

Revealing Atmospheric von Kármán Vortex Street past a Severe Thunderstorm

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ABSTRACT: The atmospheric von Kármán vortex street (VKVS) is a periodic series of turbulent shedding vortices that form in the wake of topographic obstacles and influence the dynamics of the lower atmosphere. The present study demonstrates that such structures can also form downwind of a strong thunderstorm, which acts as a fluid obstacle to the tropospheric wind. A detailed multiplatform observation dataset is analyzed to identify an atmospheric VKVS past an extreme convective storm that occurred over the Mediterranean Sea in 2015. The dataset is then utilized to set up a high-resolution large-eddy simulation to numerically reproduce the VKVS under simplifying assumptions, which accounts for the essential dynamical features of a convective updraft in a tropospheric cross flow. It is proven that the interaction between the thunderstorm updraft and the tropospheric wind produces a regular pattern of shedding vortices, extending in the entire troposphere and persisting for long distances downwind. The generation mechanism is described, and a peculiar entrainment of momentum and energy from the updraft is revealed. The shedding vortices exhibit high peaks of wind speed and turbulent kinetic energy in the low troposphere and enhance heat and mass transport in both vertical and downwind directions. Overall, the VKVS can be categorized as a self-sustained convectively induced turbulence structure, extending the influence of thunderstorms beyond the usual ranges.

SIGNIFICANCE STATEMENT: Here, we investigate a poorly documented feature in the structure of a convective storm. Using a combination of multiplatform observations and idealized high-resolution numerical simulations, we identified and described the occurrence of a regular series of shedding vortices in the wake of an intense thunderstorm occurred over the Mediterranean Sea in 2015. These vortices are organized as an atmospheric von Kármán vortex street, resulting from the interaction between the strong convective updraft and the tropospheric wind. Wake vortices extend throughout the entire troposphere and persist long distances downwind, exhibiting high peaks of velocity and turbulent kinetic energy. This sheds new light on the impact of convective processes on the dynamics of the atmosphere, adding new insights into elusive phenomena hitherto overlooked.

KEYWORDS: Vortices; Severe storms; Radars/Radar observations; Idealized models; Primitive equations model

1. Introduction

The atmospheric von Kármán vortex street (VKVS) is a common feature detectable downwind of a topographical relief, like a prominent island, usually made visible from satellite by the presence of marine stratocumulus. It consists of a periodic generation of shedding vortices with an alternating rotation that perturbs the lower atmosphere dynamics by modifying the wind field and cloud patterns (Young and Zawislak 2006; Vadasz 2023; Horváth et al. 2020).

Before being observed in the atmosphere, the VKVS has been described and analyzed in fundamental studies in fluid dynamics, where this process is of interest for both theoretical

purposes and practical applications. In this context, the vortex street is produced by the separation of an unsteady flow from a solid body, typically having cylindrical symmetry. Also in oceanic and atmospheric flows, the generation of the vortex street has usually been detected past a solid isolated obstacle (Derakhshandeh and Alam 2019). However, there are indications that VKVSs may also be generated as a result of purely dynamic processes, specifically in the presence of transverse flows intense enough to act as a fluid obstacle. For example, recent high-resolution large-eddy simulations (LES) have proven that the overshooting top of severe thunderstorms can act as a topographic obstacle to the low stratospheric wind (O'Neill et al. 2021). In the last decades, the generation of VKVS in the wake of fluid obstacles has been mainly studied by analyzing vertical jets (or buoyant jets) in cross flow in a controlled environment, utilizing both observational (Huq and Stewart 1996; Keramaris and Pechlivanidis 2016) and numerical approaches (Devenish et al. 2010; Cintolesi et al. 2019). The typical experimental setup consists of a (buoyant) jet exiting from a nozzle placed at the bottom surface of the domain, directed perpendicularly to a constant cross flow. The jet-cross-flow interaction gives rise to various turbulent structures, including the steady counterrotating vortex pair and the upright wake vortices. The latter, extensively investigated by

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Fric and Roshko (1994) and Yuan et al. (1999), may be considered one of the first extensive investigations of the generation of VKVS behind fluid obstacles: The upright wake vortices consist of a series of unsteady vertical vortices in the jet wake, which show a VKVS-like pattern but are energetically very weak compared to the classical vortices behind a solid cylinder. Despite these fundamental studies, applying the results of the aforementioned laboratory experiments to the real atmosphere is challenging due to the much higher complexity of physical phenomena.

In atmospheric science, the hypothesis that a strong updraft can act as a fluid obstacle to the environmental wind was developed in the late 1950s and 1960s (Newton and Newton 1959; Fujita and Grandoso 1968). Although the analogy with the flow past a solid cylinder at high Reynolds numbers is a clear oversimplification (Markowski 2002), preliminary numerical studies have shown that the environmental wind around convective storms produces similar features, including the flow separation on the downwind side of the thunderstorm center (Rotunno and Klemp 1982). From real-field observations, the presence of shedding vortices in the wake of a supercell storm has been suggested by Lemon (1976), who analyzed radar reflectivity data. In this case, only anticyclonic vortices have been observed, in contrast with the classical von Kármán pattern. This peculiarity was explained using the similarity between the rotating supercell updraft and laboratory experiments on rotating cylinders.

To the best of the authors' knowledge, no high-resolution numerical simulations have been applied directly to study von Kármán vortices past a convective updraft yet, and no observation datasets have shown a clear regular pattern of them at present. This may be due to the characteristics of VKVSs, which are relatively low-energy structures that develop over long spatial distances. Moreover, they are an elusive feature to be detected by observing instruments: Satellite passive measurements are unable to probe directly the inner cloud structure (in the visible-infrared part of the spectrum) or dynamics (in the microwave), while ground-based radar can often investigate limited portions of the phenomenon, sometimes not including the large extent of wake flow (Günther et al. 2021) or suffering from beam attenuation. In addition, dynamical quantities (e.g., the wind field in upper levels) can be mainly estimated indirectly using remote sensing techniques, with significant uncertainty. Actually, most of the studies on thunderstorms have focused on the convective core region to investigate its complex internal dynamics (Coffer and Parker 2018; Montopoli et al. 2021), while little attention has been given to the wake region. The use of numerical simulations also poses some challenges: On the one hand, the lack of complete and accurate measurements raises the problem of setting up reliable simulations, which need to be based on realistic datasets. On the other hand, these processes are usually not captured by weather forecast models, which operate at lower spatial resolution and rely on empirical parameterizations to represent small-scale atmospheric phenomena, resulting in loss of accuracy (Mykolajchuk et al. 2023). For these reasons, the atmospheric VKVSs may often have been overlooked by studies not specifically designed

to detect them, even though these structures could be more common than is now believed.

Although the literature on this topic is limited, the presence of a VKVS is expected to play a significant role in atmospheric dynamics. The VKVS can extend the impact area of the thunderstorm several kilometers downwind, resulting in the transport of turbulence, heat, and mass, the consequences of which need to be identified and quantified. Turbulent wake effects are included among the possible sources of near-cloud turbulence (Lemon 1976; Lane et al. 2012) and convectively induced turbulence (Sharman and Trier 2019), which can be hazardous to aviation due to their unpredictable occurrence even in clear air far from the cloud active region.

The present work focuses, for the first time, on the identification and model description of atmospheric VKVSs in the wake of severe thunderstorms. First, the fundamental hypothesis of whether the updraft of a severe thunderstorm can act as a fluid obstacle to atmospheric wind and generate VKVS, similar to what occurs with topographical obstacles, is verified. Second, the genesis mechanisms, the characteristics, and the structure of VKVS shedding vortices are described, also quantifying their impact in the atmosphere. The study is based on the analysis of a real thunderstorm, through observation datasets and simplified high-resolution numerical simulations. Results from observations and simulations, which are not conclusive when considered individually, are synergistically paired to obtain a meaningful description of these structures.

The case study selected is an intense thunderstorm that occurred on 5 September 2015, for which a detailed multiplatform measurement dataset is available (Marra et al. 2017). Radar reflectivity observations reveal the presence of a regular VKVS pattern downwind of the storm updraft during its mature stage. ERA5 reanalysis data, together with microwave satellite and radar observations, are then used to set up a simplified high-resolution LES to reproduce the dynamical processes dominating the thunderstorm system. The simulations use accurate models for atmospheric turbulence and thermofluid dynamics (without the empirical parameterizations typically used in the meteorological models) to reproduce the dominant physical processes. The use of simplified simulations for investigating complex atmospheric phenomena is a well-established study strategy that has proven to be fruitful (Rotunno and Bluestein 2024; Giove et al. 2025). Starting from the classical configuration of a buoyant jet in cross flow (Cintolesi et al. 2019), the simulation successfully reproduces the essential features of a strong convective updraft developing in an environmental crosswind and the VKVSs in the wake of the thunderstorm system, with satisfactory accuracy compared to the real phenomenon. This allows for an accurate study of the genesis and dynamics of VKVSs behind severe thunderstorms.

The paper is organized as follows: Section 2 illustrates the multiplatform observation datasets and describes the severe convective storm characteristics and the atmospheric VKVS; section 3 reports the simulation design and implementation, including details of physical assumptions and the numerical model used; section 4 shows the results of the simulations focusing on the description of the VKVS generation and development, as well as of the distinctive characteristics identified;

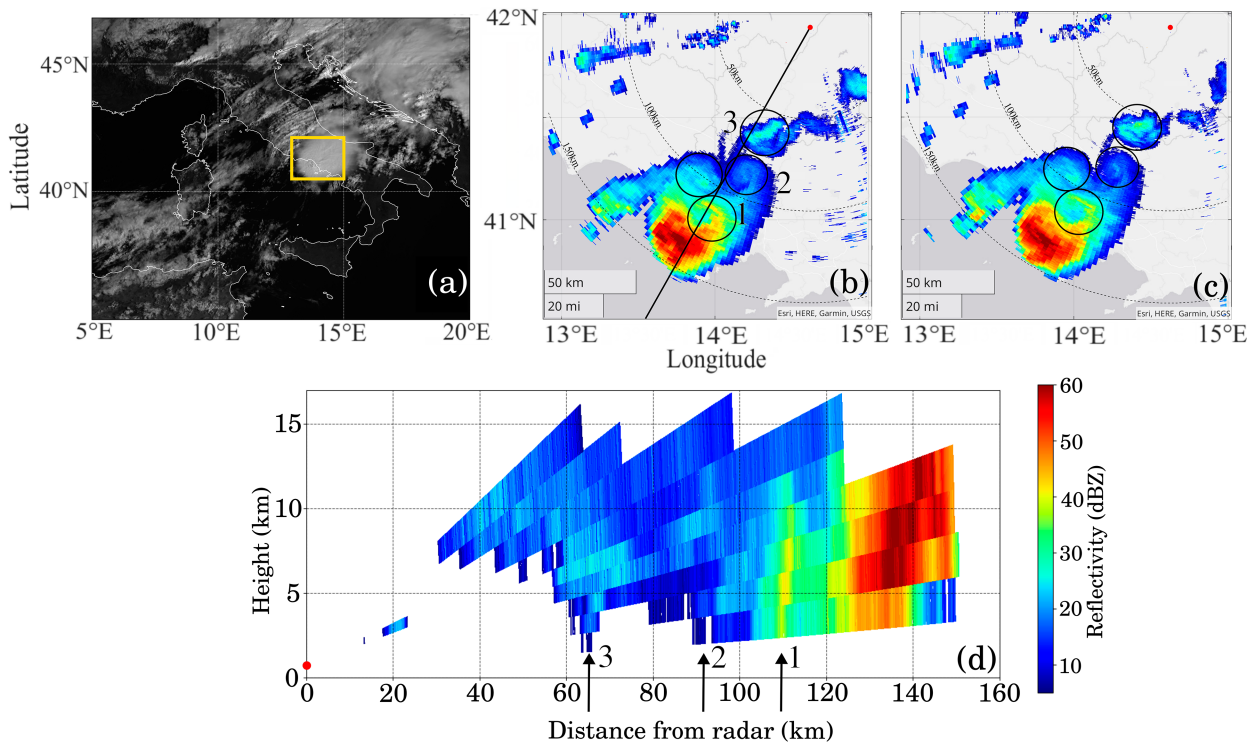


FIG. 1. The structure of the N15 storm by satellite and radar observations. (a) High-resolution visible channel image at 0842 UTC; (b) horizontal reflectivity at 0845 UTC from the Il Monte radar at 2.5° elevation, where circles indicate the vortex location; (c) as in (b), but at 0850 UTC; (d) reflectivity pseudo-RHI at 0845 UTC, 210.5° azimuth, between 0.5° and 13.5° elevation. The yellow box in (a) indicates the area of the radar in (b) and (c). The straight line in (b) indicates the direction of pseudo-RHI in (d). Arrows and numbers on the x axis in (d) indicate the location of the three vortices intersected by the pseudo-RHI; indices refer to the ones in (b). Red dots in (b)–(d) indicate the radar location.

the summary and conclusions report the main findings of the study.

2. Characteristics of the N15 storm and VKVS observation

On the morning of 5 September 2015, an extremely powerful thunderstorm grew over the Tyrrhenian Sea, impacting the Gulf and the city of Naples (Italy), lasting for more than 5 h and extending for more than 3×10^4 km² (see Fig. 1a). This severe event (hereafter the N15 storm) produced an extremely high number of lightning and hailstones with a size of 10–12 cm in diameter. Moreover, it caused severe damage to cars and buildings and some injuries among the population of the metropolitan area of Naples, which is one of the most densely populated in Europe.

The N15 storm showed extremely rare intensity for the Mediterranean area, being also one of the most globally significant thunderstorms over more than 2 years, reaching in microwave satellite channels some of the lowest brightness temperatures measured over 26 months of NASA Global Precipitation Measurement observations (Marra et al. 2017). Reflectivity data from the operational C-band dual-polarimetric weather radar “Il Monte” (41.94°N, 14.62°E, 710 m ASL) located inland benefiting from a favorable view (even if quite far from the event), revealed a singular shedding of six vortices

in the wake of the thunderstorm toward the coast approximately between 0800 and 1000 UTC, four of them shown in two consecutive radar images in Figs. 1b and 1c. A VKVS-like pattern is visible, with vortices detaching alternatively from the right and left flanks of the thunderstorm core (see animation of radar reflectivity sequence in the online supplemental material). The vortices are detected over an area that spans around 40 km in the crosswind direction and 80 km downstream of the convective core, being advected by the southwesterly tropospheric wind. The pseudo-range height indicator (pseudo-RHI) in Fig. 1d shows the vertical development of the vortices, particularly evident in the vortex marked with 1. The vortices are observed at different radar elevations and extend vertically throughout the troposphere, gradually losing vertical coherence while advected in the wake several tens of kilometers downstream of the thunderstorm core (see vortices 2 and 3 in Figs. 1b,c).

A clear evidence of the vortex structure can be made by looking at the Doppler images reported in Figs. 2a and 2b, where four vortices are highlighted by white circles in two different moments of the N15-storm evolution at the elevation of 3.5°, while the corresponding reflectivity is reported in Figs. 2c and 2d. The observed vortex speed and rotation velocity are fully compatible with the values obtained by other observations (see section 2e, taking into account the motion of the storm and the motion of the vortices with respect to the storm).

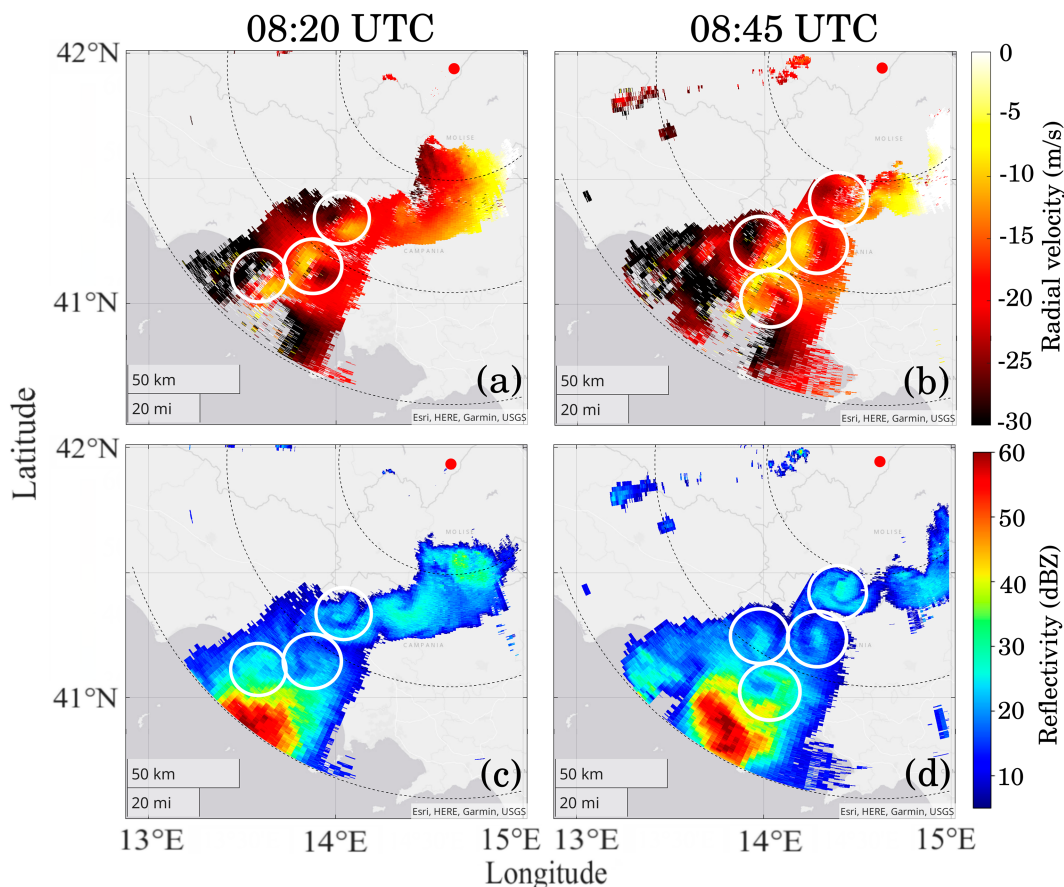


FIG. 2. Signatures of vortex rotation observed from the Doppler radar. (a),(b) Radial velocity at 0820 and 0845 UTC at 3.5° elevation; (c),(d) radar reflectivity at 3.5° elevation at 0820 and 0845 UTC, respectively. White circles indicate the location of the vortices. Red dots indicate the radar location.

The alternate rotation direction of the vortices, a fundamental attribute of VKVS, is further confirmation of the nature of the vortices. Unfortunately, the quality of the Doppler signal was not optimal along the event, especially at lower radar elevations, so that a multivariable approach to estimate VKVS attributes was necessary.

The key step in the methodology used in this study is the quantitative evaluation of the vortex characteristics, as well as the determination of boundary and initial conditions to set up the simulations at a spatial scale usually not resolved by conventional ground-based networks. Following a multiplatform approach, which takes into account different available ground- and space-based observations to analyze the N15 storm (Marra et al. 2017), the key dynamical variables of the thunderstorm are estimated. Specifically, the updraft velocity, size, and temperature, together with the atmospheric wind, are inferred by cross checking different sources, such as ERA5 reanalysis, radar, and satellite data (Grenzi et al. 2024; Vulpiani and Baldini 2024). From our analysis, no specific supercell signatures, like updraft rotation, were found in radar data. The quality of Doppler velocity data is particularly deteriorated in the updraft region, due to the distance of the convective storm from the radar, in addition to beam blocking and attenuation caused by

intense precipitation. To avoid any speculation, we present the system as it can be detected from observations, as a strong, nonrotating convective updraft. Finally, also the VKVS is identified, and the vortices attributes are discussed.

a. Updraft velocity

Since the radiosounding site closer to the N15 storm is located more than 100 km north, this is not considered representative of the thunderstorm environment. Therefore, a pseudoradiosounding is reproduced using the ERA5 reanalysis data on 15 pressure levels, from 1000 to 100 hPa, hourly available on a regular lat–lon grid at $0.25^\circ \times 0.25^\circ$ resolution. Pressure levels are denser near the surface and become coarser with height. Vertical profiles of relative humidity (%), horizontal wind components (m s^{-1}), vertical velocity component (Pa s^{-1}), temperature (K), and specific humidity (kg kg^{-1}) are extracted at 0600 UTC, over an ERA5 grid point at (41°N , 13.25°E) along the path of the thunderstorm in proximity of the Italian coast, as reported in Fig. 3. The maximum updraft speed w_{max} is estimated by computing the convective available potential energy (CAPE), including the contribution of the entrainment of cold air into the updraft [entraining CAPE (ECAPE)]. Following the approach of Peters et al. (2023), it

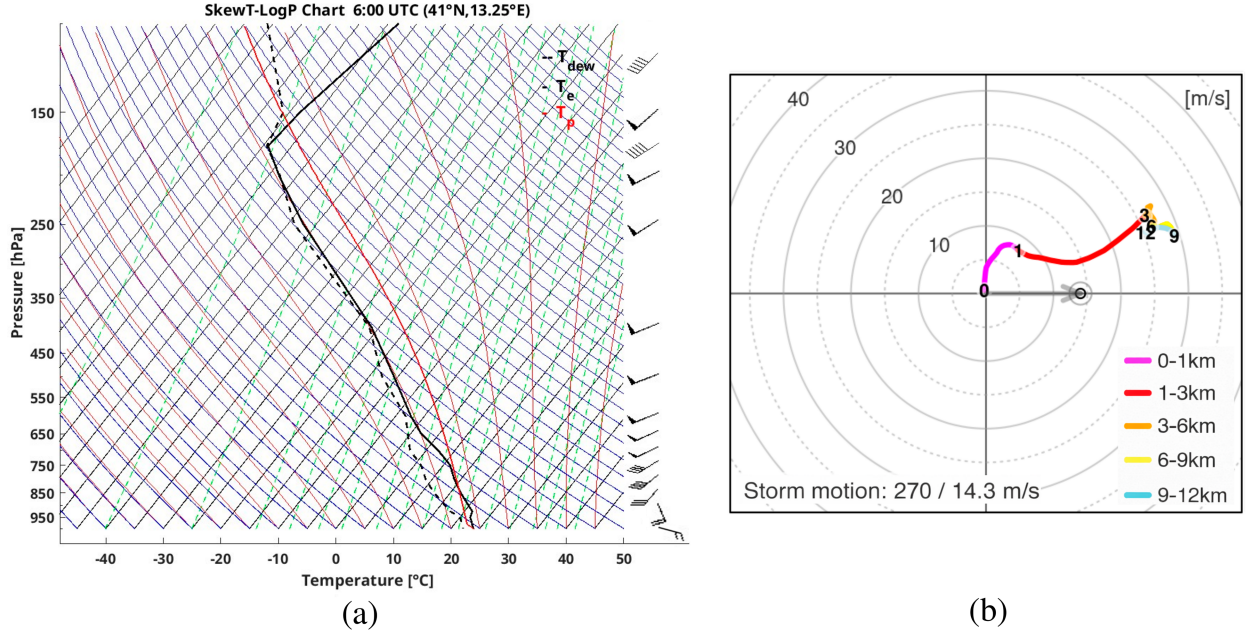


FIG. 3. (a) Skew T -log p chart at (41°N, 13.25°E) at 0600 UTC: temperature (°C) on the horizontal axis, while on the vertical axis, the logarithm of pressure is plotted. In the background, isotherms are plotted as black solid lines; dry adiabatic curves are plotted as blue solid lines; saturated adiabatic curves are plotted as red solid lines; and constant humidity curves are plotted as green dashed lines. In the foreground, the black solid bold line is the vertical atmospheric temperature profile T_e ; black dashed bold line is the dewpoint temperature vertical profile T_{dew} ; and red solid bold line is the convective rise of an air parcel T_p . On the right side, wind barbs of the tropospheric wind from ERA5 pressure levels are reported. LFC and EL are highlighted in yellow. (b) Hodograph of environmental wind profile from ERA5 model level data at (41°N, 13.25°E) at 0800 UTC. Different colors refer to different layers in the atmosphere, indicated by numbers in kilometers along the curve. The storm motion is shown as a gray line, as estimated from radar reflectivity sequence. Note: Panel (a) shows the atmospheric conditions before the passage of the convective system, analyzed in section 2a; panel (b) depicts the wind profile during the convective system transit, discussed in section 2d.

is estimated a value of $ECAPE \sim 2250 \text{ J kg}^{-1}$. Under the assumption that all potential energy is converted into kinetic energy, the maximum updraft speed can then be estimated as $w_{\max} = \sqrt{2ECAPE} \sim 67 \text{ m s}^{-1}$. This computation neglects the contributions of vertical pressure perturbation, hydrometeors' drag, and frictional forces.

The diameter d of the largest hailstones measured in the area can be used as a proxy to confirm the estimation of the maximum updraft vertical velocity (Foster and Bates 1956). Following Heymsfield and Wright (2014), a parametric formula is used to estimate the terminal velocity V_T of the hailstones as $V_T = 1207d^{0.64} \text{ cm s}^{-1}$, valid for wet-grown particles with density $\rho = 0.9 \text{ kg m}^{-3}$. For the largest hailstones observed in the N15 storm, $d = 12 \text{ cm}$, we obtain $V_T \sim 59 \text{ m s}^{-1}$, in fairly good agreement with $w_{\max} \sim 67 \text{ m s}^{-1}$ estimated from ECAPE.

b. Updraft temperature

The pseudoradiosounding also provides an estimation of the temperature difference T_{diff} between the environment and the updraft over the whole troposphere. The updraft vertical temperature profile is assumed to be approximated with the saturated adiabatic profile of a lifted parcel. The value of T_{diff} at specific pressure levels is usually utilized to compute different convective indices that describe a convective environment, e.g., the lifted index (Galway 1956) and the Showalter index (Showalter 1953). The average temperature difference

between the updraft and the environment, from the level of free convection (LFC) to the equilibrium level (EL), in the N15 storm is $\overline{T_{\text{diff}}} = 4.7 \text{ K}$. The mean buoyancy force per unit mass acting on the rising parcel during the ascent from the LFC is evaluated as $b_{\text{up}} = g\beta\overline{T_{\text{diff}}} = 0.1913 \text{ m s}^{-2}$, where $\beta = 4.15 \times 10^{-3} \text{ K}^{-1}$ is the thermal expansion coefficient of the air and $g = 9.81 \text{ m s}^{-2}$ is gravity acceleration. The value of β varies with temperature; however, the value adopted is the average of β computed over the whole troposphere, given that the observed sea surface temperature is $T_{\text{SST}} \sim 26^\circ\text{C}$ and the temperature at the tropopause from ERA5 data is $T_{\text{trop}} \sim -65^\circ\text{C}$.

The environmental thermal stratification of the atmosphere is also investigated. The unperturbed atmospheric vertical temperature profile is obtained from the pseudoradiosounding computed from ERA5 data (see Fig. 3). The profile shows the presence of three thermal layers: From the sea level to $z \sim 12.7 \text{ km}$, the troposphere is characterized by a fairly regular unstable thermal stratification with lapse rate $\Gamma_{\text{tropos}} = -7.0 \text{ K km}^{-1}$, followed by a quasi-isothermal layer at the tropopause level ($12.7 \text{ km} < z < 13.9 \text{ km}$), and a stable thermal stratification with lapse rate $\Gamma_{\text{strat}} = 2.5 \text{ K km}^{-1}$ in the stratosphere ($z > 13.9 \text{ km}$).

c. Updraft horizontal size

A robust correlation is observed between the midlevel updraft diameter D_{up} and the horizontal overshooting top area

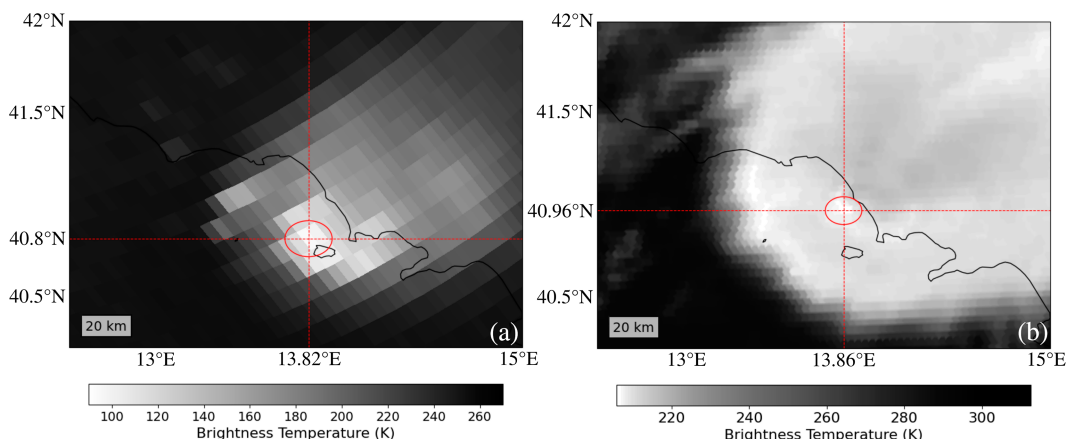


FIG. 4. Satellite images used for updraft diameter estimation: (a) GPM-GMI BT in (183.31 ± 3) GHz channel at 0847 UTC showing the BT depression associated with the N15 storm in its mature stage; (b) SEVIRI BT in channel 9 at $10.8 \mu\text{m}$ at 0857 UTC. In both panels, the estimated OTA is shown with a red circle. The light gray box length corresponds to 20 km.

(OTA) measurable from satellite observations (Marion et al. 2019); hence, this correlation is exploited to estimate the N15-storm updraft diameter.

The OTA can be obtained from the brightness temperature (BT) depression in certain frequency bands, where the BT is sensitive to cloud-top properties (such as the presence of thick ice layers) and generally not strongly affected by the background surface. The Global Precipitation Measurement mission *Core Observatory* (GPM-CO) overpassed the N15 storm at its mature, quasi-stationary stage at 0847 UTC. The channel at (183.31 ± 3) GHz of the GPM Microwave Imager (GPM-GMI) is selected to compute the overshooting top size given its sensitivity to high cloud-top ice and negligible signal from the ground (Fig. 4a). The overshooting top radius is defined as the distance between the minimum brightness temperature pixel and the inflexion point of the BT field along a radial direction. By averaging over eight different radial directions at intervals of 45° , the estimate of updraft diameter is $D_{\text{up}}^a = 22 \pm 3$ km. To be consistent with the approach of Marion et al. (2019), the same computation is also done for the channel at $10.8 \mu\text{m}$ of the Meteosat Second Generation Spinning Enhanced Visible and Infrared Imager (SEVIRI) satellite instrument (Fig. 4b). Being sensitive to thermal infrared radiation, SEVIRI provides data at a higher resolution compared to GPM-GMI. The resulting estimation of the diameter is $D_{\text{up}}^b = 17 \pm 1$ km, slightly smaller than the estimation with the GPM-GMI.

The final estimation of the updraft diameter is $D_{\text{up}} = 20$ km, approximately the average between the two above-reported estimations to exploit the information taken from both.

d. Mean horizontal wind

The predominant environmental wind impacting the N15 storm is estimated from the ERA5 database. The tropospheric wind speed and direction can be variable in the atmosphere. However, for the needs of the simplified approach adopted in our numerical simulation, the predominant tropospheric wind is assumed to be constant in the horizontal

direction and in intensity along the height; thus, the vertical wind shear will not be considered. Although this is a first-order approximation, it is considered acceptable for the aims of the present study. To support this statement, a hodograph of the environmental wind profile is shown in Fig. 3b. Above 3-km altitude, the wind intensity and direction are practically constant up to the tropopause, without significant wind shear. Instead, below 3 km, significant shear is observed: The wind shear for 0–3 km is 26.2 m s^{-1} , the wind shear for 0–6 km is 26.4 m s^{-1} , and above 6 km the wind profile does not vary substantially. Hence, the wind shear is considered negligible above 3-km height. The numerical simulation is carried out in the reference system of the thunderstorm center of mass; therefore, it is also necessary to estimate the predominant tropospheric wind relative to the center of mass (hereinafter the cross flow), which consists of the vector sum of the predominant tropospheric wind U_{wind} and the thunderstorm velocity U_{CM} , as estimated below.

First, the system of reference of the thunderstorm is defined as a right-handed triad of axes, where the origin is located in the center of mass (CM) of the thunderstorm. The vortices are transported by the predominant tropospheric wind and are utilized to define the streamwise (x axis) and spanwise (y axis) directions of the system of reference of the thunderstorm. They are, respectively, the direction parallel and orthogonal to the direction of development of the VKVS as depicted in Fig. 5. The direction of motion of the shedding vortices observed from ground radar reflectivity data is estimated. At 0800 UTC, the VKVS is almost perfectly aligned in the direction of the radar: The streamwise direction points NE with an angle of 45° with respect to the east direction; consequently, the spanwise direction points at 135° , toward southeast.

Using data at 0800 UTC, three ERA5 grid points aligned in the spanwise direction are selected [i.e., points of coordinates $p_1 = (40.75^\circ\text{N}, 13.25^\circ\text{E})$, $p_2 = (40.5^\circ\text{N}, 13.5^\circ\text{E})$, and $p_3 = (40.25^\circ\text{N}, 13.75^\circ\text{E})$] that identify a vertical plane immediately upstream from the N15 storm core, and the wind field is

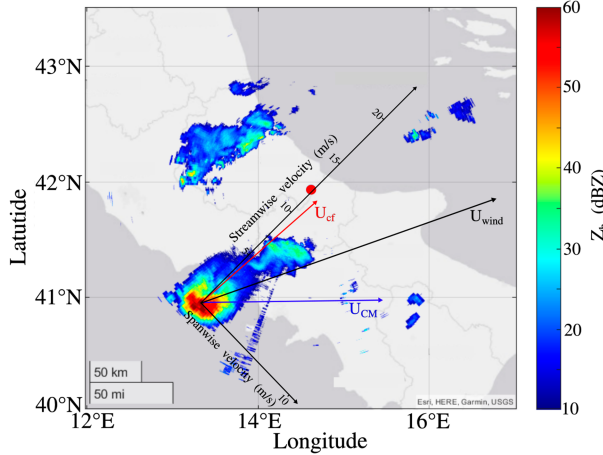


FIG. 5. Vector plot of the center-of-mass velocity U_{CM} , environmental wind velocity from ERA5 U_{wind} and environmental wind velocity in the CM reference frame U_{cf} . In the background is the reflectivity plot from ground radar data at 0800 UTC. The red dot in the middle of the figure represents the location of the ground radar.

averaged over the three pseudoradiosounding locations to obtain a single wind vector that is decomposed onto the streamwise and spanwise components. Since the vortices are clearly visible in the midtroposphere from radar reflectivity, the average value of the wind velocity at $z_1 = 400$, $z_2 = 500$, and $z_3 = 600$ hPa is computed, and it is considered the best estimate of the predominant tropospheric wind vector U_{wind} impacting the thunderstorm which is fairly constant with height in the free troposphere as shown above with the hodograph in Fig. 3b. To compute the value of the cross flow effectively experienced by the thunderstorm U_{cf} , the predominant tropospheric wind is converted into the thunderstorm reference frame, given that the convective system between 0800 and 0900 UTC is moving eastward with a fairly constant velocity of $U_{CM} = 14.3 \pm 1.0 \text{ m s}^{-1}$. The final estimation of the cross flow is $U_{cf} = 12.0 \pm 1.0 \text{ m s}^{-1}$. The direction of the wind is $\alpha_{cf} = (41 \pm 6)^\circ$ with respect to the east direction. Hence, it is practically directed along the x axis, i.e., aligned with the VKVS axis (see Fig. 5). The storm motion toward the east direction is shown in the hodograph in Fig. 3b. Storm-relative flow is mainly to the northeast above 3-km height where the tropospheric wind is almost constant, in the absence of wind shear. This is in agreement with the VKVS direction. Below 3 km, we observe instead a larger wind shear, and the storm-relative flow changes with height.

The N15-storm velocity U_{CM} has been estimated using radar reflectivity data between 0800 and 0900 UTC, from the C-band ground radar located in Il Monte. The quantity that is measured by the radar is the reflectivity factor Z related to the precipitation mass concentration M through a power law in the form $Z = aM^b$, where the values of the parameters used for the rain are $a = 2.4 \times 10^4$, $b = 1.82$, and M in grams per cubic meter, assuming a Marshall and Palmer size distribution. To determine the center-of-mass velocity, Z is used to convert reflectivity data at 2.5° PPI elevation into precipitation mass content. The resulting images have been manually

cleaned, removing noise and retaining only the thunderstorm, and processed to keep only the region of the highest mass content (i.e., the convective core). The coordinate in pixels of the center of mass of the system, defined as the average of the (x, y) coordinates of all pixels weighted on the precipitation mass content, is calculated for each radar scan every 5 min and converted into latitude–longitude coordinates. An error of 1 pixel is attributed to this operation. Having the coordinates of the center of mass at each scanning time, the velocity U_{CM} is then easily computed.

e. Vortex characteristics

The estimation of the geometrical and structural parameters of the vortices is performed with reflectivity data at 2.5° PPI elevation from the C-band ground radar, collected at time intervals of 5 min (see Fig. 6). The radar is on average 100 km far from the VKVS. Six vortices are observed (three on the right and three on the left flank of the updraft) and are detectable up to approximately 80 km downstream of the N15-storm core, forming a VKVS of about 40-km width.

Figure 6 shows a VKVS composed of four consecutive vortices with alternate cyclonic–anticyclonic rotation—marked as V1 (blue), V2 (light blue), V3 (light green), and V4 (yellow)—that are observed from 0800 to 0900 UTC. This time window has been selected because the system shows an almost stationary evolution on the sea. The analysis of the four vortices leads to the estimation of the following characteristic parameters:

- distance between vortices of the same sign $\ell_{obs} = 33.0 \pm 1.2 \text{ km}$;
- distance of the vortices from the central axis of the wake $h_{obs} = 9.4 \pm 0.6 \text{ km}$;
- vortex diameters range between $\Delta = 10\text{--}15 \text{ km}$ and maximum reflectivity of $\sim 35 \text{ dBZ}$;
- velocity of shedding vortices with respect to the storm core $U_{obs} = 12.6 \pm 1.3 \text{ m s}^{-1}$; and
- shedding period $T_{obs} = 44 \pm 5 \text{ min}$.

The distance ℓ_{obs} is computed as the average of all the distances between vortices having the same rotation (vortices V1–V3 and vortices V2–V4), computed with the available PPI scans. The calculation of h_{obs} requires an estimation of the VKVS central axis at each available PPI scan: The center of the vortices is computed at each scan as the mean value of the (lon, lat) coordinates of the vortices. The central axis is then obtained as the line passing through the pixel of maximum reflectivity and the centroid of the vortices previously computed. The pixel of maximum reflectivity is preferred to the center of mass because it is considered a better estimation of the updraft position. Subsequently, the distance from the VKVS axis h_{obs} is computed as the mean of the distances h_i between the i th vortex and the VKVS central axis at each scan. The vortex diameter Δ is estimated considering the area where the reflectivity near the vortex core is higher than 20 dBZ. The vortex velocity with respect to the storm core U_{obs} is computed as the ratio between the distance covered by each vortex between two consecutive scans and the time interval between two captured scans. Knowing U_{obs} and the motion

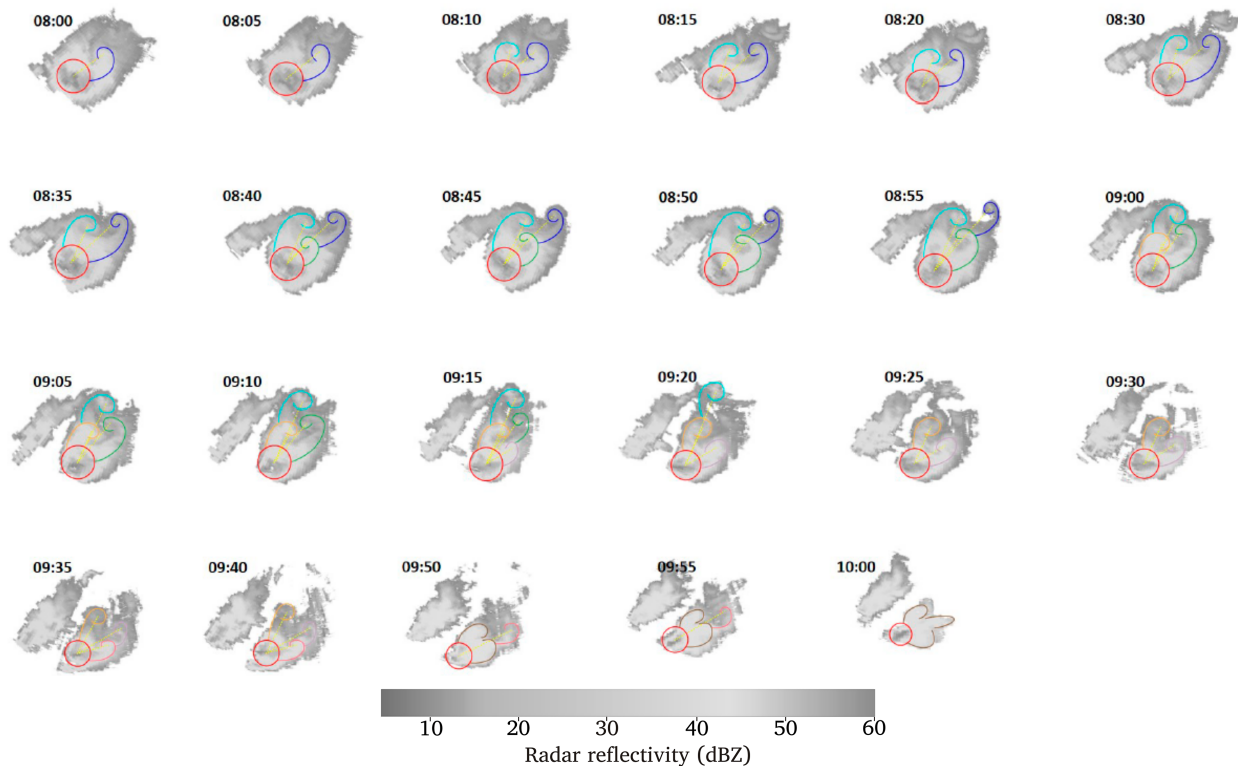


FIG. 6. Postprocessed radar reflectivity images at 2.5° elevation. All identified vortices from 0800 UTC to the dissipation at 1000 UTC are shown. The colors mark the vortices which can be visible from the generation near the thunderstorm flank to the dissipation in the wake region.

of the system at $U_{CM} = 14.3 \pm 1.0 \text{ m s}^{-1}$ in the east direction estimated in section 2d, it is possible to compute the velocity of the vortices along the radar direction in the fixed reference frame. This equals $U_{obs}^{rad} = 17.7 \pm 1.8 \text{ m s}^{-1}$, which is compatible with the observations from radar Doppler in Fig. 2.

3. Simulation approach and setup

Numerical simulations are utilized to gain insights into the structures of the VKVS. The simulations are designed not to reproduce the complex physics of an atmospheric thunderstorm but to isolate and reproduce the key dynamical processes that lead to the rise of the VKVS. This strategy relies on the key hypothesis that the generation of wake vortices is triggered by a main dynamical process: the interaction between the ascending flow of the updraft and the environmental wind. Other factors that characterize thunderstorm events (e.g., the microphysics processes, water change of phase, tropospheric wind shear) are essential for the development of thunderstorms in the real atmosphere, but their numerical reproduction is assumed not to be crucial for generating VKVS. Therefore, the idealized simulation is explicitly designed to reproduce a steady, nonrotating, buoyant updraft in a constant cross flow. The aim is not to reproduce the complete evolution of the system but to focus on the period when the vortices are detectable (between 0800 and 0900 UTC), when the storm reaches its mature stage with a fairly uniform intensity, direction, and velocity

of motion. Hence, the simulation reproduces a statistical steady-state configuration of the convective system as it is observed within this time window. The challenge of this study is also to demonstrate how a simulation of only the main dynamic processes can successfully reproduce macroscopic features such as VKVS. Topographic effects can be assumed negligible because the storm, during early and mature stage, develops only over the sea where no mountains and high islands are present along the path. The interaction with the orography becomes definitely relevant during the dissipation stage of the storm, but we are not focusing on this phase in this work. The interaction with the coast is also not considered a significant ingredient for the vortex development because the shedding of vortices in the updraft wake starts before the storm approaches the coastline.

a. Problem assumptions and mathematical model

The simplified simulation of the N15 storm is designed as a buoyant jet in cross flow (Cintolesi et al. 2019), where the updraft is reproduced by imposing a vertical velocity and positive buoyancy on a circular area at the ground, and the physical parameters are chosen on the basis of meteorological observations. Specifically, the simulation reproduces the development of a strong vertical buoyant updraft in a uniform environmental cross flow over the entire column. The updraft's development is limited in height by the stable stratification of the stratosphere

above the tropopause. The updraft is triggered by a temperature gradient in a circular area, corresponding to the projection of the thunderstorm core area to the ground. Coriolis force, cloud microphysics, water vapor, and latent heat release are neglected in the model.

The mathematical model consists of the nonhydrostatic Navier–Stokes equations, describing the time evolution of an incompressible fluid using the Boussinesq approximation. The potential temperature $\theta = \theta_B + T$ is decomposed into a base state θ_B and a perturbation T components (Mayes and Oertel 2004). Assuming that the background temperature is a linear function of the height, one can define the square of the Brunt–Väisälä frequency $N^2 = \beta d\theta_B/dz$ as a constant (β is the thermal expansion coefficient of air). The temperature is expressed in terms of the buoyancy force $b = g\beta T$, where $g = 9.81 \text{ m s}^{-2}$ is the gravitational acceleration. The governing equations are as follows:

$$\frac{\partial u_i}{\partial x_i} = 0, \quad (1)$$

$$\frac{\partial u_i}{\partial t} + u_j \frac{\partial u_i}{\partial x_j} = -\frac{\partial p^*}{\partial x_i} + \nu \frac{\partial^2 u_i}{\partial x_j^2} - b\delta_{iz}, \quad (2)$$

$$\frac{\partial T}{\partial t} + u_i \frac{\partial T}{\partial x_i} = -N^2 u_i \delta_{iz} + \alpha \frac{\partial^2 T}{\partial x_i^2}, \quad (3)$$

where u_i are the velocity components, p^* is the pressure deviation from the kinematic hydrostatic pressure, $\nu = 1.48 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ is the air kinematic viscosity, and $\alpha = 2.11 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ is the molecular thermal diffusivity. The Kronecker symbol δ_{iz} indicates that the buoyancy force acts along the vertical direction z .

The primitive equations of motion are directly resolved without the use of intermediate submodels for specific phenomena, except for subgrid turbulence. The LES numerical approach is utilized to account for the turbulent processes and perform high-resolution, unsteady, three-dimensional simulations. This technique is successfully used in the field of computational fluid dynamics, and it is found particularly suitable for reproducing the dynamical processes involved in turbulent convection (Zahn and Bou-Zeid 2024; Žužul et al. 2024; Cintolesi et al. 2015; Sullivan et al. 2014). The LES approach considers the implicit spatial filters imposed by the computational grid. The dynamical variables are divided into a resolved part and an unresolved subgrid scale component, which is modeled with the Smagorinsky model (Sagaut 2000).

b. Algorithm and numerical schemes

The numerical simulation is performed with OpenFOAM Foundation (2021). The basic solver “buoyantBoussinesqPimpleFoam” has been customized to perform the LES, as in Cintolesi et al. (2019). The governing equations are discretized in time using an implicit Euler-backward scheme and in space (except for the advection terms) using a second-order central difference Gauss linear scheme. The buoyancy-advective term is discretized with the MUSCL scheme (van Leer 1979), while

for the momentum advection, the Gauss–gamma scheme (Jasak et al. 1999) is used with $\gamma = 0.2$. The time step is dynamically computed to satisfy the Courant–Friedrichs–Lewy (CFL) condition ($\text{CFL} < 0.5$), which ensures the stability and accuracy of the simulation.

c. Case geometry and computational mesh

The computational domain is chosen as a rectangular area of $600 \text{ km} \times 320 \text{ km} \times 160 \text{ km}$, and the system of reference is centered on the N15-storm center of mass (see Fig. S11 in the online supplemental material). The updraft motion is triggered by suitable boundary conditions imposed in a circular area at the ground, corresponding to the projection of the thunderstorm core. To enforce numerical stability, this circular area is modeled as a disk of diameter $D = 20 \text{ km}$ and tiny height of 1 km , which is considered to have a negligible effect on the overall system.

The computational grid is an unstructured and staggered mesh composed of 13.2 million cells with increasing resolution near the thunderstorm core, enough to directly resolve the first part of the Kolmogorov energy spectrum (not shown). To capture the features of interest, the grid is refined near the thunderstorm core and wake regions. The background mesh resolution is of 8 km , on which two refinement regions are nested (Fig. S11). The first refinement region covers the entire thunderstorm system with cells of 1-km width. This is a rectangular region with dimensions $10D \times 4D \times 2D$, placed at a distance of $x/D = 2.5$ from the inflow. The second refinement region covers the thunderstorm core and the wake region where the VKVS is observed, leading to the resolution of 250 m . This rectangular region has dimensions $5D \times 3.5D \times 1.25D$, placed at a distance of $x/D = 3$ from the inflow. The height of $5D$ is chosen to cover the whole troposphere and the low stratosphere, while the length is set coherently with the distance from the updraft where the real vortices are observed.

d. Initial and boundary conditions

The initial and boundary conditions of the simplified simulations have been chosen to reproduce the meteorological conditions of the thunderstorm under analysis. The core of the thunderstorm is reproduced by imposing a vertical velocity $U_{\text{up}} = 10 \text{ m s}^{-1}$ and a buoyancy force $b_{\text{up}} = g\beta T = 0.1913 \text{ m s}^{-2}$ at the disk-area boundary at 1-km height. This weak vertical mass and heat flux have no direct atmospheric analog, and it is configured as an idealized setup requiring a meteorological justification. First, the vertical mass flux at disk exit is imposed to help the development of the convective updraft, and the imposed momentum flux is rapidly overcome by buoyancy forces. This weak vertical velocity is precisely intended to account for convection initiation, which is triggered by a number of meteorological near-surface processes that are not reproduced explicitly by the simulation. This setup ensures a maximum updraft vertical velocity in the upper troposphere (67 m s^{-1}), which is consistent with the observations. Second, it should be noted that a stationary source of buoyant air with no horizontal momentum in a constant cross flow produces a region of low (with respect to the cross-flow) horizontal momentum in the updraft center at upper levels, through vertical advection. Such a model

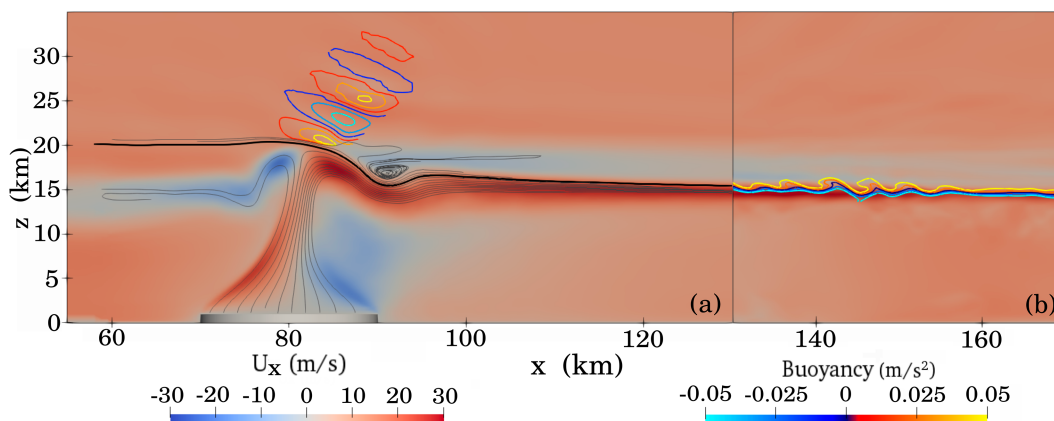


FIG. 7. Key features of the N15 storm reproduced by the simulation. Horizontal velocity (color shades) and streamlines (black lines), together with selected buoyancy isolines (colored lines), on a vertical plane along the streamwise direction and passing through the thunderstorm center. (a) Time-averaged fields; (b) instantaneous velocity and buoyancy fields. The horizontal black line approximately divides the troposphere and lower stratosphere dynamics. The gray rectangular area on the ground depicts the disk where boundary conditions trigger the updraft.

configuration also generates a vertical vorticity dipole at the updraft flanks. In the absence of environmental vertical wind shear, this configuration can be interpreted as a surrogate for possible effects of vertical wind shear, namely, the vertical advection of low-momentum air at the updraft center and the generation of a vertical vorticity dipole (Rotunno and Klemp 1982; LeBel and Markowski 2023).

The cross flow is generated by imposing a constant velocity $U_{cf} = 12 \text{ m s}^{-1}$ at the domain inflow boundary. The atmospheric stable thermal stratification above the tropopause is obtained by setting the Brunt–Väisälä parameter $N^2 = 1.225 \times 10^{-5}$ at $z > 12.7 \text{ km}$ (estimated from the observed temperature profile), while $N^2 = 0$ is set in the low troposphere.

In summary, the boundary conditions are as follows: inlet, constant velocity U_{cf} in the x direction and buoyancy $b = 0$; lateral and top boundaries, constant velocity U_{cf} in the x direction and zero gradient for buoyancy; bottom boundary, no-slip conditions for velocity and zero gradient for b are applied; disk area boundary, vertical velocity U_{up} and constant buoyancy b_{up} ; disk lateral surfaces, free-slip condition on velocity and zero gradient for b are applied to minimize their impact on the overall flow; and outlet, zero gradient for velocity and buoyancy. Pressure is set to $p^* = 0$ at the outlet, while the zero-gradient condition is set elsewhere.

Simulation lead time is defined as 2.5 times the characteristic time $t_c = 30D/U_{cf} = 50000 \text{ s}$ of the system, i.e., the time required by the cross flow to travel the whole length of the domain in a streamwise direction. Lead time is much longer than the actual lifetime of the N15 storm, but this has been adopted to collect sufficient statistics during the statistical steady state of the simulations. The resulting Reynolds number of the system, based on the core diameter and cross-flow intensity, is $Re = 1.62 \times 10^{10}$.

4. Results and discussion

An assessment is first conducted to determine whether the simplified simulation could reproduce the main characteristics

of N15 storm and the atmospheric VKVS in particular. Given the complexity of the phenomenon and the limited data available from observation (specifically, good-quality Doppler velocity radar data are not available), performing a direct comparison between the real event and the simulated one is rather challenging. The assessment strategy adopted consists of two aspects: a quantitative comparison based on geometric and dynamic parameters available in the observational dataset and a qualitative assessment to evaluate if the main structures and expected features of the convective storm have been correctly reproduced.

Figure 7 shows velocity and buoyancy fields on a vertical section oriented along the cross-flow direction and passing by the thunderstorm center (plane $y = 0$). Overall, the simulation successfully reproduces the essential dynamical features of a convective storm structure. It compares satisfactorily with the observed data, supporting the reliability of the simulation results. Specifically, the positive buoyancy imposed at the ground sustains a strong vertical updraft, with a maximum vertical velocity of $w_{\max} = 67 \text{ m s}^{-1}$, in agreement with the estimations obtained from ECAPE. The impact of the updraft on the tropopause generates a large overshooting top (simulated height $z = 19 \text{ km}$, observed $z = 16 \text{ km}$) in the lower stratosphere that mimics the presence of a topographic obstacle to the stratospheric wind, generating a persistent recirculation in the downstream region O'Neill et al. (2021). It also triggers vertically propagating gravity waves (colored contour lines in Fig. 7a) with a wavelength of $\lambda \sim 7.0 \text{ km}$, in good agreement with the orographic gravity waves (Scorer 1949). Such gravity waves can be interpreted as lee waves appearing downstream of solid topographical obstacles in a thermally stable atmosphere. The black streamlines at the tropopause level highlight the presence of the thunderstorm outflow generating the typical anvil cloud and the back-sheared anvil, respectively, downstream and upstream of the updraft. The short internal gravity waves far downstream of the overshooting top are also reproduced (Fig. 4b), having a wavelength of $\lambda = 4.7 \text{ km}$. These waves, depicted with buoyancy isolines, are vertically trapped

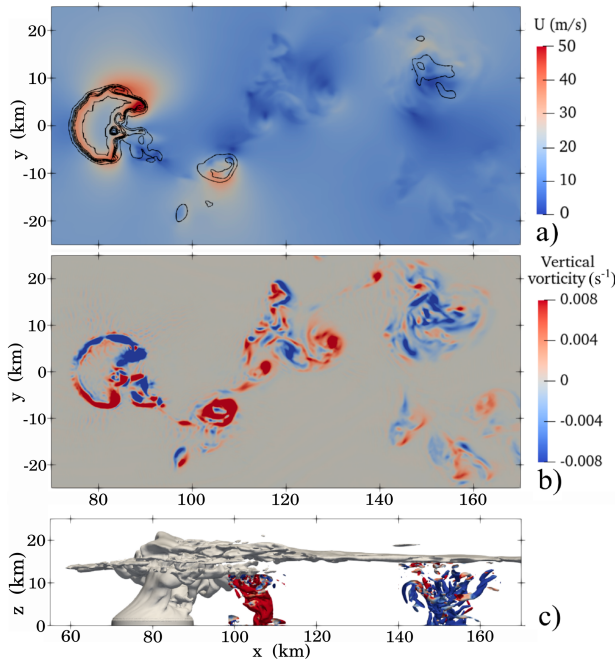


FIG. 8. VKVS identification and visualization. (a) Horizontal and (b) vertical velocity components (background color and black isolines) on a horizontal plane in the midtroposphere (7-km height). (c) Three-dimensional view of the N15 storm core through a buoyancy isosurface (in gray), together with two wake vortices visualized by the Q -criterion isosurfaces and colored with vertical vorticity (red, cyclonic; blue, anticyclonic vortices).

within the thin layer of the anvil cloud at the tropopause. They are commonly observed traveling radially from the overshooting top of large convective systems (Trier and Sharman 2018), using visible channels on geostationary satellites.

a. Identification and description of the VKVS

The presence of shedding vortices in the tail of the thunderstorm was subsequently investigated. These structures are elusive even at first glance in the simulation, and their identification required a targeted analysis of the wake flow.

Figure 8 reports the velocity and vorticity field distribution at a horizontal plane, along with a three-dimensional visualization of the VKVS turbulent structures. A regular series of shedding vortices is detectable in the wake flow, persisting more than 80–100 km downstream of the updraft (see animation of velocity field in the supplemental material). They detach from the lateral flanks of the storm updraft, showing an average propagation velocity of $U_{\text{sim}} = 11.6 \pm 0.4 \text{ m s}^{-1}$ and a distance from the wake central axis of $h_{\text{sim}} = 9.8 \pm 0.2 \text{ km}$ (slightly increasing downwind). These features are in good agreement with the observation of the N15 storm ($U_{\text{obs}} = 12.6 \pm 1.3 \text{ m s}^{-1}$ and $h_{\text{obs}} = 9.4 \pm 0.6 \text{ km}$). The detaching period of the vortices from the updraft is overestimated ($T_{\text{sim}} = 150 \text{ min}$ and $T_{\text{obs}} = 44 \pm 5 \text{ min}$), possibly as a consequence of the simplified configuration of the simulation.

The shedding vortices are clearly identified in the wake flow. They appear as coherent and persistent structures, whose core

area is characterized by peaks of positive and negative vorticity, correlated with local maxima of horizontal rotation velocity and positive buoyancy values (see Figs. 8a,b). The presence of positive buoyancy in the vortex's core clearly distinguishes them from the other turbulent structures appearing in the updraft wake (see Fig. 8c, the horizontal distribution of buoyancy is not shown for the sake of conciseness).

The positive and negative vorticity values in Fig. 8b indicate an alternation of cyclonic and anticyclonic rotation, which is a typical feature of the VKVS. It is worth mentioning that in previous work on a similar topic (Lemon 1976), only anticyclonic vortices were observed. Lemon (1976) hypothesized that this peculiarity was due to the rotation of the storm updraft. To the best of the authors' knowledge, the N15 storm is the first convective system clearly showing a classical VKVS-like pattern in the downstream region. Such a storm did not show any supercell signature in our analysis, and the numerical simulations reproduced a nonrotating convective updraft. Since the simulation is able to generate a VKVS in good agreement with observations, it is hypothesized that a weak or absent rotation of the updraft can be required for the development of a regular VKVS, with alternating vortices, as the one observed in the N15 storm. However, this point needs further investigation.

Figure 8c reports the three-dimensional structure of shedding vortices, visualized by means of the Q -criterion isosurfaces (Dubief and Delcayre 2000). The isosurfaces show the vortices with alternate signs of rotation spanning vertically throughout the entire troposphere, from the ground to the anvil cloud. The average horizontal extension of the vortex cores is 10–15 km, consistent with the observations for the N15 storm. Interestingly, the vertical structure of the vortices and their alternating pattern are similar to the upright wake vortices observed in laboratory experiments on transverse jet in cross flow, e.g., see Fric and Roshko (1994). This reinforces the hypothesis of analogy with the simplified case of a buoyant jet in cross flow, despite having very different Reynolds numbers, spatial dimensions, and thermal stratification.

To quantify the significance and impact of shedding vortices on the atmosphere, the peak rate of the mean kinetic energy $\text{MKE} = 0.5\overline{u_i^2}$, where overbar denotes the time average, and the (resolved) turbulent kinetic energy $\text{TKE} = 0.5(u_i - \overline{u_i})^2$ of the shedding vortices are calculated.

Table 1 reports the maximum values of MKE and TKE in the core of the vortices at four vertical levels within the troposphere (i.e., $z = 2, 4, 6, 8 \text{ km}$) and at four downstream distances in the wake flow (i.e., $X = 20, 40, 60, 80 \text{ km}$). To reduce instantaneous statistical variability, the MKE and TKE maxima are computed by averaging the instantaneous maxima detected for three consecutive vortices traveling in the wake flow. The maxima are normalized by the global maximum of MKE and TKE (detected near the overshooting top region), to give a measure of the relative intensity of energy carried by the vortices. The vortices are relatively highly energetic at 20 km from the updraft core and then slowly reduce their energy content while advected in the wake. The MKE peak ratio reaches values up to one-third of the global energy maximum, while the TKE can reach almost two-thirds; this suggests that

TABLE 1. Maximum MKE and TKE in the core area of the shedding vortices at four altitudes within the troposphere (i.e., $z = 2, 4, 6, 8$ km) and at four locations downstream the thunderstorm center (i.e., $X = 20, 40, 60, 80$ km). The values are normalized by the maximum of MKE and TKE detected in the whole system, i.e., $\max_{\text{MKE}}^{\text{tot}} = 2.365 \text{ m}^2 \text{ s}^{-2}$ and $\max_{\text{TKE}}^{\text{tot}} = 29.25 \text{ m}^2 \text{ s}^{-2}$.

	20 km (%)	40 km (%)	60 km (%)	80 km (%)
Maximum MKE				
2 km	36	18	18	16
4 km	34	17	16	15
6 km	25	14	14	14
8 km	20	13	13	13
Maximum TKE				
2 km	63	38	36	31
4 km	56	35	35	28
6 km	46	34	30	26
8 km	43	30	26	22

the shedding vortices are highly energetic turbulent structures of the low troposphere. Indeed, the vortices are more energetic near the ground, while they slightly lose energy in the upper levels of the atmosphere. This can be due to the interaction with the tropopause and the high horizontal velocity that develops nearby, which can increase the advection and the entrainment, destabilizing the structure of the vortices (see also Fig. 8c). Further downstream, the MKE and TKE content of the vortices decreases rapidly, still maintaining energy peak ratios (above 10% and 20%, respectively) and indicating a highly turbulent motion in the troposphere.

In addition, the same MKE and TKE maximum values at level 6 km in the vortices core are normalized by the maximum values of MKE and TKE at 6 km in the updraft. This allows for a single-level comparison between the vortices and the updraft intensity. At the reference height of 6 km (i.e., middle

troposphere), the mean (turbulent) peak ratio reaches 34% (56%), highlighting the presence of relatively high velocities and turbulence (table not shown).

These results confirm that the VKVS extends the influence of the convective system several tens of kilometers far from the updraft region, being the source of intense convectively induced turbulence (Sharman and Trier 2019), especially in the low troposphere.

b. Unveiling the generation mechanism of VKVS

Beyond the evolution and development of the VKVS, the aim of this study is also to understand the generation mechanism of the shedding vortices.

Figure 9 shows the streamwise velocity and buoyancy force at a horizontal plane in the midtroposphere (7-km height). This is the same vortex shown in Figs. 8a and 8b located at the coordinates (110; -10) km. It can be noticed that a negative pressure gradient arises in the downwind side of the updraft and generates a recirculation zone detectable by negative velocity values. This is a known feature of the updraft interacting with the environmental wind (Knupp and Cotton 1985; Rotunno and Klemp 1982) and also highlights that the strong updraft mimics the presence of a solid obstacle for the atmospheric wind (Wu et al. 2017; Denev et al. 2009). Near the lateral flanks of the storm's core (see the white boxes), the atmospheric crosswind is accelerated and subsequently deviated toward the central axis ($y = 0$) by the negative pressure gradient. The combined effects of lateral wind acceleration and deflection, along with the turbulence produced by the shear between updraft and wind, trigger a rotational motion that detaches from the updraft, eventually generating a shedding vortex. During this process, the updraft warm air is incorporated in the vortex core and advected downwind by the tropospheric crosswind. The detachment process lasts about 40–60 min. Meanwhile, the subsequent vortex on the updraft opposite side develops. Once detached, the new vortex is advected by the crosswind and is sustained in the wake by the vorticity gained during the

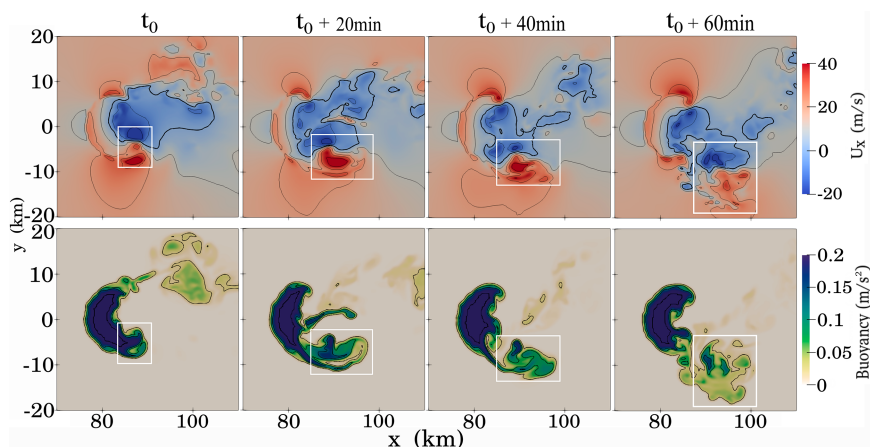


FIG. 9. Time frames of a shedding vortex generation mechanism (highlighted by white boxes). Four time frames sampled every 20 min showing the (top) horizontal velocity U_x and (bottom) buoyancy force magnitude b on a horizontal plane in the midtroposphere (7-km height). The thick black lines in the top row are the isoline $U_x = 0$.

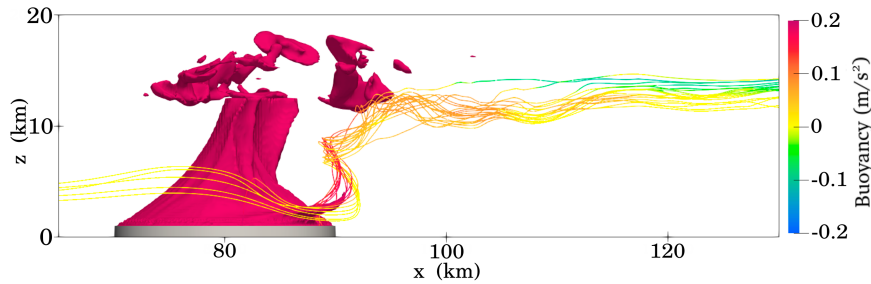


FIG. 10. Three-dimensional visualization of the cross-flow streamlines and the updraft region. The core is identified by a buoyancy isosurface (red), while the streamlines passing through a newly generated vortex are colored with the buoyancy values.

formation and by the warm air incorporated from the updraft (Fig. 9, bottom row), inducing a vertical buoyancy force.

The process in Fig. 9 is shown on a reference horizontal plane, but similar dynamics are detectable in the whole vertical extension of the updraft in the troposphere. The exact symmetric formation mechanism applies to the vortices originating on the opposite flank, showing opposite rotation and negative vertical vorticity.

Figure 10 shows a three-dimensional visualization of the updraft region and the streamlines passing through a developing vortex on the right flank of the updraft. The vertical structure of the vortices is given by the combination of the horizontal rotational motion with the vertical velocity of the buoyant updraft, which leads to a helicoidal vertical flow up to the tropopause height. This process is in agreement with the model for the wake vortices proposed by Meyer et al. (2007) and suggests that shedding vortices contribute to the vertical transport of temperature and mass from near-ground levels to the upper part of the troposphere in the storm's wake, also tens of kilometers downwind from the core.

The entrainment of updraft air in vortices is consistent with the wake entrainment process described in Knupp and Cotton (1985) and more recently investigated with high-resolution numerical simulations by LeBel and Markowski (2023). The wake entrainment has been associated with the formation of a steady counterrotating vortex pair in the thunderstorm cloud throughout the depth of the updraft. Counterrotating vortices have also been observed in other natural phenomena like wildfires (Lareau et al. 2022) and smoke plumes (Cunningham et al. 2005). Despite the similarity, the wake entrainment process occurring in our system has a different origin from the one described by Knupp and Cotton (1985) and LeBel and Markowski (2023). In those studies, the wake entrainment has been ascribed to the tilting of horizontal vorticity associated with environmental vertical wind shear. In the present case, vertical wind shear is neglected and the generation of vertical vorticity in the wake region must be attributed to other sources. First, this vorticity is mainly associated with the tilting of horizontal vorticity within the shear layer between the updraft and the environmental wind, where a strong horizontal buoyancy gradient is present. This interpretation is in accordance with the previous analysis on counterrotating vortices in vertical jets by Kelso et al. (1996). Second, an additional

contribution to vertical vorticity formation is associated with the imposition of a stationary source of vertical momentum and heat flux, with negligible horizontal momentum relative to the cross flow. This idealized model configuration, which has no direct atmospheric analog, can be interpreted as a surrogate to account for one possible effect of environmental vertical wind shear, namely, the formation of a vertical vorticity dipole on the updraft flanks (Rotunno and Klemp 1982; LeBel and Markowski 2023), as explained in the simulation setup.

5. Summary and conclusions

The present study identifies and analyses overlooked turbulent structures in the wake flow of a strong thunderstorm, which are classified as atmospheric von Kármán vortex street (VKVS). The study is conducted and based on a real meteorological event: an extreme thunderstorm detected on the morning of 5 September 2015, on the Mediterranean Sea off the coast of Naples (the N15 storm). Using a multiplatform observation dataset combined with an ad hoc numerical simulation at high resolution, a regular shedding of vortices in the wake flow of the thunderstorm has been detected. This is one of the first observations of such a regular atmospheric structure from a convective updraft. In a few previous studies, only sporadic, anticyclonic wake vortices have been observed, limited to supercell thunderstorm (Lemon 1976; Lareau et al. 2022). However, no clear supercell signatures like updraft rotation have been detected in the N15 storm.

One of the major challenges in this study is the quantitative evaluation of the vortex characteristics, as well as the determination of boundary and initial conditions to set up the simulations at a spatial scale usually not resolved by conventional ground-based networks. Exploiting the observational dataset, a simplified large-eddy simulation has been carried out to numerically replicate the predominant thermodynamic processes that generate the VKVS, without the ambition to completely resolve the multiple complex physical processes that characterize the thunderstorm. The simulation successfully reproduces (qualitatively and quantitatively) the feature of the observed N15 storm and provides the basis for a detailed description and study of the generation of the atmospheric VKVS.

Results demonstrate that the interaction between the intense updraft, behaving as a fluid obstacle, and the atmospheric wind

generates a VKVS. The VKVS involves an extensive wake region downwind of the thunderstorm, characterized by periodic transit of shedding vortices. This region is estimated 40–50 km wide and over 100 km downwind. Simulated vortices persist hundreds of kilometers downwind, while they are dissipated by the interaction with the complex terrain topography in the real environment. The VKVS triggers localized spots of relatively high kinetic and turbulent energy content compared to the values detected in the thunderstorm system. The VKVS-induced turbulence can be categorized as convectively induced turbulence (Barber et al. 2019; Sharman and Trier 2019) that impacts the entire height of the troposphere, being more energetic in the low levels near the ground and localized in the vortex core area. The VKVS-induced turbulence may represent a hazard that has not yet been well assessed for low-altitude flight or aircraft taking off and landing, catching unprepared aircraft crews and increasing the likelihood of turbulence-related injuries or functional issues (Barber and Mullendore 2020; Lane et al. 2012). The current climate change trend is expected to intensify these issues (Alberti et al. 2024). The vortices enhance the vertical transport of momentum, heat, and mass in the wake from the lower to the upper-tropospheric levels. The study suggests that shedding vortices also impact the atmospheric boundary layer, with consequences for ground-based activities (e.g., maritime transport, agricultural activities, harbor assets). This aspect, however, needs further investigation.

The generation mechanism of the shedding vortices is also investigated: On the one side, a negative pressure perturbation on the lee side of the updraft deflects the cross flow and triggers a wake entrainment process similar to the one described in [LeBel and Markowski (2023); but not equal, due to the absence of vertical wind shear in our simulation]; on the other, the shear layer instability between the updraft and the cross flow generates a vortical motion. The combination of these two forcing triggers a vertical helicoidal motion that leads to shedding vortices. This is in agreement with the study by Meyer et al. (2007). It is worth noticing that since the environmental vertical wind shear is neglected in our simulation, which was able to reproduce the atmospheric VKVS with sufficient accuracy, this component may not play a major role in the VKVS formation from a convective updraft.

These results provide new insights into singular features of thunderstorm phenomenology, which affect the local weather dynamics and extend the influence of severe convective systems beyond the usually recognized range. Finally, improving the understanding of the mechanisms governing the severe thunderstorms dynamics is of particular relevance in the context of current climate change. Although uncertainties persist in climate projections of future frequency and intensity of deep convective systems, it has been suggested that the occurrence of severe thunderstorms has increased and will continue to do so in the next century in regions where historically they rarely occur, like over Europe (Rädler et al. 2019) and particularly in the South Mediterranean region (Laviola et al. 2022; Battaglioli et al. 2023).

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Data availability statement. Numerical simulations are carried out utilizing the open-source software *OpenFOAM Foundation* (2021), and the modifications to the solver used are documented in Cintolesi et al. (2019). The radar observation dataset is publicly available in Vulpiani and Baldini (2024). Satellite and ERA5 reanalysis data are publicly available in Grenzi et al. (2024). The simulation output is publicly available in Grenzi and Cintolesi (2024).

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