A Numerical Study of a Nontornadic Supercell over France

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ABSTRACT

A case of a nontornadic supercell over France is simulated with a three-dimensional nonhydrostatic cloud model. The simulation starts from an operational data analysis without any superimposed perturbation. The initial convective cell is triggered by a mesoscale convergence line associated with a preexisting thermal boundary. The model succeeds in simulating right- and left-moving storms arising from a splitting process of the initial convective cell. The right-moving storm exhibits the characteristics of a supercell with a hooklike structure in the precipitation field, a midlevel rotating updraft, and a low-level cyclonic vortex.

A vorticity budget analysis is performed along backward parcel trajectories for the initial storm and for the supercell phases, with emphasis on the impacts of the preexisting thermal boundary and the associated low-level variations of shear. Similar mechanisms as those found for the cases over homogeneous environment operate globally. Indeed, the simulation exhibits a couplet of cyclonic and anticyclonic vortex on the flanks of the updraft leading to the split of the initial storm. The supercell derives its low-level cyclonic vorticity from tilting of baroclinically produced horizontal vorticity within the storm’s forward flank region. Nevertheless, some differences in the rotational characteristics and in the formation of the initial cell arise from the heterogeneous environment compared to those of homogeneous environment. The vorticity analysis shows that the veering with height of the cyclonic and anticyclonic vertical vorticity cores at midlevels is not symmetric due to the heterogeneous field. It is also found that an additional mechanism operates in the low-level cyclonic vertical vorticity generation.

1. Introduction

Supercell thunderstorms are well known as producers of severe weather such as large hail and damaging winds. Supercells are long-lived thunderstorms that in most cases propagate to the right of the mean winds and contain a cyclonic circulation at midlevels (Browning 1964; Lemon and Doswell 1979). They have been extensively studied in the last few decades, spurred by their ability to spawn tornadoes. But not all supercells produce tornadoes; Burgess et al. (1993) estimated that only 30%–50% of the supercells observed in Oklahoma were nontornadic. Severe tornadoes are less numerous in France, compared to the Great Plains of the United States. Dessens and Snow (1989) estimated about two significant tornadoes (≥F2) occurred per year based on statistics of a hundred French tornado cases. They suggested that the risk for a significant tornado occurring at a given location in France was 15 times lower than that in the Great Plains of the United States. More frequent in France, are the right-moving storms. For example, in 1999, the year of the case studied herein, there were at least six right-moving storms associated with severe weather in France (V. Ducrocq 2002, personal communication). Because of the severe weather generated by supercells, it is socially of critical importance to better understand these events in France and to improve their forecast. Most of the results on supercells have been gained from observations and numerical simulations of U.S. cases and it must be clarified how these results apply to the French supercells. The same issue was addressed for severe thunderstorms in Switzerland by Houze et al. (1993), who identified some differences with Great Plains supercells.

Since the first designation of “supercell” by Browning (1964) for a special type of severe long-lived thunderstorms, a lot has been learned. It has been identified that supercells originate in most cases from a splitting of an initial convective cell, which produces a rightward-propagating cyclonic storm and a leftward-propagating anticyclonic storm (Fujita and Grandoso 1968; Achtemeier 1969). This splitting process arises from the interactions between the storm’s updrafts and the environmental vertical wind shear. Indeed, a vortex pair circulation is produced through a tilting into the vertical of the horizontal vorticity associated with the ambient wind shear. These midlevel vortices induce upward pressure gradients on the updraft flanks, which reinforce the growth of new updrafts on the opposite flanks of the...
initial convective cell (Rotunno and Klemp 1982). In response to that nonlinear forcing, the updraft splits progressively into two cells that move apart laterally.

The curvature of the hodograph has been recognized by Klemp and Wilhemsson (1978) as promoting either the right- or left-moving storms: for example, a cyclonic right-moving storm is favored if the ambient wind shear vector turns clockwise with height. Rotunno and Klemp (1982) proposed that this enhancement of the right-moving storm may be explained by the linear theory, which predicts that interaction of the mean shear with the updraft produces an enhanced upward pressure gradient force on the right flank. Thus, the processes contributing to updraft maintenance and propagation are significantly nonlinear, independent of hodograph curvature, while the linear forcing terms supply an increasing forcing bias as hodograph curvature is added (Weisman and Rotunno 2000).

As the splitting storm propagates across the mean flow, the storm-relative inflow is also modified (Rotunno 1981). As a result, the horizontal vorticity has a streamwise component, in addition to the transverse component that has produced the original vortex pair. Davies-Jones (1984) stressed the importance of this streamwise vorticity in producing the rotating updraft of the supercell because the vertical vorticity generated through its tilting tends to be in phase with the vertical velocity (contrary to the crosswise component that generates vertical vorticity maxima on the updraft flanks).

Furthermore, a low-level mesocyclone frequently accompanies a supercell storm. Its presence appears to be a necessary condition for a production of tornadoes, but not all low-level mesocyclones generate tornadoes (Wakimoto and Cai 2000). While the midlevel rotation is strongly linked to the environmental shear, the low-level mesocyclone is generated by a different process. As demonstrated by Klemp and Rotunno (1983) and Rotunno and Klemp (1985), this low-level rotation finds its origin in the tilting of baroclinically generated horizontal vorticity within the forward-flank downdraft region. A baroclinic generation of horizontal vorticity within the rear-flank downdraft can also contribute to the low-level rotation (Davies-Jones and Brooks 1993). Recent works, based on findings that most of the significant tornadoes appear in proximity of preexisting boundaries (Wakimoto et al. 1998; Markowski et al. 1998; Rasmussen et al. 2000), suggest that the horizontal vorticity generated at these boundaries, such as thermal boundaries, could also be an important vorticity source for low-level mesocyclones via tilting and stretching. Atkins et al. (1999) studied numerically the link of the low-level mesocyclone with such a preexisting boundary by varying its orientation and strength. They showed that a significant portion of the air composing the low-level mesocyclone originates from lower levels in the cold side of the boundary. In that case, the production of low-level vorticity by the forward flank region appears to be less important.

The purpose of this paper is to study, through numerical modeling, a case of a nontornadic supercell over France in order to give insights on how supercells in France compare to or differ from those over the U.S. Great Plains. Most of the studies for the latter has been obtained from numerical simulations starting from a homogeneous environment, derived from a proxy sounding, superimposed on a warm or cold bubble that initiates the storm. Even the numerical studies on the interactions between supercells and preexisting boundaries assume idealized initial conditions (e.g., Atkins et al. 1999). Contrary to these studies, our numerical simulation starts from an inhomogeneous initial state arising from an operational data assimilation system. Indeed, the assumption of a homogeneous environment, that may be acceptable for a uniform topography of the Great Plains, is more questionable for the French environment subject to numerous mesoscale forcing (e.g., sea/land contrasts along the Atlantic Ocean coast, the Mediterranean sea coast, and the North Sea coasts, orographic effects induced by the Alps or the Pyrenees range). These more local effects may partially explain the differences in supercell behavior, particularly, a lower number of significant tornadoes observed in France, as suggested by Dessens and Snow (1993). Few numerical experiments have been performed starting from an operational data analysis. Hence, we adopt such an experiment including heterogeneities more representative of French environments.

The paper is organized as follows. Observational aspects of the studied case on 30 May 1999 are presented in section 2. Section 3 discusses the computational methodology employed in this study. Section 4 presents an overview of the entire supercell evolution, while a vorticity analysis is performed in section 5. Section 6 summarizes the results and discusses the implications.

2. Case description

A nontornadic supercell on 30 May 1999 accompanied by strong winds caused three deaths and much damage in Paris. The wind gusts locally reached 30 m s⁻¹ in Paris. The rainfall intensities reached 200 mm h⁻¹ during 5 min, accompanied by hailstones.

During the night of 29–30 May 1999, a first convective system (referred as S₁ hereafter) developed over the Atlantic Ocean and progressed inland. At 0430 UTC 30 May, a new convective cell (referred as S₂ hereafter) initiated on the southeastern flank of the first storm in decaying phase. Two hours later, a splitting of the S₁ cell took place with a prevailing right-moving storm (referred as R₁). Figure 1 presents the cumulative radar rainfall associated with this storm. It clearly shows the storm-splitting process with the issuing right-moving storm R₁ reaching the suburbs of Paris near 0800 UTC. The R₁ moved at a speed of about 18 m s⁻¹ and deviated from the mean flow by about 18°. The mean flow direction and intensity, which are of 210° and 13.5 m s⁻¹,
Fig. 1. Observed cumulative radar rainfall (mm) at 0930 UTC for the preceding 3.5-h period on the window shown in Fig. 3. The mean wind direction, estimated as the 0–6-km mean wind (Davies and Johns 1993) from the sounding of Fig. 4, is indicated by the arrow in the left-bottom corner. The paths of the right- (R2) and left- (L2) moving storms are marked by the dashed lines. The break in the precipitation accumulation southwest of Paris is due to reflectivity attenuation associated with the radar.

The 500-hPa analysis (Fig. 2) for 1200 UTC 29 May shows an upper-level cold low extending from Ireland to off the coast of Portugal. It induced a divergent, upper-level southwesterly flow over the convective region that was in favor of large-scale ascent. The upper-level conditions stayed almost unchanged for the entire convective period with only a slight rotation and an increase of the flow. Figure 3a presents the low-level conditions that prevail at the time of the triggering of the S2 storm. This picture has been obtained from a mesoscale analysis of mesonet surface observations [see Calas et al. (2000) for details on the mesoscale analysis system]. Clearly, the S2 storm evolved in a highly inhomogeneous environment, which is characterized by mesoscale variations in wind direction and a preexisting low-level thermal gradient with colder air in western France in connection with the large-scale upper dynamic forcing. The convective cell S2 formed just ahead of this preexisting thermal boundary and was fed by a warm southerly to southeasterly flow. Four hours later (Fig. 3b), when the storm S2 had split into L2 and R2, the low-level cold air progressed inland and the low-level wind field experienced some changes. In particular, the stronger low-level winds on the Atlantic along the Atlantic
coasts began to veer southwest. In the area of the L2 and R2 storms, the thermal gradient was also reinforced due to the cooling induced by the storms. The low-level flow in the vicinity of R2 is characterized by south-westerly winds to the north and easterly winds to the east.

The sounding observed 8 h prior to the convection over Paris (Fig. 4) showed a weakly unstable atmo-
sphere with CAPE of 385 J Kg$^{-1}$ for the most unstable parcel. The associated convective inhibition energy was not negligible (50 J kg$^{-1}$). The wind hodograph was nearly straight up to 700 hPa, with a southwesterly vertical shear. The storm-relative environmental helicity (SREH; Davies-Jones et al. 1990) computed with the observed right-moving storm motion is 67 m$^2$ s$^{-2}$. The SREH and CAPE values are within the statistical range for the supercells without significant tornadoes obtained by Rasmussen and Blanchard (1998) from the 0000 UTC U.S. soundings. However, these values do not allow for the discrimination between the environments of nonsupercell thunderstorms and those of supercells.

3. The numerical simulation

A three-dimensional, nonhydrostatic anelastic model, called MESO-NH, was used to perform the simulation. A detailed description of this model can be found in Lafore et al. (1998). The simulation was performed using two nested grids interacting with each other according to a two-way interactive grid-nesting method (Clark and Farley 1984; Stein et al. 2000). The coarser grid provides the lateral boundary conditions to the finer grid, while the variables of the coarser grid are relaxed with a short relaxation time toward the finer grid’s value on the overlapping area. The model domains used were 900 km x 1200 km and 450 km x 360 km (Fig. 5) with the horizontal grid intervals set to 10 and 2.5 km, respectively. In the following, only the results for the fine-resolution domain will be presented. The horizontal resolution may be considered as coarse compared to most previous numerical studies on supercells. However, by comparing results between 2-km and a 1-km resolution simulations, Klemp et al. (1981) found that the coarser resolution simulation displayed the same basic features as the finer resolution simulation. A higher resolution would be required, if a tornadic phase had to be simulated, which was not the case here. The vertical grid is defined by a stretched vertical coordinate (Gal-Chen and Somerville 1975), with 40 vertical levels spaced by 75 m in the lowest levels to 900 m at the model top, which is at 19 km. Raleigh damping is applied to the upper 5 km of the model. A mixed radiation–relaxation method operates on the six outermost grid points of the outer domain with temporal interpolation between 6-hourly coupling fields, which are provided by operational analyses.
The turbulence parameterization (Cuxart et al. 2000) is based on the prognostic equation for the turbulence kinetic energy with a 1.5-order closure. For the inner domain, the mixing length is directly proportional to the grid size (Redelsperger and Sommeria 1981) and three-dimensional turbulent fluxes are modeled. We have found on another case of supercell that results may be sensitive to the turbulence parameterization (Chancibault 2002). Using a 3D turbulence scheme allows us to simulate the supercell and its dynamics, which are not reproduced if only the vertical turbulent flow (1D scheme) is taken into account. For the outer domain, the mixing length follows the method of Bougeault and Lacarrere (1989) and the horizontal gradients are not considered. A modified version of the Kain and Fritsch (1990) scheme (Bechtold et al. 2001) is used as convection parameterization for the outer domain, while no convective scheme is utilized for the inner domain. In both domains, a bulk microphysical scheme (Caniaux et al. 1994; Pinty and Jabouille 1998) governs the prognostic equations of the six water species: vapor, cloud water, rainwater, graupel, snow, and primary ice. Most of the former numerical studies on supercell dynamics were performed with a warm microphysics scheme. An example of recent studies that examined the impact of improved cloud microphysics is Johnson et al. (1993), who compared supercell simulations with and without ice microphysics. They found that, in a case of highly glaciated supercell, the cold microphysics had a significant impact on the supercell features with the rotating ascent or the hooklike structure more pronounced and exhibiting a longer lifetime.

A sensitivity study to the initial conditions was performed using the operational large-scale analyses of the 4DVAR global ECMWF suite (Rabier et al. 2000; Klinker et al. 2000) and of the French 3DVAR global ARPEGE suite (Thépaut et al. 1998) for 1800 UTC 29 May 1999 and for 0000 and 0600 UTC 30 May 1999. Only the simulations starting with the 3D VAR ARPEGE analysis for 0600 UTC 30 May produced precipitation in northwestern France. The areas of precipitation in the simulations are strongly linked with large initial humidity zones in the operational analyses. In particular, the precipitation of the simulations starting with the ARPEGE analysis for 30 May at 0000 and 0600 UTC occurs inside a moist tongue extended from Atlantic coast to northern France, which is not present in the other operational analyses. Both the 0000 and 0600 UTC analyses exhibit also a low-level thermal boundary. Some differences are yet found for the SREH fields between the two analyses: the SREH computed with the observed right-moving storm motion is large in the area of the observed triggering of S2 storm for the 0600 UTC analysis (about 220 m² s⁻²) whereas it is almost zero for the 0000 UTC analysis. This may explain why only the simulation starting from the 0600
UTC ARPEGE analysis is able to reproduce a splitting process leading to the supercell. For this reason, we focus on the results of this simulation. Notice however that it starts one-and-a-half hours after the triggering of the observed storm $S_2$, leading to differences between the simulated and the observed storms. In the following, the time is counted generally in minutes from the initiation of the integration at 0600 UTC.

### 4. Simulation overview

Several convective cells are triggered on the 2.5-km resolution domain. One of them exhibits features close to the observations, as can be seen in Fig. 6 that presents the cumulative rainfall over a 2.5-h period. Clearly, as in the observed cumulative rainfall displayed in Fig. 1, the splitting process is marked with a preponderant right-moving storm. The precipitation path of the right-moving storm is however shorter in the simulation than in the observations. This is attributed to a shorter life and a slower storm motion of the simulated storm. Indeed, the simulated storm propagates at a speed of 12 m s$^{-1}$ for a deviation from the mean flow of 18°. The duration of the simulated right-moving storm is about 2.5 h while it is about 3.5 h for the observed supercell.

Figure 7 presents time series of the maximum and minimum of vertical velocity and of the average accumulated surface rainfall associated with the simulated storm. The initial increase of maximum vertical velocity corresponds to the growing of the initial cell (phase I). Downdrafts begin to intensify with the first rainfall reaching the surface after 360 min (the beginning of phase II). The initial updraft impulse reaches its maximum strength around 390 min, followed by a slight decrease at 400–420 min, indicating that the initial cell is going through its splitting stage (phase III). Such history of the maximum vertical velocity has been evidenced previously by Wilhelmson and Klemp (1978) and Rotunno and Klemp (1985). The splitting phase is also characterized by weaker downdrafts. Then, the updraft maintains its strength in a quasi-steady fashion for the next 75 min before it starts to decay gradually (phase IV). Figure 8 presents the temporal evolution of the precipitating hydrometeors (rainwater, graupel, and snow) averaged over the storm window. These temporal evolutions differ from one specie to the next. The initial increase of rainwater undergoes the steepest growth at the middle of the splitting phase (i.e., after 410 min), the rainwater reaches the maximum at 450 min during the supercell stage. The evolution of the graupel follows that of the maximum vertical velocity but lagged 20–30 min in the early stages (phases I and II). The initial increase of graupel reaches its maximum at 420 min, followed by a slight decrease. Then, the graupel mixing ratio increases again to reach its absolute maximum at about 480 min during the supercell stage. The snow specie has a smoother evolution, less correlated with the maximum vertical velocity evolution; the maximum being reached around 495 min.

The evolution of the storm at midlevels is presented in Fig. 9. At 390 min, the initial cell is mainly consti-
tuated by an updraft core with the associated precipitation core slightly lagged westward. By 420 min, the precipitation area extends northward and downward motions are now present. The appendage on the left flank of the updraft core reveals that the splitting process is initiated. By 435 min, the storm has split into a northwestern and a southeastern storm as identified by the two distinct upward cores. Nevertheless, they still share the same precipitation zone. This latter extends well northeastward, as well as the downward motions. By 465 min, the right-flank cell takes the characteristics of a supercell, with a hooklike structure in the precipitation field. A vertical cross section through the right updraft also reveals classic features of the supercell (Fig. 10): the storm-top height overshoots the tropopause, the graupel and snow fields display a weak-value vault between 3.5 and 7 km, and an extensive overhanging anvil extends more than 60 km northeast of the cyclonic updraft. Downward motions are now also evident on the southwestern side of the precipitation area (Fig. 9d). On the other hand, the left storm takes more the characteristics of a classical convective cell, but with a lifetime as long as the right-moving storm one.

A horizontal view of the low-level storm structure during the late supercell stage (i.e., at 480 min, Fig. 11) also reveals classic features of supercells: the low-level upward motions lie along the cold air boundary at the rear-flank front, whereas the low-level potential temperature perturbation shows early signs of occlusion of the low-level updraft by the gust front. The warm-inflow air tends to spiral around the center of the storm. Downward motions are found in the forward flank, and to a lesser extent in the rear flank. The hooklike structure is well pronounced at that time in precipitating hydrometeor fields at midlevels (2.5 km).

5. Vorticity analysis

Figure 12 presents the time series of maximum midlevel vorticity and of maximum low-level vorticity associated with the storm. The midlevel vorticity is first produced during the initial storm stage (phases I and II), reaching the first peak around 380 min. Then it slightly decreases before experiencing an increase again at the end of the splitting process (phase III). The midlevel vorticity during the supercell phase (phase IV) is then stronger than in the initial storm phase. The maximum of low-level vorticity increases rather monoto-
nously both during the initial-storm and splitting phases, reaching the peak during the supercell phase. As mentioned earlier, both the midlevel and low-level vorticity have large contributions to the storm splitting as well as the supercell dynamics, which may lead to a tornado. While most prior results have been obtained from simulations of events over the U.S. Great Plains using homogeneous environments our simulation offers a possibility of performing a vorticity analysis under heterogeneous French meteorological environments.

The vorticity equations that are generally used in interpreting supercell dynamics are obtained by taking the curl of the Boussinesq equations of motion and can be written for the vertical vorticity and the horizontal vorticity vector, respectively, as

\[ \frac{d\zeta}{dt} = \omega_H \cdot \nabla w + \frac{\partial w}{\partial z} + F_\zeta, \]

\[ \frac{d\omega_H}{dt} = \omega \cdot \nabla V_H + \nabla \times (Bk) + F_{\omega_H}, \]

where \( V_H = (u, v) \), \( w \), \( \omega_H = (\eta, \xi) \), and \( \zeta \) are the horizontal wind vector, the vertical velocity, the horizontal vorticity vector, and the vertical vorticity, respectively (\( \eta \) and \( \xi \) are the two components along \( x \) and \( y \) axes of the horizontal vorticity vector). Here \( F_\zeta \) and...
Fig. 10. Vertical cross section at 465 min of the snow (thick lines), graupel (gray tones), and rainwater (dashed lines) mixing ratio, along with the storm-relative wind vectors. The axis of the cross section is indicated in Fig. 9. The contours are 0.1, 0.5, 0.75, 1, 2 g kg\(^{-1}\) for the snow and the rainwater mixing ratios, whereas it is given by the gray scale at the right side of the figure for the graupel mixing ratio.

\( F_{\text{mix}} \) are the mixing terms and \( B \) is the buoyancy. The first two terms on the rhs of (1) represent the generation of vertical vorticity by tilting of horizontal vortex lines into the vertical and the vertical stretching of vortex tubes, respectively. The first term on the rhs of (2) may be written as the sum of \([\eta(\partial u/\partial x), \xi(\partial v/\partial y)]\) and of \([\xi(\partial u/\partial y) + \xi(\partial u/\partial z), \eta(\partial u/\partial x) + \xi(\partial v/\partial z)]\), representing the horizontal vorticity generation by stretching of horizontal vorticity and by tilting of vortex tubes, respectively. The second term on the rhs of (2) is the baroclinic (or solenoidal) generation due to a buoyancy gradient.

The tilting and stretching terms of Eq. (1), as well as the baroclinic generation and stretching/tilting terms of Eq. (2) have been evaluated along parcel trajectories and at specific times during the initial-storm and the supercell phases.

### a. Vorticity that leads to the splitting process

The vertical vorticity taken near its maximum (at 375 min) during the initial growth phase (Figs. 13 and 14) shares similar features as in the previous studies of conditions that lead to the splitting process. We identify clearly cyclonic vertical vorticity on the right flank of the convective updraft and anticyclonic vertical vorticity on the left flank (Figs. 13c and 14). We also find, as shown by Rotunno and Klemp (1982) (see their Fig. 4), a pressure perturbation couplet aligning perpendicular to the axis of the vortex couplet (Fig. 13d). The vortex couplet turns with height, veering 50° from 2000- to 6000- m height (Figs. 13a,c). Nevertheless, the veering is not symmetric for the two cores: the anticyclonic core veers extensively between 2000 and 6000 m, whereas the cyclonic core stays approximately at the same location. This asymmetry is a consequence of the inhomogeneous environment as will be shown next.

The triggering of the initial convective cell in the simulation occurs almost 7 h after the triggering of the observed S2 storm. During that elapsed time, the observed low-level conditions have evolved and in particular the thermal gradient has increased and the southwesterly flow along the Atlantic coast ahead of the thermal boundary has progressed inland. These evolutions are reproduced in the simulation as can be seen in Fig. 15, which presents the low-level conditions prior to the triggering of the initial convective cell. So, both observed convection and simulated convection occur in connection with a thermal gradient; the thermal bound-
Fig. 11. Plan view of the storm structure at 480 min depicting the low-level (0.5 km) updrafts and downdrafts (thick solid lines for 0.75 m s\(^{-1}\) contours, and dashed solid lines for \(-0.5\) m s\(^{-1}\) contours), together with the low-level (0.5 km) storm relative horizontal winds (a vector length of one grid point is equivalent to a vector length of 10 m s\(^{-1}\)). The sum of the mixing ratio for rainwater, graupel, and snow is shaded gray (with the gray scale given at the upper-right side of the figure). The cold pool beneath the storm is marked by the dot-dashed thick line (low-level potential virtual temperature perturbation with respect to ambient environment of \(-2\) K at 0.5 km). The letters also mark the maximum of midlevel (3 km) vertical velocity from 390 to 480 min for the initial cell (I), and then for the left- (L) and right- (R) moving cells. Only a 85 km \times 85 km portion of the full domain is shown.

Fig. 12. Time series of the maximum midlevel vorticity (solid line) and the maximum low-level vertical vorticity (dashed line) between 240 and 540 min over the storm window in the unit of 10\(^{-4}\) s\(^{-1}\). See Fig. 7 for the definition of periods I–IV.

ary for the simulated convection is however more prominent. The initial convective cell in the simulation is formed above a low-level convergence line, which induces the low-level upward motion line of about 60 km long in Fig. 15b. This upward motion line is located in the warm side at the leading edge of the preexisting thermal boundary. A low-level southerly to southwest-erly flow prevails on the warm side of the thermal boundary. The low-level horizontal temperature gradient is 0.15 K km\(^{-1}\) in the colder air behind the convergence zone, whereas it is 0.25 K km\(^{-1}\) in the convergence zone. Hodographs taken from the simulation on both sides of the convergence zone by 40 km apart show high variability in the low-level curvature. The hodograph in the cold side (location A) shows a clockwise shear vector, whereas the hodograph in the warm side (location B) exhibits an almost unidirectional shear vector.

Determination of where the air in both anticyclonic and cyclonic cores originates is accomplished by calculating backward trajectories for two different times (at 345 and 375 min). At 345 min (beginning of the initial-storm phase), the cyclonic and anticyclonic cores are still limited to the low levels (i.e., values larger than 10\(^{-3}\) s\(^{-1}\) are all below 3000 m), whereas at 375 min (just before the splitting phase), they are well developed. The backward trajectories have been computed following the method of Gheusi and Stein (2002). For that purpose, passive tracers, defined at each grid points of the 2.5-km domain, are advected during the model time integration by the explicit and eddy diffusive model transports. Then, backward trajectories are retrieved from the final positions of the tracers. In order to facilitate the trajectory analysis, the locations of the parcel trajectories are estimated in the storm relative frame.

Figure 16 shows the backward trajectories for parcels in the cyclonic and anticyclonic cores at 2000 m, superimposed on the low-level horizontal vorticity vector field and the low-level virtual potential temperature at 345 min. The convective cell is located above the low-level convergence area, so that the low-level horizontal vorticity vectors have a different direction on the west and east sides of the convective cell: the horizontal vorticity vectors point toward the north on the east of the cell and westward on the west of the cell. Figure 16 shows that the parcels in the cyclonic and anticyclonic cores originate all from the low-level warm side. The tilting and stretching terms are evaluated along these trajectories. The parcels in the cyclonic cores at 345 min (trajectory 1 on Fig. 17) experience a positive tilting term as the low-level horizontal vorticity vector and the vertical velocity gradient vector point in the same direction. It increases significantly when the parcel enters the upward convective cell, being larger than the stretching term. For parcels in the anticyclonic core, the evolution along the trajectory of the source terms of vertical vorticity evolution (trajectory 3 on Fig. 17) shows a positive tilting term, as the parcel encounters first the low-level convergence area and the convective
cell that generates positive tilting on the warm side. Then, when the parcel arrives on the cold side of the updraft, according to Eq. (1), negative tilting is produced as the horizontal vorticity vector and the vertical velocity gradient point in opposite directions. The stretching term is weak compared to the tilting term. Hence, as shown by previous studies, the cyclonic and anticyclonic core generation is mainly explained by the tilting of the horizontal vorticity by the upward motions associated with the convective cell. The upward motions associated with the convergence area also contribute to this process through a positive tilting on its warm side and a negative tilting on its cold side.

Figure 18 shows the backward trajectories for parcels in the anticyclonic and cyclonic cores at 375 min (i.e., close to the first maximum of midlevel vertical vorticity during phase II). Parcels a, b, c, and d are at 2000-m height at 375 min (Fig. 18a), whereas parcels e, f, g, and h are at 4000-m height (Fig. 18b). Most of the parcels are below 1000 m at 315 min and the majority of them originated from the low-level warm side. The tilting and stretching terms of Eq. 1 are evaluated at
Among the parcels for backward trajectories shown in Fig. 18, we draw on Fig. 19 only those that are at 360 min at the same height as the tilting and stretching fields; their locations at 360 min are marked by crosses. Presence of the low-level convergence area and a different veering of horizontal vorticity between the warm and the cold sides clearly modify the classic vorticity analysis carried out by previous studies that used a homogeneous clockwise hodograph. Parcels a, b, and f all in the cyclonic core at 375 min encounter a positive tilting at 1000 m (Figs. 19a,b) southeast of the upward motions due to the southerly horizontal vorticity vectors on the warm side. They also experience significant stretching afterward, when they pass over the upward motion core. A positive tilting is also produced north of the updraft due to the northeasterly horizontal vorticity vectors on the cold side. There experience significant negative tilting after they pass over the upward motion core.

At 2000 and 3000 m (Figs. 19c,e), the field of horizontal vorticity vectors is more homogeneous. A positive tilting is still produced on the southeastern side of the upward motion core at 2000 and 3000 m. Parcel e at 4000 m is inside the cyclonic core at 375 min, passing through this positive tilting core at 2000 m. As the positive tilting cores at different heights are all above the same location, the cyclonic cores at different heights are also above the same location. The negative tilting area shifts to the northwest from 1000 to 3000 m due to the veering of the horizontal vorticity vectors that are associated with the clockwise hodograph prevailing in the cold side (see hodograph A of Fig. 15b). Parcels c and d, which are at 2000 m inside the anticyclonic core at 375 min, experience this negative tilting core at 2000 m. Parcels g and h, which are inside the anticyclonic core at 4000 m, though do not meet the same tilting, encounter the negative tilting core at 3000 m later during their ascent. Stretching does not contribute to the vertical vorticity evolution at 2000 m, whereas it tends to attenuate the cyclonic vertical velocity at 3000 m.

Consequently, we conclude from the above vorticity analysis that the asymmetric veering of the anticyclonic core with height is due to the different curvatures of the hodographs in both sides of the low-level thermal boundary. This asymmetry may have an impact on the intensity and longevity of the left- and right-moving
FIG. 15. The horizontal cross section for 500-m-level at 315 min after the beginning of the simulation (i.e., at 1115 UTC) over the whole 2.5 domain of the (a) simulation and (b) for a zoom. (a) Same contour values and scales as in Fig. 3. The rectangle delineates the zoom used for (b). The virtual potential temperature $\theta_v$ is contoured at 0.5-K intervals with solid lines for (b), the vertical velocity is cross-hatched for 0.2, 0.4, and 0.6 m s$^{-1}$ levels, and the wind vectors are drawn for every two grid points. The arrow points to the location of the triggering of the initial convective cell, whereas A, B, and C indicate the location of the hodographs shown in the corners of the figure. The heights and wind intensities for the hodograph are given in hPa and in kts. The rounded cross is for the simulated storm motion.
storms. A straight hodograph on both sides would give mirror-image splitting supercell storms, whereas a clockwise hodograph on both sides would introduce a bias to the right-moving storm; as mentioned in the introduction, this bias is explained by the linear theory (Rotunno and Klemp 1982; Klemp 1987). In our simulation, if we base on the linear theory, the clockwise shear on the left side of the initial cell will tend to inhibit the development of the left-moving storm, but in a same time, the straight shear on the right side will not add a positive bias to the right-moving storm. We can also notice from Fig. 11 that the deviation from the mean wind direction (i.e., the initial cell motion direction) for the left-storm motion is larger (about 40°) compared to the right-storm motion direction (18°); the left-moving storm propagates along the thermal boundary, whereas the right-moving storm propagates toward the warmer air. So that the left-storm evolves continuously in connection with the low-level convergence induced by the southwesterly to westerly flow on the cold side of the thermal boundary opposing the cold pool outflows. These two mechanisms may explain why the left-moving storm has a duration as long as the right-moving storm.

b. The supercell phase

Figure 20 presents horizontal cross sections of vertical vorticity at 500 and 2000 m along with a vertical cross section along the vortex pair axis during the supercell phase at 480 min. The midlevel cyclonic vorticity core and the updraft coincide at this moment (Figs. 20b,c). The anticyclonic vorticity is also associated with the downdraft. At low levels, the cyclonic vorticity has also been produced. Previous studies found that the low-level vorticity maximum exceeds the midlevel vorticity maximum during the tornado phase (Klemp 1987). In our simulation, the value of low-level vorticity maximum is always below the value of midlevel vorticity maximum (Fig. 12). The maximum values attained by the vertical vorticity are of the order of 0.01 s⁻¹, which is slightly smaller than those observed for supercells over the U.S. Great Plains. With no observational measurement available for the vertical vorticity in our case, nevertheless, no definite assertion can be made on this point. The model may also underestimate the vertical vorticity, due to a coarse grid resolution used in our experiment (Tartaglione et al. 1996).

At that time, the supercell is still located inside the warm side of the thermal boundary (Fig. 21). The supercell tends to propagate with a weak component transverse to the thermal boundary toward the warmer air, whereas the left-moving storm propagates along the thermal boundary. The horizontal temperature gradient is strengthened and a low-level westerly flow prevails now behind the thermal boundary (cf. Fig. 21 to Fig. 15b). Strong temperature gradients are also formed by the downdrafts associated within both the rear and the
forward flanks of the supercell. The progress toward the northeast of the low-level southwesterly flow ahead of the thermal boundary mentioned in section 5a has continued and low-level southerly flow now prevails ahead of the supercell. The flow is still southwesterly at mid to upper levels.

Atkins et al. (1999), by means of idealized numerical experiments, examined the effects of a preexisting thermal boundary on a low-level mesocyclone. They found that the storm motion relative to the boundary orientation had a large impact. When the supercell propagated along the boundary or in a direction more toward the warmer side, stronger low-level mesocyclones were observed than when the storm motion had a component transverse toward the colder air. The vorticity analysis carried out on one of their boundary runs showed that the mechanism by which low-level mesocyclogenesis occurred was different in the absence of a boundary. They found, in particular, that the preexisting boundary contributed to the low-level mesocyclone by solenoidal generation of streamwise vorticity. However, this vorticity analysis was carried out for one particular configuration of the boundary with respect to the supercell: the thermal boundary intersected the supercell and the forward flank downdraft was behind the boundary in the colder air. It is therefore possible that the nature of the interactions between the supercell and the boundary is different for our simulated supercell. In order to examine these aspects, determination of where the air in the low-level cyclonic core originates was accomplished by calculating backward trajectories for parcels at 500 m at 480 min. As a sample of the low-level cyclonic core at 480 min, 21 parcels are taken at 500 m, for which backward trajectories are computed from 480 to 420 min. They are positioned in the storm relative frame (Fig. 22). All the parcels come from the warm side of the supercell. For the southern trajectories (except trajectory labeled C), the parcels originate all from the low levels and travel at the same height. These parcels are found at 480 min in the eastern side of the low-level core. For the northern trajectories, some of the parcels originate also from low levels, whereas others emanate from midlevels. The midlevel parcels descend in downdrafts to lower levels in the forward-flank region. The parcel C, initially in the rotating updraft at midlevels, descends also in the forward flank region. The horizontal vorticity vector has a southeasterly streamwise component compared to Fig. 16 for the initial-storm phase in which a crosswise southerly component was associated with the low-level inflow trajectory.

The tilting and stretching terms in Eq. (1) and the solenoidal and stretching terms in Eq. (2) have been evaluated at 500 m after 465 min. These terms are presented in Fig. 23, along with the two representative trajectories A and B from Fig. 22. Here, the locations of parcels A and B, both at 500 m at 465 min, are marked by the points in Fig. 23. Parcel A, which originates from the low-level inflow, is associated with a streamwise component of horizontal vorticity. The parcel encounters a positive tilting core generated by the interaction of the upward motions with the horizontal vorticity vector. The vertical vorticity induced by tilting is, then,
amplified by vorticity stretching (Fig. 23b). Parcel B reaches the forward-flank region to the northwest of the low-level mesocyclone, where it encounters a positive tilting core. The generated cyclonic vorticity is further enhanced by stretching. It is clear from Fig. 23a that the tilting core is produced by the horizontal vorticity vectors pointing southward in quite the same direction as the horizontal gradient of vertical velocity in this area. This strong southerly horizontal vorticity component is produced through solenoidal generation along the low-level forward-flank gust front (Fig. 23c) and further amplified by horizontal stretching (Fig. 23d).

The above vorticity analysis is resumed in Fig. 24, which illustrates schematically the mechanisms that generate the low-level cyclonic vorticity. In summary, the primary mechanism for generating the low-level cyclonic vorticity is rather similar to that in a homogeneous environment: tilting of baroclinically produced

![Fig. 20. Horizontal cross section at 480 min for vertical vorticity and vertical velocity at (a) 500 m and (b) 2000 m, and vertical cross section along the axis in (b) for vertical vorticity and (c) vertical velocity. The vertical vorticity is hatched (scale given at the right of the panels in $10^{-3} \text{s}^{-1}$), while the vertical velocity is shown in solid line for positive values and in dashed lines for negative values (contours of (a) $-0.5$, $0.5$, and $1 \text{ m s}^{-1}$; (b) $-5$, $-3$, $-1$, $1$, $3$, and $5 \text{ m s}^{-1}$; and (c) $-7$, $-5$, $-2$, $2$, $5$, and $7 \text{ m s}^{-1}$). (a) The dot-dashed thick contours are for the $-2$-K perturbation potential virtual temperature at $0.5 \text{ km}$, defined relative to the ambient environment. FFD and RFD indicate the forward-flank and rear-flank downdrafts, respectively. The full domain has been windowed to a $42.5 \text{ km}$ by $52.5 \text{ km}$ region.](image-url)
horizontal vorticity within the storm’s forward flank region [see Klemp and Rotunno (1983) or Rotunno and Klemp (1985)]. In contrast to Atkins et al. (1999), we have not found a contribution from the preexisting thermal boundary to the low-level vorticity. This may be explained by the position of the boundary with respect to the supercell: the supercell and the forward flank downdraft region remain in the warmer side contrary to the configuration of Atkins et al. (1999). Nevertheless, there is also an additional source of low-level cyclonic vorticity in our simulation: the tilting of the streamwise low-level inflow contributes significantly to the low-level vertical vorticity generation. This latter contribution is absent in the simulations of events over the U.S. Great Plains using homogeneous environments (Rotunno and Klemp 1985; Atkins et al. 1999), because the low-level environmental horizontal vorticity vector is perpendicular to the inflow in these situations. In contrast, in our simulation, the horizontal vorticity field (Fig. 22) is parallel to the storm-relative low-level inflow.
flow (Fig. 11) due to the rotation between the initial storm and the supercell phases of the low-level flow to a southwesterly component southeast of the supercell. This suggests that not only spatial but also temporal variations have to be taken into account when the potential for a supercell to develop a low-level mesocyclone is analyzed.

6. Conclusions and discussions

A case of a nontornadic supercell over France has been simulated with a three-dimensional nonhydrostatic cloud model with a 2.5-km horizontal resolution. The numerical simulation succeeds in producing left- and right-moving storms arising from a storm-splitting process. The simulation starts from an inhomogeneous initial state taken from a French ARPEGE operational analysis, with a configuration similar to that of numerical weather prediction. For this reason, the results are encouraging for next-generation operational numerical weather predictions, which are planning to use a comparably high horizontal resolution.

The simulated storm compares well with the observed and modeled supercells over the U.S. Great Plains. We
globally find the same mechanisms by which the initial storm splits and by which low-level cyclonic vorticity is created as those previously shown by the studies of Rotunno and Klemp (1982, 1985) and Klemp and Rotunno (1983). However, the simulation started from a real condition allows us to examine the impacts of low-level heterogeneities on the storm evolution. Particularly, a preexisting thermal boundary influences both storm genesis and development in its warm side. Low-level hodograph curvatures vary from the cold to the warm side over the thermal boundary. They also vary inside the warm area. The impacts of the thermal boundary and of the variability of the low-level shear on the formation, intensity, and longevity of the storm are summarized as follows:

1) The initial convective cell forms inside the convergence line located just ahead of the preexisting thermal boundary, in support of the role of boundaries in the initiation of convective storms as widely recognized (e.g., Wilson and Schreiber 1986). As the left-moving storm propagates along the thermal boundary, the convergence induced by the storm outflows and the westerly flow behind the thermal boundary may have sustained the left-moving storm updraft.

2) The different low-level hodograph curvatures on both sides of the convergence line have an impact on the vertical structures of the cyclonic and anticyclonic vorticity cores. The straight low-level hodograph on the warm side of updrafts favors a cyclonic vorticity generation in the same location independent of height. On the other hand, the clockwise low-level hodograph on the colder side of the updrafts implies a shift of the anticyclone with height. From the analysis of the pressure gradient forces of Rotunno and Klemp (1982), it can be postulated that this will tend to attenuate the bias toward the right-moving storm compared to an environment with a clockwise low-level hodograph everywhere. An examination of the pressure gradient forces has to be performed to confirm this hypothesis.

3) Temporal and spatial variations of the low-level mesoscale flows have an impact on the low-level vertical vorticity core. As the low-level horizontal vorticity vector on the southern and eastern sides of the supercell is nearly parallel to the low-level inflow, it induces tilting on the eastern side of the supercell. Hence, in addition to the solenoidally produced horizontal vorticity within the storm’s forward flank, streamwise horizontal vorticity appears to also play a role in generating the low-level cyclonic vorticity by tilting.

Overall, the previous theoretical and numerical works on U.S. Great Plains supercells using homogeneous environments also applies to the dynamics of the supercell over inhomogeneous environments as encountered in France. Nevertheless, some differences due to ambient heterogeneities are pointed out: impacts on the right-moving storm intensity as well as on the amplitude of low-level cyclonic vortex (and implicitly on the tornado production). Results obtained here by a case study are to be confirmed by other French supercell cases, especially those over southeastern France, a region where supercells are quite frequent and prone to mesoscale circulations induced by the surrounding topography.

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