastal Range – $F_m \sim 0.3$ and $B \sim 15$. Values for Coastal Range are close to the largest blocking response described by Overland and Bond (1995) that is, a stepped flow with a $F_m \sim 1$ and $B > 1$.

5.1.5 Forcing analysis

Figure 7a shows the motion equation terms at 16:00 UTC when the simulated front is still mature and has not reached Barcelona yet. The largest gradient on the height field and the convergent wind determines the front position. In the bottom left of the plot, where the CiAM is present, the advection term is dominant and is pointing eastwards. The residual term is opposed to advection and almost balances the flux. In the bottom right of the plot, corresponding to the MedAM, the balance of the wind is quasi-geostrophic (the PGF counterweights Coriolis force cancelling the acceleration). The edge of the outflow is mainly accelerated by the PGF towards the south-west due to the strong pressure gradient at the density current front. Advective and residual components compensate some PGF forcing. Behind the front, weak Coriolis force provides a small rotation onshore. Scale analysis of the Rossby number around 1 (V ~ 10 m s^{-1} and L ~ 100 km), suggests the rotational forces affect the flux to some extent.

At 19:00 UTC (Figure 7b), behind the front, where the outflow is well-established, the balance within the TrAM becomes almost antitriptic (Schaefer and Doswell, 1980), with the residual balancing the PGF. This analysis suggests that forcings at low levels in this event are very similar to those in the CTWR studied by Rahn and Parish (2008), and therefore we could conclude that motion is little scale-dependent and well described by a density current ageostrophic acceleration.

5.2 Pattern B: 30 March 2013

5.2.1 Synoptic setting

Temperature and geopotential field analysis at 500 hPa on 30 Mar 2013 at 12:00 UTC (Figure 8a) shows a broad trough with its axis located eastward of the Iberian Peninsula. NW Mediterranean coast is therefore located below the upstream branch of the trough and eastward of a small ridge located west of Spain. The cold air advected by the trough creates the proper environment to support deep moist convection. At low levels (Figure 8b), relative high pressures southwest of Iberian Peninsula and the low located over the Genoa Gulf, generate a MSLP gradient leading the flow from Atlantic Ocean directly to Italy. The Tramontane-Cierzo wind system is already well developed over the NW Mediterranean as indicated by the associated meso-low leeward of the Pyrenees (Figure 8b).

5.2.2 Mesoscale evolution

The HARMONIE simulation sequence indicates that MedAM was removed from NW Mediterranean some hours before the onset of the event by the Cierzo and the Tramontane (not shown). The simulation is consistent with the wind field derived from ASCAT observations at 9:21 and 10:06 UTC (not shown). Unlike the previous event, there are not three well-defined air masses at 13:00 UTC (Figure 9). According to the low-level air features (Figure 10), air masses north and south of the Pyrenees are identified as TrAM and CiAM respectively. Cloudiness in TrAM allows us to differentiate it from CiAM in HRVIS at 13:00 UTC (Figure 9a) being separated by a shear line (see the discussion by Jansá 1987).

Close to the easternmost edge of Pyrenees, diurnal heating and low-level convergence produced by the leeward meso-low released atmospheric convection. Inside the TrAM, the cool pool associated with several convective cells is evidenced by a disturbance in the pressure field coming out of the convective zone and progressing to the south (not shown). Those outflows cooled down low-level air near the convergence zone enhancing the thermal contrast with the CiAM. According to the simulation at 15:00 UTC (Figure 9e), when the temperature difference between the cold air mass and the CiAM achieves 5 K, the convergence line starts to advance south-westwards as a density current, showing a CDT at the coastal current edge (Figure 9f).

At this stage, cold air in the density current is not directly related to the Tramontane since TrAM remains separated from the cool convective outflow of the storms. That is a significant difference with the previous case, since the source of the cold outflow on 28 May 2013 is not convective. This makes this case more similar to the event studied by Gonzalez and Pascual (2013) (see Table 1), where a similar convective outflow was observed by radar affecting Barcelona airport as a CTD. Both cases have the same synoptic pattern defined as B at Section 2.

Figure 9 d and g shows that this outflow takes place in clear air and no signal at HRVIS reveals its presence. Unfortunately, unlike earlier in the morning, there are no ASCAT observations available for the period when the CTD is developed so no further comparisons are possible. Therefore, we should relieve in instrumental observations to localize the outflow edge. As shown in Barcelona AWS (Figure 11), the outflow reached at 16:40 UTC when the wind suddenly changed from west to east-northeast as the mean wind speed increased. As a consequence of the outflow, the temperature dropped 5 °C in 40 minutes but wind gusts did not show a significant increase since previous western wind, the Cierzo stablished at Barcelona, was already gusty. On this occasion, unlike the 28 May 2013, the CTD simulated by HARMONIE reached Barcelona slightly ahead of the observation time.

5.2.3 Scale and Cross Sections analysis

The scale analysis of this case shows a higher stability ($N \sim 2\cdot10^{-2}$ s^{-1}) than the previous case. Parameters F_m and B are 0.8 and 12 respectively for the Coastal Range, and 0.3 and 20 for Pre-Coastal Range (Table 2), providing the right environment for a large blocking response. These results, as well as similar cross-sections to the previous event (not shown), suggest that the convective outflow of this event has similar buoyancy features than Tramontane in the case of 28 May 2013.

5.2.4 Forcing analysis

The motion terms analysis (Figure 12) at 16:00 UTC when the simulated front reaches Barcelona, shows that outflow edge acceleration is mainly forced by PGF and advection. PGF provides the primarily alongshore forcing, while advection is directed onshore. As a result, the wind has some inland component primarily caused by the advection while Coriolis acceleration adds additional forcing. The onshore component slightly modifies the main along-shore balance that is slightly antitriptic. Part of the residual term opposes PGF and the other part is used to compensate some onshore forcing. On the whole, we can conclude that convection outflow plays the same role than Tramontane in the case of 28 May 2013, behaving as a density current driven by sharp differences of temperature and pressure.

6. Sensibility of Mediterranean CTDs to diabatic heating

In Section 2, we suggested a relation between diabatic warming and CTDs at NW Mediterranean, since they tend to occur in the warm season and between the afternoon and evening. In this section, we further explore this possibility by performing a sensitivity analysis of the short-wave (SW) radiation, as a proxy of diabatic warming with the HARMONIE simulations, as described in Section 3.

Figure 13 compares the SW12 experiment with the SW06 (increased vs. decreased SW radiation, see Section 3), on 30 Mar 2013 at 15:00 UTC (pattern B). The main features that change when diabatic forcing is modified are the intensity, the extension and the location of the main convergence zone associated with the CTD. Indeed, when the short-wave radiation is increased (SW12 experiment), CTD extends and accelerates, and when the short-wave radiation is decreased (SW06 experiment), it has less extension and decelerates (see Figure 13c and 13d compared to Figure 9f, and Figure 13a and 13b compared to Figure 9e). This is not surprising since when the short-wave radiation is increased, the sensible heat on dry CiAM warms more than on TrAM. This yields to a greater temperature gradient between the outflow and the environment, that leads to an increase of the density current speed according to the idealized relation (Markowski and Richardson, 2010)

$$U_c \sim \sqrt{-\frac{\theta'}{\bar{\theta}_v} gH}\,, \qquad (2)$$

where U_c is the speed of the density current, θ' is the density potential temperature perturbation at the surface, $\bar{\theta}_v$ is the mean virtual potential temperature of the environment, g is the gravity acceleration and H the depth of the outflow.

This expression agrees with the fact that at night, when temperature of CiAM drops as a response to the diabatic cooling, the mesofront becomes stationary (not shown). So, the diabatic warming of the CiAM may dramatically impact on the ability of the CTDs to reach further south, in particular to the Barcelona airport. Hence, when the diabatic effects are low, the gradient of potential temperature between the outflow and the environment is reduced and the movement of the CTD stops sooner. Indeed, in the SW06 experiment, the mesofront never reaches Barcelona (not shown). Similar results were obtained on 28 May 2013 (pattern A, not shown). This could explain why Mediterranean CTDs are more often in both the warm season and in the evening, when the diabatic forcing is larger.

6. Concluding Remarks

Every year, several NE wind surges associated with CTDs affect the NW Mediterranean area, causing potentially hazardous situations to low-level aircraft operations, affecting for example the Barcelona airport (Gonzalez and Pascual, 2013). In this paper, we first report these events as CTDs and show their main climatological and meteorological features.

CTDs at NW Mediterranean take place when synoptic northern flow impinges air with low Froud number around the Pyrenees and is directed towards the Mediterranean generating the Tramontane-Cierzo wind system. Due to the meso-low developed leeward of the Pyrenees and the Coriolis effect, Tramontane tends to curve anticyclonically producing an on-shore component of the wind that can be eventually blocked by the orography if the buoyancy of the air mass is small. This air mass flows as a density current as it collides with a much warmer air mass with a higher buoyancy, like CiAM or MedAM. CTDs occurs when $B > 1$ and $F_m \sim 1$, described by Overland and Bond (1995) as the largest blocking response environment. The origin of the cool air mass may be the Tramontane itself or may have a convective origin and probably it depends on the synoptic framework. We have identified two different synoptic patterns producing CTDs:

1. A mid-level geopotential trough with an associated surface front crossing north Iberian Peninsula in warm conditions. The cold surge over Gulf of Lion spreads

over the Mediterranean Sea into a meso-alpha front that tends to curve anticyclonically, converging with MedAM and CiAM. In this case there is no convection associated and TrAM acts as a density current overrunning the previously stablished sea-breeze.

2. A mid-level geopotential trough has already crossed north Iberian Peninsula and has swept the MedAM away. Diurnal heating and low-level convergences in the easternmost edge of Pyrenees triggers convection. In this case, the cool outflow from storms enhances the thermal gradient with the CiAM generating a density current. The hazardous event at Barcelona airport studied by Gonzalez and Pascual (2013) is associated with this pattern.

Their conceptual model is described in Figure 14. Although both cases have a very different cool air source, the interaction between the density current and the orography along the coastal range is similar, creating an antitriptic balance where the wind is accelerated ageostrophically. The mechanism that drives CTDs in NW Mediterranean is therefore quite similar to CTWR largely studied in the California coast (Rahn and Parish, 2008; Parish, Rahn and Leon, 2015), but the hazardous weather effects produced are different. Since CTDs in NW Mediterranean area establish a relatively cool and dry air over a warm sea, no low and thick stratus or fog are produced abundantly as in California CTWR cases, although the amount of cloudiness may increase. Instead, as we showed in this paper, the principal hazards are the sudden speed increase and direction shift of the wind, that may affect aeronautical operations, for example at the Barcelona airport. In addition, NE wind surges are associated with a dramatic drop of the temperature as well as a sudden increase of the humidity. This pattern occurs along the cold mesofront, but its effect is larger by the coast where the wind is locally accelerated, especially in the area of Barcelona, where gap effects due the Llobregat river valley could play a role in this enhancement (see the discussion of valley influence on CTD at Reason et al. 2000).

Through sensitivity of HARMONIE-AROME NWP model simulations we have shown that short-wave radiation warming CiAM largely influences on the CTD development and motion by increasing the potential temperature gradient between the density current and the environment air. This explains the annual and daily distribution in the climatology of the events with a maximum frequency in the warm season and between afternoon and evening.

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Table 1. Climatology of NE surges exceeding 14 m s^{-1} (50 km h^{-1}) at Barcelona between 2010 and 2016. Superindex (1) indicates the Gonzalez and Pascual (2013) case study and (2) show the cases studied in Section 4 (ES10 and ES13).

ID	**Date**	**Time**	**Gust** [m/s]	**Direction** [deg]	**Synoptic Pattern**
ES01	17/10/2010	15:40:00	15.0	68	Other
ES02	17/07/2011	11:30:00	16.3	83	A
ES03	26/08/2011	18:40:00	15.3	95	A
ES04	19/10/2011	21:00:00	14.1	76	A
ES05	05/03/2012	16:20:00	15.7	74	Other
ES06	19/04/2012	19:10:00	14.3	74	A
ES07 (1)	12/06/2012	16:10:00	23.2	72	B
ES08	21/06/2012	19:10:00	14.8	86	A
ES09	24/02/2013	16:40:00	15.0	87	Other
ES10 (2)	30/03/2013	17:10:00	15.6	81	B
ES11	02/04/2013	16:50:00	15.5	69	B
ES12	08/04/2013	17:00:00	14.4	77	B
ES13 (2)	28/05/2013	15:20:00	18.8	69	A
ES14	26/2/2014	18:00:00	15.2	73	B
ES15	29/6/2014	19:10:00	14.7	87	B
ES16	20/7/2014	18:10:00	17.0	81	B
ES17	29/7/2014	14:10:00	20.2	84	B
ES18	13/8/2014	09:40:00	15.8	86	A
ES19	1/12/2014	00:30:00	15.7	84	Other
ES20	15/5/2015	14:20:00	18.4	68	B
ES21	15/8/2015	16:30:00	14.9	85	B
ES22	24/8/2015	16:50:00	14.0	64	A
ES23	2/3/2016	20:30:00	14.1	100	A
ES24	23/4/2016	16:10:00	14.9	76	B
ES25	22/5/2016	18:30:00	14.1	98	A
ES26	29/5/2016	19:00:00	14.9	75	B
ES27	17/9/2016	20:10:00	14.9	90	B
ES28	25/9/2016	17:30:00	17.0	40	B

Table 2. Scale analysis for each case study, including the characteristic incident wind (U), the Brunt-Väisälä frequency (N), the mountain range characteristic height (H), and length (L), the mountain Froude number (Fm,) the Burger number (B) and the Rossby Radius(L_R).

Case study	**Mountain Range**	**U**	**N**	**H**	**L**	**F_m**	**B**	**L_R**
						$UN^{-1}H^{-1}$	$HNf^{-1}L^{-1}$	HNF_mf^{-1}
		[ms^{-1}]	[s^{-1}]	[m]	[m]			[m]
28 May 2013	Litoral	5	1.50E-02	300	5000	1.1	9.0	50000
	Prelitoral	5	1.50E-02	1000	10000	0.3	15.0	50000
30 Mar 2013	Litoral	5	2.00E-02	300	5000	0.8	12.0	50000
	Prelitoral	5	2.00E-02	1000	10000	0.3	20.0	50000

Figure 1. a) Area of study with the principal locations used in this paper and the conceptual model of the Tramontane-Cierzo System. Yellow arrows show the principal fluxes around the mountains. Red ‘A’ and blue ‘L’ indicate the orographic pressure dipole. Red dashed line indicates the recurrent convergence line. b) The domain used by the numerical simulation (in colours, showing the height) and the regions displayed in panels a) and c) (in red). c) Position of the Catalan Coastal Range (in green) and Pre-coastal Range (in violet). Red dots and black numbers show respectively the position and the height (in m) of the highest mountains for each Range.

Figure 2. ECMWF synoptic analysis on 28 May 2013 at 12:00 UTC. a) Isohypses (solid lines, in gpm) and temperature (dotted, in °C) at 500 hPa. b) SLP (solid, in hPa) and temperature at 850 hPa (dotted, in °C).

Figure 3. Evolution of the 28 May 2013 episode at 12:00, 16:00 and 18:00 UTC. Panels a, d, g show the MSG HRVIS images. Notice the mesofront as an enhanced cloud structure (marked with pink arrows). Panels b, e, h show the simulated pseudo-equivalent temperature (shaded colours in red), potential temperature (contour lines in black) and convergence zones over $0.3 \cdot 10^{-3}$ s^{-1} (shaded in grey) at 1000 hPa. Panels c, f, i show the simulated wind direction (arrows, in black), speed (shaded colours in green) and geopotential height at 1000 hPa (contour lines in orange). Location of Barcelona is marked as a blue dot, and elevations over 1000 m according to the model ground are shaded in black. *Cross-sections depicted in Figure 5 are marked in panels e and h.*.

Figure 4. a) Potential Temperature and b) specific humidity simulated by HARMONIE at 16:00 UTC. Blue dot shows the location of Barcelona. Notice the large gradient at the boundaries of each air mass.

Figure 5. Cross-sections of pseudo-equivalent temperature (shaded colours in red) and potential temperature (black contour lines) along the AB segments (panels a, c) and A’B’ segments (panels b, d) shown in Figure 3e, h at 16:00 UTC and 18:00 UTC.

Figure 6. a) Temperature (red) and dew point (green), b) wind (yellow) and wind gust (brown) speed and, c) wind (yellow) and wind gust (brown) direction from 10-minutely instrumental observations at AEMET station 0201D in Barcelona on 28 May 2013.

Figure 7. Wind barbs (black, in m s^{-1}), 1000 hPa height contours (grey lines, in gpm) and vectors representing the terms in the momentum square (see legend) simulated by HARMONIE on 28 May 2013 at a) 16:00 UTC and b) 19:00 UTC. Red dotted line delimits the different airmasses labeled at the sides. The location of Barcelona is depicted as a red point.

Figure 8. As in figure 2 but for 30 March 2013 at 12:00 UTC.

Figure 9. As in Figure 8 but for 30 March 2013 at 13:00, 15:00 and 17:00 UTC.

Figure 10. As in Figure 4 but for 30 March 2013 at 15:00 UTC.

Figure 11. As in Figure 6 but for 30 March 2013.

Figure 12. As in Figure 7 but for 30 March 2013 at 15:00 UTC.

Figure 13. As in Figure 9e,f but for the experiment SW12 (panels a, c) and the experiment SW06 (panels b, d).

Figure 14. Conceptual model of the CTD in the NW Mediterranean coast showing the Tramontane air mass (TrAM), the Cierzo air mass (CiAM) the Mediterranean air mass (MedAM) and the orographic interaction area for a) Pattern A, and b) Pattern B.