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URBAN LAND USE AND COMPLEX TOPOGRAPHIC EFFECTS ON HIGH-IMPACT WEATHER EVENTS

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Abstract

The prediction and understanding of high-impact weather phenomena remain particularly challenging in areas characterized by complex topography, where observations are often insufficient, and numerical weather prediction models frequently struggle to accurately reproduce and predict such events. To illustrate these challenges, two remarkable case studies are presented: the sinking of the *Bayesian* yacht at the interface between the sea and the mountainous inland of Sicily on 19 August 2024, and the Bora windstorm that affected Trieste on 11 - 12 February 2012. Although the numerical models analyzed demonstrated generally good skill in simulating these episodes, some limitations emerge in capturing near-surface strong winds. The scarcity of observations emerged as a significant shortcoming for the interpretation of the meteorological processes involved and the assessment of model performance. Subsequently, the thesis investigates the effect of urban land use on local severe storms. Urban areas can be considered as elements of topographic complexity, associated with heightened exposure to weather-related hazards. A detailed analysis of a convective windstorm that occurred in Milan on 25 July 2023 reveals clear urban-induced modifications, showing that even a fast-moving and highly energetic convective storm can be significantly influenced by a large city. To overcome the intrinsic limitations of individual case studies, idealized numerical simulations of a storm interacting with a circular urban area are also performed. These experiments indicate a statistically significant weakening of the storm with increasing city size, primarily driven by the urban dry island. Nevertheless, in the simulation with the largest city, a downwind pressure minimum induced by the urban heat island triggers a new storm. Overall, this research advances our understanding of the complex interactions between urban environments and deep moist convection, clarifying the roles of buildings, vegetation, and urban surfaces in shaping storm dynamics, with the ultimate goal of improving early warning systems.

Acronyms

ACM2 Asymmetric Convective Model. [31](#), [32](#)

AGL Above Ground Level. [vii](#), [ix](#), [xii](#), [xiv](#), [9](#), [10](#), [22](#), [26](#), [31](#), [32](#), [33](#), [34](#), [36](#), [45](#), [48](#), [52](#), [53](#), [55](#), [57](#), [58](#), [59](#), [61](#), [62](#), [66](#), [68](#), [73](#), [103](#)

AIS Automatic Identification System. [xiv](#), [26](#), [27](#), [29](#)

amsl above mean sea level. [36](#), [37](#), [40](#), [48](#), [52](#), [56](#), [57](#), [68](#), [70](#)

ARPA Agenzia Regionale per la Protezione dell’Ambiente. [ix](#), [47](#)

ARPA FVG Agenzia Regionale per la Protezione dell’Ambiente del Friuli Venezia Giulia. [32](#)

ARPAE Agenzia regionale per la prevenzione, l’ambiente e l’energia dell’Emilia - Romagna. [27](#), [29](#)

BEM Building Energy Model. [3](#), [45](#), [46](#), [53](#), [54](#), [72](#), [78](#), [87](#), [107](#), [112](#)

BEP Building Effective parameterization. [3](#), [45](#), [46](#), [53](#), [54](#), [72](#), [78](#), [87](#), [107](#), [112](#)

BouLac Bougeault-Lacarrère. [31](#), [32](#), [45](#), [46](#), [53](#), [57](#), [76](#)

BS Bulk Shear. [52](#)

CAPE Convective Available Potential Energy. [vi](#), [ix](#), [x](#), [xi](#), [xii](#), [8](#), [9](#), [11](#), [12](#), [14](#), [15](#), [16](#), [23](#), [52](#), [57](#), [59](#), [68](#), [69](#), [75](#), [76](#), [77](#), [99](#), [105](#), [107](#), [110](#), [114](#), [115](#)

CFD Computational Fluid Dynamic. [viii](#), [4](#), [5](#), [17](#), [30](#), [31](#), [34](#), [37](#), [38](#), [40](#), [42](#), [43](#), [116](#)

CHEM Chemistry. [15](#)

CIN Convective INhibition. [9](#), [12](#), [76](#)

ECMWF European Centre for Medium-Range Weather Forecasts. [27](#)

EL Equilibrium Level. [8](#)

EML Elevated Mixed Layer. [75](#)

EPS Ensemble Prediction System. [viii](#), [27](#), [28](#), [29](#)

GLOBUS GLObal Building heights for Urban Studies. [xi](#), [xiv](#), [45](#), [46](#), [78](#), [79](#)

HRES High RESolution. [27](#)

I-CHANGE Citizen Actions on Climate Change and Environment. 13

ICON ICOSahedral Nonhydrostatic. viii, 27, 28, 29, 113

IFS Integrated Forecast System. 27, 45

KENDA kilometre-scale ensemble data assimilation. 27

LCL Low Condensation Level. xi, 8, 23, 75, 76, 77, 78

LCZ Local Climate Zone. 45, 46, 48

LES Large Eddy Simulation. 3, 15, 17, 76, 112, 115

LETKF local ensemble transform Kalman filter. 27

LFC Level of Free Convection. 8, 9

LHS Left Hand Side. 118, 120

LIDAR Light Detection and Ranging. 15

LLJ low-level jet. 37, 56

LU Land Use. ix, x, 46, 47, 49, 55, 56, 59, 64, 65

MCS Mesoscale Convective System. 11, 23, 29, 111

METROMEX METROPolitan Meteorological EXperiment. 13

ML Mixed Layer. vii, 22, 23, 116

mslp mean sea level pressure. vii, viii, ix, 21, 25, 26, 35, 51, 100

MU Most Unstable. 52, 57, 68

MYJ Mellor Yamada Janjic. xi, 31, 32, 34, 40, 45, 46, 53, 57, 76, 77

NYC New York City. 85, 87, 99, 107, 111

PBC Periodic Boundary Condition. 72

PBL Planetary Boundary Layer. ix, xi, 3, 8, 13, 23, 31, 32, 34, 45, 46, 53, 56, 57, 58, 59, 74, 75, 76, 77, 102, 103, 110, 114, 115

PV photovoltaic. vi, xii, 85, 87, 95, 96, 99

PVU Potential Vorticity Unit. vii, 21

PW Precipitable Water. 23, 52

QLCS Quasi Linear Convective System. 9, 10, 11, 15, 16, 17, 44, 50, 66, 71, 111, 112, 114, 115, 116

rh relative humidity. xiv, 32, 33, 54

RHS Right Hand Side. 10, 118, 120, 121

RIJ Rear Inflow Jet. [10](#), [11](#), [113](#)

RISKADAPT Asset Level Modelling of RISKS in the Face of Climate Induced Extreme Events and ADAPtation. [4](#), [17](#), [30](#), [43](#), [116](#)

RKW Rotunno Klemp Weisman. [10](#), [16](#)

RM Right Mover. [122](#)

RMSE Root Mean Square Error. [xiv](#), [32](#), [33](#), [34](#), [53](#), [54](#)

RRTM Rapid Radiative Transfer Model. [31](#), [32](#)

RRTMG Rapid Radiative Transfer Model for Global climate models. [31](#), [32](#)

SB Surface Based. [23](#)

SB CAPE Surface Based CAPE. [xi](#), [77](#), [78](#)

SEVIRI spinning enhanced visible and infrared imager. [vii](#), [23](#), [24](#)

SFC surface. [9](#)

SIMC Struttura IdroMeteoClima. [27](#), [29](#)

SKEBS Stochastic Kinetic Energy Backscatter Scheme. [xi](#), [xiv](#), [79](#), [80](#), [82](#)

SPOUT Statistical and Programmable Objective Updraft Tracker. [xi](#), [xv](#), [80](#), [81](#), [82](#)

SRH Storm Relative Helicity. [vi](#), [ix](#), [xi](#), [xiii](#), [9](#), [14](#), [16](#), [23](#), [52](#), [57](#), [58](#), [59](#), [69](#), [76](#), [77](#), [78](#), [107](#), [108](#)

T temperature. [xiv](#), [32](#), [33](#), [54](#)

TEAMX Multi-scale transport and exchange processes in the atmosphere over mountains. [12](#), [13](#), [17](#), [116](#)

TIM Thunderstorm Intensification from Mountains to Plain. [13](#), [17](#), [116](#)

TKE Turbulent Kinetic Energy. [ix](#), [34](#), [40](#), [41](#), [59](#), [62](#), [69](#), [76](#)

TRACER Tracking Aerosol Convection Interactions Experiment. [14](#)

UDI Urban Dry Island. [xiii](#), [13](#), [14](#), [57](#), [100](#), [102](#), [105](#), [107](#), [109](#), [110](#), [111](#), [112](#), [114](#), [115](#), [116](#)

UH updraft helicity. [x](#), [xi](#), [xii](#), [48](#), [61](#), [62](#), [63](#), [64](#), [82](#), [83](#), [88](#), [89](#), [90](#), [92](#), [93](#), [96](#), [99](#)

UHI Urban Heat Island. [xiii](#), [13](#), [14](#), [16](#), [44](#), [48](#), [50](#), [57](#), [69](#), [100](#), [102](#), [103](#), [105](#), [107](#), [109](#), [110](#), [111](#), [112](#), [114](#), [115](#), [116](#)

UPI Urban Pollution Island. [13](#)

URE Urban Rainfall Effect. [13](#), [14](#)

US United States. [56](#), [57](#)

USD United States Dollars. [1](#), [2](#)

WD wind direction. [xiv](#), [32](#), [33](#), [34](#)

WG wind gust. [xiv](#), [32](#), [33](#)

WIND RISK Thunderstorm outflows measurement and modeling for strongWIND nowcast and RISK mitigation. [17](#), [116](#)

WMO World Meteorological Organization. [52](#)

WRF Weather Research and Forecasting. [ix](#), [xiv](#), [3](#), [4](#), [5](#), [15](#), [30](#), [31](#), [32](#), [33](#), [34](#), [37](#), [43](#), [44](#), [45](#), [47](#), [68](#), [72](#), [73](#), [76](#), [80](#), [107](#), [113](#)

WS wind speed. [xiv](#), [32](#), [33](#), [34](#)

WSW west-northwest. [88](#)

WUDAPT World Urban Database and Access Portal Tools. [45](#), [46](#)

YSU Yonsei University. [xi](#), [31](#), [32](#), [76](#), [77](#)

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Chapter 1

Introduction

High-impact weather phenomena can cause billions of [USD](#) in losses, as well as injuries and fatalities, representing a major threat to our societies and economies. These hazardous weather phenomena include tropical cyclones, severe local storms, windstorms, floods, droughts, heat waves, and wildfires. Scientific research seeks to acquire a deeper understanding of these events in order to improve early warning systems, design effective emergency plans, inform the public about associated risks, and strengthen socio-economic resilience, for example, through more robust infrastructure and buildings.

This preliminary chapter of the thesis introduces the topic. First, the societal and economic impacts of high-impact weather events are discussed, establishing the motivation for this work. Next, the study area and related challenges are presented. The scientific questions guiding the thesis are then outlined, and finally, the overall structure of the thesis is described.

1.1 Motivation of the thesis

Extreme weather and climate change are responsible for about 300 billion [USD](#) in economic losses every year worldwide (Bower et al., [2025](#)). High-impact weather events, such as tropical cyclones, floods, and local severe storms, account for greater losses than earthquakes (Bower et al., [2025](#)). Moreover, losses caused by local severe storms, flooding, and wildfires are increasing (MunichRe, [2025](#)). This upward trend is partly related to the increasing frequency and intensity of certain phenomena, such as hailstorms (Battaglioli et al., [2023](#)) and floods (IPCC, [2023](#)) due to climate change, and partly to rising exposure and vulnerability of societies (Bower et al., [2025](#); Strader et al., [2024](#)). In addition to economic impacts, high-impact weather events also cause a considerable number of injuries and deaths. Low-income countries are particularly vulnerable due to the lack of early warning systems and mitigation strategies (e.g., effective emergency plans and resilient infrastructure). For example, in 2024 Typhoon Yagi caused 842 deaths in Southeast Asia, while the monsoon season in India was responsible of 1,842 fatalities in the same year (Bower et al., [2025](#)). At the same time, European countries have also recently experienced deadly high-impact weather events, such as the Valencia flood of 28-30 October 2024, which caused 232 fatalities (Martin-Moreno et al., [2025](#)).

High-impact weather is a broad term encompassing a variety of very different meteorological phenomena. This thesis focuses on two specific types: downslope windstorms and local severe storms, with a special emphasis on convective windstorms. A single local severe storm typically has a smaller impact than a flood or tropical cyclone, resulting in fewer losses, injuries, or deaths, and less media coverage. Consequently,

scientific research on this topic has historically been less extensive than for other high-impact weather phenomena. However, local severe storms occur much more frequently than floods and tropical cyclones in certain regions (e.g., Europe) and they are the third leading cause of global economic losses from natural disasters. According to major reinsurance companies, they caused 80 billion USD in losses in 2024 alone (Bower et al., 2025). The substantial losses caused by these phenomena is pushing scientists to fill the gaps in our understanding and forecasting skills of these events.

The most severe socio-economic impacts of storms typically occur when they strike urban areas, where large populations, dense infrastructure, and numerous vehicles are at risk of damage or injury. For instance, the Valencia flood struck a medium-to-large city, highlighting the vulnerability of urban areas to high-impact weather (De Sherbinin et al., 2012). Wurman et al. (2007) showed that if a large, intense tornado were to cross a residential area of Chicago instead of rural areas in the Great Plains, it could result in 4,500-45,000 deaths and massive economic losses. Urban areas also contain highly vulnerable infrastructure, such as hospitals, that can be damaged by wind or hail, compromising emergency response (Thomas, 2011). Notably, the highest number of injuries (400) and the second-largest economic loss (3,000 M €) from a hailstorm in Europe were reported in 1984, when a severe storm struck Munich, the third-largest city in Germany (Heimann and Kurz, 1985; Púik et al., 2019). The fact that cities are hotspots of exposure and vulnerability to local severe storms underscores the need for a thorough examination of the interactions between urban land use and severe storm behavior. This is the main subject of this dissertation.

1.2 Introduction to the scientific problem

Urban land use is considered an element of topographic complexity. Therefore, this investigation is framed within the broader discussion of the interactions between high-impact weather events and complex topography. Understanding and predicting high-impact weather events is more challenging in complex terrain regions than over flat, homogeneous surfaces. Complex topography influences the dynamics of mesoscale phenomena and can even induce new phenomena that do not occur over flat terrain (e.g., orographic precipitations, Miglietta and Rotunno 2009). Moreover, obtaining accurate and detailed observations in these regions is often more difficult, either because the phenomena display strong spatial variability that requires denser observation networks, or because the terrain itself complicates instrument deployment (e.g., weather stations and radars are difficult to install over water, while mountains cause beam blocking of radar signals, Germann et al. 2022). Numerical models also face limitations: grid spacing is often too coarse to fully resolve environmental heterogeneity (e.g., mountains are smoothed, Prein et al. 2016); many processes must be parameterized due to insufficient resolution (e.g., turbulence generated by local obstacles, Fernando et al. 2010); and surface parameterizations exert stronger effects (e.g., at the land-sea interface or in urban environments, Salamanca et al. 2011). In regions such as Italy, characterized by steep mountain ridges surrounded by seas and frequently affected by high-impact weather (Miglietta et al., 2023), improving physical understanding and predictability of these phenomena is crucial for protecting lives and properties.

As previously mentioned, urban land use can be considered as an additional element of topographic complexity. It has its own roughness length, as well as distinct sensible and latent heat fluxes. Although cities are relatively small, large cities have dimensions comparable to those of local severe storms (~10-100 km). Furthermore, it is well estab-

lished that the presence of buildings and urban materials significantly modifies the [Planetary Boundary Layer \(PBL\)](#) (Barlow 2014). Since local severe storms are fueled primarily by air masses located within the PBL (Coffer et al., 2023), it is reasonable to hypothesize that they may be influenced by urban land use (Shepherd, 2013). The main uncertainty lies in whether urban perturbations to the atmosphere are sufficiently strong to alter the behavior of highly energetic mesoscale phenomena such as local severe storms. A second key question concerns which physical processes are most important in driving these modifications. Understanding whether and how urban land use can alter intense severe storms (i.e., how it modifies the hazard component of the risk equation) is essential for improving early warning systems and developing effective mitigation and adaptation strategies.

This scientific problem is associated with a significant scale challenge: in urban areas, turbulence plays a crucial role, and it acts at the microscale (~ 1 m). In contrast, local severe storms are typical mesoscale phenomena (~ 10 -100 km). Therefore, to properly simulate all processes involved in the interaction between urban areas and severe storms, meter-scale simulations would need to be performed over domains of at least 200 km. Such simulations are far beyond the computational capabilities available at the University of Bologna and in general of most of the Research Centers worldwide. In this thesis a compromise was adopted by using the [Weather Research and Forecasting \(WRF\)](#) (Skamarock et al. 2019) model coupled with the [Building Effective parameterization-Building Energy Model \(BEP-BEM\)](#) schemes (Martilli et al., 2002; Salamanca et al., 2010). This configuration represents the state of the art for mesoscale simulations of urban processes at kilometer-scale resolutions. [BEP-BEM](#) are multi-layer models that solves more accurately the atmospheric vertical profile than single-layer urban models. In an idealized framework, the CM1 model (Bryan and Fritsch, 2002) could be employed to simultaneously represent both building-induced turbulence and severe storm dynamics with a [Large Eddy Simulation \(LES\)](#) approach (see, e.g., Rozoff et al. 2023). However, in CM1 the parameterizations of specific and latent heat fluxes between the urban environment and the atmosphere are highly simplified (George Bryan, personal communication). In addition, such simulations would require enormous computational resources, making them unfeasible, particularly when using an ensemble approach as adopted in this work.

The effect of the urban land use on severe storm is investigated using numerical simulations on either a case study and on an idealized setting. The case study analyzed occurred in Milan, on July 25, 2023, when thousands of trees were downed in a few minutes by convective wind gusts in the city, compromising the urban mobility. A case study permits to deeply investigate a realistic setting, and numerical simulations can be compared to observations. Nevertheless, its outcome can not be generalized. On the other side, idealized simulations permit to achieve more general results. Their limitation is that they rely on the outcomes of a numerical model, that must be trusted. To enhance the trustworthiness of the idealized simulations, in this thesis the same model settings used in the real case simulations of the Milan case, assessed with observations, were employed in the idealized setting.

Since Milan is surrounded by mountain ridges, with new nearby seas (the Ligurian and Adriatic Sea), the effect of the complex terrain is also considered in the analysis. Having a full picture of all the interactions between the severe storm, mountains, seas and urban land use is important to understand the dynamics of these high-impact weather events and to improve their prediction. Within this framework, the sinking of the *Bayesian* yacht in Porticello, Sicily, on August 19, 2024 was investigated. 7 people died due to the sinking, and the event had a remarkable media coverage, also because uncertainties surrounded the nature of the weather phenomenon that caused the sinking. The sinking

occurred in a region characterized by complex topography, at the interplay between a sea, and inland mountainous terrain. This event represents an emblematic case highlighting the lack of sufficient observations in complex topography regions to properly understand weather phenomena. It also presents the challenge of predictability in such settings. Furthermore, it represents a valuable example of the benefits of research on mesoscale meteorology for our society: our investigation helped to understand the meteorological conditions that lead to the sinking, providing suggestions to how avoid these tragedies.

Lastly, a further case study was conducted to investigate a downslope windstorm in Trieste. That study is framed within the [Asset Level Modelling of RISks in the Face of Climate Induced Extreme Events and ADAPtation \(RISKADAPT\)](#) project, funded by the European Union (Project number: 101093939), that aims to support systemic, risk-informed decisions regarding adaptation to climate change induced compound events at the asset level, focusing on the structural system¹. Downslope windstorms are a high-impact weather phenomenon generated by complex orography with still some relevant scientific gaps. The analysis carried out addresses the challenge of accurate simulations of downslope windstorms, and it is part of a larger effort of improving our understanding of the microscale effects of buildings on windstorms.

1.3 Scientific questions of the thesis

Summarizing the challenges mentioned in the previous section, this dissertation aims to address three core scientific questions:

***Question 1:** How reliable are simulations of high-impact weather events in complex topographic regions?*

***Question 2:** Can urban land use influence the development or behavior of local severe storms?*

***Question 3:** Which physical processes are responsible for these modifications?*

These core questions can be further divided into five more specific ones:

- A) What physical phenomenon caused the sinking of the *Bayesian* yacht? Was the weather event predictable?
- B) Can the [WRF](#) model reproduce the extremely intense downslope wind gusts in Trieste during the Bora windstorm of February 2012? Can these simulations be used to initialize a [Computational Fluid Dynamic \(CFD\)](#) model?
- C) Was the convective windstorm that struck the city of Milan in July 2025 influenced by urban land use? How was the pre-storm environment modified by the city?

¹<https://riskadapt.eu/>

- D) How sensitive is a supercell to urban morphological parameters? Is the storm weakened or intensified when passing over a city?
- E) Which physical processes cause the observed urban modifications to the supercell in the idealized simulations?

These questions are addressed in the chapters of this thesis, as outlined in the next section.

1.4 Structure of the thesis

The thesis is structured as follows. First, in Chapter 2, the scientific literature relevant to the topics of this dissertation is introduced. In particular, the physics of downslope windstorms and local severe storms is presented. The interaction of severe storms with complex topography and urban land use is then thoroughly reviewed. The last section of the chapter summarizes six current scientific gaps regarding the interaction of high-impact weather, especially downslope windstorms and local severe storms, with topography and urban land use.

Chapter 3 highlights the current strengths and limitations in the study of windstorms in complex topographic settings by analyzing two distinct case studies. In Section 3.1, the meteorological conditions during the sinking of the *Bayesian* yacht are discussed, making use of all available observations. Outputs from an operational convection-permitting ensemble are also analyzed to answer Specific Question A of Section 1.3. In Section 3.2, the skill of the WRF model in reproducing the Bora downslope windstorm in Trieste is evaluated (Specific Question B of Section 1.3). The choice of vertical profiles from the WRF simulations used to initialize the CFD simulations is also discussed.

The focus of the thesis then shifts more specifically to the effect of urban land use on severe storms. In Chapter 4, the case study that occurred on July 25, 2023, in Milan is investigated in detail through numerical simulations. The influence of urban land use on both the pre-storm environment and the convective windstorm is assessed, addressing Specific Question C of Section 1.3.

After evaluating the numerical model in that real-world event, Chapter 5 presents idealized simulations carried out to systematically investigate the effect of urban morphological parameters on a supercell, addressing Specific Question D of Section 1.3. The physical processes responsible for the observed modifications are then analyzed in depth, addressing Specific Question E of Section 1.3.

Finally, Chapter 6 summarizes the results achieved in the thesis, provides tentative answers to the core scientific questions outlined in Section 1.3, and offers suggestions for future works.

Chapter 2

Literature Review

This chapter provides an overview of the state of the art of scientific literature regarding the interaction between high-impact weather events, in particular downslope windstorms and local severe storms, and complex topography.

First, an introduction to the features and physical processes that generate these phenomena is provided in Sections 2.1 and 2.2. The effect of complex topography on the dynamics of local severe storms is discussed in Section 2.3. This discussion is not necessary for downslope windstorms, as they are a direct consequence of mountainous terrain. The prediction of downslope windstorms and local severe storms in complex topography is also briefly reviewed in Sections 2.1 and 2.3, respectively. Subsequently, urban land use is considered as an additional element of topographic complexity, and its effect on local severe storms is examined in detail in Section 2.4. Finally, Section 2.5 highlights the current scientific gaps regarding the interactions between the high-impact weather events, complex topography and urban land use, with references to the sections where some of these gaps are addressed.

2.1 Downslope windstorms

A downslope windstorm is an orographically induced mesoscale phenomenon associated with very intense wind gusts (sometimes exceeding 40 ms^{-1}) in the lee of mountains. It is dynamically driven and differs from katabatic winds (i.e., density currents generated by the cooling of sloped terrain). Well-known examples of downslope windstorms include the Chinook in the Rockies, the Foehn in the Alps, and the Bora in Croatia. Downslope windstorms occur when a deep layer of air is forced over a mountain in the presence of a strong stable layer near the crest of the barrier. A hydraulic jump feature usually emerges in conjunction with the downslope windstorm. These events are also commonly associated with leeside warming. The Bora, however, is an exception, since in that case the upstream air is so cold that compressional warming is insufficient to offset the cold advection (Markowski and Richardson, 2011).

Shallow water equations can provide a first physical explanation for the occurrence of downslope windstorms. At this point, it is useful to introduce the Froude number, $Fr = U(NH)^{-1}$, where U is the intensity of the upstream flow, N is its Brunt-Väisälä frequency, and H is the fluid depth. It can be shown that if $Fr \sim 1$, a solution of the shallow water equations produces a hydraulic jump, and thus denser streamlines appear in the lee of a mountain. This solution occurs when $Fr < 1$ upstream of a mountain; in this case, the fluid is in a *sub-critical* regime, and potential energy is transformed into kinetic energy. Since U increases as potential energy is converted into kinetic energy, Fr

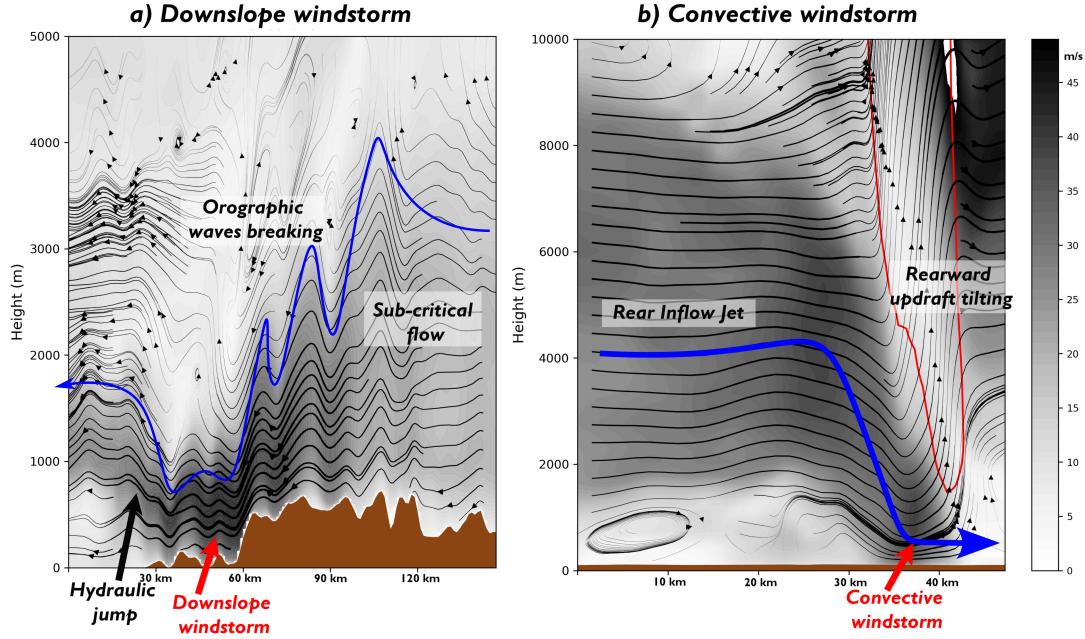


Figure 2.1: a) Streamlines, wind speed (grey shading) and topography (brown shading) of a downslope windstorms simulated in Section 3.2 (simulation *MYJ_Tom_rad*) at 00:00 UTC on 11 February 2012. The *dividing streamline* is highlighted with a blue line. b) as in a) but for a convective windstorm simulated in Chapter 4 (simulation *GLOBUS* at 03:30 UTC on 25 July 2023). The *rear inflow jet* (blue arrow) and the updraft (red contour) are highlighted. Some relevant features mentioned in the text are shown in the Figure.

can exceed 1 at the mountain top. Hence, the flow transitions to a *super-critical* regime in the lee of the mountain, characterized by further conversion of potential energy into kinetic energy. In this way, kinetic energy is extracted from the flow both upstream and downstream of the mountain. The re-adjustment of the flow to the initial *sub-critical* regime occurs through a hydraulic jump in the lee of the mountain (Durrant, 1986).

The shallow water equations provide an understandable explanation of hydraulic jumps and downslope windstorms. However, this framework assumes the presence of a free surface, which is absent in the atmosphere; thus, the fluid depth H is not clearly defined. Moreover, internal waves are not considered in this theory, even though they are of crucial importance in the atmosphere. To overcome these limitations, two theories have been proposed to describe downslope windstorms: the *resonant amplification theory* and the *hydraulic theory* (Lin, 2007).

The *resonant amplification theory*, proposed by Peltier and Clark (1983), is mainly based on results from idealized nonlinear numerical simulations. They showed that downslope windstorms occur when an upward-propagating mountain wave breaks above a mountain. The wave-breaking region is characterized by strong turbulence and wind reversal. This region acts as an internal boundary that reflects the upward-propagating waves back toward the ground, producing a downslope windstorm through partial resonance with the upward propagating mountain waves. Without this nearly perfect reflector, the wave below could not be maintained and would lose most of its energy due to dispersion.

In contrast, the *hydraulic theory* is based on the similarity between the flow configuration during downslope windstorms and the dynamics of finite-depth, homogeneous flow over a mountain ridge. This theory relies on the presence of a *dividing streamline* within the mixed region where orographic wave breaking occurs. The height of the di-

dividing streamline upstream of the mountain defines the *critical level height*. Assuming the presence of the dividing streamline, Smith (1985) obtained an analytical solution to Long's equation:

$$\frac{\partial^2 \delta}{\partial z^2} + l^2 \delta = 0 \quad (2.1)$$

where δ is the displacement of an arbitrary streamline and $l = NU^{-1}$ is the Scorer parameter. That solution characterizes a hydraulic jump, thus downslope windstorms can be described as hydraulic jumps.

Durran and Klemp (1987) showed that the response of a simulated flow to the *critical level height* is in better agreement with the *hydraulic theory*. The presence of a stable layer in the upstream flow is important to define a smoothly stratified *sub-critical* flow impinging on a terrain barrier. The flow then transitions to a *super-critical* regime at the mountain top, and the downstream breaking of orographic waves creates a well-mixed region that decouples the atmospheric flow. The flow below the *dividing streamline* accelerates, producing intense wind gusts and eventually a hydraulic jump. This dynamic is reproduced by the real-case simulation conducted for this dissertation, shown in Figure 2.1a. In that figure, a *sub-critical* flow impinging on a terrain barrier generates orographic waves that break, producing a downslope windstorm in the lee of the terrain, and finally a hydraulic jump. In any case, resonance of orographic waves, as described by the *resonant amplification theory*, likely contributes to downslope windstorms, at least in some events (Lin, 2007).

These ground-breaking studies in the 1980s explained the physical dynamics of downslope windstorms. In later years, research has focused mainly on improving the prediction of the wind gusts that occur during downslope windstorms. Numerical models struggle to provide accurate predictions of wind gusts in these events for several reasons. Grid resolution (Reinecke and Durran, 2009a), numerical methods (Reinecke and Durran, 2009a), PBL parameterizations (Smith, 2007), initial conditions (Reinecke and Durran, 2009b), and surface-atmosphere parameterizations (Cao and Fovell, 2016) have been identified as the model components most responsible for inaccurate simulations of wind gusts. More recently, however, several approaches have been proposed to improve the predictability of wind gusts. These include statistical methods (e.g., Mercer et al. 2008), multi-physics ensemble approaches (e.g., Cao and Fovell 2018), and new diagnostic tools (e.g., the "critical level," Metz and Durran 2023), all of which have shown promising results. Nevertheless, accurate prediction of wind gusts during downslope windstorms at local scales (e.g., building scale) remains a significant challenge that warrants further efforts, especially considering the severe damage that gusts exceeding 40 m, s^{-1} can produce.

2.2 Local Severe Storms

Differently from shallow convection, deep moist convection extends several kilometers into the troposphere and is associated with the condensation of water vapor. It develops when an air mass is lifted to saturation at the so-called *Low Condensation Level (LCL)*, where latent heat is released due to condensation. If the lifting continues and the mid-troposphere exhibits steep lapse rates, the lifted parcel may acquire positive buoyancy at the *Level of Free Convection (LFC)*, producing an updraft that can extend up to the *Equilibrium Level (EL)*. This process is physically described by the *parcel theory* (presented in the author's Bachelor's thesis, De Martin 2020). The *parcel theory* introduces the concept of *Convective Available Potential Energy (CAPE)*:

$$CAPE = g \int_{LFC}^{EL} \frac{T'_v}{\bar{T}_v} dz \quad (2.2)$$

where g is the gravitational acceleration, T'_v is the virtual temperature difference between the parcel and the environment, $\bar{T}_v$ is the virtual temperature of the environment, and SFC denotes the surface. An atmospheric environment can produce deep moist convection if some $CAPE$ is present and an external forcing mechanism (e.g., wind convergence) lifts the initial parcel to the LFC . **Convective INhibition (CIN)** is the work that an external force must do to lift the initial parcel. It is defined similarly to $CAPE$, but the integration is performed between the surface and the LFC .

2.2.1 Organized Deep Moist Convection

Organized deep moist convection generates local severe storms that produce all convective-related hazards, namely large hail, intense wind gusts, tornadoes, and heavy rainfalls. Local severe storms are mesoscale phenomena with a typical horizontal extent of 10-100 km. At mid-latitudes, there are two main types of severe storms: supercells and **Quasi Linear Convective System (QLCS)**, formerly referred to as "squall lines". The features and dynamics that generate and maintain these two storm types are completely different.

Supercells are thunderstorms with a deep, persistent, and rotating updraft called a *mesocyclone* (Glickman and Zenk, 2024b). The mesocyclone sustains the propagation and longevity of the storm. It is generated by the tilting and stretching of environmental streamwise vorticity by the updraft, as discussed in the Ph.D candidate's Master's thesis (De Martin, 2022). Because the atmosphere typically contains some amount of cross-wise vorticity, the tilting of environmental vorticity usually leads to an initial storm split, producing a pair of storms: a cyclonic right-moving supercell and an anticyclonic left-moving supercell. A quantitative explanation of storm splitting and the deviant motion of supercells is provided in Appendix A.3.

Supercells produce most convective hazards because the rotation associated with the mesocyclone can generate intense tornadoes (Rotunno and Bluestein, 2024), while the long-lasting, wide, and rotating updraft allows for the formation of the largest hailstones (Kumjian et al., 2021). Dynamic pressure perturbations associated with the mesocyclone can also induce strong wind gusts (De Martin et al., 2025c).

In addition to a sufficient amount of $CAPE$, which is necessary to produce an intense updraft, environmental vorticity is required to generate a mesocyclone and, consequently, a supercell. Commonly used proxies for quantifying environmental vorticity include the vector wind difference between the surface and 6 km **Above Ground Level (AGL)**, Bulk Shear 06 km, Weisman and Klemm 1982) and the **Storm Relative Helicity (SRH)**, Davies-Jones 1984), usually computed in the lowest 3 km of the troposphere:

$$SRH_{0-3} = \int_{surface}^{3km} (\vec{v} - \vec{c}) \cdot \vec{\omega}_h dz \quad (2.3)$$

where $\vec{v}$ is the environmental wind, $\vec{c}$ is the storm motion, $\vec{\omega}_h$ the horizontal component of the environmental vorticity.

On the other hand, a **QLCS** is a line of convective storms that can propagate either parallel to the orientation of the line (*parallel stratiform* type, Parker and Johnson 2000) or orthogonal to it (*trailing stratiform QLCS*). The *trailing stratiform QLCS*s can produce extremely intense straight-line winds (Weisman, 1993), whereas the *parallel stratiform* type can be responsible for extreme precipitation events (Schumacher and Johnson, 2005).

In this dissertation, we focus on the *trailing stratiform QLCSs*. The theory proposed to explain their maintenance is the *RKW* theory (Rotunno et al., 1988). In short, this theory states that the baroclinic vorticity generated by cold pools along the leading edge of the gust front can be balanced by the environmental vorticity. Consequently, a steady state can be achieved, in which updrafts remain erected along the gust front, allowing the convective system to persist for several hours. However, limitations of this theory have been highlighted in subsequent studies, mainly because the *optimal* environmental wind shear predicted by the *RKW* theory is higher than the values commonly observed in association with *QLCSs*. One possible explanation for this mismatch is that the *RKW* theory may describe the conditions for convective initiation along the gust front, but it becomes less relevant for explaining the maintenance of updrafts in the mid and upper troposphere, since cold pools are confined to the lowest 1-3 km *AGL*. Another possibility is that the *RKW* theory seeks an *optimal* state for *QLCS* maintenance, whereas *QLCS* that produce intense winds are generally sub-optimal from that perspective, being rear-tilted, as will be described in the next Section.

2.2.2 Convective windstorms

Convective windstorms are caused by outflows associated with deep moist convection. They can be produced by meso- γ -scale (~ 10 km) downdrafts, called *downbursts*, or by meso- β -scale (~ 100 km) cold pools and *QLCSs* (Markowski and Richardson, 2011). The development of a *downburst* can be described by the vertical momentum equation (Eq. A.3):

$$\frac{dw}{dt} = -\frac{1}{\rho} \frac{\partial p'_d}{\partial z} + \left(-\frac{1}{\rho} \frac{\partial p'_b}{\partial z} + B \right) \quad (2.4)$$

where w is the vertical velocity, t is time, ρ is density, p'_d represents dynamic pressure perturbations, p'_b represents buoyancy pressure perturbations (see Appendix A for definitions of pressure perturbations), and B is buoyancy. A downward motion can be generated either by dynamic forcing (the first term on the *RHS* of Eq. 2.4) or by thermodynamic forcing (the second and third terms on the *RHS* of Eq. 2.4). Dynamic forcing is generally negligible, except in the case of occluding downdrafts in supercells. These occur when the low-level mesocyclone intensifies, which in turn strengthens the downward-directed dynamic vertical pressure gradient force. The case study analyzed by De Martin et al. (2025c) provides an example of this type of downbursts in Italy. More commonly, downbursts are thermodynamically induced by two mechanisms: latent cooling and hydrometeor loading. Latent cooling results from sublimation and melting of ice, as well as evaporation of liquid water. Hydrometeor loading, on the other hand, increases the density of the air parcel. Both processes enhance the negative buoyancy of the air mass, which then accelerates downward and collapses toward the ground generating a downburst.

Besides downbursts, convective windstorms can also be produced by cold pools associated with long-lasting *QLCSs* that deepen over time. The horizontal pressure gradient generated in this way can induce severe winds even in the absence of downdrafts. Moreover, the deepening of cold pools causes a rearward tilting of the updrafts because the *optimal* condition of the *RKW* theory is no longer satisfied. The tilting of the updrafts produces a pressure minimum above the cold pool due to hydrostatic effects (Weisman, 1993). This mid-level pressure minimum accelerates air into the convective system from the rear, generating a *Rear Inflow Jet* (RIJ). This dynamic is represented in Figure 2.1b, which displays the outcome of a real-case simulation conducted for this dissertation. An

updraft (shown with a red contour) is rearward-tilted, inducing a [RIJ](#) that is advected toward the ground by the downdraft, ultimately producing intense straight-line winds at the surface. The tilting of baroclinic vorticity by the leading updraft can also produce a couplet of *bookend vortices* (Weisman and Davis, 1998). These vortices can further intensify the [RIJ](#), promoting the bowing of the [QLCS](#) (*bow echo*, Fujita 1978). *Mesovortices* can also form along a [QLCS](#) through other mechanisms, such as the upward tilting of environmental vorticity (Wakimoto et al., 2006). The superposition of the horizontal pressure gradient associated with a deep cold pool, the descent of a [RIJ](#) toward the surface, and the wind field of a mesovortex produces the most intense convective straight-line winds.

Both high values of [CAPE](#) and strong wind shear are important for generating intense convective windstorms. High [CAPE](#) values strengthen the [RIJ](#) and increase the amount of hydrometeors within the updrafts, while environmental wind shear helps maintain strong updrafts despite the presence of deep cold pools.

Convective windstorms are fundamentally different from tornadoes and waterspouts. This distinction is briefly emphasized here, especially regarding the difference between waterspouts and downbursts, since this aspect is of particular relevance for the analysis presented in Section 3.1.

A waterspout is an intense columnar vortex of small horizontal extent that forms over a body of water, suspended from a cumuliform cloud (Glickman and Zenk, 2024c). Two major categories of waterspouts can be identified: mesocyclonic and non-mesocyclonic (Miglietta, 2019). Mesocyclonic waterspouts form in essentially the same way as mesocyclonic tornadoes over land. They are associated with supercell thunderstorms and can produce wind gusts up to 80 m s^{-1} (Rennó and Bluestein, 2001). Non-mesocyclonic waterspouts are generally weaker and can form through the stretching of shear instabilities along convergence lines (Lee and Wilhelmson, 1997) or within a mesovortex produced by a [QLCS](#) (Fujita, 1981; Trapp et al., 2005).

Whereas tornadoes are characterized by convergent and upward flow, convective windstorms are associated with descending and divergent flow. Typical downburst wind speeds range from 30 to 45 m s^{-1} , and they can cause damage of similar intensity to weak tornadoes (Fujita, 1990). Convective windstorms generally affect a larger area and last longer than waterspouts, although some downbursts can remain localized to only a few hundred meters in horizontal extent (*microbursts*).

2.3 Effect of topography on severe storms

In recent decades, several studies have addressed the interaction between local severe storms and topography, focusing particularly on the effects of mountains and valleys. Most of these studies relied on numerical simulations, since observational studies are limited by the fact that it is not possible to know the storms evolution in the absence of terrain (Bosart et al., 2006). These works have shown that orography can influence severe storms in multiple ways, sometimes enhancing them and at other times weakening them. Some of these interactions are briefly summarized below.

Mountain ridges can alter the propagation of outflows, modifying their speed and sometimes inducing hydraulic jumps on the lee side of mountains (Reeves and Lin, 2007). The altered propagation of outflows also affects the structure of [Mesoscale Convective Systems](#) ([MCSs](#)) and the triggering of new convective cells (Frame and Markowski, 2006). Supercell storms are highly sensitive to the underlying orography (e.g., Curic et al. 2007). A systematic numerical investigation of the effects of orography on supercells found that most storm modifications can be attributed to terrain-induced heterogeneity in sound-

ings and hodographs (Markowski and Dotzek, 2011). For instance, downslope winds in the lee of mountains systematically reduce [CAPE](#), that promotes a weakening of the storm. Similarly, Scheffknecht et al. (2017), studying a real case in the Alps, showed that storm intensity was influenced by the underlying topography. They noted that thermally induced plain-to-mountain flows near the ground created favorable conditions for supercell development.

Wind channeling within valleys can also promote tornadogenesis, as hypothesized for several events in both Europe (Peyraud, 2013) and the United States (Bosart et al., 2006). More recently, Feldmann et al. (2024) conducted a numerical investigation of supercells over complex terrain, characterized by mountains and a valley with a lake, based on resampled topography from the Prealps in northwestern Italy. They found that slopes, high-altitude terrain, and moisture sources created a strongly heterogeneous storm life cycle, with alternating phases of intensification and weakening. In general, valleys containing lakes were found to foster more supportive environments for supercells.

In summary, complex orography induces thermally driven flows, wind channeling, downslope winds, and variations in outflow propagation. All these elements influence the fundamental ingredients for severe storm development. Depending on the setting, severe storms may be either weakened or intensified, making their prediction particularly challenging in complex terrain.

The presence of the sea further complicates the interaction between storms and the surrounding environment (Lericos et al., 2007; Lombardo and Colle, 2012). Because the sea has a much higher heat capacity than land, it heats and cools at a slower rate, leading to temperature contrasts between land and offshore areas. This thermal gradient induces sealand breezes, which locally modify wind patterns and, consequently, the environmental vorticity. Moreover, the sea's higher heat capacity and greater surface moisture strongly influence [CAPE](#) and [CIN](#), producing substantial differences between offshore and inland regions. In addition, surface drag is generally lower over the sea than over land, resulting in stronger winds over water. Therefore, the presence of a nearby sea in combination with inland mountain ridges adds further complexity to the dynamics of local severe storms.

Italy provides an emblematic case, as the peninsula is bordered by the Mediterranean Sea and flanked by two steep mountain ranges, the Alps and the Apennines. Severe storms are frequent in this setting. Orography plays a key role in storm initiation (Miglietta et al., 2017b), while storm intensity can be modulated by the sea surface temperature of the Mediterranean (Ricchi et al., 2023; González-Alemán et al., 2023; Miglietta et al., 2017a). The topography and atmospheric stability creates contrasting flow regimes (*flow over* and *flow around*), which govern storm initiation and propagation (De Martin et al., 2024; Miglietta and Davolio, 2022; Pichelli et al., 2017). Additionally, tongues of high- Θ_e air can extend from the Adriatic Sea into the Po Valley, fostering the development of supercells and even tornadoes (De Martin et al., 2024; Miglietta et al., 2016).

These combined complexities make the prediction of convective high-impact weather phenomena in the region particularly challenging (Pilguy et al., 2022; Miglietta et al., 2016). While some recent improvements have been achieved (De Martin et al., 2025b; De Martin et al., 2025c), forecast skill for severe storms in complex terrain remains limited. Understanding these processes is further hindered by the scarcity of observational data. Most previous results are derived from numerical simulations, as observational networks are often too sparse to provide a comprehensive view of storm evolution. The strong spatial variability of atmospheric conditions in mountainous regions requires particularly dense observation networks, which are rarely available. To address this gap, dedicated field campaigns are underway (e.g., [Multi-scale transport and exchange pro-](#)

cesses in the atmosphere over mountains - TEAMX, Pfister et al. 2024) or planned (e.g., Thunderstorm Intensification from Mountains to Plain - TIM, Fischer et al. 2025).

2.4 Effect of urban land use on severe storms

Mountains and seas can be considered as first-order perturbative terms for the atmosphere, relative to a base state characterized by a flat surface with homogeneous land use. By contrast, a city can be considered a second-order perturbative term (or even lower, depending on its characteristics). Although cities exert a notable influence on the atmosphere (Oke and East, 1971; Bornstein, 1975), the perturbative scale of buildings (characteristic heights of 10-100 m) and urban land use (characteristic horizontal scale of 10 km) is smaller compared with that of mountains (characteristic heights of 1000 m) or seas (characteristic horizontal scale of 100-1000 km). Nonetheless, cities have received increasing attention in recent years because they represent hotspots for weather-related risks, encompassing hazards, vulnerabilities, and exposures, particularly in the context of climate change (IPCC, 2024). Furthermore, cities are expanding in both number and size (UN, 2019), thereby exposing a growing population to the impacts of urban weather. Within this context, the influence of urban areas on precipitation and storms has only recently begun to be investigated systematically (Yang et al., 2024), and limited attention has so far been devoted to the most extreme and damaging events.

Urban areas influence the PBL, for instance causing the well-known Urban Heat Island (UHI) effect (Oke, 1982; Ward et al., 2016), namely higher temperature over cities with respect to the surrounding rural areas. Several factors can explain the presence of the UHI (Deilami et al., 2018; Markowski and Richardson, 2011): decreased outgoing long-wave radiation because of the reduced sky view factor by buildings; increased short-wave absorption due to the thermal properties of urban materials; anthropogenic heat emissions from, e.g., transports and industries; decreased evapotranspiration (i.e., reduced loss of energy due to latent heat flux). Moreover, urban canopies significantly alter near-surface airflow, generally reducing wind intensity due to increased roughness (Bornstein and Johnson, 1977; Britter and Hanna, 2003; Di Sabatino et al., 2008), and increasing turbulence and vertical motions above buildings (Christen et al., 2009; Cintolesi et al., 2021). Additionally, urbanization often decreases moisture levels due to reduced vegetation, leading to the formation of an Urban Dry Island (UDI, Dou et al. 2015; Hao et al. 2023). Cities generate also the Urban Pollution Island (UPI, Ulpiani 2021), that was recently investigated by the Citizen Actions on Climate Change and Environment (I-CHANGE) project, funded by the European Union (Project number: 101037193).

Urbanization's influence extends beyond the PBL, potentially altering rainfall patterns through the Urban Rainfall Effect (URE). The URE phenomenon is studied since the 1950s. A large-scale field campaign named METROpolitan Meteorological EXperiment (METROMEX) was conducted in North America in the 1970s to test the hypothesis that urban areas can enhance precipitation. The results showed urban-induced increases in precipitation 25-75 km downwind of St. Louis, Missouri, during the warm-season months (Changnon and Ackerman, 1977). Cumulative amounts were enhanced by 5-25% (Changnon et al., 1991). However, Lowry (1998) raised doubts about the methodology used in that campaign. Subsequent observational studies using satellite data (Shepherd et al., 2002) and radar data (Bornstein and LeRoy, 1990; Niyogi et al., 2011) further supported the hypothesis that urban areas influence the development of convective rainfall, typically intensifying precipitation downstream of cities. The URE phenomenon can be driven by multiple factors: (a) UHI can favor vertical motions (Thielen and Gadian, 1997;

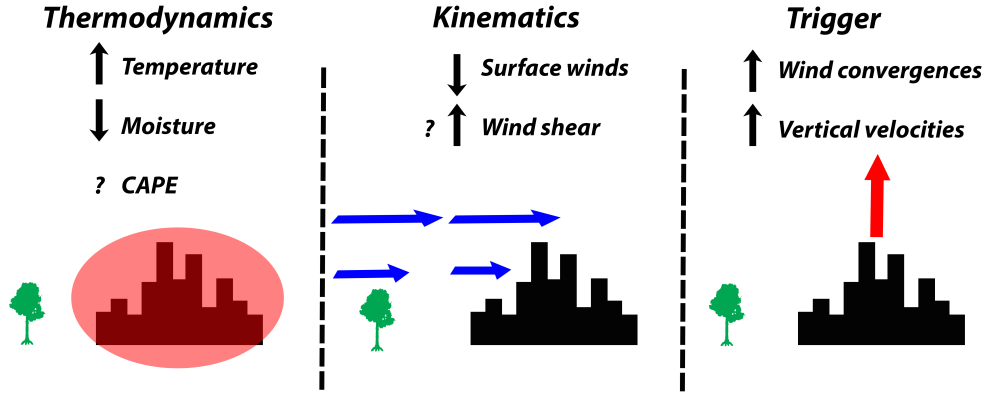


Figure 2.2: Sketch of urban effects on severe storm ingredients.

Rozoff et al., 2003; Shepherd, 2013); (b) increased roughness leads wind convergences and subsequent convection initiation (Yang et al., 2021a); (c) buildings can alter the movement of convective systems (Bornstein and LeRoy, 1990); and (d) urban aerosols might affect rainfall efficiency and cloud formation (Lin et al., 2011; Fan et al., 2020). Some studies have noted that the URE may depend on factors such as the intensity of the incoming flow, the strength of the UHI, and the surrounding topography (Yang et al., 2019; Shen and Yang, 2023). In recent years several studies based on numerical simulations supported the thesis that cities influence the development of convective rainfalls, for instance in Beijing, China (Yang et al., 2014), Paris, France (Forster et al., 2024), Indianapolis, Indiana (Moraglia et al., 2024) and Phoenix, Arizona (Yang et al., 2019). The URE phenomenon is much more evident in compact urban areas, typical in Europe, Eastern US, and China, than in more dispersed cities such as those in Africa or South America (Yang et al., 2024). A systematic meta-analysis of peer-reviewed articles on URE revealed that precipitation increases by approximately 18% downwind of cities and by 16% over cities, as found in both observational and numerical studies, although the latter type of studies tends to underestimate the precipitation changes (Liu and Niyogi, 2019). The study also noted that urban feedbacks on rainfall events are stronger during the day than at night. The recent [Tracking Aerosol Convection Interactions Experiment \(TRACER\)](#), conducted in Houston, Texas, in 2021-2022, may provide new insights into the URE phenomenon from an observational perspective (Jensen et al., 2025).

While there is consensus that urban areas significantly influence rainfall patterns, albeit without a universally accepted theory on the URE phenomenon (Shepherd, 2013), little research has focused on the effect of urban areas on the most extreme convective events, such as hailstorms, convective windstorms, and tornadoes. Urban land use can modify all the ingredients necessary to produce severe storms, as summarized in Figure 2.2. Wind convergences are enhanced over the cities, favouring vertical motions and convective initiations (Yang et al., 2021a; Rozoff et al., 2003), even if it is not clear if those wind convergences are dynamically or thermodynamically induced. UHI and UDI influence instability: some authors found a decrease of it over cities (Yang et al., 2021b; Reames and Stensrud, 2018; Yang et al., 2014), while others found an increasing amount of CAPE over urban areas (Moraglia et al., 2024). Increased aerodynamic roughness affects wind shear by altering surface winds, usually leading to higher values of SRH over cities (Fan et al., 2023; Reames and Stensrud, 2018). Consequently, it is reasonable to expect that severe storms could be influenced by urban areas. However, severe storms are highly energetic atmospheric phenomena (Lilly and Jewett, 1990), making it theoretically

challenging to determine whether urban-induced perturbations are strong enough to significantly impact on storm dynamics. Moreover, it is not clear *a priori* whether cities may intensify or weaken a severe storm, given the uncertainties in the urban effect on some parameters, such as [CAPE](#).

Numerical experiments have begun to address this question in the last period. Reames and Stensrud (2018) studied a supercell event in the US Great Plains using the [Weather Research and Forecasting \(WRF\)](#) model with hierarchical clustering analysis, examining whether the location of Dallas-Forth Worth, Texas, would influence the development of a supercell. They found that the low-level mesocyclone was strongly dependent to the city's location. Similarly, Naylor (2020) and Naylor and Mulholland (2023) observed an intensification of a [QLCS](#) crossing an idealized city, characterized by higher temperatures and surface roughness length, though not considering other factors like moisture, using the CM1 cloud model. A similar outcome was reached studying a supercell (Naylor et al., 2024). Finally, Lin et al. (2021) examined the impact of Kansas City, a medium-sized US city, on a supercell event using the [WRF-CHEM](#) model, which also accounts for urban aerosols. They found a 20% increase in large hail occurrence in the presence of the city, with land use having a greater effect on the storm than aerosols. A follow-up study explored tornado potential in the same event, revealing stronger low-level vorticity production over urban land use compared to croplands (Fan et al., 2023).

While the previous works analyzed the mesoscale effects of cities on storms, some studies have analyzed the microscale interactions between severe storms and buildings. In particular, several engineering works have investigated the wind loads of downbursts (Chen and Letchford, 2004) and tornadoes (Yang et al., 2011a; Haan Jr et al., 2010) on buildings. Much less literature addresses the opposite perspective, namely, the effect of buildings on downbursts or tornadoes. Kawaguchi et al. (2019) conducted a numerical study using a hybrid meteorological model coupled with an engineering [LES](#) to study the effect of buildings on a tornado event in Japan. They found that the tornado structure was significantly deformed by buildings, splitting into finer vortices. Overall, the tornado appeared to be weakened by the presence of buildings, although some intense wind peaks still occurred due to wind channeling in the urban canyons. The weakening of tornadoes due to urban structures had already been suggested by Fujita (1973), who studied the frequency of tornadoes in Chicago and Tokyo. Regarding convective windstorms, Bélair et al. (2025) analyzed the passage of a thunderstorm outflow over Montreal (Canada) using a [LIDAR](#) and found that the urban environment did not significantly alter the dynamics of thunderstorm outflows.

2.5 Summary: current gaps in literature

The high-impact weather phenomena studied in this dissertation are physically quite different, as introduced in the previous sections of this Chapter. Downslope windstorms are dynamically induced when a *sub-critical* flow is forced over a terrain barrier in the presence of a stable layer, generating hydraulic jump-like behavior in the lee of the mountain (Figure 2.1a). In contrast, convective windstorms may arise from either thermodynamic effects, induced by the negative buoyancy of a storm downdraft (*downburst*), or dynamic effects driven by pressure gradient forces associated with the evolution of a [QLCS](#) (Figure 2.1b). Furthermore, supercells, which can also produce intense windstorms, differ from [QLCS](#) in that they are sustained by mesocyclone propagation rather than by the balance between cold pools and environmental vorticity.

Although there is broad agreement on the physical processes that generate these high-

impact weather phenomena, many research questions remain open. In particular, the interactions of high-impact weather with mountains, water bodies, and land use have received increasing attention in recent years, largely because advances in numerical modeling now allow more accurate investigations that were not feasible two decades ago. The following paragraphs summarize the open scientific gaps relevant to this dissertation, with references to the chapters where they are addressed.

Understanding the physical processes producing high-impact weather

Regarding downslope windstorms, discrepancies still exist between the *hydraulic theory* and the *resonant amplification theory*, as reviewed by Lin (2007) and in section 2.1. Similarly, research on the internal dynamics of QLCSs is ongoing, due to the limitations of the RKW theory (see section 2.2.1), which have motivated researchers to further explore the complexity of QLCS dynamics (Bryan and Rotunno, 2014; Coniglio et al., 2012; Weisman and Rotunno, 2004). Given the high impact of supercells, active research is also devoted to improving our understanding of their dynamics (Coffer et al., 2023; Peters et al., 2020) and the hazards they can produce (Fischer et al., 2024; Kumjian et al., 2021). Although these topics are of great interest, they fall beyond the scope of this dissertation, which is primarily focused on the interaction of severe storms with topography and land use at the mesoscale.

Predictability of high-impact weather events

Complex topography and land-use heterogeneity (e.g., the presence of water bodies and cities) make the forecasting of high-impact weather particularly challenging, as reviewed in Sections 2.1 and 2.3. This is due to several factors, including insufficient grid resolution, inaccuracies in numerical methods, and limitations in physical parameterizations. Data assimilation (Wheatley et al., 2015), ensemble approaches (Sobash et al., 2016), and post-processing statistical methods (including machine learning, Sobash and Ahijevych 2024) are being used to address these limitations. The predictability of convective windstorms in complex topographic regions will be briefly discussed in Section 3.1, where the performance of a high-resolution convection-permitting ensemble will be analyzed. In Sections 3.2 and 4, the skill of a numerical model in reproducing intense wind gusts during both downslope and convective windstorms will be assessed.

Urban Footprints on the Pre-Storm Environments

Few studies have analyzed the modification of the atmospheric environment prior to convective storms. Those that did report divergent results, summarized in Fig. 2.2. For example, both increases and decreases in CAPE over urban areas have been found, depending on the study, with the majority reporting a decrease. Moreover, the effect of urban land use on wind shear requires further investigation, as only a few case studies have addressed this issue, noting a decrease in SRH over urban areas. Furthermore, while there is agreement that urban land use generates wind convergences, it is not clear whether these are dynamically induced by buildings or are a thermodynamic effect related to the UHI. Another important limitation of the studies carried out so far is that they focus mainly on US or Chinese cities, while European cities have been poorly investigated. This aspect is examined for a European city in Chapter 4 and systematically for idealized cities in Chapter 5.

Mesoscale effects of the urban land use on severe storms

Three main uncertainties surround the effects of urban areas on severe storms at the mesoscale ($\sim 10\text{-}100$ km): i) What modifications does urban land use induce in the ingredients necessary to produce severe storms? ii) Are these modifications strong enough to perturb highly energetic severe storms? iii) What are the physical processes driving the urban-induced modifications?

The literature addressing these aspects is very limited, as discussed in section 2.4. The first point relates to the previous paragraph, while regarding the second, the studies conducted so far have mainly found an intensification of supercells and QLCSs due to urban land use. However, these works are based on case studies of US cities. Thus, the generalization of this outcome to other convective events and cities remains unclear. Even less work has been carried out on the third point (e.g., Lin et al. 2021). The mesoscale effects of urban land use on severe storms are investigated in detail in Chapters 4 and 5.

Microscale effects of buildings on windstorms

The microscale ($\sim 1\text{-}1000$ m) effects of urban surfaces on windstorms and tornadoes have not been extensively investigated so far, with rare exceptions (e.g., Kawaguchi et al. 2019; Bélair et al. 2025), mainly due to computational limitations and the lack of dedicated observational campaigns. The new generation of meteorological models (e.g., CM1, Bryan and Fritsch 2002, and FastEddy, Sauer and Muñoz-Esparza 2020) makes it possible to study these interactions. For instance, Rozoff et al. (2023) studied the interactions between intense winds caused by a tropical cyclones and buildings with the CM1 model. Furthermore, specific field campaigns are planned to investigate this aspect, such as the [Thunderstorm outflows measurement and modeling for strongWIND nowcast and RISK mitigation \(WIND RISK\)](#) project (Burlando et al., 2024). The mesoscale simulations presented in Section 3.2, conducted within the framework of the [RISKADAPT](#) project, were used to perform [CFD](#) simulations of the interaction between a downslope windstorm and buildings. These results will be published soon by the research group but were not included in this dissertation, as they fall beyond its scope. Regarding the broader application of [LES](#) simulations to high-impact weather events in urban areas, we believe there is considerable potential for significant advancements in the coming years.

Detailed observations of high-impact weather

A major chronic limitation in high-impact weather research in complex terrain or urban areas is the lack of dense observational networks, as discussed in Sections 2.3 and 2.4. In such settings, the atmospheric environment is highly heterogeneous, requiring dense observations to capture its complexity. Furthermore, high-impact weather is often characterized by sudden, abrupt changes in certain variables (e.g., wind) that last from seconds to minutes, which also requires a high sampling frequency. Although advancements in numerical models are enabling more insightful studies of high-impact weather events in complex terrain and urban areas, the lack of sufficient observations still hinders the understanding and prediction of these phenomena, with consequences for early warning systems and citizens safety, as will be discussed in Section 3.1. Ongoing ([TEAMX](#), [WIND RISK](#) and [RISKADAPT](#)) and planned ([TIM](#)) field campaigns may help fill this major gap.

Chapter 3

High-Impact Windstorms in Complex Terrain Settings

The study of atmospheric phenomena over complex terrain poses several additional challenges compared to flat and homogeneous regions, as introduced in Chapter 1 and reviewed in Section 2. In summary, complex topography strongly influences the dynamics of mesoscale phenomena (e.g., wind channeling, thermally driven flows). Second, obtaining accurate and detailed observations of high-impact weather events is generally more difficult, both because of the greater spatial variability of atmospheric variables and the practical limitations associated with instrument deployment. Third, numerical models often struggle to provide reliable simulations and forecasts in such environments, mainly due to insufficient grid resolution or inaccuracies in physical parameterizations.

For these reasons, despite the remarkable advances achieved in recent decades (as discussed in Section 2), both the understanding and prediction of high-impact weather events in complex terrain remain challenging. Two case studies are analyzed in this chapter to highlight the current strengths and limitations in this area of study, with a particular focus on Italy. Both events are windstorms: in the first case, the windstorm was caused by a convective storm that hit Sicily in August, while the second event was a downslope windstorm that occurred in northeastern Italy in February. Both events also took place in coastal areas bordered by inland mountains. Finally, both had significant human impacts: in the first, seven people died due to the sinking of a yacht, while in the second, there was one fatality and at least ten people were injured by the Bora winds. These two cases are therefore representative of summer and winter high-impact weather events along the coasts of the Mediterranean Sea. The available observations are analyzed to shed light on the characteristics of these events. We also evaluate the performance of numerical models in reproducing the windstorms, in order to assess the reliability of state-of-the-art meteorological models for accurate forecasting.

The first case, discussed in Section 3.1, concerns the sinking of the yacht *Bayesian* in Porticello, which occurred at 02:06 UTC on 19 August 2024. All the available observations are analyzed to study what weather phenomenon caused the sinking, as well as the outputs of a convection-permitting ensemble to investigate its predictability. Hence, scientific question A stated in section 1.3 is addressed. This section is adapted from De Martin et al. (2025b).

The second analyzed case, discussed in Section 3.2, is the violent Bora wind event in Trieste on 11-12 February 2012. Here, the focus is on understanding the influence of Trieste's local orography on the intensity of the Bora wind gusts. The ability of numerical models to reproduce these gusts is also investigated, and question B of section 1.3 is then

addressed.

Finally, in Section 3.3, we discuss the strengths and limitations of understanding and predicting high-impact weather events in complex terrain, in light of the two case studies.

3.1 The *Bayesian* sinking in Porticello: a predictable convective windstorm?

3.1.1 Introduction

At 02:06 UTC on 19 August 2024 the *Bayesian* yacht sank approximately 600 meters off Porticello, a small village east of Palermo, Sicily (Figure 3.1). The sinking occurred during the passage of a thunderstorm, which was accompanied by heavy rain and severe wind gusts. Of the 22 people on board, seven (four British, two Americans and one Canadian) lost their lives, making this, to the authors knowledge, the deadliest thunderstorm event in Italy, excluding flood events, since the Venice and Padua tornado of 1970 (Janeselli, 1972; Borghi and Minafra, 1972; Miglietta and Matsangouras, 2018). This tragedy captured public international attention due to the unclear circumstances surrounding the sinking. The yacht was of considerable size: 56 meters long, with a mast reaching 75 meters in height. Since the event occurred at night, there is no clear footage, and no direct witness saw the sinking. Additionally, no radar data were available for the area at that time. While some witnesses claimed to have seen a waterspout offshore near Porticello, no concrete evidence was provided.

This has led to a debate about whether a waterspout or a convective windstorm caused the sinking. These two phenomena, although capable of producing extremely severe wind gusts, are quite different, as pointed out in section 2.2.2. Both waterspouts and convective windstorms are common in the Mediterranean Sea, especially between late summer and autumn (Burlando et al., 2018; Miglietta and Matsangouras, 2018). The scientific question regarding which meteorological phenomenon caused the tragedy is relevant to clarify the actual reason for the sinking, since pressure forces exerted on the yacht differ significantly between a waterspout and a downburst (Sengupta et al., 2008). Another controversial aspect of the *Bayesian* accident is whether the storm was predictable. Severe storms are phenomena known for their poor predictability; however, ensemble modelling systems can provide an indication of the areas with higher likelihood of storm occurrence (Sobash et al., 2016; Johns and Doswell III, 1992). A yellow warning for thunderstorms was issued one day ahead on 19 August 2024 on the northern coasts of Sicily¹. In Italy warnings are issued by the Civil Protection and are impact-based, with four colour levels, the yellow being the second lowest, referring to events responsible for localized damage. The colour level relates to the expected coverage of the impacts, rather than the degree of confidence in the occurrence of severe weather. This chapter is devoted to shedding light on the atmospheric phenomenon that caused the *Bayesian* yacht sinking, exploiting all the available observations. The predictability aspect is investigated as well, using outputs from deterministic and probabilistic simulations initialised one day in advance.



Figure 3.1: Map of Sicily, in which the sounding location (Trapani) is highlighted; b) zoom in the red box shown in a), including the locations of three weather stations (Aspra, Bagheria and Solanto), as well the position of the *Bayesian* sinking area, and of the videos #1, #2 and #3 (field of view shown with red bows). Background maps copyright Google Maps. The storm motion retrieved from satellite data is shown with a red arrow.

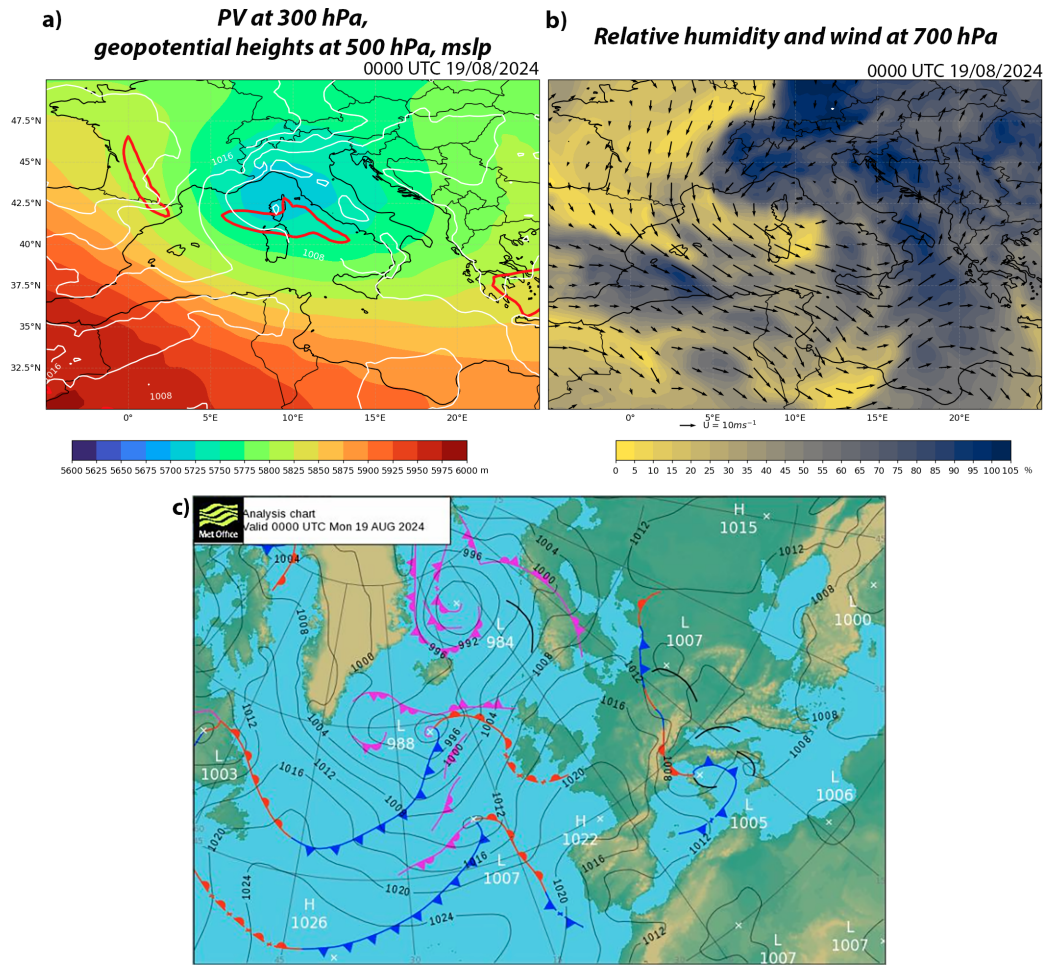


Figure 3.2: ERA5 reanalysis data at 00:00 UTC on 19 August 2024: a) Potential Vorticity at 300 hPa (PVU units, red contour for Potential Vorticity equal to 2 PVU), geopotential heights at 500 hPa (colour shading) and mean sea level pressure (mslp, white contours); b) relative humidity (colour shading) and wind speed at 700 hPa (arrows); c) surface pressure map (with fronts) of the Met Office at 00:00 UTC on 19 August 2024.

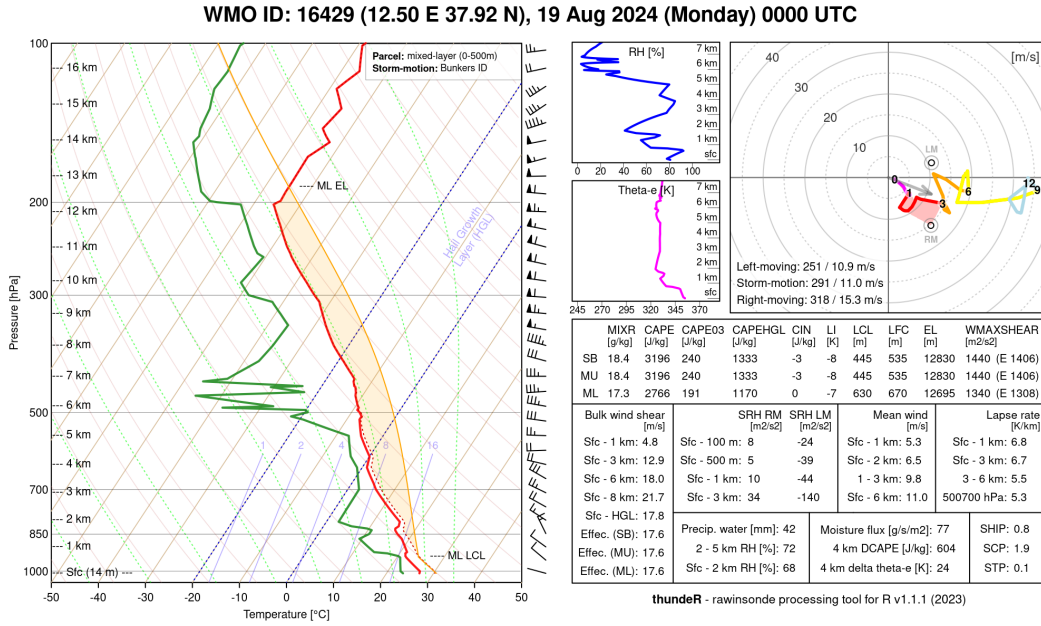


Figure 3.3: Sounding launched at Trapani (location shown in Figure 3.1a) at 23:00 UTC on 18 August 2024. Some relevant parameters are shown. The Mixed Layer (ML) is computed for the first 500 m AGL. Source of the plot is http://www.rawinsonde.com/thunder_app. An explanation of plotted parameters is available at: http://www.rawinsonde.com/thunder_app/thunder_v1_1_parameters.pdf (Accessed on 23 November 2024).

3.1.2 Synoptic and mesoscale analysis

Starting from a synoptic analysis it is observed that between 18 and 19 August 2024 a cut-off low² moved eastward from southern France toward Central Italy. It was well-defined in the upper levels at 00:00 UTC on 19 August with a closed minimum of geopotential height at 500 hPa between Sardinia and the Ligurian Sea and a positive anomaly of Potential Vorticity at 300 hPa (Figure 3.2a). On the southern side of the upper-level cyclone, there was a strong gradient of geopotential height (about $25 \text{ gpm}(100\text{km})^{-1}$, Fig. 3.2a) between the cyclone and a subtropical ridge over the western Saharan desert, resulting in an intense upper-level north-westerly flow (peak of 36 m s^{-1} at 326 hPa in the sounding in Trapani shown in Fig. 3.3). At the surface, a cyclonic circulation was centred close to the Corsica Island (Figure 3.2c). Within this configuration, a cold front (Fig. 3.2c) moved eastward across the Tyrrhenian Sea during the night of 19 August, pushed by intense westerly winds (20 m s^{-1} at 700 hPa over Sardinia, Fig. 3.2b), followed by a drier air mass (Figure 3.2b). This front triggered deep moist convection over the sea, which subsequently affected the northern coasts of Sicily.

A sounding was launched in Trapani (90 km west of Porticello; Fig. 3.1) at 23:00 UTC on 18 August 2024 (3 hours before the sinking), just before the cold front approached northern Sicily, in the prefrontal side. Thus, it can be considered representative of the mesoscale environment in which the storm leading to the *Bayesian's* sinking developed. The sounding profile (Fig. 3.1) revealed a very moist air mass in the first 500 m AGL, with an average mixing ratio of 17.3 g kg^{-1} in this layer (18.4 g kg^{-1} at the surface), result-

¹<https://github.com/pcm-dpc/DPC-Bollettini-Criticita-Idrogeologica-Idraulica> (accessed 23 November 2024).

²A cold low that has grown out of a trough and become displaced out of the basic westerly current and lies equatorward of this current (Glickman and Zenk, 2024a)

ing in a low **LCL** at 630 m and in high values of **Mixed Layer (ML) CAPE** (2766 Jkg^{-1}), without a capping inversion. Dry layers were located at 2 km and between 6 and 8 km. In the first 1 km, the wind speed was weak. However, above that elevation, a westerly flow steadily intensified up to 12 km, causing moderate-to-high values of wind shear (18.0 ms^{-1} between the surface and 6 km). **SRH** 0-3 km was weak ($34 \text{ m}^2\text{s}^{-2}$) because speed shear was only moderate in the 0-3 km layer (12.9 ms^{-1}), and there was weak directional shear. Thus, the environment was conducive to the development of severe storms: there were high values of moisture in the **PBL**, significant instability, moderate-to-strong deep-layer shear, and an approaching cold front. The environment appeared supportive of severe wind gusts, given the presence of dry layers in the mid-troposphere (Atkins and Wakimoto, 1991), the high moisture content (**Precipitable Water PW** as high as 42 mm), especially near the surface, the high instability, and strong mid-level unidirectional flow (Taszarek et al., 2020; Squitieri et al., 2023). The low **SRH** (either in the 0-3 km, 0-1 km and 0-500 m layers) limited the potential for mesocyclonic tornadoes (Coffer et al., 2020; Nixon and Allen, 2022), although local scale processes (such as the interaction with coastal topography and outflow boundaries) may have locally altered the **SRH** in the proximity of the storm. Conversely, splitting-storm supercells were still possible due to substantial wind shear (Weisman and Klemp, 1986); also, the low **LCL** and the substantial **Surface Based (SB) CAPE** in the first 3 km of the troposphere (**SB CAPE** 0-3 km equal to 240 Jkg^{-1}) are known to be favourable conditions for tornado occurrence (Rasmussen, 2003; Shabbott and Markowski, 2006).

3.1.3 Observational description of the severe storm

The severe storm that hit the *Bayesian* yacht developed in the central Tyrrhenian Sea at approximately 23:00 UTC on 18 August 2024, three hours before the sinking. It moved east-south-eastward, exhibiting characteristics of a **MCS**. Since radar images were not available for that storm, data from the **spinning enhanced visible and infrared imager (SEVIRI)** instrument on the geostationary Meteosat-11 satellite were used to analyse the storm evolution. All the images studied here are from the thermal infrared band centered at $12.0 \mu\text{m}$ (Figure 3.4; an animation, between 00:12 UTC and 02:57 UTC on 19 August 2024 is provided in Video S1). The animation shows the growth and decay of several convective cells within the **MCS**. Notably, a convective cell developed between 01:27 UTC and 01:57 UTC in the southwestern corner of the **MCS** (indicated by the white arrow in Figure 3.4a). This cell reached its minimum cloud top temperature at 01:57 UTC (approximately 210 K, suggesting the presence of an overshooting top) before it started to weaken. A cold U-V feature appeared at 01:57 and 02:12 UTC, which highlighted the severity of the storm (McCann, 1983; Brunner et al., 2007). That storm struck the *Bayesian* yacht. The minimum brightness temperature between 01:42 UTC and 02:27 UTC (line in Fig. 3.4b), representative of the overshooting storm top, was tracked to retrieve the storm motion (Bedka et al., 2010). The obtained storm motion is from 283° , which is close to the value estimated from the sounding averaging the wind in the 0-6 km layer (291° , Fig. 3.3). Conversely, the expected direction for a right-mover supercell from the sounding (following Bunkers et al. 2000) is from 318° (Fig. 3.3). Thus, the convective cell was not following the direction expected for a right-mover supercell (see black and blue arrows in Fig. 3.4c), suggesting that the storm causing the *Bayesian* sinking was not a supercell.

Regarding surface observations, official weather stations are not present close to the area affected by the sinking. However, three ground weather stations from the amateur network "Weather Underground" are located less than 5 km away from the location of the *Bayesian*, in the villages of Solanto (1.4 km away), Bagheria (3 km away), and Aspra (4.5

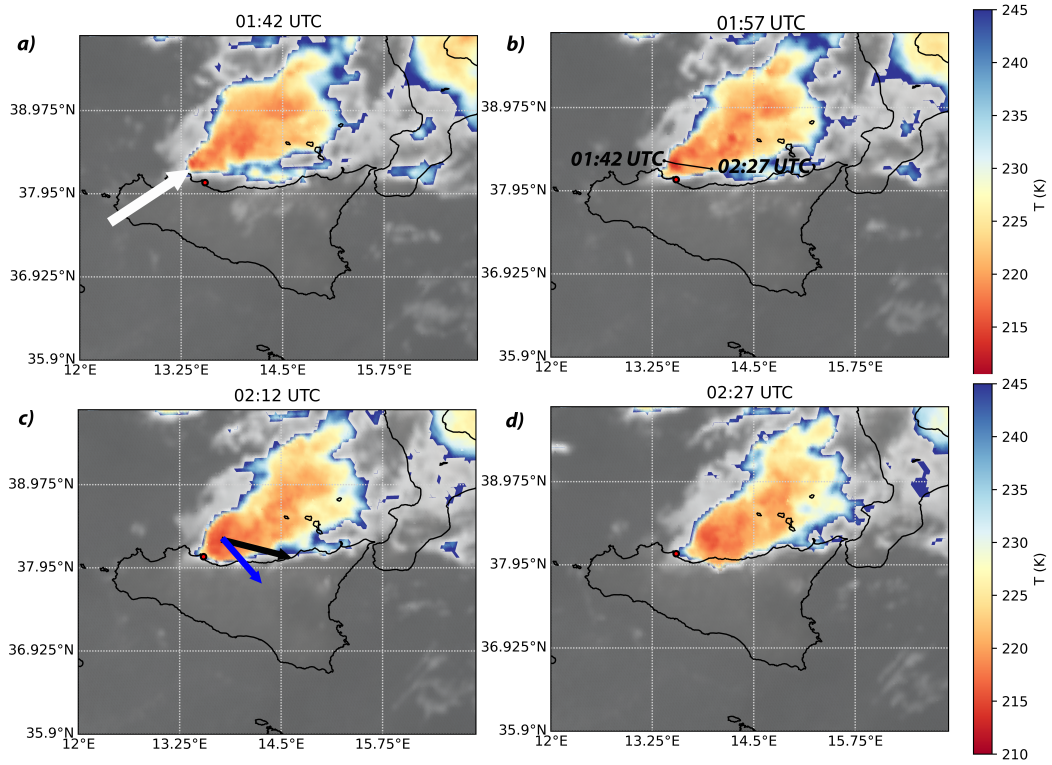


Figure 3.4: Equivalent Brightness Temperature inferred from infrared band $12.0\ \mu\text{m}$ of **SEVIRI** instruments at: a) 01:42 UTC, b) 01:57 UTC, c) 02:12 UTC, d) 02:27 UTC on 19 August 2024 (parallax-corrected). The *Bayesians* yacht location is shown with a red dot. The time shown is the end time of the 12-min period of acquisition of **SEVIRI**. In (b) the track of the minimum of Brightness Temperature between 01:42 UTC and 02:27 UTC is shown. In (c) the storm motion retrieved from the overshooting storm top (black arrow) and the one estimated from the sounding for a right-mover supercell with the Bunkers method (blue arrow) are shown. The parallax was corrected.

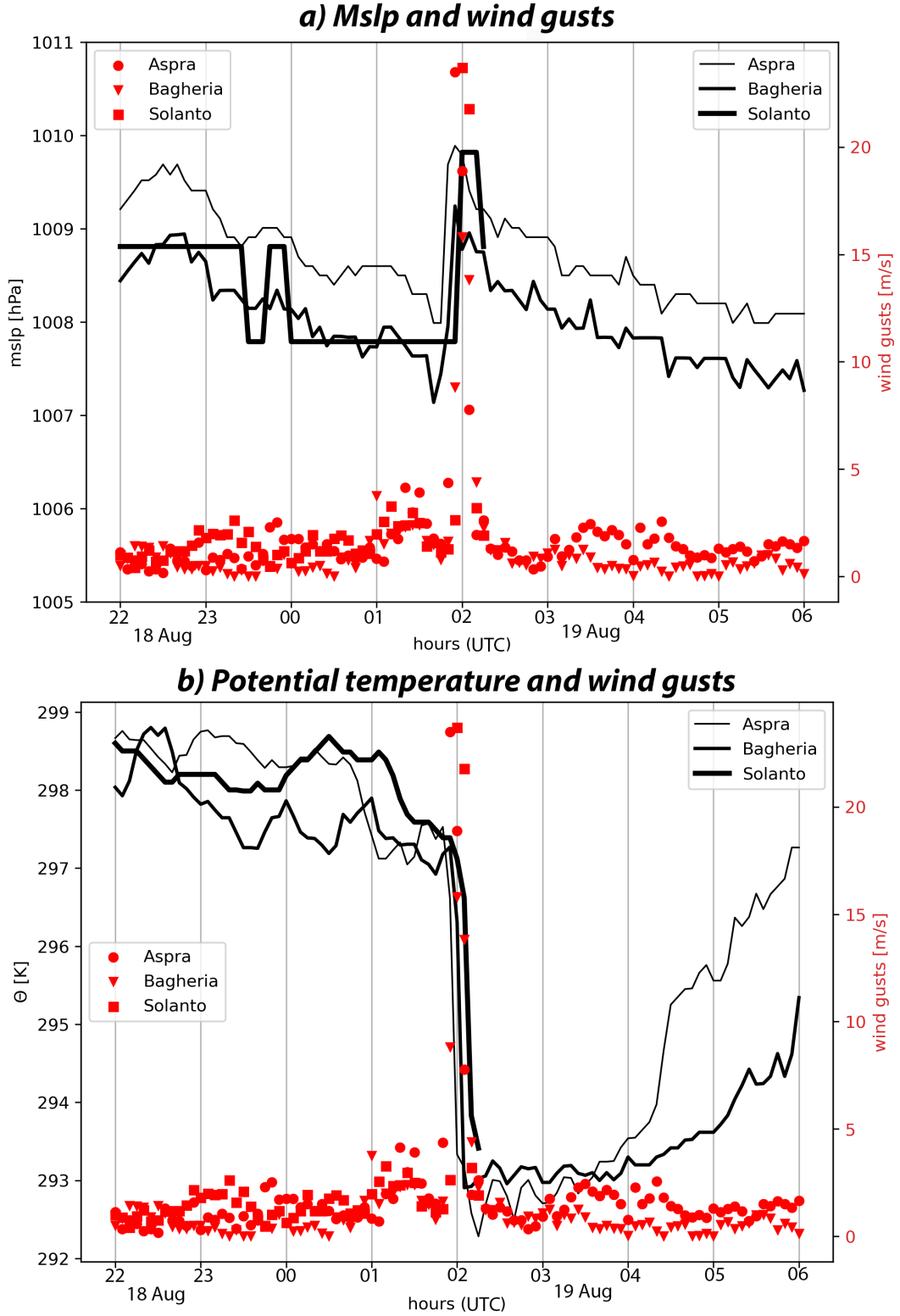


Figure 3.5: Data from ground weather stations (locations shown in Figure 3.1b) between 22:00 UTC on 18 August 2024 and 06:00 UTC on 19 August, 2024: a) **mslp** (black lines, thickness depending on the location, resolution of ± 0.1 hPa in Aspra and Bagheria, ± 1 hPa in Solanto) and wind gusts (red dots, square and triangles, resolution $\pm 0.1 \text{ ms}^{-1}$); b) potential temperature (black lines, thickness depending on the location, resolution of ± 0.1 K) and wind gusts (red dots, square and triangles, resolution $\pm 0.1 \text{ ms}^{-1}$). Sampling frequency of all the weather stations is 5 minutes.

km away). Comparison with the satellite data shows that these three weather stations were struck by the southwestern side of the severe storm. The data plotted in Figure 3.5 were compared with those collected by the official weather stations in the surroundings to verify that the measured quantities were in reasonable agreement. The coherent evolution in the three weather stations, compatible with the passage of the gust front as described hereafter, further assure the reliability of these data for investigating surface atmospheric evolution. Nevertheless, we cannot exclude that they could be biased, and, thus, the absolute values of the measurements should be taken with caution. In all three stations, [mean sea level pressure \(mslp\)](#) steadily decreased by about 1.5 hPa between 22:30 UTC and 02:00 UTC. A slightly more pronounced decrease of [mslp](#) (0.3 hPa in Aspra and 0.5 hPa in Bagheria) is observed at 01:45 UTC. Then, a sharp increase of 2 hPa was observed in correspondence with the passage of the gust front at 02:00 UTC, with wind gusts up to 24 ms^{-1} . The pressure increase was slightly delayed between the three weather stations because Aspra, the westernmost location, was affected by the storm first, and Solanto, the easternmost, was hit last (Figure 3.1b). Subsequently, the pressure slightly decreased. In correspondence with the pressure jump, a potential temperature drop of about 4.5 K was observed (Figure 3.5b). The sharp increase in pressure, the sudden drop in potential temperature, and wind increasing at the same time are well-known signatures of gust front occurrence (Iwashita and Kobayashi, 2019). The recorded wind gusts were not extremely high (24 ms^{-1}), possibly because amateur weather stations were not ideally located (i.e., at 10 m [AGL](#)), and, thus, underestimated their intensity. It is likely that offshore, where the surface drag was reduced and the storm possibly stronger, the wind was more intense. Furthermore, before the pressure increase, the fast pressure drop that is typically associated with mesocyclones was not observed (Straka et al., 1996). However, a drop of 0.3-0.5 hPa was recorded, possibly the effect of the meso-low ahead of a linear convective system (Johnson and Hamilton, 1988). In agreement with the satellite analysis, surface weather stations data suggest that a supercell was not present in Porticello that night.

Three videos (locations #1³, #2⁴ and #3⁵ in Figure 3.1b) were recorded during the event along the coast, between 500 and 1000 m away from the *Bayesian* yacht. Considering the storm's trajectory, these locations could have potentially been affected by a hypothetical tornado, which would have subsequently caused the *Bayesian* sinking (Fig. 3.1b). These videos do not show the presence of a waterspout or of rotating winds. They do show 40-50 seconds of very intense wind gusts, with significantly reduced visibility, preceded by a few minutes of strong wind gusts (possibly the gust front). One of these videos also shows some minor damage (chairs, tables, and umbrellas being moved by the wind). In Table 3.1, wind data from the weather stations, a description of video #2 (Figure 3.1b), and the timing of the *Bayesian* yacht movements during the storm, as reported by the [Automatic Identification System \(AIS\)](#) published in the media⁶, are combined. The yacht began to swing at about 01:50 UTC. The first wind gusts in Aspra were recorded at 01:54 UTC, and in Bagheria between 01:54 UTC and 01:59 UTC. The swinging of the *Bayesian* was likely caused by the waves and wind associated with the gust front. According to [AIS](#) data, the *Bayesian* likely lost its anchor at 01:59 UTC, when the strongest gust was measured in Aspra (23.5 ms^{-1}), and in the location of video #1 the wind was already quite strong (possibly stronger than 15 ms^{-1} , because tarpaulins were flying). The yacht then moved south-eastward, while at 02:04 UTC the strongest wind gust was

³<https://x.com/PavanFederico00/status/1825799907573113097> (accessed 27 November 2024).

⁴<https://www.youtube.com/watch?v=6MQL-eiVKbs> (accessed 27 November 2024).

⁵<https://x.com/PavanFederico00/status/1825798132132671582?t=fvY1V2u4G1xaZscX91w1mg&s=19> (accessed 27 November 2024).

⁶<https://www.dailymail.co.uk/news/article-13769241/Bayesian-shipbuilder-claims-crew-16-minutes-warn-passengers.html> (accessed 30 November 2024).

measured in Solanto (23.7 ms^{-1}), and at 02:05 UTC a downburst was likely occurring in video #1. The yacht sank at 02:06 UTC, while the wind was significantly decreasing in the nearby Solanto. Thus, it appears that the initial gust front, likely associated with gusts of up to $15\text{--}20 \text{ ms}^{-1}$, caused the yacht swinging and the loss of its anchor. Then, when a downburst likely occurred, the *Bayesian* was adrift and sank at the end of the storm.

Time (UTC)	Wind gusts Aspra (ms^{-1})	Wind gusts Bagheria (ms^{-1})	Wind gusts Solanto (ms^{-1})	Video location #1 (Figure 1b)	<i>Bayesian</i> yacht
1:50					Swinging
1:54	12.8	2	2.9		Swinging
1:59	23.5	13.8	5.1		Anchor free
2:01		15.8		Tarpaulins flown	Moved toward SE
2:04	18.9		23.7	Strong wind	Moved toward SE
2:05			20.8	Strong wind	Moved toward SE
2:06				Moderate wind	Sinking
2:08		10.7			
2:09	14.3				
2:14	3.6	6.6	3.2		

Table 3.1: Data of weather stations of the "Weather Underground" network in Aspra, Bagheria and Solanto, and some description of video #1 (locations shown in Figure 3.1b). The behavior of the *Bayesian* yacht as inferred from AIS track is also reported.

3.1.4 Predictability of the event

For the purpose of this study, we employed the ICON model (Zängl et al., 2015), operationally run at ARPAE-SIMC, the HydroMeteorological and Climate Service of the Emilia-Romagna region in Italy. Outputs of both deterministic and probabilistic weather forecasts over Italy at a horizontal grid spacing of 2.2 km were used. Specifically, the deterministic model, called ICON-2I, is initialized twice a day at 00:00 UTC and 12:00 UTC using analyses generated by the kilometre-scale ensemble data assimilation (KENDA; Schraff et al. 2016) system, in which the local ensemble transform Kalman filter (LETKF; Hunt et al. 2007) scheme is implemented. The KENDA-LETKF system is employed to assimilate conventional observations as well as radar reflectivity and radial wind data from the Italian radar network (Gastaldo et al., 2021; Gastaldo et al., 2022). The boundary conditions for ICON-2I are provided by the High RESolution Integrated Forecast System (IFS-HRES), run operationally at the European Centre for Medium-Range Weather Forecasts (ECMWF). The probabilistic model, called ICON-2I-EPS, consists of 20 ensemble members and has a forecast range of 51 hours. ICON-2I-EPS is initialized once a day at 21:00 UTC, using analyses generated by KENDA-LETKF, with boundary conditions provided by the IFS ensemble members (IFS-ENS).

Some relevant outputs from these operational simulations are presented in Figure 3.6 and are used to assess the predictability of the downburst. The ICON-2I-EPS run initialized at 21:00 UTC on 17 August 2024 (29 hours ahead of the sinking) shows probabilities between 10% and 25% for wind gusts exceeding 24.5 ms^{-1} in the southern Tyrrhenian Sea (Fig. 3.6a), while some ensemble members simulate wind gusts exceeding 32.7 ms^{-1} (Fig. 3.6b). The deterministic run initialized at 00:00 UTC on 18 August 2024 (26 hours ahead) simulates a storm crossing the southern Tyrrhenian Sea. The simulated gust front impacts the Porticello area at 04:00 UTC (two hours later than the actual time; Fig. 3.6c), with wind gusts up to 25 ms^{-1} . The ICON-2I run initialized at 00:00 UTC on 19 August

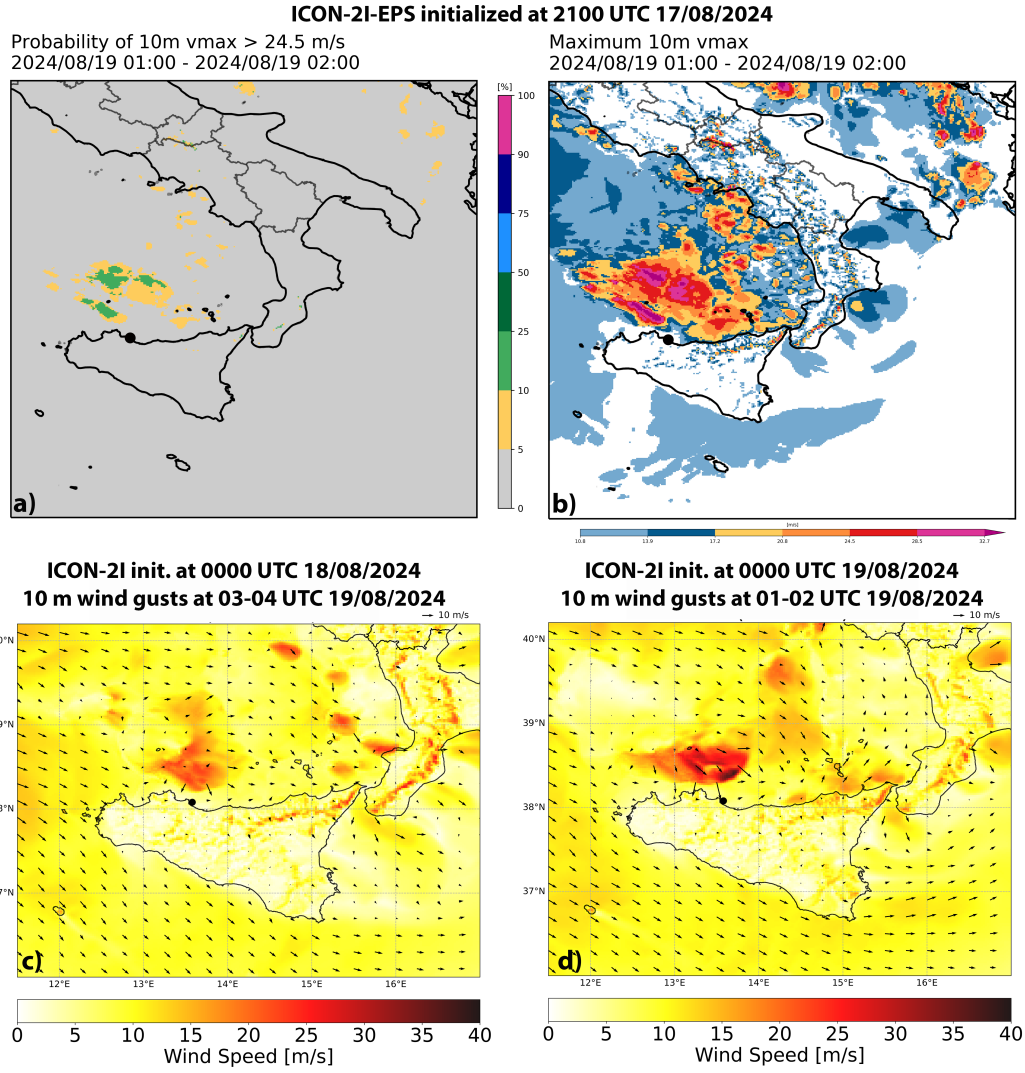


Figure 3.6: a) Probability of maximum wind gusts at 10 m stronger than 24.5 m s^{-1} between 01:00 and 02:00 UTC on 19 August 2024 computed by **ICON-2I-EPS** initialized at 21:00 UTC on 17 August 2024; b) same as (a) but for maximum wind gusts at 10 m; c) wind speed at 10 m at 04:00 UTC (arrows) and max wind gusts (colour shading) at 10 m between 03:00 and 04:00 UTC on 19 August 2024 simulated by **ICON-2I** initialized at 00:00 UTC on 18 August 2024; d) wind speed at 10 m at 02:00 UTC (arrows) and max wind gusts (colour shading) between 01:00 and 02:00 UTC on 19 August 2024 simulated by **ICON-2I** initialized at 00:00 UTC on 19 August 2024. The location of Porticello is highlighted with a black dot.

2024 (Fig. 3.6d) simulates the gust front in the same location as the run initialized 24 hours before, but at the correct time and with greater intensity (wind gusts up to 40 ms^{-1} offshore). Although the latter run was too close to the event to be useful for issuing warnings, it provides a plausible reproduction of the storm that caused the sinking, offering a benchmark to compare with the previous simulations.

In conclusion, the run initialized at 00:00 UTC on 18 August 2024 and the [ICON-2I-EPS](#) discussed earlier identify the possibility of severe wind gusts in the southern Tyrrhenian Sea between 01:00 UTC and 04:00 UTC on 19 August 2024. It is relevant that convective windstorms were simulated both in the deterministic run and in some members of the probabilistic simulations already 29 hours ahead. Thus, although the locations and timing were not predicted exactly, the simulated convective wind gusts should be considered as an indication of the possible occurrence of a severe convective event responsible for a significant impact in the area.

3.1.5 Summary

A detailed analysis of the available observations suggests that the sinking of the *Bayesian* yacht was likely caused by a convective windstorm. The reasons for this conclusion are summarized as follows:

1. Satellite images showed an [MCS](#) over the southern Tyrrhenian Sea between 23:00 UTC on 18 August 2024, and 03:00 UTC on 19 August 2024. Convective cells had a duration shorter than 1 hour and the storm motion that was not in agreement with that expected for a right-mover supercell, suggesting that supercells were not present during this event.
2. A few videos taken along the coast captured very intense north-westerly wind gusts, with no evidence of rotating winds or waterspouts. Those videos were taken upwind the *Bayesian* yacht. Before sinking, the yacht drifted southeastward pushed by the northwesterly wind.
3. Data from weather stations revealed classic downburst features, such as an increase in pressure and a drop in potential temperature corresponding to the strongest gusts. No signs of mesocyclones (e.g., sudden pressure drop) were detected.

Whilst these observations suggest that the storm responsible may not have been a supercell, in the absence of radar observations, we cannot be entirely confident in this conclusion. While we are not aware of any tornado damage along the coast, and all the observations analyzed in this study do not suggest the presence of a tornado, its occurrence cannot be completely ruled out. Even if the presence of a supercell is dismissed, a non-mesocyclonic tornado could have developed. Thus, the possibility of a hidden, rain-wrapped tornado that formed just offshore and struck the *Bayesian* yacht cannot be entirely ruled out. In any case, [AIS](#) data show that the yacht was dragged by northwesterly winds for several minutes before sinking, suggesting that straight-line winds were the most plausible weather-related cause of the incident.

The presence of an upper-level cyclone over central Italy, an approaching cold front, high values of moisture, potential instability, and deep layer shear contributed to the presence of an environment favourable to the development of severe storms in the southern Tyrrhenian Sea. While such highly localized events cannot be predicted with precise timing and location even a few hours in advance, both high-resolution deterministic and ensemble simulations run operationally by [ARPAE-SIMC](#) with the [ICON](#) model predicted strong wind gusts in the southern Tyrrhenian Sea as early as 29 hours before the

event. Consistently, the Italian Civil Protection issued a thunderstorm warning for 19 August 2024, over the northern coasts of Sicily. Thus, the possible occurrence of severe storms in that area was expected, even if with the limitations mentioned above.

3.2 Bora windstorm in Trieste

3.2.1 Introduction

After having analyzed the summer-time convective windstorm in Sicily, in this section a winter Bora downslope windstorm in Trieste is investigated. Trieste is located in north-eastern Italy near the border with Slovenia. The city is built between the Karst plateau in the eastern side, whose hills reach an elevation of about 400 m, and the Gulf of Trieste in the western side (Fig. 3.7b). Both topographic features influence the climate of Trieste, which is categorised as humid subtropical (Köppen: Cfa) with warm summers and cool winters.

The city is frequently affected by the Bora wind: a cold, usually very dry, and gusty wind blowing from the northeast (Orli et al., 1994), which usually occurs during the cool season (November to March). The speed of the Bora wind varies along the shore as it is affected by the orography of the mountain chain along the eastern Adriatic coast (Beg Paklar et al., 2001). Bora winds are a type of intense downslope wind, occurring downwind the Dinaric Alps (Grisogono and Belui, 2009; Klemp and Durran, 1987). Extensive research has been conducted on the Bora winds along the Croatian coast, or on the effect of Bora on the Adriatic Sea circulation (Crise et al., 2006). The generation and development of this wind in the Trieste region has been only marginally considered in the literature. Despite Bora being a mesoscale phenomenon that results from forces and drivers at the regional scales, it is also very sensitive and influenced by the local topography of the terrain (Grisogono and Belui, 2009). Hence, it is of interest to investigate how the wind from the inland interacts with the specific topography surrounding the city of Trieste.

One of the most extreme Bora events in recent times occurred in February 2012, when wind gusts reached 47.2 m/s in the city of Trieste, while the air temperature was as low as -5°C . This exceptionally strong event lasted about 12 days, the longest since 1979 (Raichich et al., 2013; Davolio et al., 2015). This event was selected to study how the local topography influence the generation of the Bora winds in Trieste by means of high-resolution mesoscale simulations. The simulations were carried out in the framework of the **RISKADAPT** project. In particular, Pilot target 3 of the project is focused on the Cattinara hospital, located above a hill in Trieste (Fig. 3.7b). The project aims to assess the force pressure exerted by Bora winds on the hospital's windows, in order to improve building regulations. Thus, the mesoscale numerical simulations were also used to initialize **CFD** simulations. Their results are not discussed in this thesis because they fall beyond its scope. In section 3.2.2 the setting of the mesoscale numerical simulations run is presented and in section 3.2.3 they are assessed with surface observations. In the same section the synoptic and mesoscale conditions that lead to the intense Bora wind event are analyzed, as well as the effect of the local topography and the choice of the input profiles for the **CFD** simulations.

3.2.2 Numerical simulations

Version 4.5.1 of the **Weather Research and Forecasting** (WRF) model (Skamarock et al., 2019) is used to run the simulations. Those are initialised at 00:00 UTC on 10 February 2012, with the 38 vertical levels of ERA5 (Hersbach et al., 2020). The WRF simulations

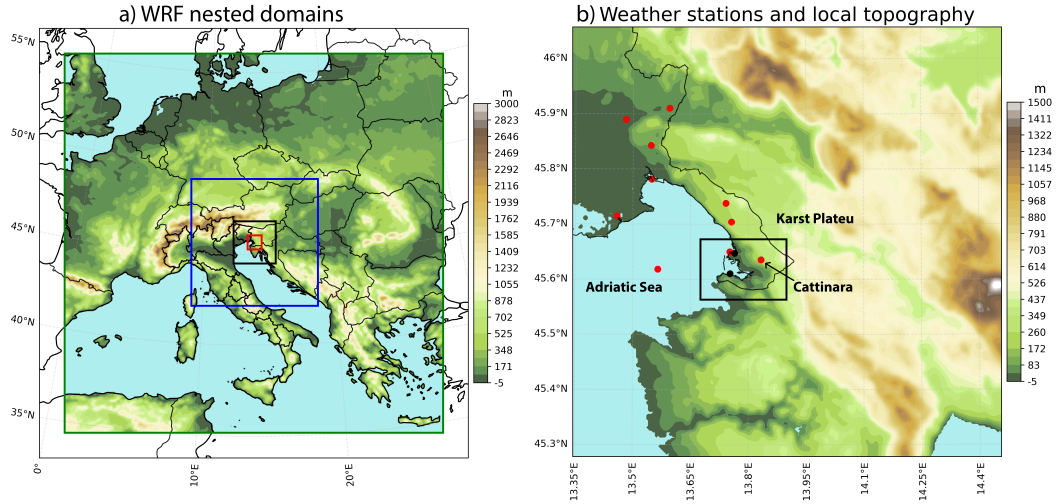


Figure 3.7: a) Nested domains: green (9 km grid-spacing), blue (3 km grid-spacing), black (1 km grid-spacing), red (333 m grid-spacing); (b) Orography in the innermost domain d04 and geographical localisation of weather stations by ARPA FVG (red dots, black dots are weather stations that recorded only temperature and relative humidity).

Name	PBL scheme	Microphysics scheme	First level	model	Radiation scheme	Land surface model
<i>Yon_Tom</i>	YSU	Thompson	5 m AGL		RRTMG	Noah-MP
<i>MYJ_Tom</i>	MYJ	Thompson	5 m AGL		RRTMG	Noah-MP
<i>MYJ_Tom_vl</i>	MYJ	Thompson	10 m AGL		RRTMG	Noah-MP
<i>acm2_px</i>	ACM2	Thompson	5 m AGL		RRTMG	Pleim-Xiu
<i>MYJ_Dou6cl</i>	MYJ	WRF Double-momentum 6-class	5 m AGL		RRTMG	Noah-MP
<i>MYJ_Tom_rad</i>	MYJ	Thompson	5 m AGL		RRTM-Dudhia	Noah
<i>Yon_Tom_noah</i>	YSU	Thompson	5 m AGL		RRTMG	Noah
<i>BouLac_Tom</i>	BouLac	Thompson	5 m AGL		RRTMG	Noah-MP
<i>Yon_Tom_rad</i>	YSU	Thompson	5 m AGL		RRTM-Dudhia	Noah

Table 3.2: Configurations of the WRF model tested, including details on the different sub-models and schemes adopted.

are performed with 64 vertical levels, the first one of which is at about 5 m above ground level (AGL). The sensitivity to the choice of the first model level is assessed by comparing against a simulation with a higher vertical level of 10 m AGL (simulation named *MYJ_Tom_vl*, Table 3.2).

Four one-way nested WRF simulations are carried out to dynamically downscale ERA5 down to a sub-kilometre resolution that will allow to better capture small-scale features of the Bora wind and initialise the CFD simulations. These four simulations have 9 km, 3 km, 1 km and 333 m of grid spacing and 300×300 grid points each (Fig. 3.7a). The simulation domains are shifted slightly eastward with respect to the Cattinara hospital because Bora is a westward wind, thus more space is given east of the hospital to allow WRF to properly reproduce the wind dynamics (Fig. 3.7b).

Choice of physics parametrisation

The initial physical parametrisation suite (*Yon_Tom*, Table 3.2) is selected based on previous numerical experiments conducted on similar events (e.g., Park et al. 2022; Cao and Fovell 2016). The PBL is parametrised with the [Yonsei University](#) scheme (YSU, Hong et al. 2006); the Thompson scheme is used for microphysics processes (Thompson et al., 2008); the [Rapid Radiative Transfer Model for Global climate models](#) (RRTMG) scheme for long-wave and short-wave radiation (Iacono et al., 2008); Noah-MP as a land surface model (Niu et al., 2011; Yang et al., 2011b) and the revised MM5 scheme (Jiménez et al., 2012a) is used for surface physics.

To assess the dependence of the results on the choice of the physical suite, several sensitivity tests are performed (Table 3.2). Downslope windstorms are strongly influenced by surface exchanges between the surface of the Earth and the atmosphere (Cao and Fovell, 2018), thus three experiments are performed to assess the sensitivity of the model to the PBL scheme: the [Mellor Yamada Janjic](#) (MYJ, Mellor and Yamada 1982; Janjic 1994) PBL scheme is implemented in the *MYJ_Tom* simulation; the [Asymmetric Convective Model](#) (ACM2, Pleim 2007) PBL scheme, which requests the Pleim-Xiu (Pleim, 2006) land surface model, is used in the simulation called *acm2_px*; and the [BouLac](#) scheme (Bougeault and Lacarrere, 1989a) is tested in the *BouLac_Tom* run. Since during the event there was some precipitation on the upwind side of the Karst plateau, a different microphysics scheme, namely the [WRF Double-momentum 6-class](#) (Hong and Lim, 2006), is tested in the numerical experiment *MYJ_Dou6cl*. A different land surface model, Noah (Mukul Tewari et al., 2004), is used in the experiment *Yon_Tom_noah*. Finally, also a different radiation scheme, namely [RRTM](#) (Mlawer et al., 1997) for long-wave radiation and [Dudhia](#) (Dudhia, 1989) for short-wave radiation are tested with either the MYJ or YSU PBL schemes (simulations *MYJ_Tom_rad* and *Yon_Tom_rad*, Table 3.2).

Model validation

The results of the innermost domain *d04* with 333 m of grid spacing are compared against 11 ground weather stations of the Environmental Agency of the Friuli-Venezia Giulia region ([ARPA FVG](#), dots in Fig. 3.7b). The model validation is performed only on the Italian side of the domain as our aim is to assess the model accuracy on the downwind side of Bora (westward of the Karst Plateau). At the locations of the weather stations hourly values of 2 m AGL temperature (T), 2 m AGL relative humidity (rh), 10 m AGL maximum wind gust (WG), 10 m AGL instantaneous wind speed (WS), and 10 m AGL wind direction (WD) are considered. Three statistical metrics, namely Pearson correlation coefficient, bias and [Root Mean Square Error](#) (RMSE), are computed in the time period from 00:00 UTC on 10 February 2012 to 00:00 UTC on 12 February 2012. Then, the values obtained are averaged across the 11 weather stations.

3.2.3 Results

Numerical simulation assessment

Table 3.3 presents the computed statistical metrics for 2 m air T and rh. Overall, the results show that the simulations reproduce both variables accurately, with only small discrepancies compared to the observations. Specifically, the simulations tend to slightly underestimate air temperature, with RMSE values around 1 °C, low biases (mostly less than 1 °C), and relatively high correlation coefficients, typically greater than 0.6. For relative humidity, the RMSE values are consistently below 15%, with average biases of

Name	r (T)	Bias (°C, T)	RMSE (°C, T)	r (rh)	Bias (% , rh)	RMSE (% , rh)
<i>Yon_Tom</i>	0.63	-0.9	1.2	0.51	7.7	9.6
<i>MYJ_Tom</i>	0.57	-0.6	1.2	0.47	12.0	13.3
<i>MYJ_Tom_vl</i>	0.60	-1.0	1.4	0.45	13.3	14.4
<i>acm2_px</i>	0.55	-0.5	1.0	0.56	9.4	10.9
<i>MYJ_Dou6cl</i>	0.60	-0.8	1.3	0.48	9.9	11.6
<i>MYJ_Tom_rad</i>	0.75	-0.6	1.1	0.46	11.6	13.1
<i>Yon_Tom_noah</i>	0.58	-1.1	1.4	0.43	11.7	12.7
<i>BouLac_Tom</i>	0.66	0.8	1.1	0.45	4.1	9.1
<i>Yon_Tom_rad</i>	0.69	-0.7	1.0	0.52	7.0	9.2

Table 3.3: Pearson correlation (r), bias and RMSE of 2m temperature (T) and 2m relative humidity (rh) of the simulations performed. The two best values for each column are bolded.

Name	r (WG)	Bias (ms^{-1} , WG)	RMSE (ms^{-1} , WG)
<i>Yon_Tom</i>	0.56	-6.9	8.2
<i>MYJ_Tom</i>	0.61	-5.9	7.6
<i>MYJ_Tom_vl</i>	0.60	-6.4	7.8
<i>acm2_px</i>	0.64	-7.3	8.2
<i>MYJ_Dou6cl</i>	0.58	-6.2	7.7
<i>MYJ_Tom_rad</i>	0.60	-6.3	8.0
<i>Yon_Tom_noah</i>	0.61	-3.9	6.3
<i>BouLac_Tom</i>	0.71	-6.8	7.6
<i>Yon_Tom_rad</i>	0.58	-6.6	7.9

Table 3.4: Pearson correlation (r), bias and RMSE of wind gusts (WGs) at 10 m AGL. The two best values for each column are bolded.

about 10% and correlation coefficients around 0.5. Among the tested cases, the *Yon_Tom_rad* simulation shows the best agreement with the observations.

Tables 3.4 and 3.5 presents the statistical metrics computed for the wind variables, i.e. WG, WS and WD at 10 m AGL. Results show that the simulations tend to underestimate wind gusts with an average bias of about $6 ms^{-1}$. This is in agreement with Cao and Fovell (2018) who observed a similar behaviour when studying a downslope windstorm in the United States; they found that WRF was unable to resolve the very fast and local turbulent features of the wind gusts, causing a systematic underestimation of their magnitude. The model also struggles to accurately predict wind direction with quite large RMSE values and low Pearson correlations. This could be attributed to the fact that in reality the wind direction changes locally due to the presence of small-scale obstacles, such as buildings or trees, that are not considered in the model. Looking closely though, the simulations exhibit a slightly better agreement with observations near the hospital with a lower RMSE value of about 10° . A weather station above the hill likely leads to better agreement with observations because it is more isolated from surface obstacles that can influence the wind direction locally. The instantaneous wind speed demonstrates lower RMSE values (about $4 ms^{-1}$) and biases (about $1 ms^{-1}$) than the other two variables, suggesting that the simulations are overall capable of reproducing the strong Bora wind speeds, which were often above $20 ms^{-1}$ near the surface with gusts exceeding $30 ms^{-1}$.

In Tables 3.4 and 3.5, the values of the two best performing model configurations for each metric are presented in bold. It is evident that no simulation outperformed

Name	r(WS)	Bias (ms^{-1} , WS)	RMSE (ms^{-1} , WS)	r(WD)	Bias ($^{\circ}$, WD)	RMSE ($^{\circ}$, WD)
<i>Yon_Tom</i>	0.48	0.5	3.5	0.10	8.4	38.1
<i>MYJ_Tom</i>	0.49	1.6	3.9	0.20	5.5	27.6
<i>MYJ_Tom_vl</i>	0.49	1.6	4.1	0.20	3.1	24.8
<i>acm2_px</i>	0.54	1.6	3.9	0.09	4.7	26.9
<i>MYJ_Dou6cl</i>	0.44	1.6	4.3	0.20	1.2	24.1
<i>MYJ_Tom_rad</i>	0.49	1.3	4.0	0.13	4.9	30.2
<i>Yon_Tom_noah</i>	0.47	3.0	4.8	0.15	7.3	34.4
<i>BouLac_Tom</i>	0.55	2.7	4.1	0.27	3.7	16.4
<i>Yon_Tom_rad</i>	0.48	0.6	3.6	0.11	10.6	41.3

Table 3.5: Pearson correlation (r), bias and **RMSE** of instantaneous **wind speed (WS)** and **wind direction (WD)** at 10 m **AGL**. The two best values for each column are bolded.

the others. Some simulations exhibit better agreement on wind direction (*MYJ_Dou6cl*), others on wind speed (*Yon_Tom_rad*) or on maximum wind gusts (*Yon_Tom_noah*). The simulation with the first model level at 10 m **AGL** (*MYJ_Tom_vl*) worsens the skill scores of the wind gusts and wind speed compared to the simulation with the first model level at 5 m **AGL** (*MYJ_Tom*), suggesting that a model level closer to the ground leads to better results. In this study, the *MYJ_Tom_rad* simulation was selected as reference run since it employs **MYJ**, the only **PBL** scheme that provides as output the parametrised **Turbulent Kinetic Energy (TKE)**, necessary to initialise the **CFD** simulations, and since is the best-performing physics suite compared to the other that employ **MYJ**.

Synoptic and mesoscale analysis

After having assessed the quality of the **WRF** simulations carried out, in this section outputs from numerical simulation *MYJ_Tom_rad* are used to understand the synoptic and mesoscale environment that lead to the intense Bora wind event.

Between 10–11 February 2012, an upper-level cut-off low, pushed by a ridge located over northern Europe, moved westward from eastern Europe toward the central Mediterranean (Fig. 3.8a,b). This cut-off low was associated with a very cold air mass, characterised by widespread temperatures lower than -15°C at 850 hPa (Fig. 3.8c,d). The Alpine and Dinaric ridges partially blocked the westward movement of this cold air mass at the low levels. This peculiar baroclinic situation caused the development of an extra-tropical cyclone over central Italy (Fig. 3.8b, Buzzi and Tibaldi 1978). The surface cyclone moved eastward, causing an abrupt intensification of the pressure gradient over the eastern Adriatic coasts (1 005 hPa over central Italy; 1 030 hPa over eastern Slovenia), triggering violent Bora winds in the Adriatic region (Fig. 3.8b).

The peak intensity of the Bora winds was reached during the night between 10 and 11 February. In Figure 3.9, the wind field at 00:00 UTC on 11 February at different heights is shown, as reproduced by the domain *d03*. At 10 m **AGL** the model simulates the strongest north-easterly winds over the Adriatic Sea (Fig. 3.9a), with a pattern of alternate banners of stronger and weaker winds, which is a known consequence of the topography of the Dinaric Alps (Grisogono and Belui, 2009). The relatively weak winds over the land downwind to the Dinaric Alps crests are related to the much stronger roughness and drag of the inland terrain with respect to the maritime surface.

At 950 hPa, the wind field has some notable differences with respect to the wind at 10 m **AGL** (Figure 3.9b). In general, the winds are much more intense over Trieste and

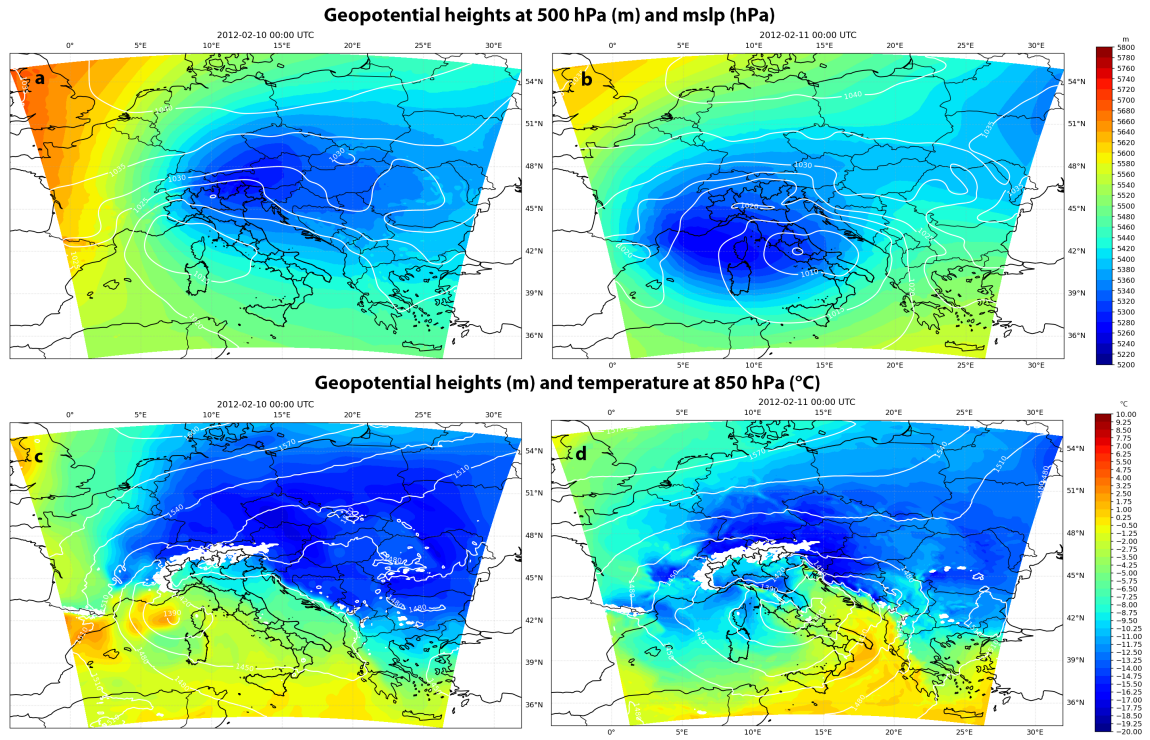


Figure 3.8: (a) Geopotential heights at 500 hPa (m, colour shading) and **mslp** (hPa, contours) at 00:00 UTC on 10 Feb 2012; (b) as in (a) but at 00:00 UTC on 11 Feb 2012; (c) Temperature (°C, colour shading) and geopotential heights (m, contours) at 850 hPa on 00:00 UTC on 10 Feb 2012; (d) as in (c) but at 00:00 UTC on 11 Feb 2012. Data from the outermost domain *d01* (Fig. 3.7) of simulation *MYJ_Tom_rad*.

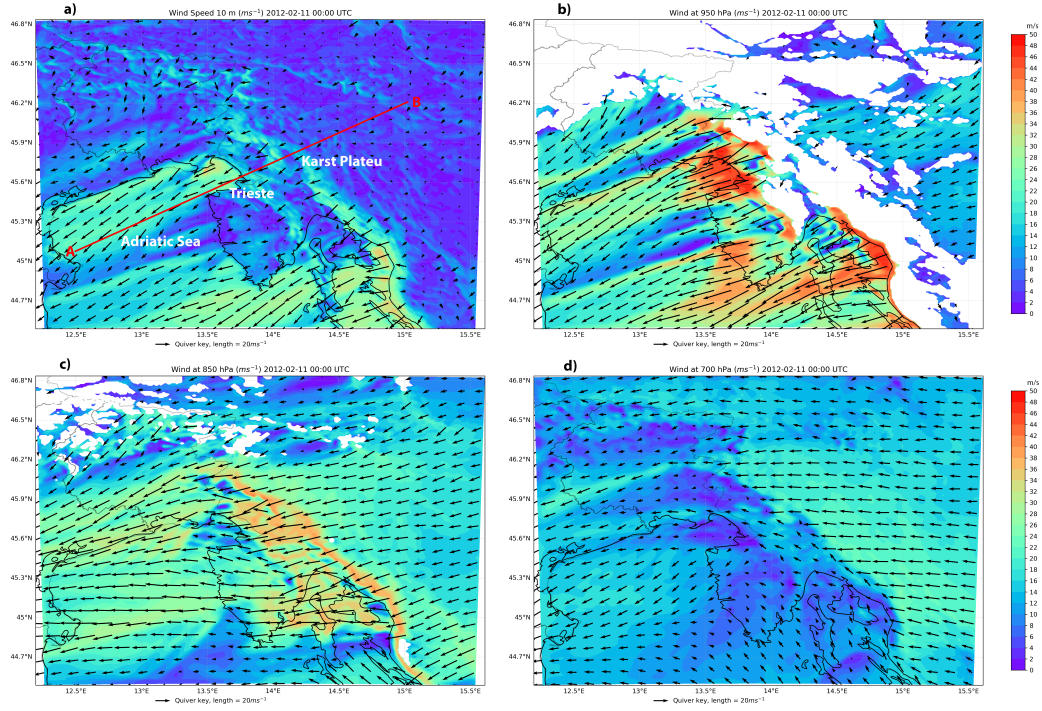


Figure 3.9: Wind speed and direction at (a) 10 m AGL, (b) 950 hPa, (c) 850 hPa and (d) 700 hPa at 00:00 UTC on 11 Feb 2012. The plotted domain is *d03* (Fig. 3.7) of simulation *MYJ_Tom_rad*. The line AB in (a) highlights the location of the cross section shown in Figure 3.10. The white shading in (b) and (c) covers the areas with elevation higher than pressure levels 950 hPa and 850 hPa, respectively.

the Krk Island, Croatia, with peaks of about 50 m s^{-1} . In addition, the wind peaks are shifted from the sea surface to the coastal area. Winds channels can be recognised also at this elevation, while a more uniform easterly flow is present east of the mountain ridge (the white colour in Fig. 3.9b denotes terrain higher than the pressure level of 950 hPa).

At 850 hPa, the wind speed is weaker and more uniform than at 950 hPa (Fig. 3.9c). In this case, the wind peaks are located over the mountain crests, and the channels generated by the topography are no longer evident, as the Dinaric Alps have an elevation lower than 1400 m (Fig. 3.7a), roughly correspondent to the 850 hPa layer (Fig. 3.8d). Finally, at 700 hPa, the Bora winds are not present anymore (Fig. 3.9d). An easterly flow is blowing eastward of the Dinaric Alps, while a convergence zone between southerly winds and weak easterly winds is present over the Istrian peninsula.

To better understand the generation of the Bora winds, a cross section along a Bora banner (red line in Fig. 3.9a) is presented in Fig. 3.10. The wind tangential to that cross section is quite uniform between the surface and 3 km AGL eastward of the highest Karst peaks (B corner in Fig. 3.10a), with wind speeds lower than 25 m s^{-1} . The Karst peaks generate orographic waves that rapidly break downstream, approximately above the city of Trieste—note the area with very weak to no wind at all between 2–3 km AGL over Trieste in Fig. 3.10a. Thus, the peak of flow intensity shifts from $\sim 1.5 \text{ km amsl}$ over the Karst plateau toward the surface over Trieste, becoming thinner and stronger. Above the Adriatic Sea, the layer with intense winds weakens and a hydraulic jump feature occurs. The layer with intense winds is shifted from about 300 m amsl over Trieste to about 1.5 km amsl on the left side of the cross section.

In Fig. 3.10b, the Brunt Vaisala frequency, N^2 , is plotted along the same cross section together with the tangential wind. N^2 is defined as:

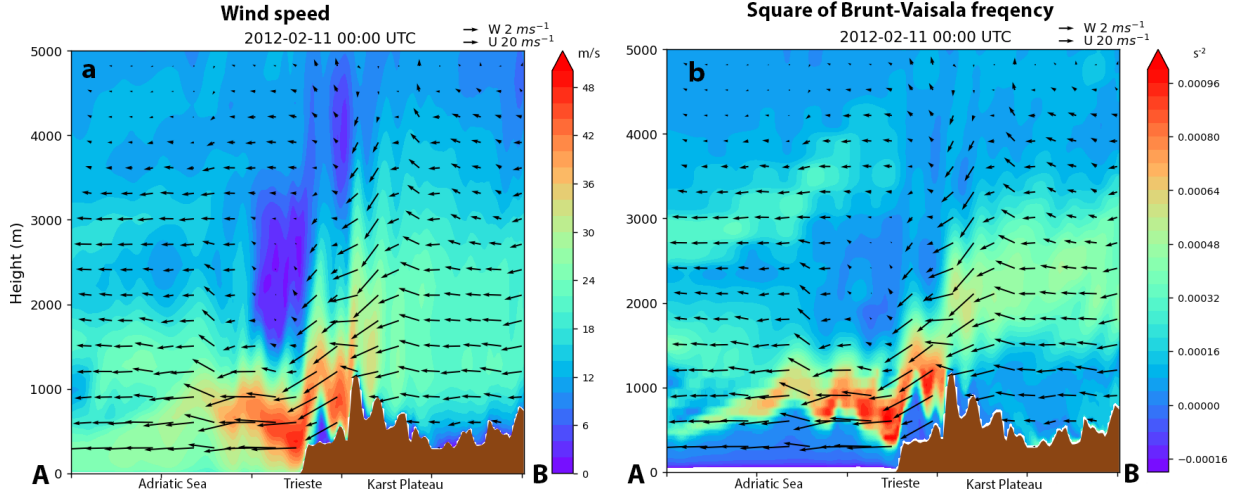


Figure 3.10: Cross section AB shown in Fig. 3.9a: (a) wind speed, (b) square of Brunt-Vaisala frequency at 00:00 UTC 11 Feb 2012.

$$N^2 = \frac{g}{\theta} \frac{\partial \theta}{\partial z}, \quad (3.1)$$

where g is the gravitational acceleration, θ is the potential temperature, and z is the vertical coordinate. The positive values of N^2 in Fig. 3.10b highlight the presence of a stable layer that behaves as an upper limit for the Bora wind. This stable layer is located at 2–3 km [amsl](#) before impacting the highest Karst peaks; then it intensifies as it descends at an elevation lower than 1 km [amsl](#). The stable layer was generated by the continental cold air mass flowing westward (Fig. 3.8c,d). As shown in previous studies (Durrán, 1986; Klemp and Durrán, 1987), the presence of a stable layer in the low-levels is the crucial feature to generate downslope windstorms. The flow trapped below the stability discontinuity behaves as a hydraulic jump when it crosses a mountain, with streamlines that are compressed close to the surface (as explained in section 2.1). This dynamic has also been confirmed for the Bora winds over Trieste by looking at the results of Fig. 3.10.

Summing up, the pressure gradient between the cyclone over central Italy and the anticyclone over eastern Europe triggers the cold low-level easterly flow. As it interacts with the Dinaric Alps and the Karsts plateau, the stable layer generated by the low-level cold air mass causes a hydraulic jump feature downstream of the mountains, responsible for a downslope windstorm along the Adriatic coasts. During the February 2012 event, the Bora winds were particularly strong due to the unusually cold low-level air mass, generating a particularly strong stable layer.

Choice of the CFD initialization profiles

To downscale the mesoscale [WRF](#) simulations at a [CFD](#) level, a realistic wind profile representative of the flow upwind of the inlet boundary of the [CFD](#) domain must be selected. This is achieved by sampling five grid points of the innermost domain *d04* on the flat Karst plateau (black dots in Fig. 3.11a). Then, the time evolution of the wind speed and direction in these five points is studied to select the peak intensity of the Bora, as well as a period of almost stationarity of the flow (to avoid transition phases). A two-hour window, between 03:00 and 05:00 UTC on 11 February 2012 is selected (Fig. 3.11b,c).

The input profile is then computed by averaging the five profiles in this time interval. Fig. 3.12 depicts the extracted profiles. A typical [low-level jet \(LLJ\)](#) profile can be

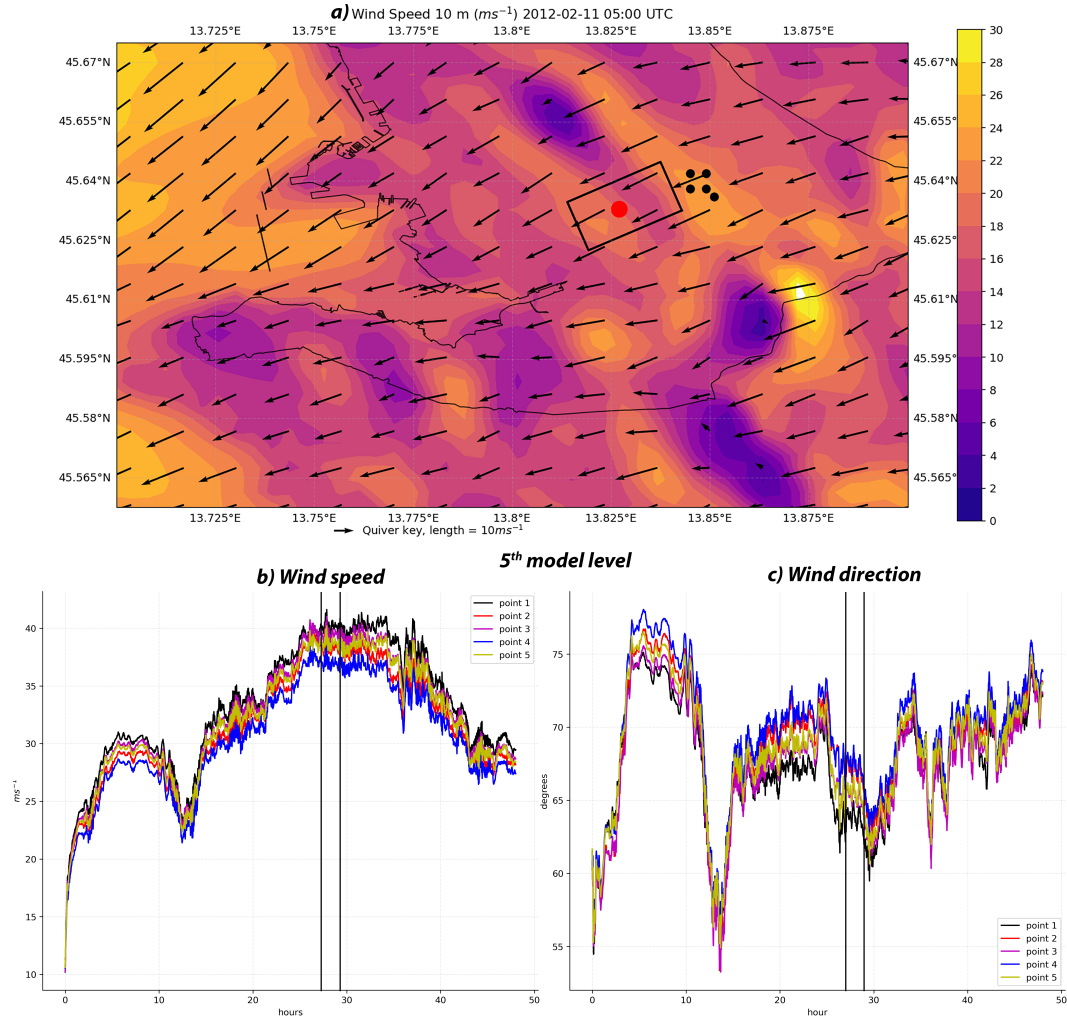


Figure 3.11: (a) wind in Trieste at 0500 UTC on 11 Feb 2012 (the plotted domain is shown with a black box in Fig. 3.7b). The black box in a) highlights the CFD domain, the red dot the location of the Cattinara hospital, the black box small dots show the location of the five grid-points used to initialize the CFD simulation. (b) timeplot of the wind speed throughout the simulation (0 is the initialization at 00:00 UTC on 10 Feb 2012) at the 5th model level in the five points shown in (a); the vertical lines highlight the period where the mean was computed to initialize the CFD. (c) as in (b) but with wind direction.

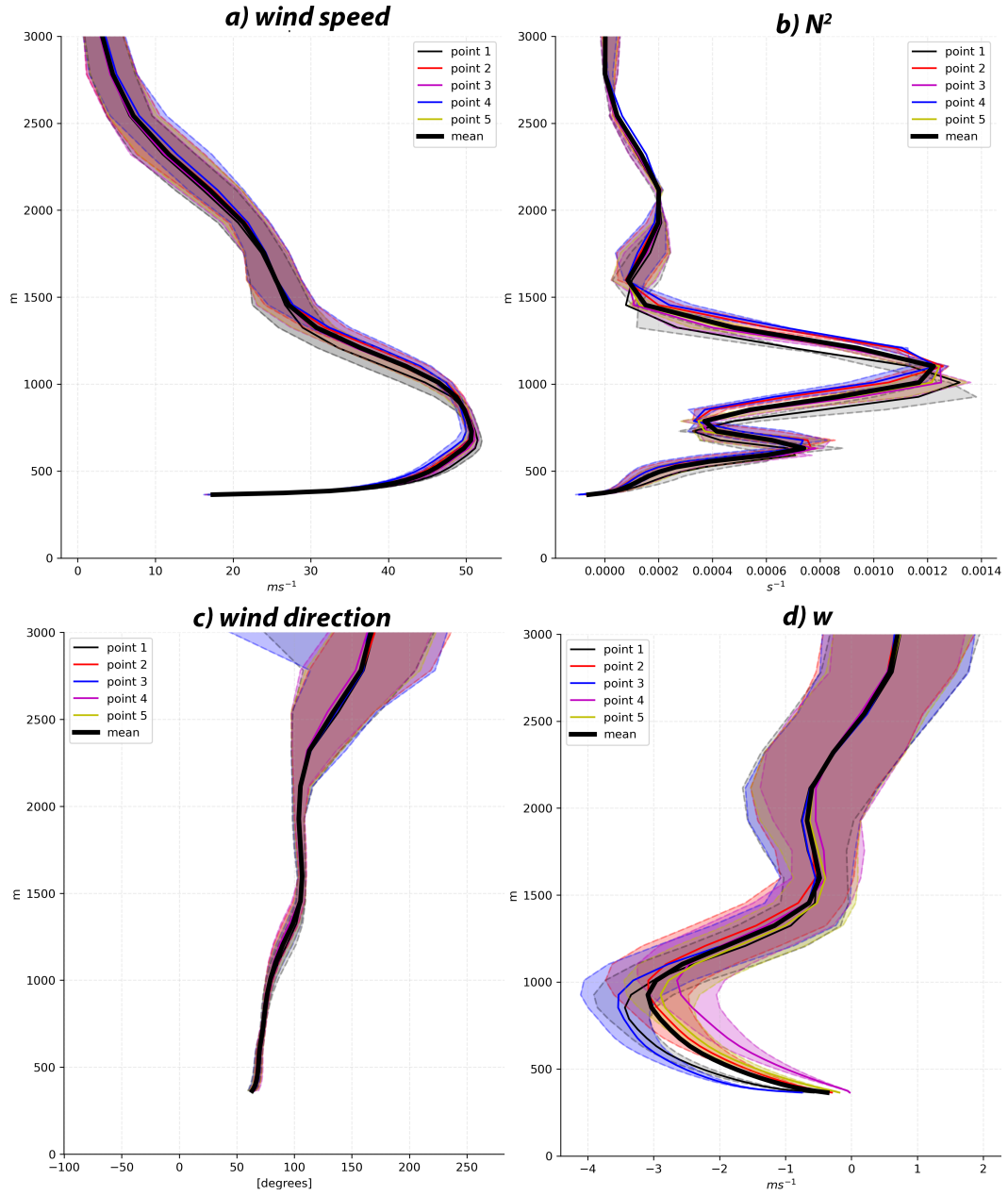


Figure 3.12: Mean (solid lines), 25th and 75th percentiles (dashed lines) between 03:00 and 05:00 UTC on 11 February 2012 in the five points shown in Fig. 3.11a of (a) wind speed; (b) square of Brunt Vaisala frequency (N^2); (c) wind direction; (d) vertical component of velocity (w).

observed for wind flow with maximum values of 50 m s^{-1} at 730 m [amsl](#) (Fig. 3.12a), and near zero values at 3 km [amsl](#). The jet is confined by the stable layer with a peak of N^2 at 1100 m [amsl](#) (Fig. 3.12b). A secondary maximum is located closer to the ground, at 630 m [amsl](#), just below the height corresponding to the maximum wind speed. The direction of the wind changes slightly southward from 65° on the surface to 100° at 2 km [amsl](#) (Fig. 3.12c). The jet is associated with a downward motion that peaks at 930 m [amsl](#). The intensity of the downward motion changes slightly between the five points considered between -2.5 and -3.5 m s^{-1} . The peak of the downward motion is 200 m below the maximum of N^2 and 200 m above the maximum of the jet. The stable layer forces the downward displacement of the streamlines just below it, and then the downward displacement causes the peak of wind speed 200 m below (Durran, 1986). In general, the profiles of the five points considered behave similarly, with small variances in the period considered. Thus, their average is a good proxy for a realistic wind profile to initialise the [CFD](#) simulations.

Some features of the computed inlet wind profile need further attention. The [CFD](#) simulations require a vertical profile of the [TKE](#) as an input variable. From a mesoscale point of view, the [TKE](#) is the sum between the resolved part (TKE_{RES}) and the parameterised by the turbulence scheme (TKE_{PAR}). TKE_{RES} is calculated directly from the model output, for each time step, and then averaged temporally over 20 minutes, as in Zonato et al. (2022). It reads:

$$\text{TKE}_{\text{RES}} = \overline{u'^2} + \overline{v'^2} + \overline{w'^2}, \quad (3.2)$$

where $\overline{u'^2}$, $\overline{v'^2}$ and $\overline{w'^2}$ are wind speed variances. If c is an instant model variable (u, v , or w), and C its time average, wind speed variances are computed as:

$$\overline{c'^2} = \overline{(c - C)^2}, \quad (3.3)$$

using the averaged wind of the last 20 minutes.

TKE_{PAR} is calculated by the [MYJ](#) scheme, using a 1.5-order closure (Mellor and Yamada, 1982):

$$\frac{D}{Dt} \text{TKE}_{\text{PAR}} = \frac{\partial}{\partial z} \left[S \ell \sqrt{2 \text{TKE}_{\text{PAR}}} \right] - \overline{w'u'} \frac{\partial U}{\partial z} - \overline{w'v'} \frac{\partial V}{\partial z} + \frac{g}{\theta} \overline{w'\theta'} - \varepsilon, \quad (3.4)$$

where ℓ is the diagnostic mixing length scale, S is a constant, $\overline{w'u'}$ and $\overline{w'v'}$ are the parametrised mechanical production terms, $\overline{w'\theta'}$ is the parametrised thermal production, U and V are the mean horizontal wind components of the flow, g is gravitational acceleration and θ is the potential temperature.

Another turbulent feature that must be considered is the turbulent dissipation coefficient ε computed as:

$$\varepsilon = \frac{\text{TKE}_{\text{PAR}}^{3/2}}{B_1 \ell}, \quad (3.5)$$

where $B_1 = 1.87799326209552761 \text{ (s}^3 \text{m}^{-2})$ is a closure constant.

Fig. 3.13 presents the turbulent features of the extracted profiles. Close to the ground there is a remarkable mechanical production of turbulence because of the interaction of the surface with the intense low-level jet (Fig. 3.13a,b). Consequently, [TKE](#) is very high near the surface with maximum values larger than $20 \text{ m}^2 \text{ s}^{-2}$ (Fig. 3.12c). At the height which corresponds to the low-level jet, near-zero [TKE](#) values are observed. Finally, the dissipation of [TKE](#), ε , logically appears higher close to the ground (Fig. 3.12d). These results are considered physically meaningful and representative of intense Bora winds

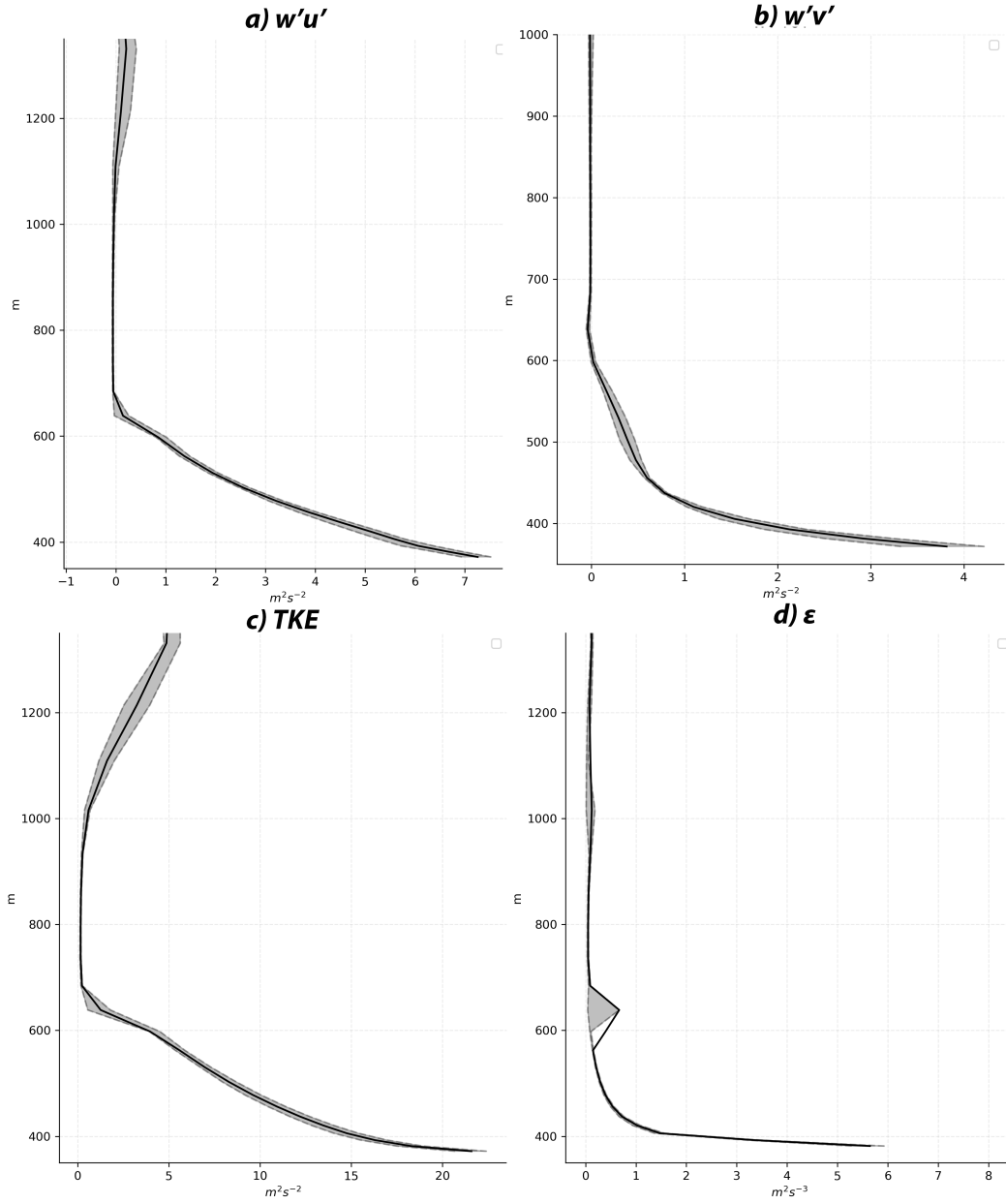


Figure 3.13: Mean (solid lines), 25th and 75th percentiles (dashed lines) between 03:00 and 05:00 UTC on 11 February 2012 in the five points shown in Fig. 3.11a of momentum fluxes (a) $\overline{w'u'}$, (b) $\overline{w'v'}$, TKE (c) and (d) ϵ .

upwind of the Cattinara hospital. Therefore, they were used to run CFD simulations, the results of which will be reported in a manuscript currently in preparation.

3.3 Discussion

This chapter has shown that numerical models are generally quite accurate in reproducing meteorological phenomena in complex terrain, and they represent a powerful tool for providing reliable forecasts of high-impact weather events. However, four main issues were identified:

1. Mesoscale numerical simulations struggle in reproducing wind gusts at the weather stations site, due to their turbulent features and to the presence of small-scale obstacles that cannot be resolved with this type of simulations.
2. Complex terrain induces mesoscale and microscale effects that require a CFD approach to be fully represented and understood. CFD simulations are not feasible for operational purposes due to their computational costs, suggesting that alternative approaches, such as machine learning, may be needed to predict small-scale topographically induced effects.
3. The lack of preparedness for high-impact weather events can lead to fatalities, as in the case of the *Bayesian* yacht. An effective early warning system that combines numerical simulations and observations could reduce the impacts of such phenomena.
4. The scarcity of observations represents a major limitation for the understanding and prediction of these events, as highlighted by both case studies. Funding should therefore be allocated to further strengthen the Italian observational network, public availability of the data is strongly encouraged, and field campaigns to study specific phenomena are needed.

The sinking of the yacht *Bayesian* highlighted the significant limitations in the physical understanding of meteorological phenomena in areas with complex topography and sparse observations. All available data (weather stations, videos, satellite, soundings, and reanalysis) were thoroughly analyzed to determine whether a waterspout or a convective windstorm struck the yacht. Although these data suggests that the sinking was more likely caused by a convective windstorm, the waterspout hypothesis cannot be ruled out, as the absence of radar data prevents a clear understanding of the storm mode and the possible presence of rotation at the yacht's location.

The ensemble of numerical simulations analyzed did predict the potential occurrence of intense convective wind gusts over the southern Tyrrhenian Sea during the night. However, the issue of local severe storms warnings remains a known challenge in Europe (Doswell III, 2015). This accident highlights the urgent need for a warning system dedicated to severe storm, as advocated by Miglietta and Rotunno (2016), covering both sea and inland areas. Specifically, the presence of an early warning system based on radar data could have avoided such a large number of fatalities. In fact, as the satellite analysis pointed out, the storm developed three hours before the accident and kept a coherent trajectory during its lifetime. Thus, there was enough time to warn people. Hence, it is encouraged the application of the European public warning system, based on cell broadcast technology and introduced by the EU Directive 2018/1972, also to local severe storms.

On the other side, efforts should be put on improving public awareness of severe storm risks. Institutions and media should increase their focus on people, working to enhance both the quality of severe storm warnings and the communication of associated risks. The 'Io non rischio' campaign, promoted by the Italian Civil Protection, has been successful in raising awareness of different kinds of natural risks such as floods, earthquakes, and fires; it is believed that time is mature to focus also on local severe storms.

Similarly to the previous work, the study of the Bora winds in Trieste showed that a numerical model was overall able to accurately simulate the Bora. The hydraulic jump over the sea and the breaking of orographic waves over the city of Trieste were successfully reproduced. Moreover, comparisons between the simulated near-surface variables and observations from ground weather stations showed good agreement. Nevertheless, the number of weather stations considered was small, and observations of the vertical profiles of Bora were lacking, which weakens the observational assessment. It was also noted that the numerical model struggled to reproduce wind direction at the station sites. This limitation is likely due to the models inability to resolve small-scale features such as buildings, trees, or fine topographic structures that can influence wind direction. The simulations also underestimated wind gusts, likely because [WRF](#) is unable to resolve the fast and turbulent features of intense winds close to the surface - even at a resolution of 333 m, which is already very high for mesoscale simulations. A [CFD](#) approach could overcome these limitations. Indeed, a manuscript is currently in preparation by the research group, reporting on a study in which a dynamical downscaling of [WRF](#) simulations with a [CFD](#) model were performed, yielding promising results. In addition, field campaigns are essential to address the scarcity of Bora observations, that are currently hindering the understanding of the phenomenon and the evaluation of numerical simulations. In fall 2025, the University of Bologna is planning a field campaign in Trieste to study the Bora within the project [RISKADAPT](#); the data collected could be used to assess numerical simulations and help address the limitations identified in this study.

Chapter 4

Effects of Urban Land Use on a Severe Convective Windstorm

In the previous chapter, the challenges of understanding and predicting high-impact weather events in complex topography were discussed. In this chapter, the same topic is extended to include the urban environment as an additional element of complex topography. The influence of urban land use on the atmosphere is generally weaker than that of mountains or the sea. However, as discussed in section 2.4, it is well established that urban land use modifies the atmosphere and generates new phenomena, such as the UHI. This chapter aims to determine whether these urban-induced atmospheric modifications are strong enough to influence the dynamics of a high-impact weather event. Specifically, we analyze the convective windstorm that struck Milan, a medium-large European city located in northwestern Italy, on 23 July 2023 at 02:00 UTC. Although the event caused no injuries or fatalities because it occurred at night, thousands of trees were uprooted across the city, severely compromising urban mobility the following day. The convective storm was generated by a QLCS, the same type of convective system that likely produced the windstorm leading to the sinking of the *Bayesian* yacht. Here the focus is on assessing the effect of the city of Milan on the storm, rather than its predictability or the possible presence of a vortex.

Few studies investigated the specific aspect of the interaction between urban areas and organized severe storms, focusing mainly on supercells and US cities. The effect of European urban areas on convective precipitations has been addressed only in a few cities (Lorenz et al., 2019; Forster et al., 2024), without focusing on severe storms. Studies on QLCS and European cities are missing. European and US cities differ in their morphological characteristics, with European cities typically being more compact and having lower buildings. Additionally, QLCS dynamics differ from those of supercells: in the first ones, the dynamics are mainly driven by cold pool propagation, while in the second ones, the mesocyclone maintains the supercell development (section 2.2.1). The July 2023 event is analyzed using the WRF model to understand the effect of Milan on the pre-storm environment and on the storm itself, by comparing simulations with and without urban land use. Therefore, this article examines how a different type of city might affect a different type of storm, with outcomes that cannot be predetermined. The study relies mainly on the results of numerical simulations. As highlighted in the previous chapter, state-of-the-art numerical models can accurately reproduce high-impact weather events, especially when using an ensemble of simulations. This provides a solid basis for trusting their outcomes in the present setting. Nevertheless, an evaluation of the simulated near-surface variables and radar reflectivity will be conducted to support the analysis.

Simulation name	PBL	Microphysics	Urban model	Urban Land Use Dataset	Name correspondent no-urban simulation
<i>BB_Bou</i>	BouLac	Thompson	BEP-BEM	WUDAPT	<i>Bou_nourb</i>
<i>Bulk_Bou</i>	BouLac	Thompson	Bulk	WUDAPT	<i>Bou_nourb</i>
<i>BB_MYJ</i>	MYJ	Thompson	BEP-BEM	WUDAPT	<i>MYJ_nourb</i>
<i>Bulk_MYJ</i>	MYJ	Thompson	Bulk	WUDAPT	<i>MYJ_nourb</i>
<i>Bou_NSSL</i>	BouLac	NSSL 2-mom	BEP-BEM	WUDAPT	<i>Bou_NSSL_nourb</i>
<i>GLOBUS</i>	BouLac	Thompson	BEP-BEM	GLOBUS	<i>Bou_nourb</i>

Table 4.1: Configurations of the WRF simulations performed and name of the equivalent non-urban simulation.

This chapter, which is adapted from De Martin et al. (2025d), is organized as follows: Section 4.1 describes the observations used, the numerical simulation setup, and the methods of analysis. Section 4.2 provides a detailed description of the case study and a verification of the model simulations. Section 4.3 presents the numerical results, and Section 4.4 discusses the main findings.

4.1 Method and data

4.1.1 Numerical Simulations Setup

Numerical simulations were performed using version 4.5.1 of the WRF model (Skamarock et al., 2019). Three one-way nested domains were defined (Fig. 4.1a), with grid-spacings of 9, 3, and 1 km and grid dimensions of 201x201, 253x252, and 301x301 points, respectively¹. The simulations utilized 64 vertical levels, with the first model level at approximately 5 m AGL, and 20 vertical levels within the first kilometer of the atmosphere to better resolve PBL exchange processes. The upper boundary of the simulation was set at 50 hPa.

Various initial conditions were tried to identify the one that best replicates the severe storm dynamics. Simulations initialized with the 38 atmospheric vertical levels of ERA5 reanalysis at 00:00 UTC, 06:00 UTC, 12:00 UTC, and 18:00 UTC on 24 July 2023 failed to reproduce the severe storm, regardless of the physical configuration of the model or the number of vertical levels in the simulation. Similarly, simulations initialized with the 137 atmospheric vertical levels of the IFS at 00:00 UTC, 06:00 UTC, and 12:00 UTC on 24 July 2023 did not successfully replicate the severe storm that impacted Milan. However, the simulation initialized at 18:00 UTC on 24 July 2023 with the IFS analysis, accurately reproduced the severe storm. Although the IFS analysis has a horizontal resolution of approximately 9 km in this area, suggesting that the outermost domain (*d01*) might be redundant due to its matching resolution with the initial condition, tests using only two nested domains (with grid-spacings of 3 km and 1 km) failed to reproduce the severe storm. The reason of that is beyond the aim of this dissertation.

Once a reasonable simulation of the severe storm was achieved, a sensitivity test on the model physics was conducted (as detailed in Table 4.1). Specifically, different model configurations were explored by altering the physical schemes that resolve the exchange processes between the urban surface and the atmosphere. The BEP (Martilli et al., 2002) and the BEM (Salamanca et al., 2010) models were toggled on and off during these sensitivity experiments. These models were utilized with the same parameterizations implemented by Zonato et al. (2020), under the assumption that the Local Climate Zone (LCZ)

¹An additional simulation with a grid spacing of 333 m, nested within the 1 km domain, did not improve skill scores and required computational resources beyond our capabilities.

of Milan share similar geometric and thermal characteristics with those of Bologna, a nearby Italian city. Urban **Land Use (LU)** data from **World Urban Database and Access Portal Tools (WUDAPT)** (Demuzere et al. 2022), with a 100 m grid-resolution, was employed taking the most widespread **LCZ** category as representative for each WRF cell (Zonato and Chen, 2021). Additionally, a simulation using the more advanced **GLobal Building heights for Urban Studies (GLOBUS)** Dataset (Kamath et al., 2024), which provides high-resolution (100 m) building data derived from a neural network trained on various sources, including satellite data and a digital elevation model, was conducted (simulation name *GLOBUS*). This dataset enables the use of area-averaged building heights, building height histograms, building surface-to-plan area fractions, and plan area fractions, in place of **LCZ** categorical values.

We also tested the sensitivity to the **PBL** scheme by switching between the **BouLac** scheme (Bougeault and Lacarrere, 1989b) and the **MYJ** scheme (Mellor and Yamada, 1982; Janjic, 1994). **BouLac** was selected due to its extensive testing with **BEP-BEM** (Ribeiro et al., 2021), while **MYJ** was chosen based on its use in previous studies of convective events in urban areas (Yang et al., 2021a; Lin et al., 2021). Thus, we run two simulations with **MYJ**: one with **BEP-BEM** (simulation name *BB_MYJ*) and the other without (simulation name *Bulk_MYJ*). The same was done for **BouLac** (simulations *BB_Bou* and *Bulk_Bou*, respectively). All these simulations employ the Thompson microphysics scheme (Thompson et al., 2008). Since we are studying a severe storm event, an additional simulation to test the sensitivity to the microphysics was performed running a double moment scheme (Mansell et al. 2010, simulation name *Bou_NSSL*).

All simulations employed the Kain-Fritsch cumulus scheme (Kain, 2004) in the outermost domain, with no cumulus schemes activated in the inner domains. Long-wave radiation was parameterized using the rapid radiative transfer model (Mlawer et al., 1997), and shortwave radiation was handled by the Dudhia scheme (Dudhia, 1989). Finally, the revised MM5 scheme (Jiménez et al., 2012b) was used for the surface layer, and Noah-MP (Yang et al., 2011b; Niu et al., 2011) was utilized as the land-surface model.

4.1.2 No-Urban Experiments

In addition to the simulations described in Table 4.1, three non-urban simulations were conducted. In these simulations, the urban **LCZs** (land categories 51 to 60 in Fig. 4.1b) were replaced with croplands (land category 12, Fig. 4.1b). These non-urban simulations were run with the same configurations as the urban ones. The non-urban simulations were limited to three, as the urban simulations employing **BouLac** and **BEP-BEM** switched off or on, and that using **GLOBUS**, share the same non-urban counterpart (simulation name *Bou_nourb*), as well the two **MYJ** simulations (simulation name of the no-urban counterpart *MYJ_nourb*). Then, we run also the non-urban counterpart of the *Bou_NSSL* simulation, called *Bou_NSSL_nourb*.

To disentangle the effects of the cities of Milan and Turin on the severe storm, two additional experiments were conducted. In these experiments, the urban land use was replaced with croplands, but only within the Milan urban area (within the black box in Fig. 4.1b, referred to as the *NoMi* experiment) or only within the Turin urban area (within the blue box in Fig. 4.1b, referred to as the *NoTo* experiment). Both experiments were performed using the same model configuration as the *BB_pbl8* experiment.

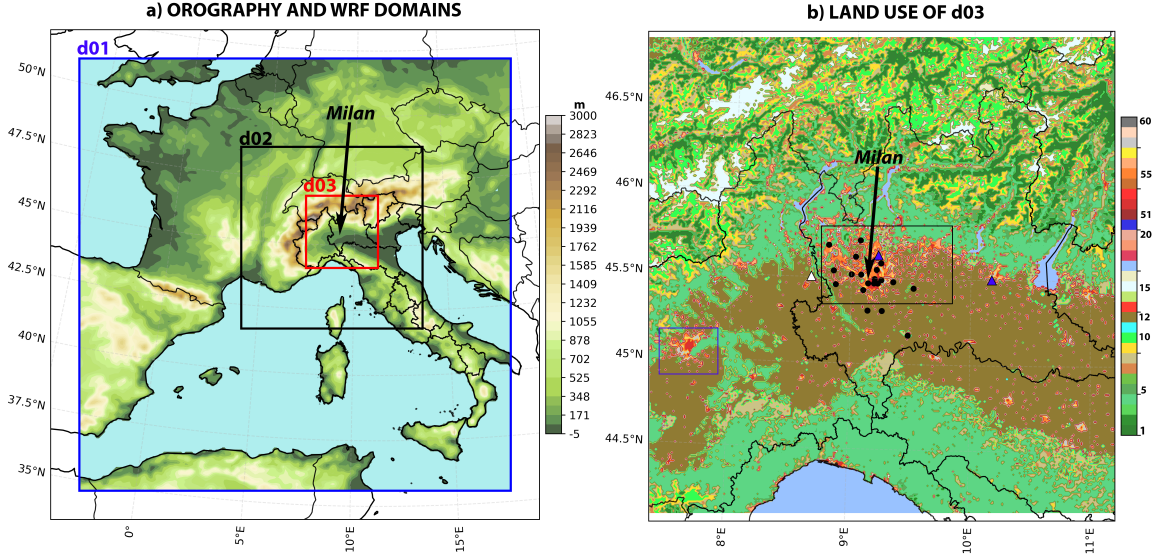


Figure 4.1: (a) Three nested domains used in the WRF simulations (colored boxes) and model orography (m, color shading). (b) LU Index of the innermost domain *d03* (red box in (a)): weather stations of ARPA Lombardia used for the model validation of surface variables (black dots), Novara Cameri sounding (white triangle), and Desio and Flero radars (blue triangles) are shown. The black and blue boxes in (b) highlight the domain where urban land use was replaced with croplands in the *NoTo* and *NoMi* experiments.

4.1.3 Observations

Surface weather station data from the Environmental Agencies of the Lombardy (ARPA Lombardia) and Piedmont (ARPA Piemonte) regions were utilized to analyze mesoscale surface variables and validate numerical simulations. A total of 69 weather stations (represented by dots in Fig. 4.5a) of ARPA Piemonte and ARPA Lombardia were used to assess cumulative rainfall, while 34 weather stations (dots in Fig. 4.5b) of ARPA Lombardia were used to evaluate maximum wind gusts. The model's performance in simulating surface variables within the urban area of Milan was also verified using 20 weather stations (black dots in Fig. 4.1b) of ARPA Lombardia.

The structure of the severe storm was examined using the vertical maximum intensity obtained from the polarimetric X-band radars located in Desio and Flero (locations shown with blue triangles in Fig. 4.1b), managed by ARPA Lombardia. Additionally, the atmospheric sounding launched in Novara Cameri (indicated by a white triangle in Fig. 4.1b) at 23:00 UTC on July 24, 2023, was employed to analyze the pre-storm vertical profile.

4.1.4 Physics ensemble

An a priori analysis showed that the different model configurations (outlined in Table 4.1) exhibited similar skill scores, with all configurations reasonably reproducing the severe storm development. Consequently, a physics ensemble (Evans et al., 2012; Yang et al., 2021c) was created to strengthen the analysis and reduce numerical noise. The ensemble mean was verified in the same manner as each individual ensemble member. Additionally, the mean difference between each urban and non-urban counterpart simulation was calculated.

To quantify the impact of the urban land use on the severe storm, three regions were

selected based on the location of the city and the storm's north-northeastward motion (Fig. 4.10): the *upstream* region (red box), the *city* region (black box), and the *downstream* region (blue box). These regions are all of equal size, and their width was chosen to encompass most of the severe storm in the direction perpendicular to the storm's motion. The *upstream* region primarily consists of croplands located upstream of Milan (where the storm inflow is likely still unaffected by the urban land use). The *city* region includes the city center and its suburbs, while the *downstream* region contains small towns and croplands downstream of the city. The *city* region is slightly shifted westward relative to the city center because the storm inflow approached from the east, causing the storm to ingest air above Milan slightly before impacting the city.

Within these regions, the average of the ensemble mean difference was calculated for several relevant variables describing the features of the severe storm: cumulative rainfall, maximum wind gusts at 10 m AGL, maximum updraft and downdraft speeds, maximum updraft helicity (UH, the integrated product of vertical velocity and vorticity between 2 and 8 km amsl, a useful proxy for mesocyclone detection in numerical simulations, Kain et al. 2008), and maximum column-integrated graupel. Since the storm's spatial extent varies across the three regions, the average was computed by filtering out values below 1 (with units depending on the variable) within the boxes. For UH, a threshold of $10 \text{ m}^2 \text{ s}^{-2}$ was applied.

4.2 Case study

4.2.1 Study area

Milan (45°28' N, 9°11' E) is a city in Northwestern Italy, located in the Lombardy region. The urban area, known as 'Città metropolitana di Milano', covers $1\,575 \text{ km}^2$ and has a population of 3 214 630 inhabitants (a density of $2,040 \text{ inhabitants/km}^2$), which increases to 5 million when including the surrounding smaller cities (ISTAT, 2023). The city center of Milan is characterized by LCZ 2 (compact mid-rise, with average building heights of about 28 m), while the suburbs are predominantly LCZ 5 (open mid-rise, Fig. 4.1b).

Milan sits at 120 m amsl in the flat Po Valley. The Alps, a high and steep mountain ridge, are located 40 km to the north and 80 km to the west, while the Apennines, another mountain ridge with lower peaks, are 50 km to the south. Milan has a humid temperate climate (Cfa, according to Köppen 1884), with an average July temperature of 23.6°C (Puche et al., 2023) and frequent thunderstorms (Cacciamani et al., 1995). The foothills north of Milan are a hotspot in Europe for severe storms, with frequent supercells and hailstorms (Feldmann et al., 2025; Giordani et al., 2024; Feldmann et al., 2023; Taszarek et al., 2020; Punge et al., 2017). This area is also affected by severe convective windstorms (Burlando et al., 2024). The UHI effect in Milan has intensified significantly over the last century due to extensive urbanization, reaching a mean intensity of 4°C between 2012 and 2020 (Frustaci et al., 2022).

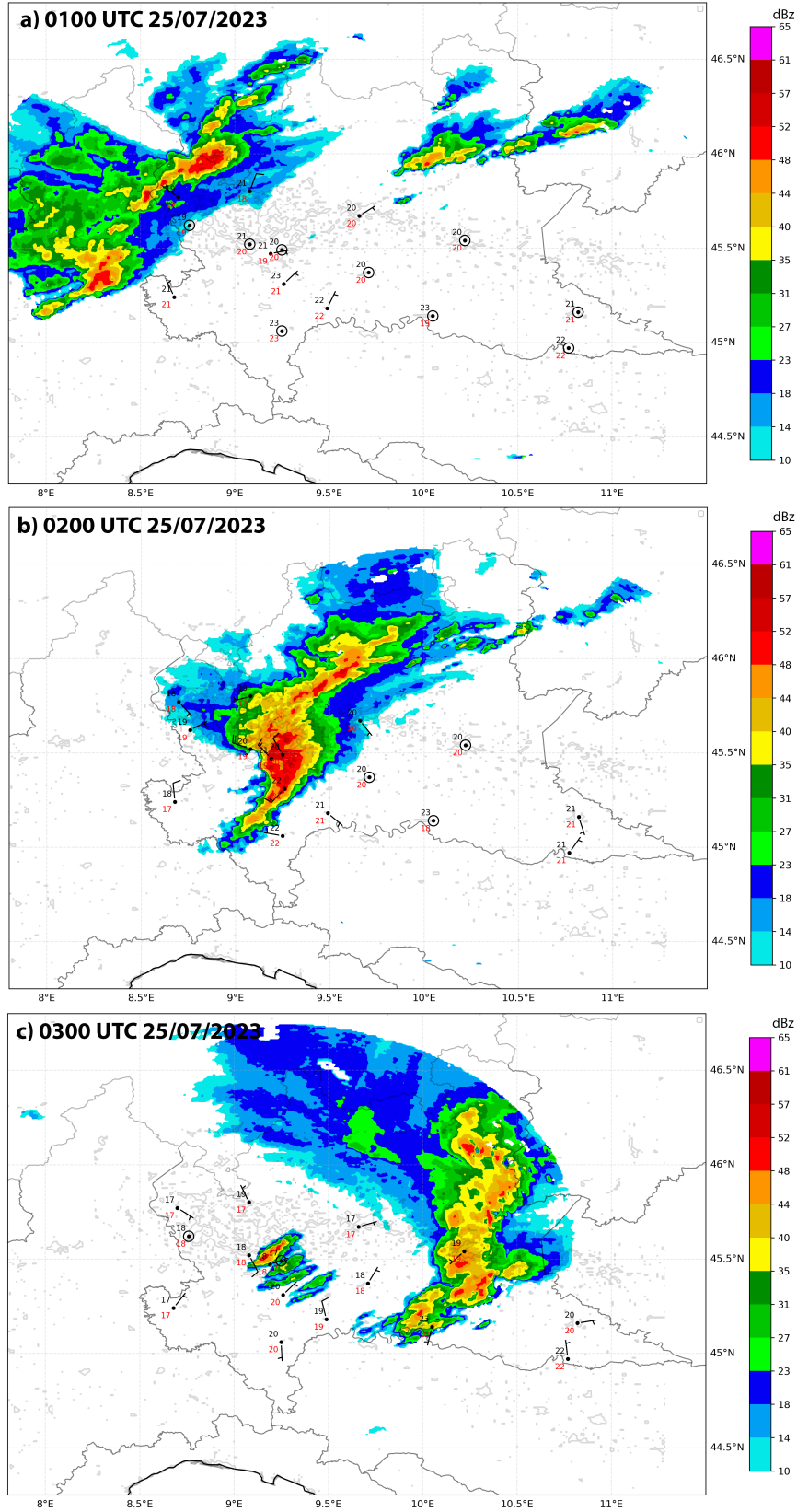


Figure 4.2: Radar reflectivity (dBZ) and surface weather station data: temperature (black numbers), dew point temperature (red numbers), wind (half barb 3 m s^{-1} , full barb 5 m s^{-1}) at (a) 01:00 UTC on 25 Jul. 2023, (b) 02:00 UTC 25 Jul. 2023, (c) 03:00 UTC 25 Jul. 2023. The plotted domain is slightly smaller than that of Fig. 4.1b (because the radar coverage is smaller). Urban areas (LU Index higher than 50) are highlighted with grey contours.

4.2.2 Windstorm description

July 2023 was a particularly active month for severe storms in Northern Italy (Panosetti and Tomassetti, 2024). Among several severe storm outbreaks, the one between 24 and 25 July 2023 was the most violent (De Martin et al., 2025a). Five severe storms crossed Northern Italy from the early afternoon of the 24th to the early hours of the 25th. The urban area of Milan was repeatedly affected by thunderstorms during this event (see the radar animation in the supplemental material Video S2). The most violent storm struck the city center at 02:00 UTC on the 25th. This storm was associated with intense wind gusts that uprooted thousands of trees in the parks and along avenues, and also brought down many light poles and electric cables (Milano, 2023). Some buildings were damaged, but fortunately, no injuries or fatalities were reported, likely because the storm occurred at night. However, urban mobility was severely compromised due to fallen trees blocking streets, and 14 schools along with many parks were closed in the following days.

Surface observations showed that just before the storm occurrence in the Milan urban area the UHI was absent, probably due to the outflows from previous storms (Fig. 4.2a). Temperatures were around 20°C, similar to those in the surrounding foothills, with slightly higher temperatures (about 23°C) south of Milan. The air mass near the surface was nearly saturated, with a weak northerly flow. The severe storm was triggered over the Western Alps (Fig. 4.2a) and moved quickly eastward. It was a QLCS that assumed a bow-shape (Weisman, 1993; Klimowski et al., 2004) when it impacted Milan (Fig. 4.2b). At that time, the strongest wind gusts were recorded in the city center (29.1 m s^{-1} at 02:20 UTC), accompanied by an intense downpour (45 mm in just 10 minutes). After striking the city, the storm's central part weakened, while new convective cells developed in the southern part of the storm (Fig. 4.2c).

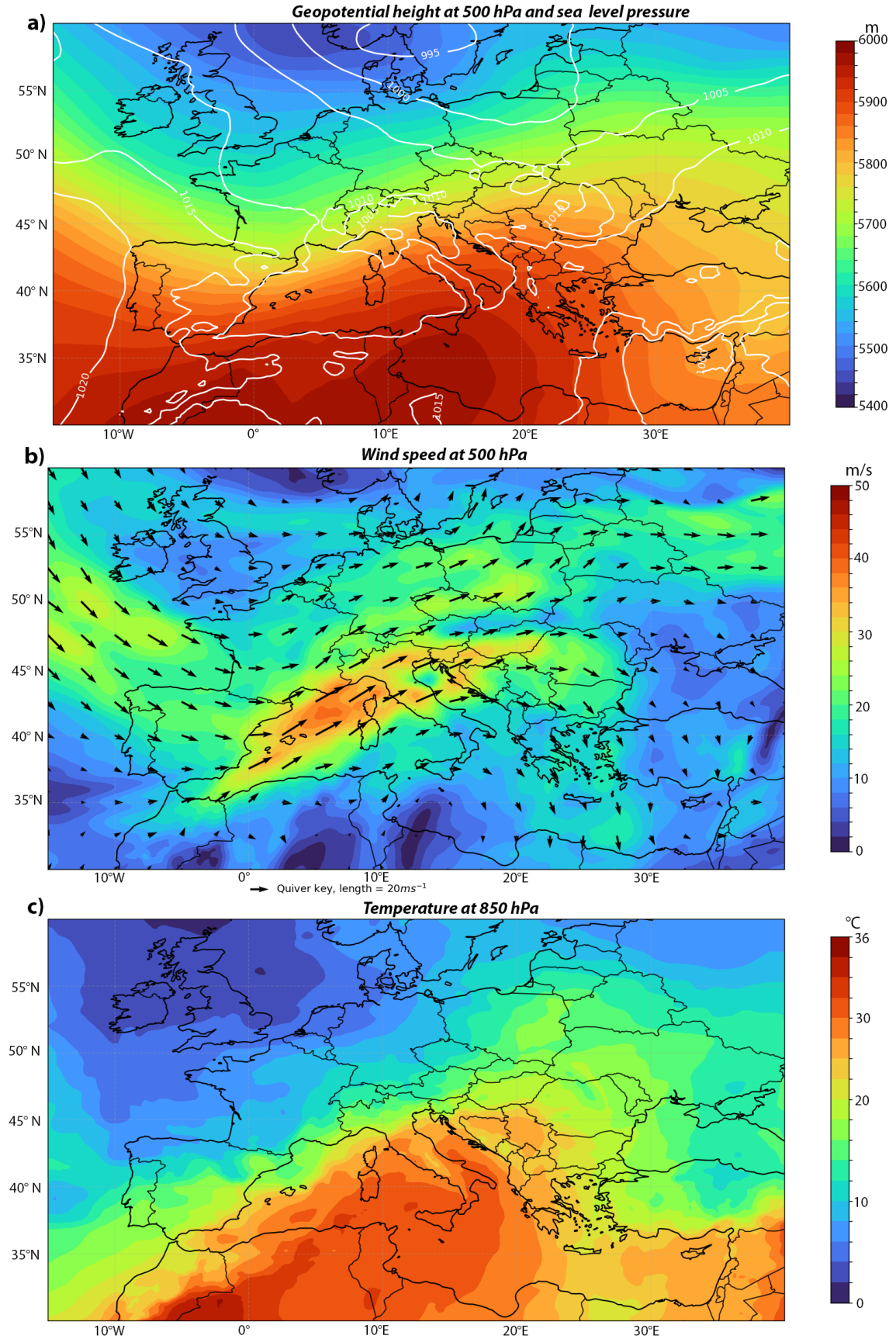


Figure 4.3: Data from ERA5 reanalysis at 00:00 UTC on 25 July 2023: (a) Geopotential heights at 500 hPa (m, colour shading) and **mslp** (hPa, contour lines); (b) wind speed (m s^{-1} , colour shading) and direction (arrows) at 500 hPa; (c) Temperature at 850 hPa (colour shading).

Index acronym	Index name	index value
h_{MUP}	elevation of the Most Unstable parcel	1229 m
MU CAPE	MU Convective Available Potential Energy	2308 Jkg^{-1}
$BS_{0-6\text{ km}}$	bulk shear between surface and 6 km amsl	30.0 ms^{-1}
$BS_{0-3\text{ km}}$	bulk Shear between surface and 3 km amsl	12.1 ms^{-1}
$SRH_{0-3\text{ km}}$	Storm Relative Helicity between surface and 3 km amsl	129 Jkg^{-1}
PW	Precipitable Water	40 mm

Table 4.2: Parameters derived from the sounding launched in Novara Cameri at 00:00 UTC on 25 July 2023, plotted in Fig. 4.4. Parameters are computed with ThundeR (Czernecki et al., 2022).

4.2.3 Synoptic and mesoscale analysis

A subtropical ridge was situated over the central Mediterranean (Fig. 4.3a), associated with a very hot air mass in the low-levels (temperature at 850 hPa up to 26°C at 40°N, Fig. 4.3c). In contrast, an upper-level trough was located over Western Europe, generating an intense thermal and geopotential gradient over Northern Italy (Fig. 4.3a,c). An intense southwesterly flow (up to 35 ms^{-1} at 500 hPa) was present between the Balearic Islands (south of Spain) and Northern Italy (Fig. 4.3b), with a closed pressure minimum at the surface in the Po Valley (Fig. 4.3a). This synoptic situation is well-known for favoring severe storms in Northern Italy (Manzato et al., 2020; Miglietta et al., 2016; Costa et al., 2001; Bechini et al., 2001).

A sounding was launched in Novara Cameri (WMO code 16064, white triangle in Fig. 4.1b) at 00:00 UTC on 25 July 2023, just north of the area that would be impacted two hours later by the most intense core of the severe storm. This sounding is therefore representative of the pre-storm environment. The vertical profile revealed an increase in equivalent potential temperature (Θ_e) from the surface to a maximum of 342 K at 1229 m (height of the **Most Unstable** (MU) parcel, Table 4.2). This condition is conducive to elevated convection, i.e., convection that develops above a near-stable surface layer (Corfidi et al., 2008). At higher elevations, Θ_e values decreased to a minimum of 320 K at 500 hPa, creating a 20 K difference between low-level and mid-level Θ_e , a condition favorable for downbursts (Atkins and Wakimoto, 1991).

Relevant sounding-derived parameters were calculated using ThundeR (Czernecki et al., 2022) and are reported in Table 4.2. The **MU CAPE**, BS_{0-6km} , and **PW** values (2308 Jkg^{-1} , 30.0 ms^{-1} , and 40 mm, respectively) were higher than the typical values observed during warm ($T_{2m} > 15^\circ C$) convective severe wind events ($> 32ms^{-1}$) in Europe. In fact, the 90th percentiles of the same parameters obtained by Taszarek et al. (2020) for the period 1979-2018 using ERA5 reanalysis data are approximately 1800 Jkg^{-1} , 27 ms^{-1} , and 42 mm, respectively. On the other hand, SRH_{0-3km} (129 Jkg^{-1}) and BS_{0-3km} (12.1 ms^{-1} , respectively) fell within the typical range for such events (medians of about 130 Jkg^{-1} and 14 ms^{-1} , respectively, Taszarek et al. 2020).

Therefore, the event under investigation was highly prone to the development of severe convective wind gusts and was characterized by anomalously high values of potential instability, moisture, and wind shear. Notably, the most unstable layer was located around 1 km **AGL**, a different condition with respect to other studies on severe storms in urban areas, where the most unstable parcel was found at the surface (e.g., Lin et al. 2021; Reames and Stensrud 2018).

4.2.4 Model verification

The simulations described in Section 4.1.1 accurately reproduce the development of the nocturnal severe storm, as evidenced by the comparison of observed radar reflectivity

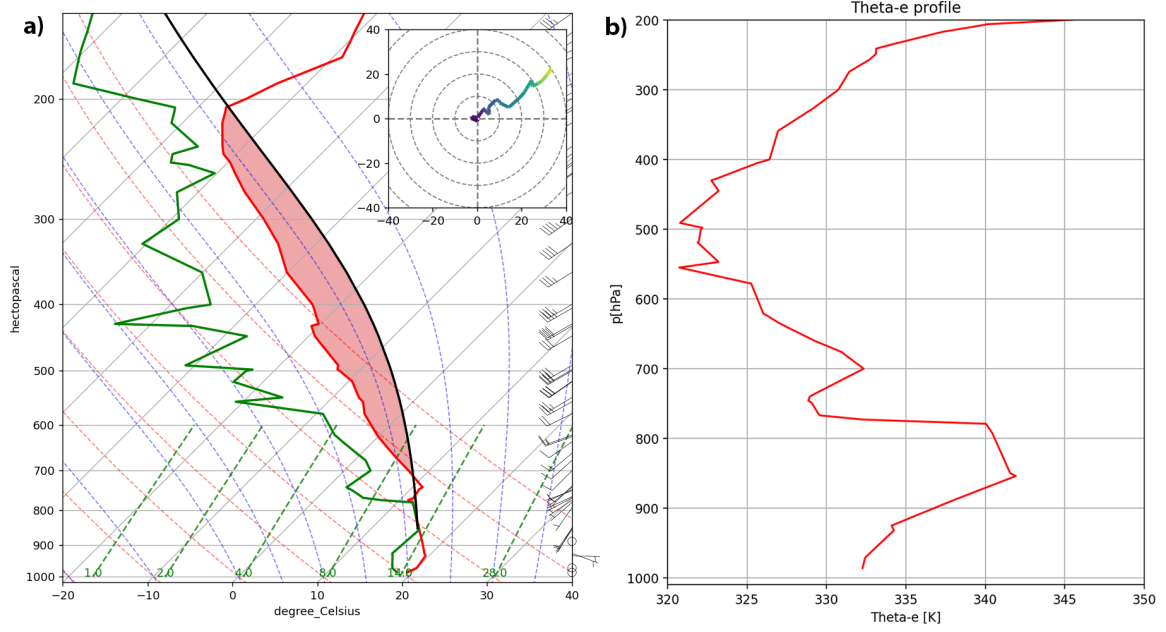


Figure 4.4: (a) Skew-t sounding launched in Novara Cameri (location shown with a cross in Fig. 4.1b) at 00:00 UTC on 25 Jul 2023: dew point temperature (green line), temperature (red line) and most unstable parcel profile (black line) are shown. On the right wind barbs are plotted (half barb 2.5 ms^{-1} , full barb 5 ms^{-1}). In the upper right part of the figure, the hodograph is shown: colors change according to height, violet (surface), green (5000 m AGL). (b) vertical profile of Θ_e of the same sounding.

Simulation name	$\text{bias}_{\text{rain}}$	$\text{RMSE}_{\text{rain}}$	$\text{bias}_{\text{gusts}}$	$\text{RMSE}_{\text{gusts}}$
<i>BB_Bou</i>	-4.4 mm	14.7 mm	-3.7 ms^{-1}	6.5 ms^{-1}
<i>Bulk_Bou</i>	-4.6 mm	13.1 mm	1.3 ms^{-1}	6.4 ms^{-1}
<i>BB_MYJ</i>	-7.8 mm	12.0 mm	-7.8 ms^{-1}	9.3 ms^{-1}
<i>Bulk_MYJ</i>	-6.9 mm	12.5 mm	-5.1 ms^{-1}	6.8 ms^{-1}
<i>Bou_NSSL</i>	-4.4 mm	12.2 mm	-3.6 ms^{-1}	7.4 ms^{-1}
<i>GLOBUS</i>	-5.3 mm	13.7 mm	-4.1 ms^{-1}	8.5 ms^{-1}
ensemble mean	-5.7 mm	12.3 mm	-5.5 ms^{-1}	9.0 ms^{-1}

Table 4.3: Bias and RMSE of cumulative rainfalls and maximum wind gusts at 10 m AGL in the 00:00-05:00 UTC 25 Jul 2023 period between simulations described in Table 4.1 (or their ensemble mean) and weather stations plotted in Fig. 4.5. Best values for each index are bolded.

(Fig. 4.2) with simulated radar reflectivity (e.g., the maximum vertical reflectivity of the *GLOBUS* run is shown in Video S3). A quantitative assessment of cumulative rainfall and maximum wind gusts between 00:00 and 05:00 UTC on July 25, 2023, reveals similar biases and **Root Mean Square Errors (RMSEs)** across the six simulations (Table 4.3). This short period was selected to focus the validation specifically on the severe storm. Precipitation biases are relatively small (around 5 mm), while **RMSEs** are larger (approximately 13 mm), because simulated rainfalls are slightly weaker in the Alps and more abundant east of Milan compared to observations. The simulations employing the **BouLac PBL** scheme (*BB_Bou* and *Bou_NSSL*) achieve the best bias (-4.4 mm), whereas the simulation based on the **MYJ PBL** scheme (*BB_MYJ*) achieves the best **RMSE** (12.0 mm).

Biases and **RMSEs** for wind gusts show more variability among the simulations, with generally better performance observed in the runs utilizing the **BouLac PBL** scheme (*Bulk_Bou* shows both the best bias, $+1.3 \text{ ms}^{-1}$, and **RMSE**, 6.4 ms^{-1}). Simulations based on the **BEP-BEM** model (*BB_Bou* and *BB_MYJ*) exhibit higher negative biases compared

Simulation name	r_T	bias _T	RMSE _T	r_{rh}	bias _{rh}	RMSE _{rh}
<i>BB_Bou</i>	0.78	1.50 °C	2.03 °C	0.43	-11.5 %	13.4 %
<i>Bulk_Bou</i>	0.76	1.91 °C	2.36 °C	0.30	-16.0 %	17.6 %
<i>BB_MYJ</i>	0.83	1.24 °C	1.76 °C	0.47	-7.6 %	10.4 %
<i>Bulk_MYJ</i>	0.84	1.33 °C	1.82 °C	0.34	-11.8 %	14.1 %
<i>Bou_NSSL</i>	0.80	1.65 °C	2.12 °C	0.20	-13.4 %	15.5 %
<i>GLOBUS</i>	0.80	1.26 °C	1.84 °C	0.44	-10.9 %	12.8 %

Table 4.4: Pearson correlations (r), bias and **RMSE** of 2m **T** and 2m **rh** between simulations described in Table 4.1 and surface weather stations in the surroundings of Milan (dots in Fig. 4.1b). The verification is performed in the 18:00 UTC 24 Jul, 2023 - 06:00 UTC 25 Jul, 2023 period. Best values for each index are bolded.

to analogous simulations with a bulk urban parameterization (*Bulk_Bou* and *Bulk_MYJ*). This discrepancy is likely because **BEP-BEM** realistically captures the shift of streamlines due to the presence of buildings (the inertial sublayer, Britter and Hanna 2003), but it does not simulate the roughness sublayer and the urban canopy layer, and therefore, it does not account for the turbulent wind gusts generated by buildings.

The verification of surface variables in the urban area of Milan was conducted every 30 minutes over the entire simulated period, from 18:00 UTC on July 24, 2023 to 06:00 UTC on July 25, 2023. The wind field was not verified, as it was generally weak during the event, except during the thunderstorm passage when gusts up to 30 m s^{-1} were recorded for a few minutes. Results show good accuracy for the 2-meter temperature, with a correlation coefficient of around 0.8, a bias of approximately 1.5°C , and an **RMSE** of about 2°C (Table 4.4). However, the performance is slightly worse for 2-meter relative humidity: there is a negative bias of about -10%, and the **RMSE** is relatively large at around 15% (Table 4.4). The negative bias in relative humidity is present in the initial conditions and propagates through the simulations. Despite this, the model satisfactorily reproduces the severe storm. Relative humidity remains almost constant during the period, except for a sharp variation associated with the passage of the storm, which occurs about 1 hour later in the simulation compared to observations. Consequently, there is a large covariance between simulations and observations during the storm's passage, resulting in a relatively low correlation coefficient (around 0.3). Among the various simulations, the *BB_MYJ* simulation achieves the best skill scores for both temperature and relative humidity.

Given that all six simulations exhibit similar skill scores for the considered variables, a physics ensemble was created (as anticipated in Section 4.1.4). In Figure 4.5, the ensemble mean of cumulative rainfall and maximum wind gusts between 00:00-05:00 UTC on 25 July 2023 is shown, along with observations from weather stations. The ensemble mean accurately captures the storm's trajectory and intensity. There is a slight negative bias in precipitation (-5.7 mm , Table 4.3), primarily due to the model's underestimation of moderate precipitation over the Alps. Similarly, wind gusts are also slightly underestimated, with a bias of -5.5 m s^{-1} and a relatively large **RMSE** of 9.0 m s^{-1} (Table 4.3), due to the underestimation of wind gusts in urban environments by the **BEP-BEM** simulations, as discussed earlier (note the gap in wind gusts over Milan in Fig. 4.5b).

In summary, both the ensemble members and the ensemble mean accurately reproduce the severe storm dynamics and the pre-storm environment. Consequently, these simulations are considered as a reference and will be used to quantify the effect of the urban land use in the next section.

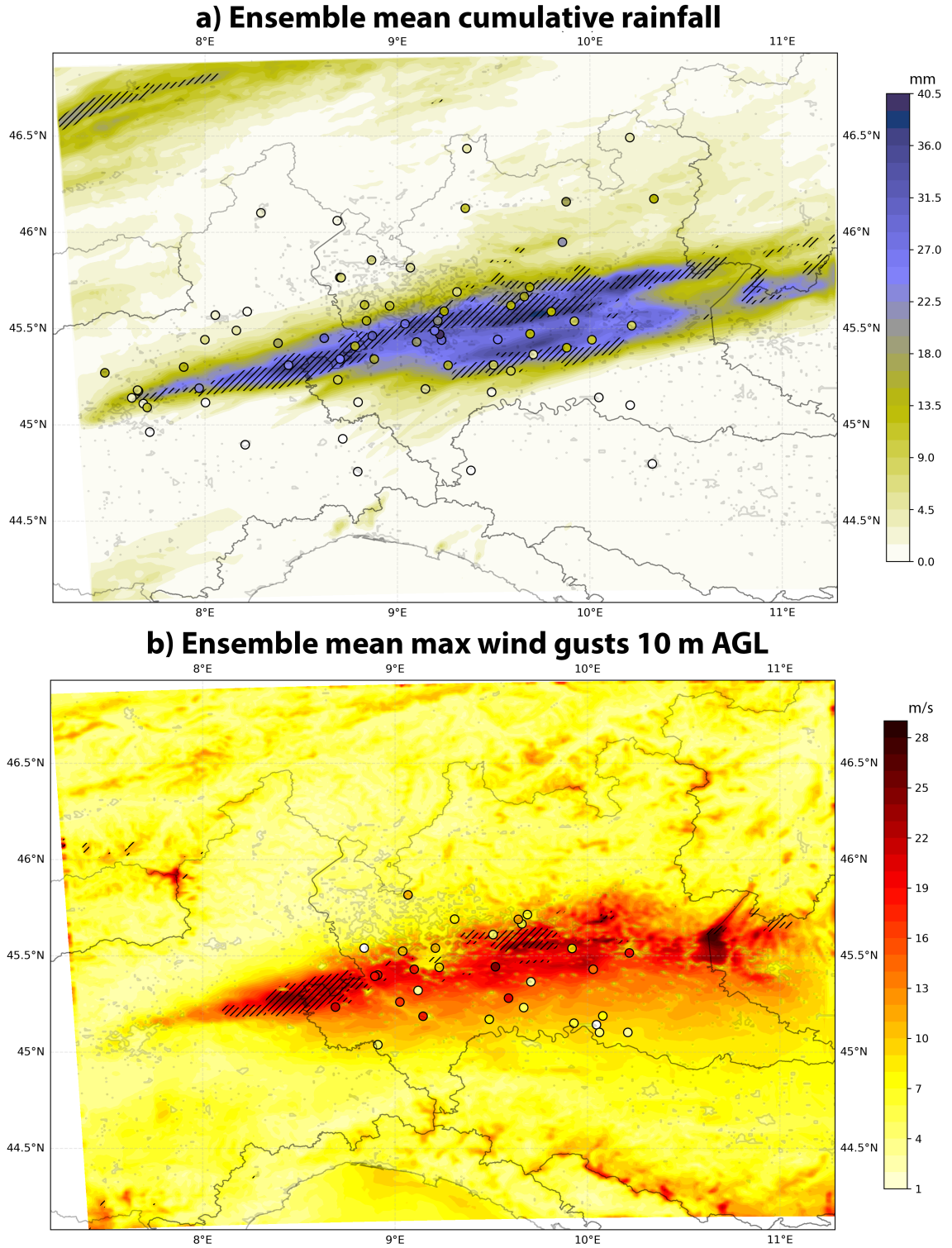


Figure 4.5: Ensemble mean of (a) cumulative rainfall (mm) and (b) maximum wind gusts at 10 m AGL between 0000 and 0500 UTC on 25 Jul 2023 (colour shading). Filled dots represent observations. Hatching highlights areas with standard deviation higher than 9 mm in (a) and 6 ms^{-1} in (b). The plotted domain is the same of Fig. 4.1b. Urban areas (LU Index higher than 50) are highlighted with grey contours.

4.3 Results

4.3.1 Pre-storm environmental differences

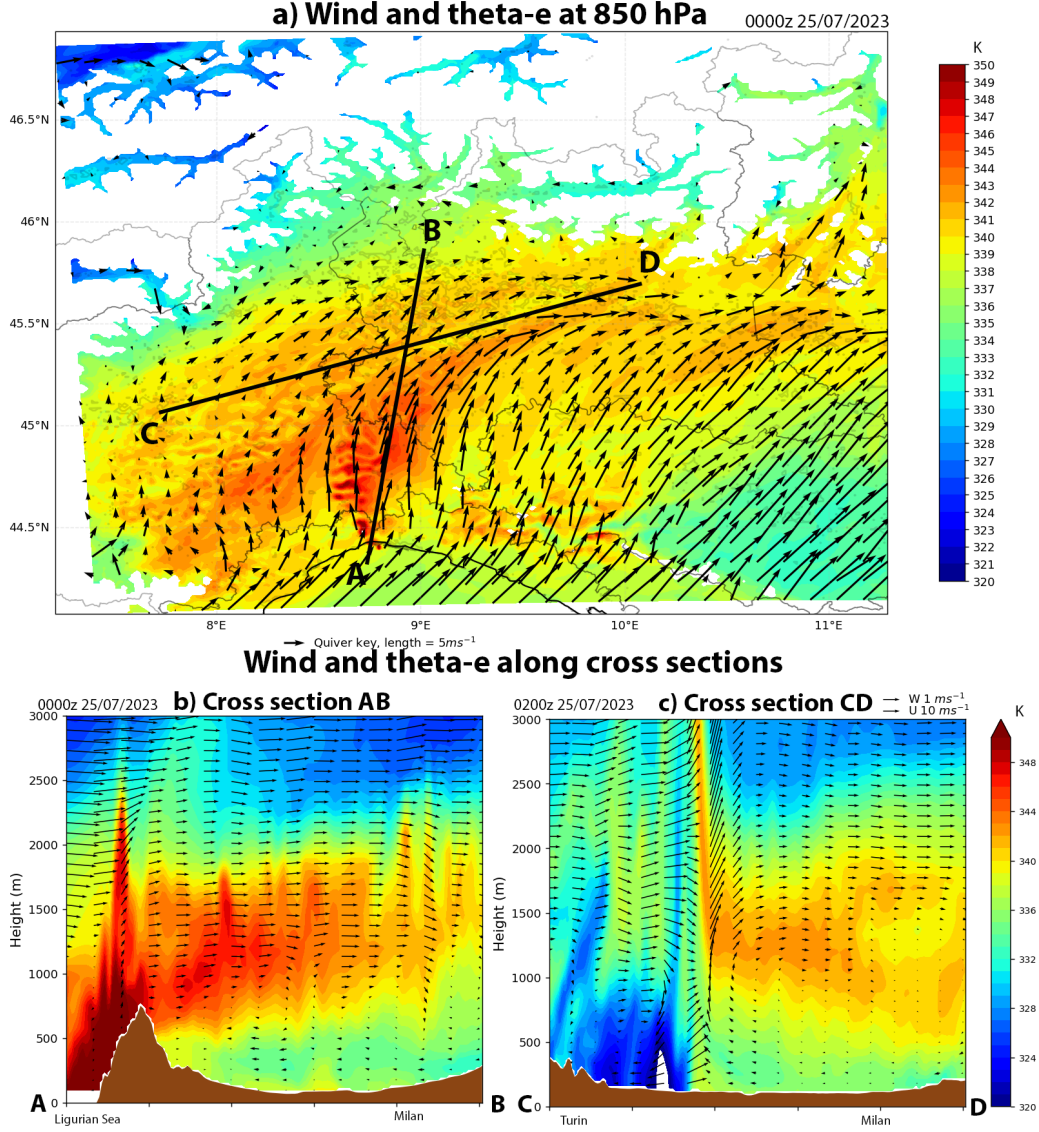


Figure 4.6: (a) Wind and Θ_e at 850 hPa at 00:00 UTC on 25 July 2023 in the *GLOBUS* simulation (the domain is the same of Fig. 4.1b). (b) Wind and Θ_e along the cross-section AB shown in (a) at 00:00 UTC on 25 July 2023; (c) as in (b) but along the cross-section CD at 02:00 UTC on 25 July 2023. Urban areas (LU Index higher than 50) are highlighted with grey contours.

As introduced in section 4.2.3, the most potentially buoyant air in the pre-storm environment was located between 1 and 2 km *amsl*. This situation is sometimes observed in the US during nocturnal severe weather outbreaks when a LLJ flows above the stable nocturnal PBL (Geerts et al., 2017). However, in the present case, there was no LLJ (Fig. 4.4a). On the contrary, the elevated potentially unstable layer was generated by a peculiar mesoscale dynamic related to the local topography of the region. A cross-section between the Ligurian Sea and Milan (location shown in Fig. 4.6a), derived from the *GLOBUS* run (Table 4.1)², reveals that the " Θ_e inversion" observed in the plain (Fig.

²A single simulation is preferred to the ensemble mean to analyze the mesoscale dynamics without the

4.4b) was caused by the lifting of a high- Θ_e air mass located over the Ligurian Sea by the Apennines (Fig. 4.6b). The intense meridional flow pushed this high- Θ_e air northward, where it became trapped between 1 and 2 km [amsl](#). The severe storm that crossed the plain a few hours later absorbed this elevated high- Θ_e tongue (Fig. 4.6c), likely leading to its intensification. The importance of high- Θ_e tongues from the Adriatic Sea for severe storms development over northeastern Italy was highlighted in previous studies (De Martin et al., 2024; Miglietta et al., 2016). This is the first time such a high- Θ_e tongue from the Ligurian Sea has been observed being lifted over the Apennines, generating an elevated potentially buoyant layer over the Po Valley and intensifying severe storms over northwestern Italy.

This peculiar dynamics explains why no significant differences in [CAPE](#) are observed between the urban and no-urban simulations (Table 4.1) at 00:00 UTC on 25 July 2023, just before the development of the severe storm (Fig. 4.7a). This result might seem unexpected given that Θ_e values at the first model level (5 m [AGL](#)) are about 4 K lower in the Milan urban area compared to a scenario without urban land use (Fig. 4.7c). To clarify this, the mean vertical profiles of Θ_e in the urban and no-urban ensembles over Milan are plotted in Fig. 4.8c. These profiles show that the large difference in Θ_e values is confined to the first 30 hPa of the troposphere. At around 850 hPa, there is the peak in Θ_e previously discussed, which is unaffected by urban land use. Since [MU CAPE](#) uses the parcel with the highest Θ_e to compute potential instability (Manzato and Morgan, 2003), this elevated parcel is not impacted by urban land use, resulting in almost similar values in the urban and no-urban ensembles. The decrease in Θ_e values in the lowest 30 hPa of the troposphere over Milan is driven by a decrease in moisture (Fig. 4.8b). The slight increase in temperature caused by the [UHI](#) (Fig. 4.8a) is insufficient to compensate for the moisture decrease, suggesting that the [UDI](#) effect can induce a *urban low- Θ_e island*. Note that in the simulation there is a weak [UHI](#) that is not present in the observations. Without the [UHI](#), Θ_e values would have been even lower.

Furthermore, [PBL](#) heights, computed according to the [MYJ](#) and [BouLac PBL](#) schemes, increase on average by about 100 m over the city center of Milan compared to the surrounding rural area (Fig. 4.7e), consistently with the well-established literature on urban boundary layer dynamics (Barlow, 2014; Brandi et al., 2021; Fenner et al., 2024).

Regarding the kinematic differences in the pre-storm environment between urban and non-urban simulations, Fig. 4.7b shows that SRH_{0-3km} decreased by 10-20 Jkg^{-1} over the Milan urban area, which is about 10% less than in the surrounding rural areas (considering the value of 129 Jkg^{-1} measured by the sounding, Table 4.2). The mean urban and non-urban hodographs plotted over Milan at 00:00 UTC on July 25, 2023, in Fig. 4.8d highlight that the differences in the wind vertical profile are located close to the surface and between 1 and 2 km [amsl](#). Differences in the 1-2 km layer are random (slightly varying results at different points), while the difference near the ground is systematic. Indeed, Fig. 4.7b shows how the [SRH](#) difference follows the urban contours, suggesting it may be closely related to what happens near the ground.

This result contrasts with other supercell events studied in the [US Great Plains](#), where an increase in [SRH](#) was observed in urban areas (Reames and Stensrud, 2018; Fan et al., 2023). Reames and Stensrud (2018) suggests that the increased [SRH](#) observed in their simulations, both over and downstream of the city, is due to a decrease in surface wind speed over the city, which in turn intensifies wind shear. Even in this case, 10 m wind speeds are decreased by about 1 ms^{-1} (Fig. 4.7d) over Milan, but, overall, the [SRH](#) is reduced. Note that, in this case, surface wind speeds were generally weak, about 3 ms^{-1}

filter of an average.

(Fig. 4.2a) before the storm's development. To determine which atmospheric layer contributes to the decrease in SRH_{0-3km} , the vertical profile of SRH_{0-z*} (z^* is the upper limit of the integration of SRH) is plotted for both the urban and non-urban ensemble means in Fig. 4.9a. SRH_{0-z*} is defined as:

$$SRH_{0-z*} = \int_{surface}^{z*} (\vec{v} - \vec{c}) \cdot \vec{\omega}_h dz \quad (4.1)$$

where $\vec{v}$ is the environmental wind, $\vec{c}$ is the storm motion computed according to Bunkers et al. (2000), and $\vec{\omega}_h$ is the horizontal component of vorticity. Near the surface, SRH_{0-z*} is larger in the urban ensemble (black arrow in Fig. 4.9a), but at about 100 m AGL, it becomes smaller in the urban ensemble than in the non-urban one, and this difference persists up to 3 km AGL. The differences in the hodographs between 1 and 2 km AGL (Fig. 4.8d) compensate for each other in the vertical profile of SRH_{0-z*} and do not influence the difference generated in the first few hundred meters of the troposphere (black arrow in Fig. 4.9a).

The difference in SRH_{0-z*} in the first hundred meters is unlikely to be caused by storm-relative winds, as the relative differences between the urban and non-urban scenarios are small near the ground (Fig. 4.9c). On the other hand, the vertical profiles of the horizontal vorticity components, defined as:

$$\vec{\omega}_h = \left(-\frac{\partial v}{\partial z}, \frac{\partial u}{\partial z} \right) \quad (4.2)$$

where v is the meridional wind component and u is the zonal component (assuming that the horizontal gradient of vertical velocity is negligible), show that in the first 30 m AGL, horizontal vorticity is much larger in the non-urban ensemble (blue arrow in Fig. 4.9b). Between 30 and 200 m AGL (with slight differences depending on the component considered), vorticity is more positive in the urban ensemble than in the non-urban one (green arrow in Fig. 4.9b). Since this positive vorticity is multiplied by negative storm-relative winds (Fig. 4.9c, Eq. 4.1), the urban ensemble experiences a more negative contribution to SRH_{0-z*} compared to the non-urban ensemble, which explains the negative difference in SRH_{0-z*} .

To summarize, simulations of the analyzed case study revealed that in the pre-storm environment there are notable differences between the urban and non-urban ensembles in the first hundred meters of the PBL over the Milan urban area. Near-ground thermodynamic differences, such as the *urban low- Θ_e island*, had likely a marginal influence on the severe storm dynamics because the most unstable layer was elevated. In contrast, kinematic differences are more remarkable, with a 10% reduction in SRH_{0-3km} due to increased vorticity in the 30-200 m AGL layer over the city.

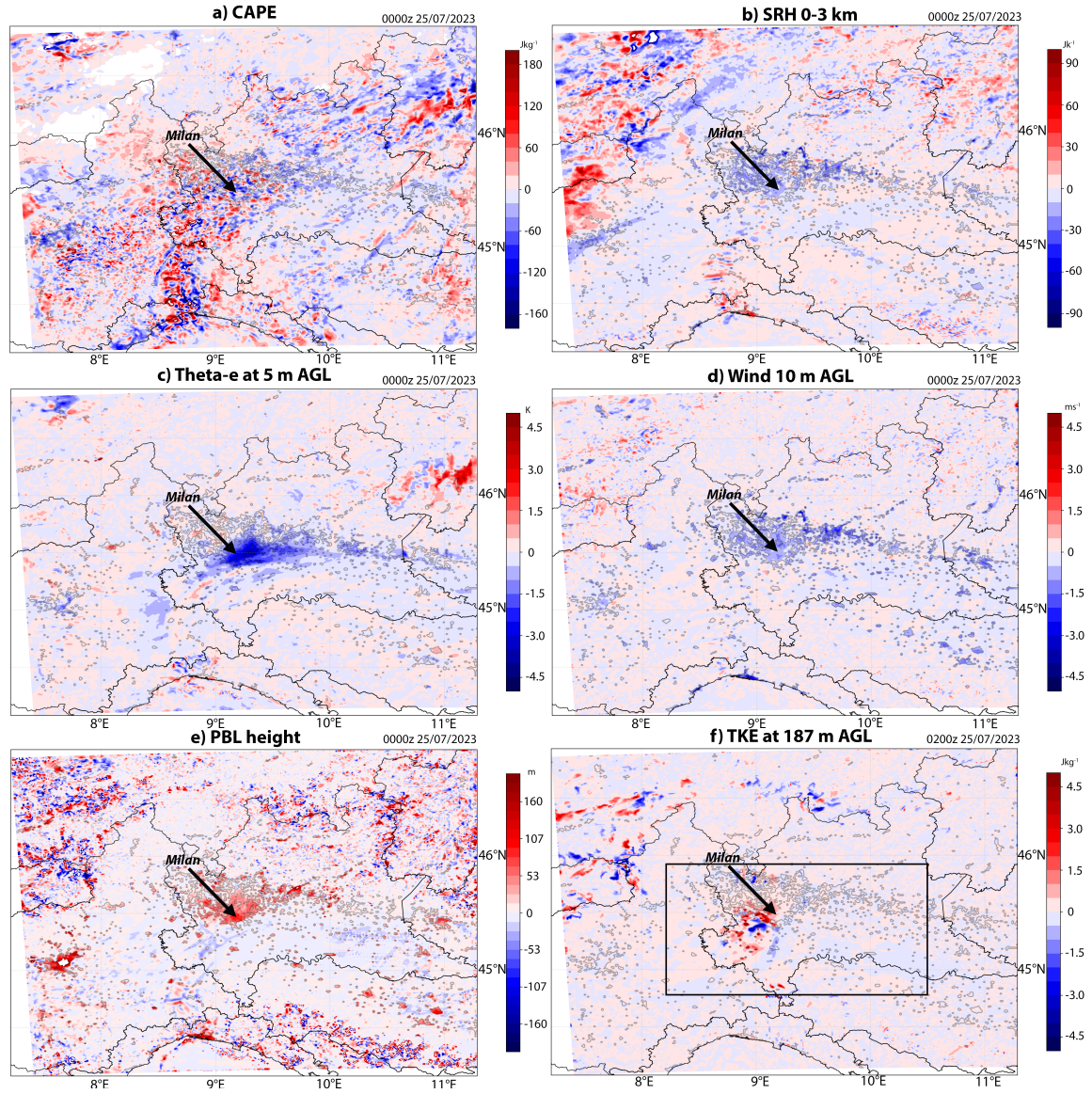


Figure 4.7: Average ensemble mean difference of (a) CAPE, (b) SRH 0-3 km, (c) Θ_e at 5 m AGL, (d) wind at 10 m AGL, (e) PBL height, (f) TKE at 187 m AGL. All plots are referred to 00:00 UTC 25 July 2023, except (f) which is plotted at 02:00 UTC 25 July 2023. The domain plotted is the same as Fig. 4.1b, urban areas (LU Index higher than 50) are highlighted with grey contours.

**Mean urban and no-urban vertical profiles at 0000 UTC 25th July 2023
over Milan (45.47°N, 9.19°E)**

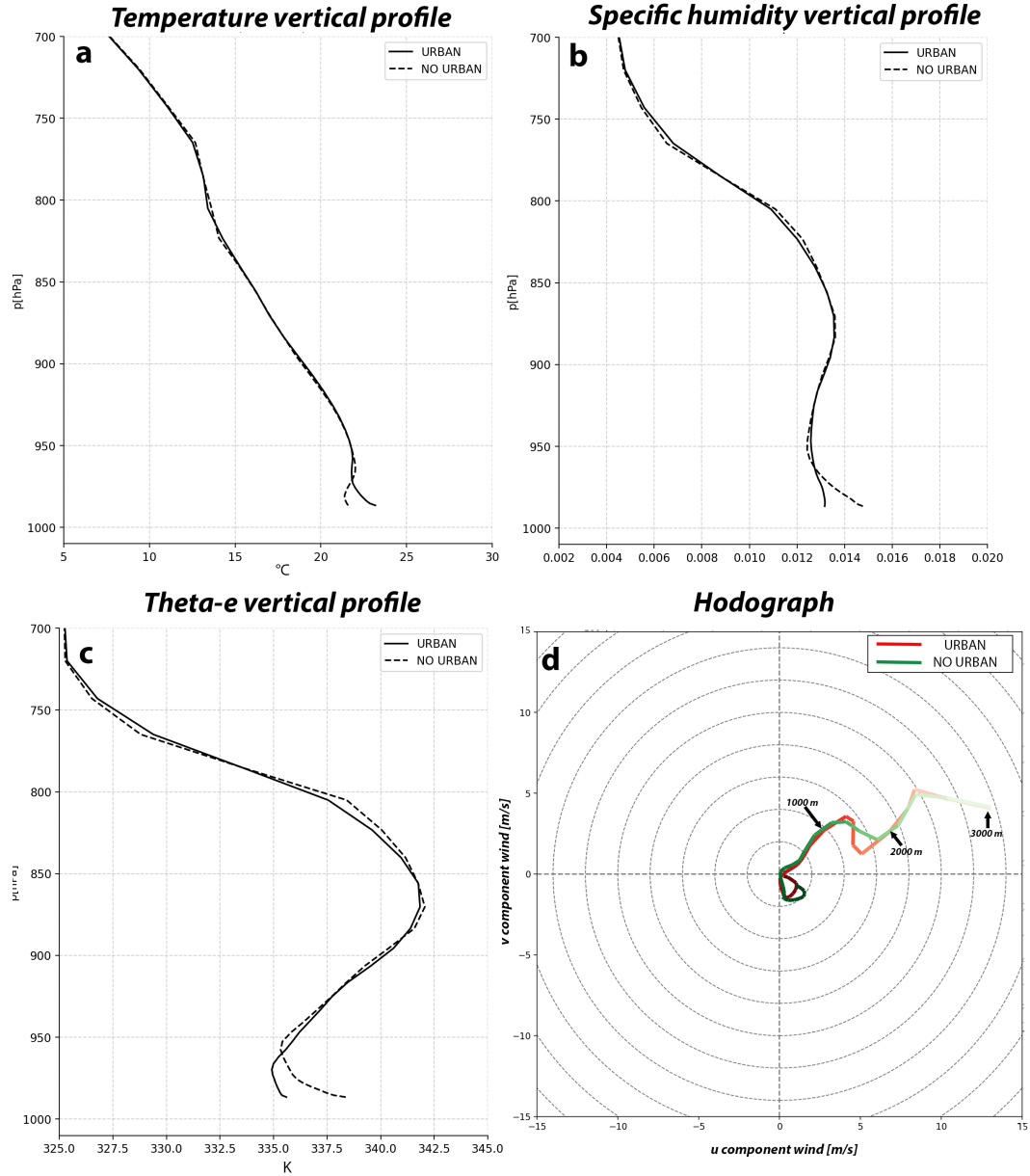


Figure 4.8: (a) Ensemble mean urban (solid line) and no-urban (dashed line) vertical profiles over Milan (45.47°N, 9.19°E) at 00:00 UTC on 25 July 2023 of (a) temperature, (b) specific humidity, (c) Θ_e ; in (d) the ensemble mean urban (red line) and no-urban (green line) hodograph in the same location and time of the vertical profile in (a) are plotted.

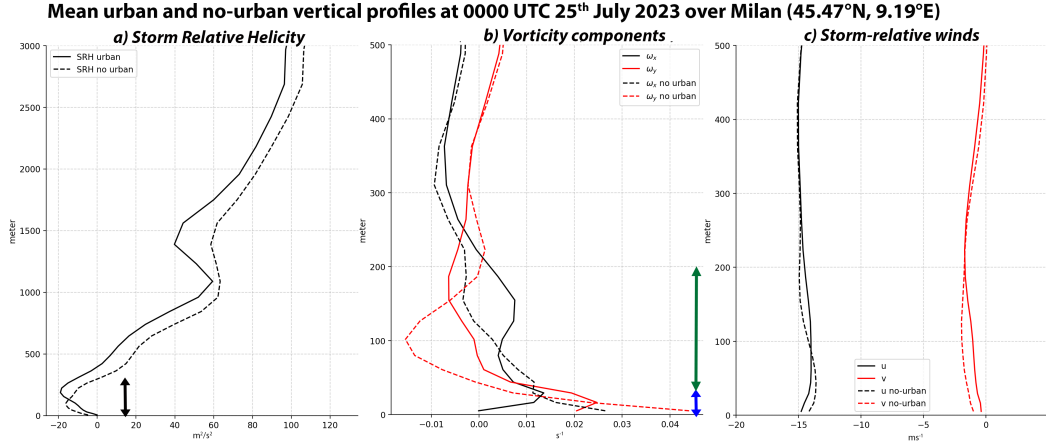


Figure 4.9: (a) Ensemble mean urban (solid line) and no-urban (dashed line) vertical profiles over Milan (45.47°N, 9.19°E) at 00:00 UTC on 25 July 2023 of (a) storm relative helicity, (b) zonal (ω_x , black) and meridional (ω_y , red) vorticity components, (c) storm relative winds u and v .

4.3.2 Storm differences

After having analyzed the differences in the pre-storm environment, this section discusses the differences observed in the simulated storms between the urban and non-urban ensembles. The analysis is carried out using the boxes described in Section 4.1.4.

Upstream region

In the *upstream* region with respect to the storm motion (red box in Fig. 4.10), the storm exhibits a slight northward shift. This shift can be inferred from cumulative rainfall patterns (Fig. 4.10a), maximum updraft speed (Fig. 4.10c), and maximum column-integrated graupel amount (Fig. 4.10f). This northward displacement does not correspond with any significant changes in precipitation intensity or 10-meter wind gusts (Table 4.5). However, there are notable decreases in updraft strength (with a maximum **UH** on average 9.2% lower than in non-urban simulations, Table 4.5) and in the amount of column-integrated graupel (-10.1%, Table 4.5).

The origin of this northward storm shift in the *upstream* region appears to be linked to the urban area of Turin, as the shift begins precisely over that city. To test this hypothesis, two additional experiments were conducted: *NoMi*, where Milan was removed, and *NoTo*, where Turin was removed. Surprisingly, the results indicate that when Milan is removed (while Turin remains), the northward shift disappears, and a slight southward shift occurs instead (Fig. 4.11a). Conversely, when Turin is removed but Milan remains, the northward shift reappears (Fig. 4.11b). This finding suggests that the northward storm shift is somehow generated by the urban land use associated with Milan rather than Turin.

To further understand the reasons behind the northward shift of the storm in the *upstream* region under the urban scenario, the low-level circulation in this area is analyzed. At 950 hPa (approximately 300 m **AGL** over Milan), a weak northerly flow with low- Θ_e values descends from the Alps before the storm development. This flow affects the western part of the Milan urban area before turning right and moving westward toward Turin (Fig. 4.11c). A tracer released at the first model level (5 m **AGL**), just westward of Milan (red dot in Fig. 4.11d), highlights an easterly flow that carries the tracer toward Turin within a few hours. By comparing tracer density at two different isobaric levels between

Variable	Upstream	City region	Downstream
Cumulative Rainfall	-0.14 mm (-1.79%)	+0.28 mm (+1.99 %)	+6.19 mm (+8.15 %)
Max 10 m AGL wind gusts	-0.1 ms^{-1} (-1.0 %)	-1.4 ms^{-1} (-13.4 %)	-0.7 ms^{-1} (-5.3 %)
Max updraft speed	-0.47 ms^{-1} (-5.1 %)	+0.64 ms^{-1} (+3.9 %)	+0.35 ms^{-1} (+1.85 %)
Max UH	-2.5 $m^2 s^{-2}$ (-9.2 %)	+4.1 $m^2 s^{-2}$ (+6.5 %)	+3.4 $m^2 s^{-2}$ (+3.7 %)
Min downdraft speed	-0.13 ms^{-1} (-3.1 %)	-0.12 ms^{-1} (+1.9 %)	-0.06 ms^{-1} (+0.9 %)
Max Graupel amount	-0.96 kgm^{-2} (-10.1 %)	+0.71 kgm^{-2} (+3.7 %)	+1.22 kgm^{-2} (+5.6 %)

Table 4.5: Average differences between urban and no-urban simulations of some relevant variables within the *upstream* (red box), the *city* (black box), and the *downstream* regions (Fig. 4.10) between 00:00 and 05:00 UTC on 25 July 2024. The average percentage differences with respect to the absolute value of the variables in the urban simulations are reported in brackets.

the *NoTo* and *NoMi* experiments, the influence of urban land use on this northerly flow becomes evident. In the experiment where Milan is included (*NoTo*), the tracer disperses more southward and westward. The increased roughness due to Milan’s urban land use slightly shifts the northerly flow in these directions, subsequently affecting severe storm development in the *upstream* region. Although there are no apparent differences in the pre-storm environment in this area (as shown in Fig. 4.7), the meridional shift in tracer density at 980 hPa (Fig. 4.11e) and 950 hPa (Fig. 4.11f) is pretty similar to the northward shift of the storm observed in Fig. 4.10a,c. Therefore, it is plausible that the shift in the low-level northerly flow, induced by Milan’s urban land use, is the cause of the observed storm shift in the *upstream* region.

City region

In the *city* region (black box), the previously observed northward shift of the storm disappears. Unlike in the *upstream* region, the urban simulations show stronger updrafts within the city of Milan, with maximum updraft speed increasing by 3.9% and maximum UH by 6.5%. Additionally, there is a slight increase in the amount of column-integrated graupel (+3.7%) and precipitation (+2%). The intensification of the storm occurs directly over Milan (as indicated by the black arrows in Fig. 4.10c,f).

This intensification is likely due to the interaction of wind gusts with the urban landscape. When wind gusts hit the buildings of Milan, they generate turbulence, which increases TKE, as shown in Fig. 4.7f. This enhanced turbulence leads to more vigorous vertical mixing, which, in turn, may contribute to the observed intensification of updrafts over the city (Fig. 4.10c,d and Table 4.5). The band of intensified updrafts extends from the city center toward the east, slightly to the right of the mean wind direction.

In contrast, wind gusts are significantly reduced over Milan (-13.4%) due to the increased surface roughness associated with the urban landscape. As discussed in section 4.2.4, this reduction in wind gusts over the city is likely overestimated because the resolution and parameterization employed in the model do not fully capture the turbulent gusts generated by interactions with buildings.

Downstream region

In the *downstream* region with respect to the storm motion (blue box in Fig. 4.10), there is a marked intensification of precipitation, with an average increase of 6.2 mm (+8.2%, Table 4.1). This intensification is consistent with the results of other urban morphology impacts studies (e.g., Yang et al. 2024; Sui et al. 2024). Furthermore, in the *downstream* region there is an increase in the maximum column-integrated graupel amount (+1.22 kgm^{-2} , +5.6%, Table 4.5) compared to the non-urban simulations. This intensification occurs

primarily in a stripe along the southern side of the box (indicated by the red arrow in Fig. 4.10a). This stripe is continuous with the band of intensified UH and column-integrated graupel amount that originates over the city center (black arrows in Fig. 4.10c,f). The intensified precipitation observed downstream of Milan is likely a direct consequence of the strengthened updrafts over the city center. Although updrafts are stronger in the urban simulations within the *downstream* region, their intensity is less pronounced than in the *city* region (Table 4.5). Moreover, a wake of reduced wind gusts extends downstream of the city center for several tens of kilometers (Fig. 4.10b), with wind gusts being reduced by an average of 5.3% compared to the non-urban simulations.

In all three regions (*upstream*, *city*, and *downstream*), there are no remarkable differences in the maximum downdraft speed between the urban and non-urban simulations (Fig. 4.10e and Table 4.5).

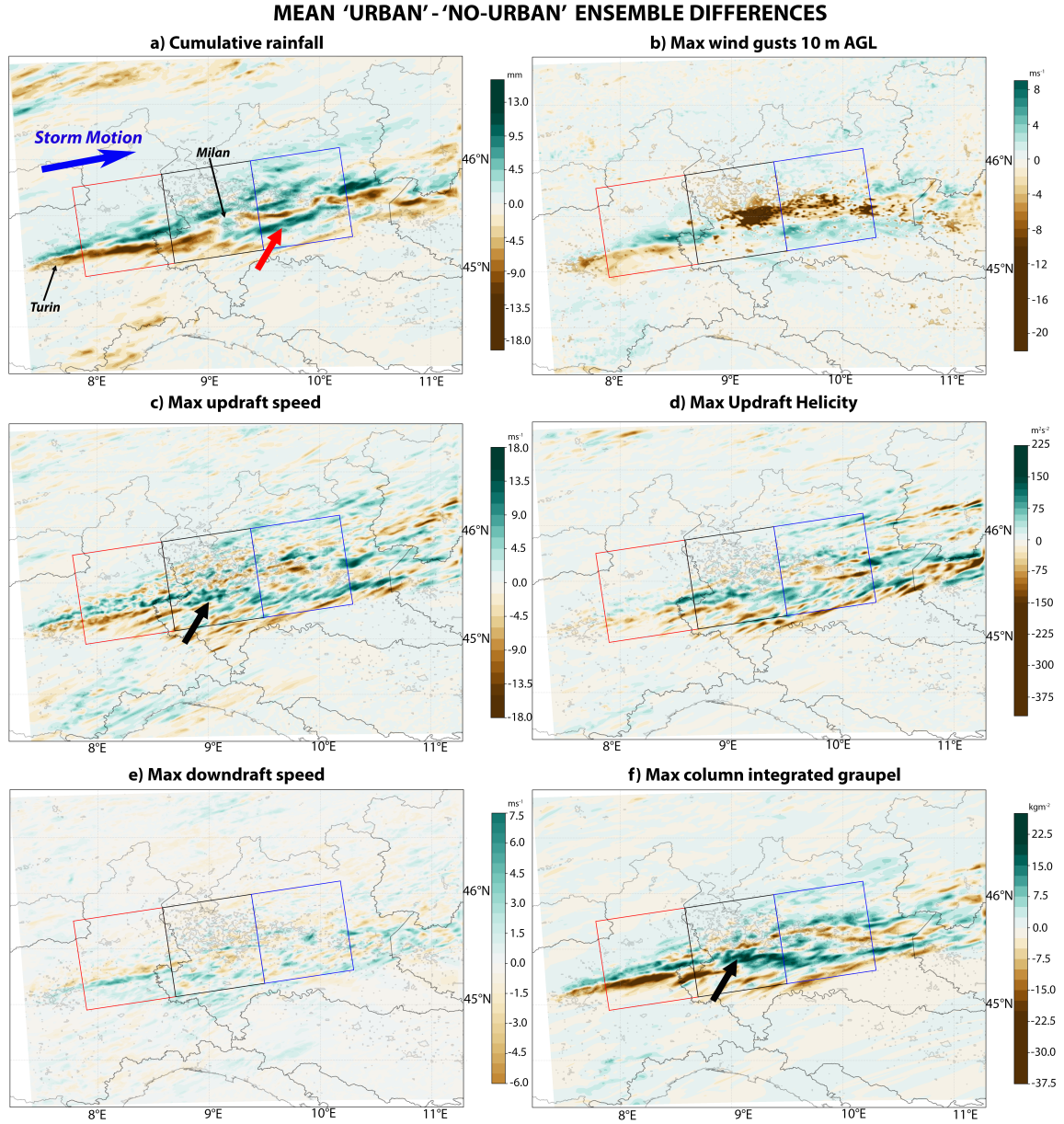


Figure 4.10: Mean ensemble difference between urban and no-urban simulations of (a) cumulative rainfalls (mm), (b) max wind gusts (ms^{-1}), (c) max updraft speed (ms^{-1}), (d) max UH (m^2s^{-2}), (e) max downdraft speed (ms^{-1}), (f) max column integrated graupel (kgm^{-2}). All these quantities are computed in the 0000-0500 UTC period on 25 Jul 2023. The three boxes highlight the *upstream* region (red box), city region (black box) and *downstream* region (blue box) within which the average differences shown in Table 4.5 are computed. Urban areas (LU Index higher than 50) are highlighted with grey contours.

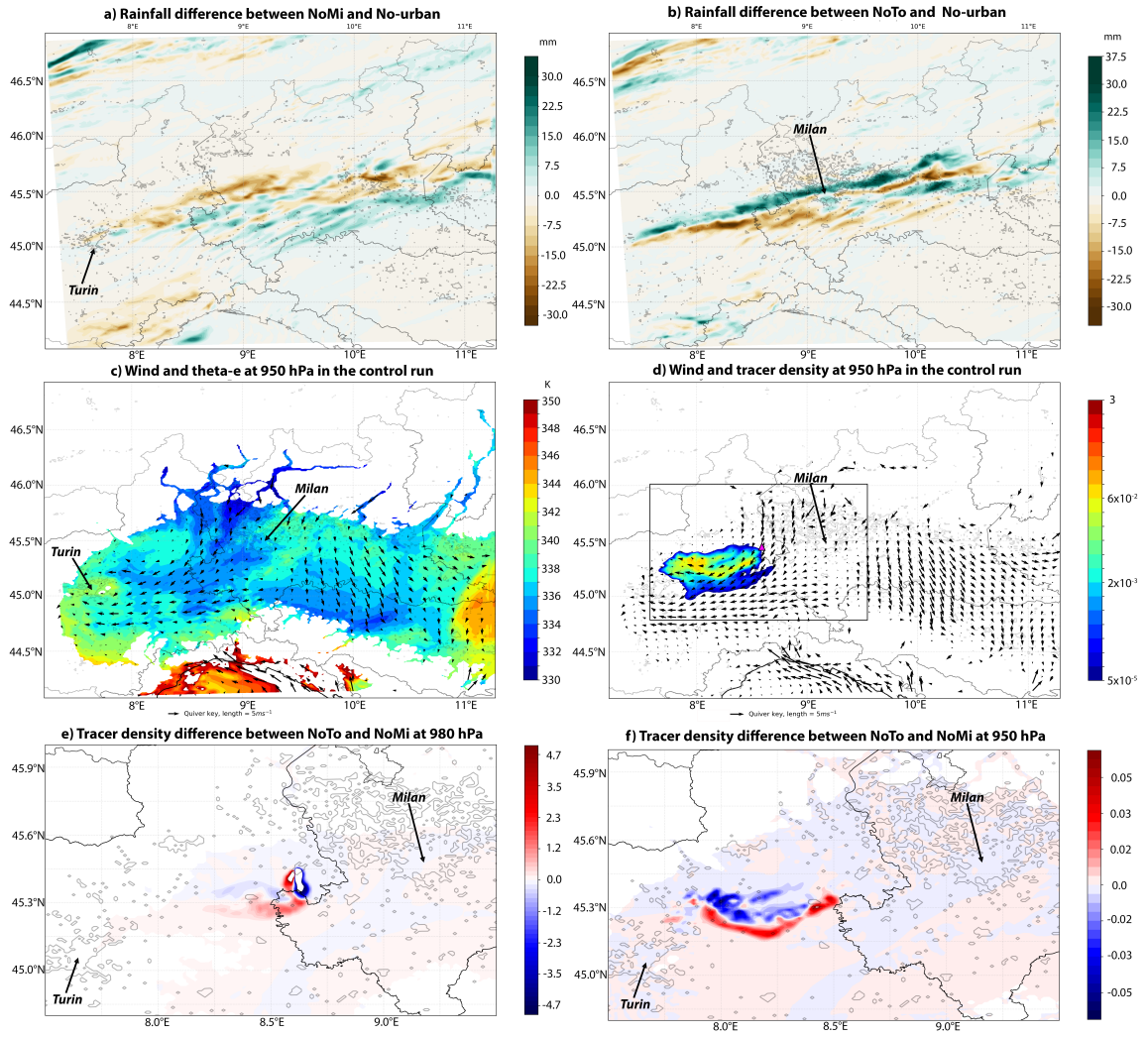


Figure 4.11: Cumulative rainfall difference between (a) *NoMi* and *Bou_nourb* experiments and between (b) *NoTo* and *Bou_nourb*; (c) Θ_e (color shading) and wind speed (arrows) at 950 hPa in the GLOBUS run at 00:00 UTC on 25 July 2023; (d) concentration (color shading) at 00:00 UTC on 25 July 2023 of a tracer released in the location shown with a red triangle and wind speed (arrows) at 950 hPa; concentration difference (color shading) at 00:00 UTC on 25 July 2023 at (e) 980 hPa and (f) 950 hPa of the same tracer of (d) between the *NoTo* and *NoMi* experiments within the black box in (d). Urban areas (LU Index higher than 50) are highlighted with grey contours.

Updraft modifications

To better clarify the effect of the city of Milan on the storm, an analysis of vertical velocities using only the *GLOBUS* run and the no-urban counterpart simulation is carried out. In Fig. 4.12, four different time steps are plotted on the same map, two before the impact of the storm on the city and two after the impact. Areas with vertical velocities above certain thresholds at three different pressure levels are shown for the urban simulation (red shading) and the no-urban simulation (green shading).

At 02:00 UTC, the shape of the updraft is similar in both the urban and no-urban simulations, with the previously discussed northward shift evident in the urban case. At 02:30 UTC, the shift is no longer present, and in the urban simulation a bow echo shape appears at 700 and 950 hPa (Fig. 4.12b,c), which is not evident in the no-urban simulation. After the impact with the city (at about 02:40 UTC), the structure of the updraft is markedly modified. By 03:00 UTC, the storm in the no-urban simulation assumes a compact bow echo shape that was not present before. On the other hand, in the urban simulation the updraft splits into three discrete cells (Fig. 4.12a,b), with a new convective cell forming south of Milan. Furthermore, the convective line is delayed compared to the no-urban experiment, particularly around the latitude of Milan (45.45°N). One hour after the clash with Milan, the updrafts again have similar shapes in both urban and no-urban simulations, but the entire convective structure is delayed in the urban simulation.

The city slows down the progression of the gust front, consistent with the weakening of wind gusts over the city (Fig. 4.10b), initially only in the city center (at 03:00 UTC, Fig. 4.12c), and subsequently along the entire *QLCS* (03:30 UTC, Fig. 4.12c). After the impact with the city, the gust front intensifies on the southern side in the urban simulation (between 45°N and 45.2°N at 03:00 UTC), likely because it is deviated southward by the city. Consequently, the *QLCS* is broken, with new convective cells generated to the right of the storm motion, while the central part of the *QLCS* is weakened.

Cross-sections AB and CD, plotted in Fig. 4.13 highlight that the shape and tilting of the updrafts in the urban and no-urban simulations are very similar. At 02:00 UTC, they are almost superimposed (Fig. 4.13a), while at 03:30 UTC, 50 minutes after the clash with Milan, the storm in the urban simulation is delayed by about 6 km compared to the storm in the no-urban simulation (Fig. 4.13b). These cross-sections confirm that the inflow layer was quite deep, up to 2.5 km *AGL* (see the depth of the air mass with high Θ_e values ingested by the updraft), suggesting that the thermodynamic effect of Milan in the first hundred meters is likely negligible. Note also the significant vertical extent of the simulated storm, up to 14 km *AGL*.

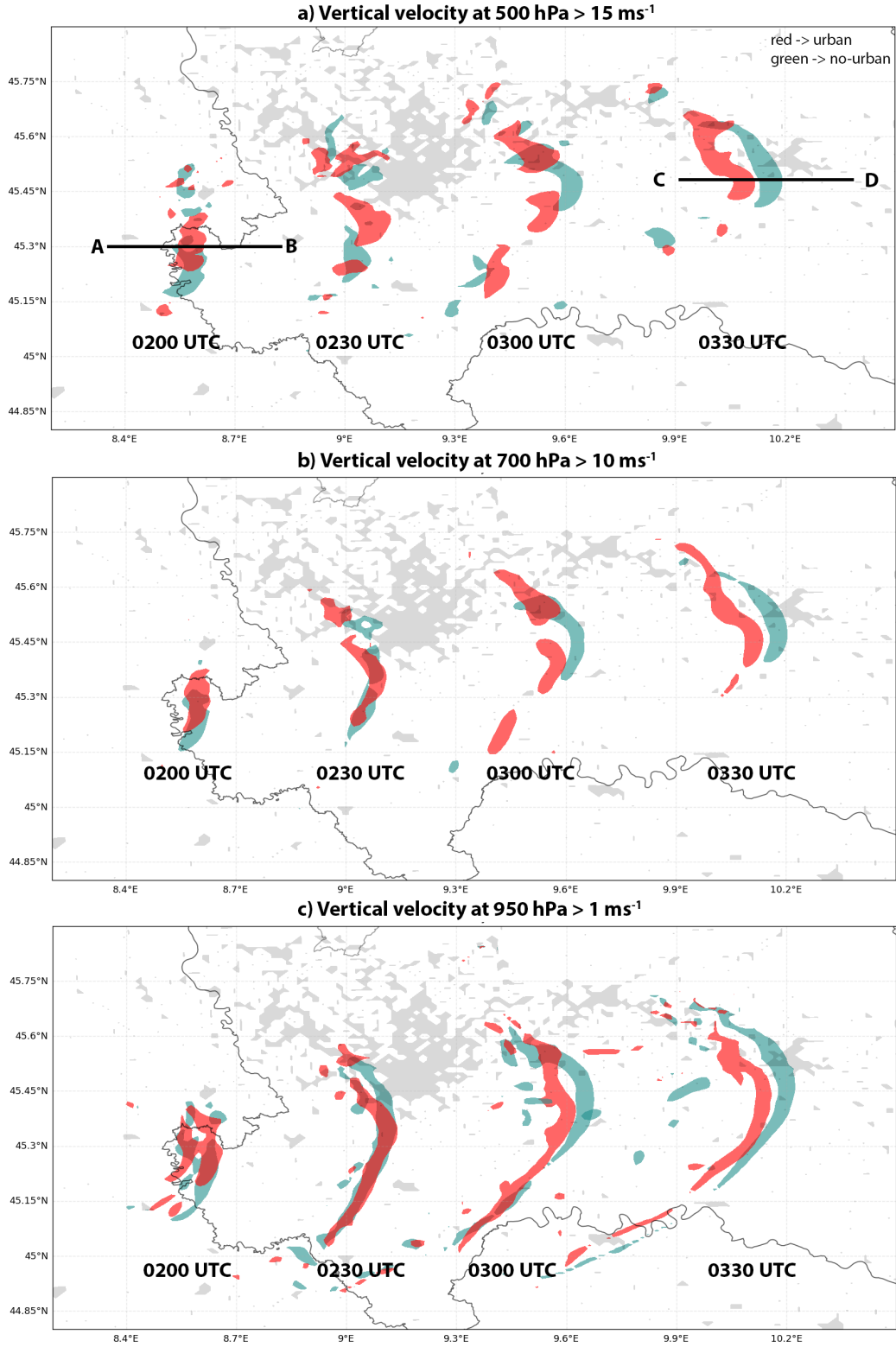


Figure 4.12: (a) vertical velocities at 500 hPa higher than 15 ms^{-1} are shaded with red (GLOBUS run) or green (Bou_nourb experiment): four different time steps (02:00, 02:30, 03:00, 03:30 UTC) are shown accordingly to labels in the Figure. AB and CD segments indicate the location of the cross-section shown in Fig. 4.13. (b) as in (a) but for vertical velocities at 700 hPa higher than 10 ms^{-1} ; (c) as in (a) but for vertical velocities at 950 hPa higher than 1 ms^{-1} . The plotted domain is shown with a black box in Figure 4.9f. Urban areas are shaded with grey.

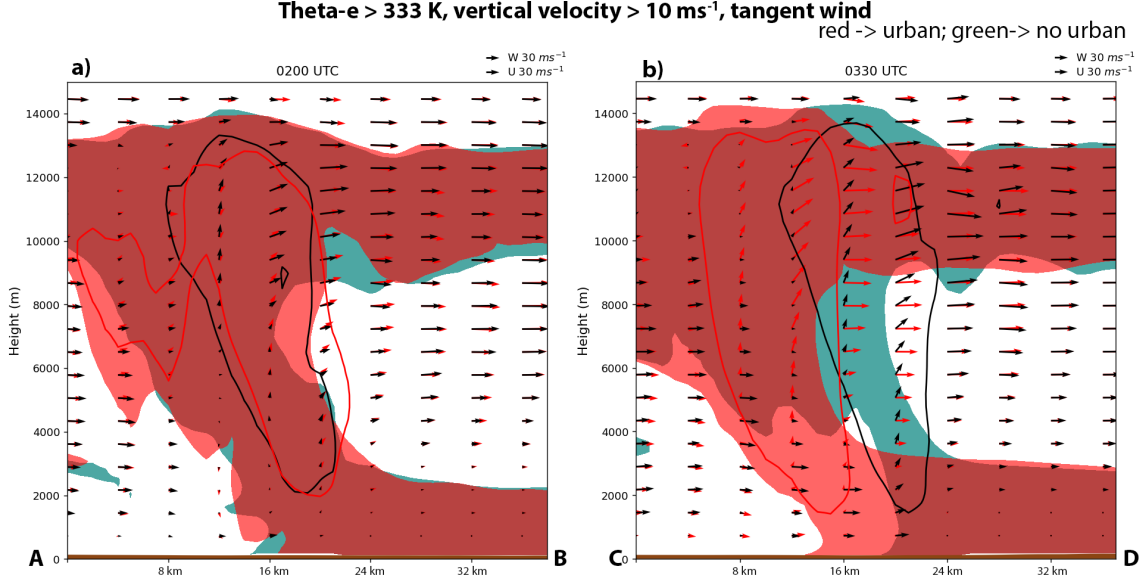


Figure 4.13: (a) Wind (arrows) and vertical velocities higher than 10 ms^{-1} (contours) of *GLOBUS* simulation (red) and *Bou_nourb* experiment (black) along the cross-section AB in Fig. 4.12a at 02:00 UTC 25 July 2023. The area with Θ_e values higher than 333 K is shaded with red (*GLOBUS* run) and green (*Bou_nourb* experiment). (b) as in (a) but along cross section CD of Fig. 4.12a at 03:30 UTC 25 July 2023.

4.4 Discussion

The convective windstorm that hit Milan on 25 July 2023 serves as a relevant case study for understanding the interaction between a medium-large city and an organized convective storm. The severe damage caused by wind gusts in the city center raises the question of whether urban land use may have influenced the storm's development. A numerical experiment was carried out: a physics ensemble and a corresponding no-urban physics ensemble obtained running the same simulations with the urban land use replaced by croplands were generated using the *WRF* model. The differences between these two ensembles were thoroughly investigated to assess the impact of urban land use on the pre-storm environment and on the storm itself. Key findings regarding the effect of urban land use on the pre-storm environment in this case study are hereafter summarized:

- Milan generates an *urban low- Θ_e island*, with a decrease of Θ_e by 4 K at the surface, extending up to 300 m *AGL*, due to reduced moisture over the city.
- Despite the surface Θ_e decrease, which would typically suggest reduced potential instability, *MU CAPE* remains unchanged because an elevated layer at about 1.5 km *amsl*, containing higher Θ_e values, is not affected by urban land use. This elevated high- Θ_e air mass originated from the Ligurian Sea, was lifted by the Apennines, and pushed northward by a strong meridional flow.
- Environmental vorticity decreases within the first 30 m *AGL* above Milan but increases in the 30-200 m *AGL* layer.
- SRH_{0-3km} decreases by about 10% over the city due to the increased vorticity in the 30-200 m *AGL* layer.

One of the most counterintuitive outcome is the reduction in SRH_{0-3km} over the city, which contrasts with the findings of Reames and Stensrud (2018) and Fan et al. (2023). Similar to them, an increasing amount of vorticity within the first hundred meters over Milan is observed. The overall decrease in SRH in this event is attributed to the specific wind profile, as discussed in section 4.3.1. Thus, our results confirm that low-level SRH and vorticity are strongly influenced by urban land use, but the sign of this influence may vary depending on the event. The impact of urban areas on $CAPE$ also varies: some studies reported increased values over cities (Moraglia et al., 2024), while others found lower values (Yang et al., 2021b; Reames and Stensrud, 2018; Yang et al., 2014). Soderholm et al. (2017) observed higher Θ_e values over the city of Brisbane just before a severe hailstorm, due to the UHI and the moisture advected by the sea breeze. In this context, the presence of an *urban low- Θ_e island* is highlighted, driven by the reduced moisture over the city. Further investigation of whether this feature is observed in other cities and events is encouraged.

After having discussed the effects of urban land use on the pre-storm environment, the main impacts of urban land use on the convective windstorm are summarized in Figure 4.14 with a sketch. Notably, a northward shift of the storm is observed in the *upstream* region. Additional numerical experiments suggest that this shift may be a result of Milan that causes a slight deviation of a low-level northerly flow with low- Θ_e values (Fig. 4.11). This is an intriguing finding, as it demonstrates that a large city can induce storm modifications even 100 km upstream by altering low-level winds. Over the city, wind gusts are reduced by an average of 13%: the gust front is slowed down by the increased roughness of the urban landscape. The ensemble mean also indicates an intensification of updrafts over the city (Fig. 4.10c), probably due to the increased TKE (Fig. 4.7f) generated by the interaction of the gust front with buildings, which enhances vertical motions. *Downstream* of the city, the storm splits into three convective cells (Fig. 4.12b), with a delay of approximately 5 km compared to a no-urban scenario. The central convective cell produces more rainfall than in the no-urban scenario (red arrow in Fig. 4.10a). The storm modifications over and downstream of the city seem mostly related to the interaction of the gust front with the urban area. Thus, the modification of the flow by the urban canopy seems to have a larger effect on the storm than the changes in the thermodynamics.

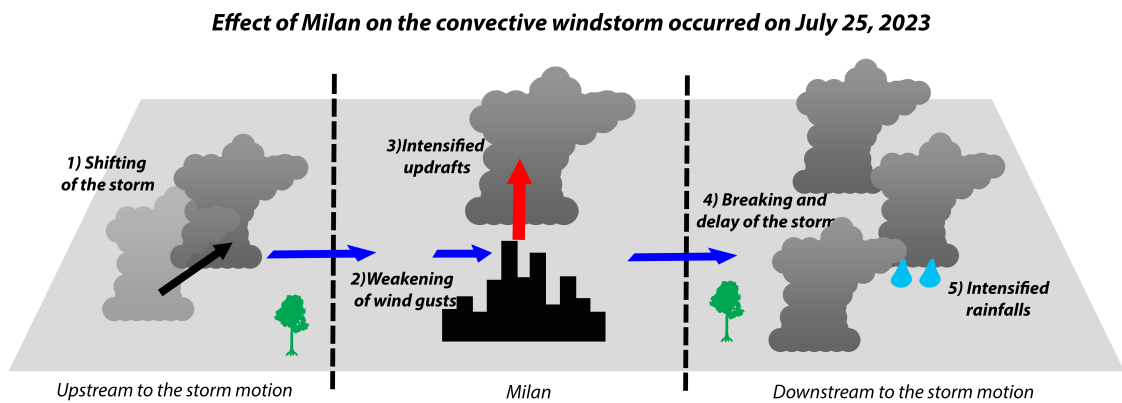


Figure 4.14: Conceptual model of the effect of the urban land use on the convective windstorm occurred on July 25, 2023, in Milan. The storm is represented with a grey cloud, the wind gust with a blue arrow, the updraft with a red arrow. The three relevant regions (*upstream*, *city* and *downstream* with respect to the storm motion) are separated by dashed lines.

The intensification of updrafts over the city (Lin et al., 2021; Forster et al., 2024; Platonov et al., 2024) and the splitting of the storm (Bornstein and LeRoy, 1990; Niyogi

et al., 2011; Lin et al., 2021; Yang et al., 2021a) have also been observed in previous studies. However, differently to some previous works (e.g., Dou et al. 2015), the downstream intensification of precipitation in the present case is an effect of the slowing down of the gust front over the city, rather than by a merging of the flow downstream of Milan. The slowing down of the gust front causes stronger vertical velocities over the city. Then, the intensified updrafts, rapidly pushed downstream by the strong upper-level flow, cause the downstream enhancement of precipitations. Furthermore, updraft intensification and storm splitting were shown for weaker thunderstorms with much less intense wind gusts. Hence, the novelty of this study is that even a fast-moving and highly energetic convective storm can be significantly influenced by a large city. This finding is particularly noteworthy because the pre-storm environment was characterized by a peak in Θ_e at about 1-2 km [amsl](#), which fueled the severe storm and masked the effects of the *urban low- Θ_e island*. A surface-based severe storm might be even more affected by the city. Furthermore, the analysis of the pre-storm environment, which shows a significant impact of the city on vorticity and Θ_e values in the first hundred meters above the surface, suggests that a city might dramatically alter the behavior of tornadic supercells, as this type of storm is highly dependent on atmospheric conditions in the very low-levels (Coffer et al., 2023; Roberts et al., 2020).

Although this robust and detailed sensitivity study demonstrates that the city has a significant impact on the severe storm, a limitation is that it is challenging to distinguish the city's impact from other atmospheric and topographic factors. While it can provide valuable insights into the physical processes occurring, it is not straightforward to precisely identify the mechanisms causing the observed storm modifications. Furthermore, the extent to which these results can be generalized to other cities and severe storm events remains unclear (Liu and Niyogi, 2019).

Chapter 5

Idealized Simulations of Supercell Storms Interacting with the Urban Environment

To address the limitations of a case study analysis highlighted in the previous Chapter and by Liu and Niyogi (2019), a new study is designed to systematically investigate the effects of cities on severe storms. In this work, particular emphasis is placed on understanding the underlying physical processes that drive storm modifications. Whereas the previous Chapter examined the impact of urban land use on a QLCS, the present study focuses on supercell storms. Supercells are the most extensively studied type of severe convective storm and can produce the full range of hazards associated with deep moist convection: lightning, hail, damaging wind gusts, heavy rainfall, and tornadoes. Moreover, several previous studies have suggested that supercells may exhibit notable sensitivity to urban land use (Fan et al., 2023; Lin et al., 2021; Naylor, 2020; Reames and Stensrud, 2018), potentially because they are fueled by air masses located very close to the ground (Coffer et al., 2023), as also suggested at the end of the previous Chapter.

In the present study, the simulation setup is idealized, featuring a completely flat domain with uniform land use (cropland) except for the presence of a city. The city is modeled as circular and characterized by uniform urban morphological parameters. Although idealized, the city's features are inspired by real European and US cities frequently affected by severe storms. Simulations are performed during the afternoon, when convective storms are more likely to be surface-based than at night, since non-surface-based storms are less likely to be affected by urban land use.

The analysis is divided into two main phases. The primary objective of the first phase is to investigate how an already developed supercell is modified by the presence of a city. To this end, the storm is initiated upstream and advected toward the city center, where an urban boundary layer has developed during the preceding hours. In particular, we aim to assess the storm's sensitivity to variations in city size, urban fraction, building height, and building density. Because cities are not expected to exert a dramatic influence on supercells, an ensemble approach is adopted to separate the physical signal of urban-induced storm modifications from the intrinsic chaotic variability of the atmosphere. A robust statistical analysis of the ensembles is then performed to quantify the impact of urban land use on supercells.

The second phase of the analysis focuses on understanding how the pre-storm environment is influenced by different urban morphological configurations. This aspect is crucial for explaining the urban-induced storm modifications identified in the first phase.

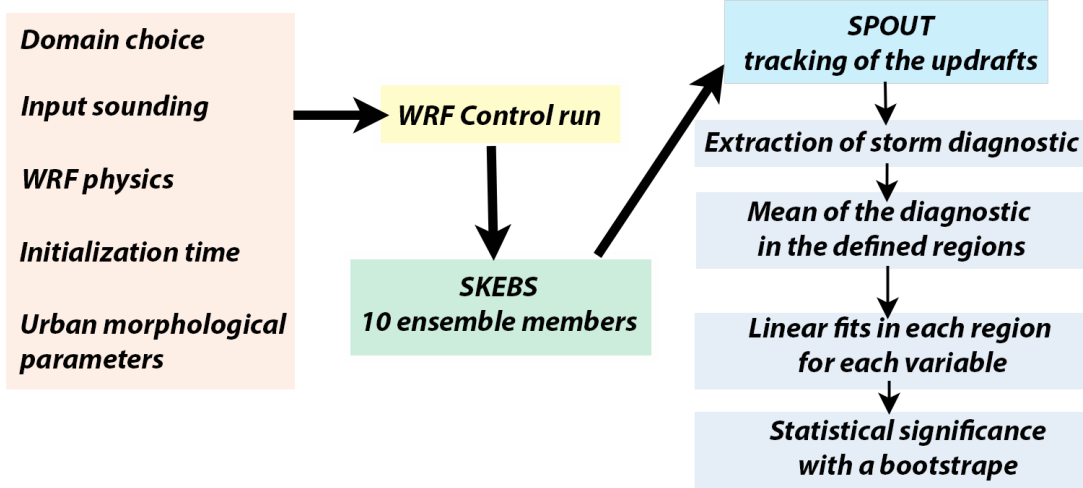


Figure 5.1: Summary of the method employed to produce and post-process every numerical experiment conducted.

The methodology used in the study is summarized in Figure 5.1 and described in detail in Section 5.1. The results of the first phase are presented in Section 5.2, while those of the second phase are shown in Section 5.3. Finally, the main outcomes of both phases are discussed in Section 5.4, with comparisons to existing literature.

5.1 Method

5.1.1 Numerical simulation setup

Numerical simulations are performed using version 4.6.1 of the [WRF](#) numerical model. In particular, the *sea_breeze_2d* idealized case, adapted to run with the [BEP-BEM](#) scheme in a three-dimensional domain (Zonato et al., 2021; Pappaccogli et al., 2020), is employed for the analysis. Numerical simulations are conducted in a single square domain with 1 km of horizontal resolution, consisting of 250×250 grid points and 64 vertical levels (the same vertical levels used in the simulations of Chapter 4). The domain is completely flat with croplands as uniform land use category. The domain is relatively large for an idealized simulation, as sufficient space is required for storm development upstream of the city. Additionally, space is needed downstream to study the effects of the city on the storm for at least one hour after it hits the urban area.

Another reason for choosing a large domain is the use of [Periodic Boundary Conditions \(PBCs\)](#). Although we tested open radiative boundary conditions, also combined with [PBCs](#) on two sides, these configurations resulted in an unphysical loss of mass once the storm developed (up to 10 hPa after a couple hours the storm initiation), a known problem with these boundary conditions. Therefore, [PBCs](#) are deemed more appropriate. A known limitation of [PBCs](#) is that, after storm initiation, gravity waves and outflow boundaries can exit through one boundary and re-enter from the opposite side, causing spurious waves interactions. To mitigate this, the domain size is chosen to ensure that such effects occur only after the period analyzed in this study.

The simulations are initialized at 10:00 am (local time). A deeper discussion regarding the choice of the timing of the simulation is provided in section 5.1.4. The supercell is initiated using a thermal bubble released at 04:00 pm near the bottom-left corner of the domain, at grid point (40, 40). The bubble is a thermal perturbation of +3 K released at

1800 m AGL. The perturbation relaxes to zero with a vertical radius of 1800 m and an horizontal radius of 10000 m. This setup allows the storm to develop for approximately two hours before reaching the city. This numerical configuration is similar to that used by Markowski and Dotzek (2011) to investigate interactions between idealized supercells and idealized orography.

The microphysics scheme used is Morrison et al. (2009), with hail as the prognostic rimed ice species, as this scheme is widely employed in idealized supercell simulations (Dennis and Kumjian, 2017; Peters et al., 2023a; Mulholland et al., 2024). Longwave radiation is parameterized using the Rapid Radiative Transfer Model (Mlawer et al., 1997), while shortwave radiation is parameterized using the Dudhia scheme (Dudhia, 1989). Radiative forcing corresponds to conditions at a latitude of 45°N (the same as Milan) on May 31st. Noah-MP is used as land surface model (Niu et al., 2011; Yang et al., 2011b), while the Revised MM5 scheme (Jiménez et al., 2012c) is used to resolve the surface layer processes.

5.1.2 Choice of the input sounding

An important step in setting up idealized simulations of deep moist convection is the selection of the sounding profile. A spin-up time is needed in the simulations to permit to the urban land use to modify the boundary layer. During this spin-up time deep moist convection that can alter the setting of the urban boundary layer is not desired. Some real soundings were initially tested, but the simulations produced either widespread uncontrolled storms or no storms at all. To achieve the desired controlled experiment, the Weisman-Klemp supercell sounding profile (Weisman and Klemp, 1982) was used, a common choice for supercell storm study (e.g., Markowski and Dotzek 2011). The advantage of this idealized profile is that the amount of latent instability can be easily changed adjusting some parameters. The vertical profile of potential temperature $\Theta(z)$ is defined as:

$$\Theta(z) = \begin{cases} \Theta_0 + (\Theta_{tr} - \Theta_0) \left(\frac{z}{z_{tr}} \right)^{\frac{5}{4}} & z \leq z_{tr} \\ \Theta_{tr} \exp \left[-\frac{g}{C_p T_{tr}} (z - z_{tr}) \right] & z > z_{tr} \end{cases} \quad (5.1)$$

where g is gravity acceleration, C_p the specific heat capacity of the atmosphere, $\Theta_0 = 296K$ the potential temperature at the surface, $z_{tr} = 12000m$ the height of the tropopause, $\Theta_{tr} = 343K$ the potential temperature at the tropopause and $T_{tr} = 219K$ its temperature. Those parameters can be modified to obtain the desired amount of latent instability.

The vertical profile of pressure $p(z)$ is also needed to initialize WRF. Obtaining $p(z)$ starting from Equation 5.1 is not straightforward: this computation is explicitly done in the following lines.

Consider the hydrostatic balance:

$$\frac{dp}{dz} = -\rho g \quad (5.2)$$

Replacing the density ρ with the equation of state yields:

$$\frac{dp}{dz} = -\frac{p}{RT} g \quad (5.3)$$

where R is the specific gas constant. Then, replacing the temperature T with the definition of potential temperature Θ gives:

$$\frac{dp}{p} = -\frac{g}{R\Theta} \left(\frac{p_0}{p} \right)^{\frac{R}{C_p}} dz \quad (5.4)$$

Rearranging:

$$dp = -\frac{g}{R\Theta} \frac{p_0^{\frac{R}{C_p}}}{p^{\frac{R}{C_p}-1}} dz \quad (5.5)$$

Thus:

$$p^{\frac{R}{C_p}-1} dp = -\frac{g}{R\Theta} p_0^{\frac{R}{C_p}} dz \quad (5.6)$$

Integrating between the levels p_1 and p_2 and the corresponding levels z_1 and z_2 :

$$\int_{p_1}^{p_2} p^{\frac{R}{C_p}-1} dp = -\int_{z_1}^{z_2} \frac{g}{R\Theta} p_0^{\frac{R}{C_p}} dz \quad (5.7)$$

Defining the mean Θ in the layer between z_1 and z_2 as $\bar{\Theta}$, it yields:

$$\frac{C_p}{R} \left(p_2^{\frac{R}{C_p}} - p_1^{\frac{R}{C_p}} \right) = -\frac{g}{R\bar{\Theta}} p_0^{\frac{R}{C_p}} (z_2 - z_1) \quad (5.8)$$

Hence:

$$p_2 = p_1 - \left[\frac{g}{R\bar{\Theta}} \frac{R}{C_p} p_0^{\frac{R}{C_p}} (z_2 - z_1) \right]^{\frac{C_p}{R}} \quad (5.9)$$

And ultimately:

$$p_2 = p_1 - p_0 \left(\frac{g}{C_p \bar{\Theta}} \right)^{\frac{C_p}{R}} (z_2 - z_1)^{\frac{C_p}{R}} \quad (5.10)$$

Since the pressure at the tropopause p_{tr} is known with the given parameters, $p(z)$ can be obtained going upward and downward from the tropopause level using $p_1 = p_{tr}$ and Equation 5.10.

At this point, the profiles of potential temperature and pressure of the input sounding are known. The profiles of specific humidity and wind still needs to be defined. The vertical profile of relative humidity $rh(z)$ is defined as:

$$rh(z) = \begin{cases} 1 - \frac{3}{4} \left(\frac{z}{z_{tr}} \right)^{\frac{3}{4}} & z \leq z_{tr} \\ 0.25 & z \geq z_{tr} \end{cases} \quad (5.11)$$

An exponent of 0.75 is used instead of the 1.25 employed in the original version of Weisman and Klemp (1982), in agreement with Markowski and Dotzek (2011). They used a smaller exponent to reduce the moisture above the PBL, to avoid complications related to a moist absolutely unstable layer. Since the aim of this work is similar to the Markowski and Dotzek (2011) study, their correction is implemented. The moisture in the boundary layer is defined fixing a maximum value for specific humidity $q_{bl} = 14.5 \text{ gkg}^{-1}$:

$$q(z) = \min(q_{bl}, q_{rh(z)}) \quad (5.12)$$

where $q_{rh(z)}$ is the specific humidity obtained from Equation 5.11.

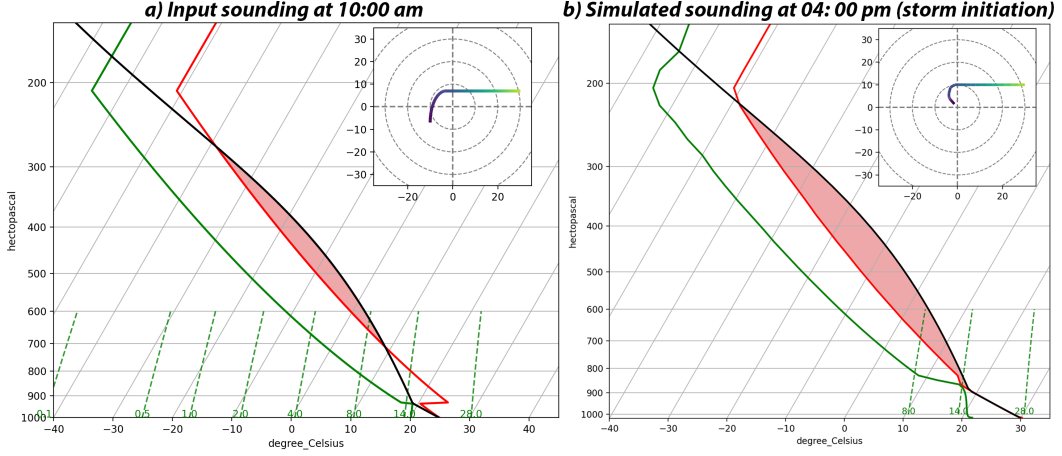


Figure 5.2: (a) input sounding at 10:00 am; (b) simulated sounding at 04:00 pm when the storm is initiated with a thermal bubble. Temperature (red line), dew point temperature (green line) and temperature of a surface parcel lifted adiabatically (black line) are shown. CAPE area is red shaded. The hodograph is shown in the upper right onset (the first 6 km of wind speeds are plotted).

Several tests were performed changing the moisture in the PBL (q_{bl}), and the other parameters controlling $rh(z)$. However, the desired setup (no spurious convection, and then a controlled isolated supercell storm) was never achieved. The issue is related to the radiative heating during the day that quickly makes the profile super-adiabatic in the lowest levels, triggering spurious convection. Indeed, the previously mentioned studies (Weisman and Klemp, 1982; Markowski and Dotzek, 2011) do not have the radiation scheme activated, while it is needed in this work to produce the urban boundary layer. To fix this issue an artificial thermal inversion is added in the profile. This modification aims to resemble the presence of an Elevated Mixed Layer (EML). High CAPE environments are often generated by the advection of dry and hot air mass from dry-lands toward moist boundary layers, or vice versa. This situation commonly precedes the development of severe storms both in Europe (Dahl and Fischer, 2016) and in the US (Andrews et al., 2024). This process generates a peculiar thermodynamic profile, characterized by a moist boundary layer, a thermal inversion at about 900-800 hPa, followed by a much drier layer with a steep lapse rate (see Figure 1 in Emanuel 2023). Thus the vertical profile of potential temperature $\Theta(z)$ is modified as follows:

$$\Theta_{EML}(z) = \begin{cases} \Theta(z) & z \leq h_{lcl} \\ \Theta(z) + 4K & z \geq h_{lcl} \end{cases} \quad (5.13)$$

where h_{lcl} is the height of the LCL. An increase of 4 K is chosen performing a sensitivity until the development of moist convection is completely avoided (on the other hand, a too strong thermal inversion results in the suppression of the convection triggered with the bubble). A thermal inversion of 4 K is compatible with real EML profiles (Tuckman and Emanuel, 2024). The relative humidity profile is also modified as follows:

$$rh(z) = \begin{cases} rh(z) & z \leq h_{lcl} \\ rh(z)/1.5 & z \geq h_{lcl} \end{cases} \quad (5.14)$$

The denominator 1.5 is chosen in order to decrease the mid-level moisture, without drying excessively the troposphere. This may cause a too strong entrainment of mid-level dry air in the updrafts, that it's preferable to avoid (Peters et al., 2023b). In Figure 5.2a is

CAPE	CIN	LCL	Bulk Shear 0-6 km	SRH 0-3 km	SRH 0-1 km
2740 Jkg^{-1}	0 Jkg^{-1}	1084 m	32.3 ms^{-1}	200 Jkg^{-1}	33 Jkg^{-1}

Table 5.1: Some relevant thermodynamic and kinematic indices computed on the simulated sounding at 04:00 pm shown in Figure 5.2b.

shown the resulting input profile, while in Figure 5.2b is shown how that profile evolves at 04:00 pm in the WRF simulation (6 hours after the starting of the simulation). The solar heating deepens the PBL, eroding the thermal inversion and creating high CAPE values, as expected by the conceptual model of Emanuel (2023). At this time the convective inhibition is negligible, and convection can grow once a lifting is provided. In the previous six hours deep moist convection is suppressed by the thermal inversion, permitting the generation of the urban boundary layer.

The environmental wind profile is defined by the quarter-circle hodograph proposed by Rotunno and Klemp (1982). The profiles of the zonal $u(z)$ and meridional $v(z)$ wind components used for the current study, and plotted in the hodograph of Figure 5.2a, are defined as:

$$u(z) = \begin{cases} u_1 - u_1 \cos\left(\frac{\pi z}{2h_1}\right) - \frac{u_2}{4} & z < h_1 \\ u_1 + (z - h_1) \frac{u_2 - u_1}{h_2 - h_1} - \frac{u_2}{4} & h_1 \leq z < h_2 \\ \frac{3}{4}u_2 & z \geq h_2 \end{cases} \quad (5.15)$$

and

$$v(z) = \begin{cases} v_1 \sin\left(\frac{\pi z}{2h_1}\right) & z < h_1 \\ v_1 & z \geq h_1 \end{cases} \quad (5.16)$$

where $h_1 = 1500$ m, $h_2 = 6000$ m, $u_1 = 7$ ms^{-1} , $u_2 = 40$ ms^{-1} , $v_1 = 10$ ms^{-1} are chosen to produce a right-mover supercell (Markowski and Dotzek 2011 used similar parameters). The diurnal mixing in the boundary layer changes the imposed wind profiles, as can be inferred comparing the hodographs at 10:00 am and 04:00 pm in Figure 5.2. This is also a consequence of the lack of a meridional thermal gradient in the domain, that hinders the presence of a thermal wind capable of maintaining the curved hodograph in long-lasting simulations. The transient evolution of the simulated profile could be an issue that it is further discuss in section 5.1.4. Values of some relevant parameters computed on the simulated sounding at 04:00 pm, when convection is initiated, are shown in Table 5.1. At 04:00 pm the atmosphere is conducive to supercell storms, since the abundant, albeit not extreme, CAPE (2740 Jkg^{-1}), strong bulk shear 0-6 km (32.3 ms^{-1}) and moderate SRH in the 0-3 km layer (200 Jkg^{-1}).

5.1.3 Sensitivity to the boundary layer scheme and soil moisture

The choice of the PBL scheme is carefully investigated. Several schemes are tested, including the 1.5 TKE closure ($km_{opt} = 2$ in the WRF namelist), coupled with either the BouLac, YSU, or MYJ schemes. Another parameterization, which is self-adaptive to grid-size changes between LES and mesoscale limits, is also tested ($km_{opt} = 5$ in the WRF namelist, Zhang et al. 2018). No remarkable differences are observed among the PBL schemes considered (e.g., Fig. 5.3), and the YSU scheme is preferred because it is widely used in severe storm studies at mid-latitudes (e.g., Reames and Stensrud 2018; Pilguy et al. 2019).

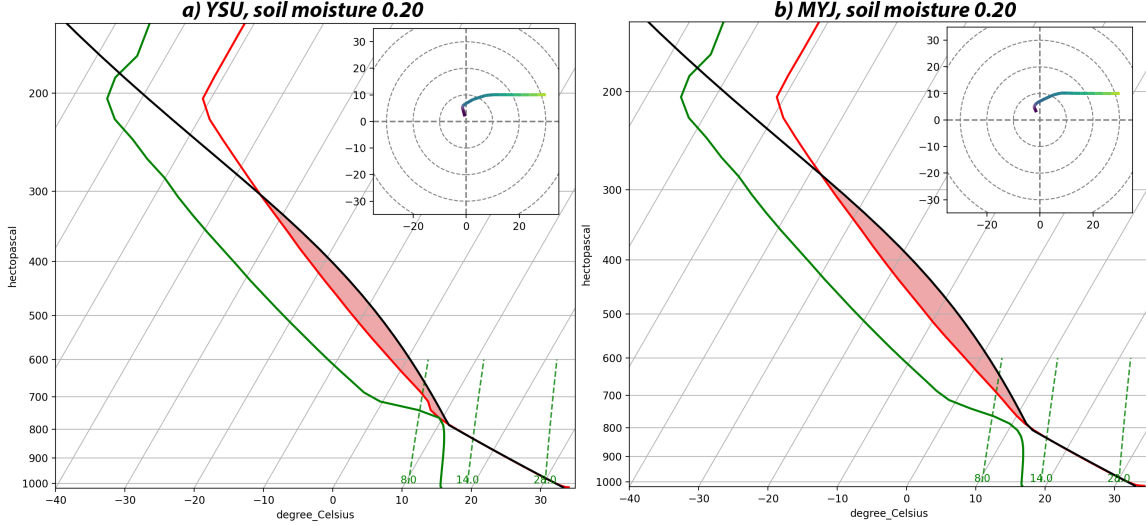


Figure 5.3: (a) Simulated sounding at 04:00 pm using the [YSU PBL](#) scheme with a soil moisture of 0.20; (b) simulated sounding at 04:00 pm using the [MYJ PBL](#) scheme with a soil moisture of 0.20. Temperature (red line), dew point temperature (green line) and temperature of a surface parcel lifted adiabatically (black line) are plotted. [CAPE](#) area is red shaded. The hodograph is shown in the upper right onset (the first 6 km of wind speeds are shown).

One parameter that significantly affects the growth of the [PBL](#) is soil moisture. Using a value of 0.20 instead of 0.34 (that corresponds to an almost saturated soil) leads to an excessively deep and dry [PBL](#), which is not ideal for supercell storms (compare Fig. 5.3a with Fig. 5.2b). The combination of the [YSU PBL](#) scheme and a soil moisture value of 0.34 provides, after six hours of spin-up time, the sounding profile that most closely resembles the one required to generate a supercell storm in an idealized study (Weisman and Klemp, 1982). For this reason, that soil moisture value and that [PBL](#) scheme are employed in this study.

5.1.4 Initialization time

The initialization time is carefully examined as well. After several attempts, it is decided to initialize the simulations at 10:00 am and to trigger the storm at 04:00 pm, a typical initiation time of convection (Manzato et al., 2022). In the first four hours of the simulation, [SB CAPE](#) rapidly increases from 1250 Jkg^{-1} to 3000 Jkg^{-1} (Fig. 5.4a) due to radiative heating. At 02:00 pm, mixing begins to reduce surface moisture, and consequently, [SB CAPE](#) gradually decreases, reaching a minimum between 05:00 pm and 08:00 pm. Due to the deepening of the [PBL](#), the [LCL](#) also rises quickly during the initial hours of the simulation, reaching a maximum between 05:00 pm and 08:00 pm (Fig. 5.4b). Regarding the kinematics, [SRH](#) decreases in both the 0-3 km (Fig. 5.4c) and 0-1 km (Fig. 5.4d) layers, reaching a minimum around 05:00 pm in the 0-3 km layer and around 03:00 pm in the 0-1 km layer. This decrease is due to the adjustment of the initial wind profile to the thermal wind.

These evolutions suggest that the simulation becomes steady between 05:00 pm and 08:00 pm. Therefore, triggering the storm at 04:00 pm and analyzing its interaction with the urban area between 05:00 pm and 07:00 pm is an appropriate choice. Any modifications observed in the numerical experiments in that time-span can be attributed to the urban morphological parameters rather than to changes in the broader environment.

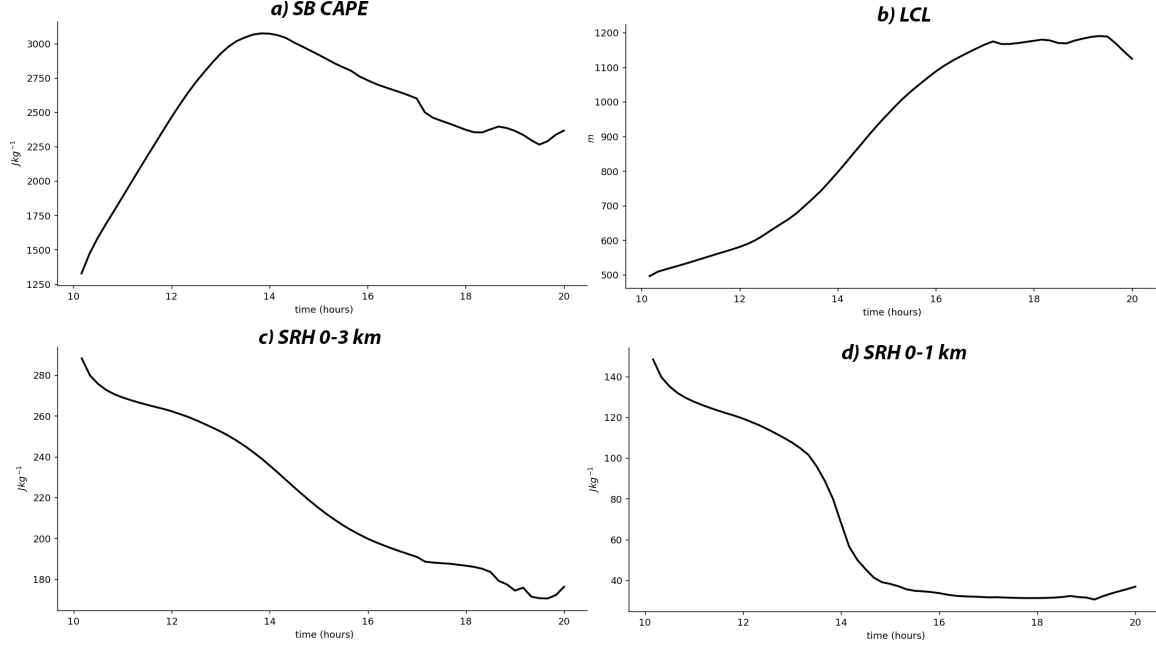


Figure 5.4: Time evolution of domain-averaged (a) **SB CAPE**, (b) **LCL**, (c) **SRH** in the 0-3 km layer, and (d) **SRH** in the 0-1 km layer. No convection is triggered in the simulation considered to produce these plots.

5.1.5 Representation of the city in the simulation domain

A circular city is centered at grid point (90,76). This location is chosen based on an analysis of the supercell evolution in the *no_urban* control run, where the land use is cropland everywhere. The goal is to locate the city such that the supercell passes over its center two hours after initiation, coinciding with the time of peak intensity. Urban surface processes are resolved using the **BEP-BEM** model, the most advanced parameterization of urban processes in mesoscale modeling (e.g., Ribeiro et al. 2021). The **BEP-BEM** model requires several urban morphological parameters to run. The sensitivity of the supercell to some of them is investigated in this work: street width (s), building width (b), building height h_b , and urban fraction (urbfrc). The **GLOBUS** dataset (Kamath et al., 2024) is used to obtain realistic values of these parameters to generate the idealized city. In that dataset, the plan area fraction (λ_p) and the building surface-to-plan area ratio (λ_b) are reported.

city	size (km)	h_b (m)	s (m)	b (m)	urbfrc	λ_b	λ_p
Chicago	30x15	9.2	17.1	8.7	0.68	0.72	0.23
Oklahoma City	30x15	8.4	26.4	9.1	0.42	0.30	0.10
New York City	26x15	19.9	30.2	14.6	0.62	0.75	0.20
London	20x20	11.8	17.7	9.7	0.55	0.66	0.19
Dallas	19x12	11.3	25.5	12.9	0.47	0.43	0.16
Berlin	15x10	20	10.4	8.7	0.52	1.32	0.23
Milan	14x7	18.6	26.8	16.6	0.63	0.78	0.24
New York City downtown	12x6	44.6	41.6	21.8	0.65	1.12	0.22
Kansas City	10x7	11.9	50.8	14.9	0.49	0.29	0.11
Atlanta	8x5	20.5	31.2	17.5	0.51	0.61	0.18
Chicago downtown	2x2	114.5	49.76	29.5	0.75	2.45	0.28

Table 5.2: Mean urban morphological parameters sourced from the **GLOBUS** dataset of some representative mid-latitude European and US cities frequently affected by severe storms.

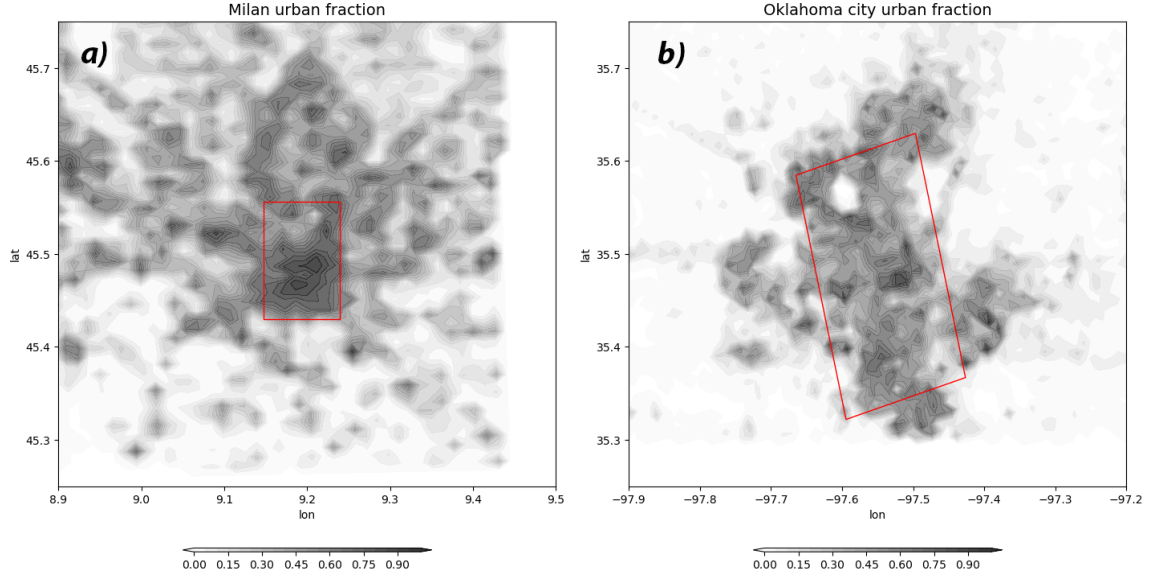


Figure 5.5: Urban fraction of (a) Milan and (b) Oklahoma city. The red boxes show the area within where the [GLOBUS](#) parameters are averaged.

These are defined as:

$$\lambda_p = \frac{b}{s + b} \quad (5.17)$$

and:

$$\lambda_b = \frac{2h_b}{s + b} + \lambda_p. \quad (5.18)$$

s and b can be obtained from λ_p and λ_b knowing h_m with the following equations:

$$b = \frac{2h_b\lambda_p}{\lambda_b - \lambda_p} \quad (5.19)$$

$$s = \frac{2h_b\lambda_p}{\lambda_b - 1} \frac{1}{\lambda_b - \lambda_p}. \quad (5.20)$$

Since the aim of this study is to perform a sensitivity analysis on supercell evolution based on urban morphological parameters, the idealized city is defined using uniform values. These constant values are obtained by averaging the [GLOBUS](#) data for selected real cities within a defined region (see two examples in Fig. 5.5). The chosen cities serve solely as a representative sample of large mid-latitude European and U.S. cities that are affected by severe storms. The averaged morphological parameters are presented in Table 5.2, while how they are used to perform the numerical study is clarified in section 5.1.8.

5.1.6 Generation of the ensembles

For each numerical experiment, eleven simulations are conducted: one is the control run, the other ten simulations are generated using the [Stochastic Kinetic Energy Backscatter Scheme](#) (SKEBS) scheme. SKEBS is employed to represent energetic contributions from unresolved physical processes by introducing stochastic perturbations to temperature and streamfunction (Berner et al., 2009). In the present study, the parameters used to

Backscatter rate for streamfunction	10^{-5}
Backscatter rate for temperature	10^{-6}
Decorrelation time of noise	3h
Power law for streamfunction perturbations	-1.83
Power law for temperature perturbations	-1.83
Variance of noise	0.0833
Maximum forcing wavenumber	250
Vertical structure for random perturbation field	0 (constant)

Table 5.3: Parameters set to generate the ensemble of WRF simulations with SKEBS.

generate the ensemble with SKEBS (Table 5.3) are set following Romine et al. (2014), who investigated ensemble dispersion in deep moist convective events. The only difference from their setup is the value of the maximum wavenumber, which is set to match the number of grid points in the current simulations (250), as recommended in the WRF User Guide¹.

The ten ensemble members are generated using ten distinct random seeds, which are kept constant across all numerical experiments. As a result, experiments using the same seeds are not fully independent. However, this does not pose a problem: the statistical analysis accounts for this by computing linear fits across simulations sharing the same seed (further details are provided in Section 5.1.9).

5.1.7 Storm tracking

The Statistical and Programmable Objective Updraft Tracker (SPOUT) algorithm is used to track the simulated updrafts. It was originally developed by Terwey and Rozoff (2014) for numerical simulations of idealized tropical convection. Several user-defined parameters can be set when running SPOUT. Fine-tuning of these parameters was necessary to achieve satisfactory tracking of the supercell updrafts. Table 5.4 presents the parameter values used in this study and compares them with those employed by Terwey and Rozoff (2014), while in the next paragraphs the algorithm’s operation is described.

SPOUT identifies updrafts by searching for vertical velocity values that exceed a defined threshold, referred to as *checkValues*, starting from a minimum vertical level (*MINUPLEVEL*) and extending upward for a user-defined number of vertical levels (*NUMLEVELS*). Thus, it analyzes the range from *MINUPLEVEL* to *MINUPLEVEL* + *NUMLEVELS*. An updraft is considered valid if the vertical velocity exceeds the *checkValues* threshold for at least *MINUPHEIGHT* vertical levels within a horizontal radius defined by *SPACETHRES*.

Once detected, the updraft at a given time step is advected using the environmental wind, with a correction applied through the *velCorrected* parameter. This correction is included since it is well-known that thunderstorms can often move at fractions of the imposed advective velocities, especially when considering supercells. If, at the advected location, the same thresholds are met within a horizontal radius defined by *RANKTHRES* for at least *POINTTHRES* vertical levels, the updraft at the new time step is considered to be the same as the previous one, and the two are linked. An updraft must persist for at least *OUTPUTTIMETHRES* consecutive time steps to be retained in the output.

The outputs of this sophisticated algorithm require further filtering to isolate the supercell track. In practice, it was observed that the algorithm frequently splits the supercell track. This occurs because SPOUT tracks updrafts individually, and supercells may cycle,

¹https://www2.mmm.ucar.edu/wrf/users/wrf_users_guide/build/html/namelist_variables.html

Parameter name	Brief description	Terwey and Rozoff (2014)	this study
MINUPLEVEL	First level to be scanned	6 (1255 m)	20 (901 m)
WTHRES	Threshold value for checking by how much a local maximum must exceed its neighbors	0	0
NUMLEVELS	How many vertical gridpoints the code will be searching for updrafts starting from MINUPLEVEL	20	30
MINUPHEIGHT	Minimum updraft height to be considered (in vertical levels)	15	10
SPACETHRES	Spatial threshold that must be satisfied for the vertical linking of updrafts	4 km	3 km
RANKTHRES	Spatial threshold that must be satisfied for the temporal linking of updrafts	10 km	7 km
POINTTHRES	Vertical spatial threshold that must be satisfied for the temporal linking of updrafts (in vertical levels)	6	3
OUTPUTTIMETHRES	Minimum number of time steps that a tracked updraft must be to be considered	3 (18 min)	2 (20 min)
velCorrect	Correction factor for the advective process in temporal linking	1.0	0.35
checkValues	Minimal vertical velocity for each level's maxima to be qualified for updraft status	1.5 m s^{-1}	1.0 m s^{-1}

Table 5.4: Values of [SPOUT](#) parameters set to track the updrafts in the Terwey and Rozoff (2014) work and in the present study.

variable	meaning
wmax	maximum vertical velocity
w500	maximum vertical velocity at 500 hPa
ζ 500	maximum vertical vorticity at 500 hPa
uh max	maximum updraft helicity
uh mean	mean updraft helicity
ζ 700	maximum vertical vorticity at 700 hPa
w850	maximum vertical velocity at 850 hPa
ζ 850 max	maximum vertical vorticity at 850 hPa
w950	maximum vertical velocity at 950 hPa
ζ 950	maximum vertical vorticity at 950 hPa
Θ_e (500 hPa) max	maximum equivalent potential temperature at 500 hPa
Θ_e (500 hPa) mean	mean equivalent potential temperature at 500 hPa
Θ (5 m) min	minimum potential temperature in the first model level
Θ (5 m) mean	mean potential temperature in the first model level

Table 5.5: Storm diagnostic variables used in the analysis. Every values is computed considering a 8 km radius following the storm track.

producing new updrafts. Additionally, other storms are triggered in the simulations due to the left-moving supercell and the outflows.

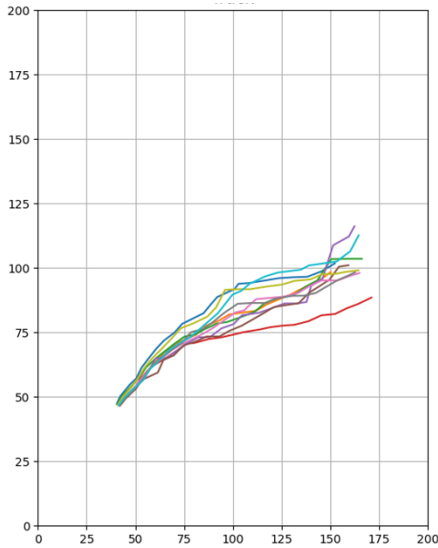


Figure 5.6: Example of the supercell tracks as detected by SPOUT and the further filtering. The ten tracks obtained with the ensemble generated by SKEBS and the track of the control run are plotted.

To filter out these updrafts, only the track that begins near the location where the thermal bubble is released is considered. If the track terminates at $x < 150$ (i.e., before three hours after the bubble release), a new track is searched for within a 10 km radius. If found, the new track is connected to the previous one. This process is repeated until the track reaches $x = 150$ or the storm dissipates. An example of the eleven tracks obtained for the eleven ensemble members using this tracking algorithm is shown in Figure 5.6.

Each point along the track is considered the storm center at a given time step. Storm diagnostics are extracted every 10 minutes, and the extracted variables are listed in Table 5.5. In the Results section special emphasis is given to the UH because it is the most representative diagnostic variable for supercell storms. For each time step, the mean (or maximum or minimum, depending on the variable) is calculated within a 8 km radius around the storm center, and this value is assigned to the storm at that time step.

Then, three regions are defined, similarly to the study in Chapter 4. However, in this case the regions are defined with a temporal criterion (instead of a spatial one), because the interest is to compare the storms at the same developing stage in different numerical experiments. In any case, the temporal criterion can be easily transformed in a spatial one since at a given time step the storm is in almost the same location in every simulation (if it is not dissipated). Those regions are defined with respect to a zero, defined as the time step when the supercell is above the city, corresponding to grid point (90,76) in the control run of the *no_urban* experiment. This time step is at 06:00 pm (see for instance Fig. 5.7d).

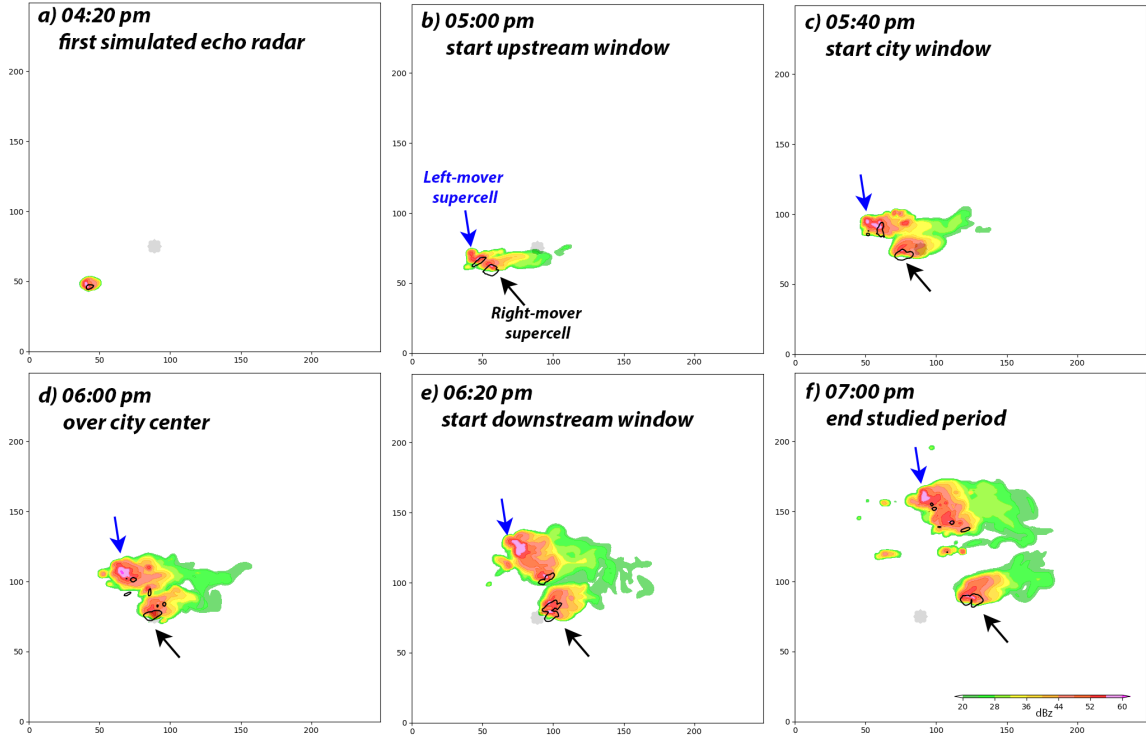


Figure 5.7: Maximum simulated reflectivity (colour shading) at a) 04:20 pm, b) 05:00 pm, c) 05:40 pm, d) 06:00 pm, e) 06:20 pm, f) 07:00 pm by the control run of numerical experiment *10km_1*. The city is shown with a grey shading. UH value higher than 100 m s^{-1} are highlighted with a black contour. The black (blue) arrow indicates the right-mover (left-mover) supercell.

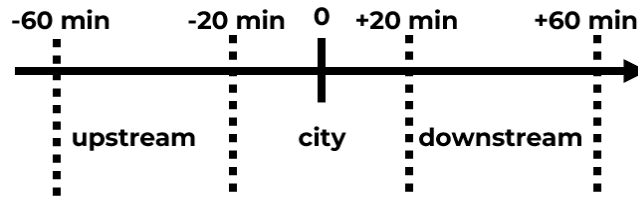


Figure 5.8: Criterion used to define the upstream, city and downstream regions.

Then, three 40 minutes-windows are defined (sketch in Figure 5.8). The first one, from -20 to +20 minutes, defines the *city region*. The *upstream region* is defined from -60 to -20 minutes. Lastly, the *downstream region* is defined from +20 minutes to +60 minutes. In these windows the variables previously computed for each time steps are averaged. Thus, considering the 10 minutes output frequency, four time steps are averaged for each window.

5.1.8 Numerical experiments conducted

First, a sensitivity analysis on city size is conducted, with city diameters ranging from the *no_urban* scenario to a city with a diameter of 30 km (numerical experiments *city size 1*, Table 5.6). In Figure 5.9 a schematic representation of the different configurations of urban morphological parameters is shown (the first row refers to the city size case). Table 5.7 presents the parameters used to define the idealized city in these numerical experiments; these are based on the Milan GLOBUS data reported in Table 5.2. Milan is

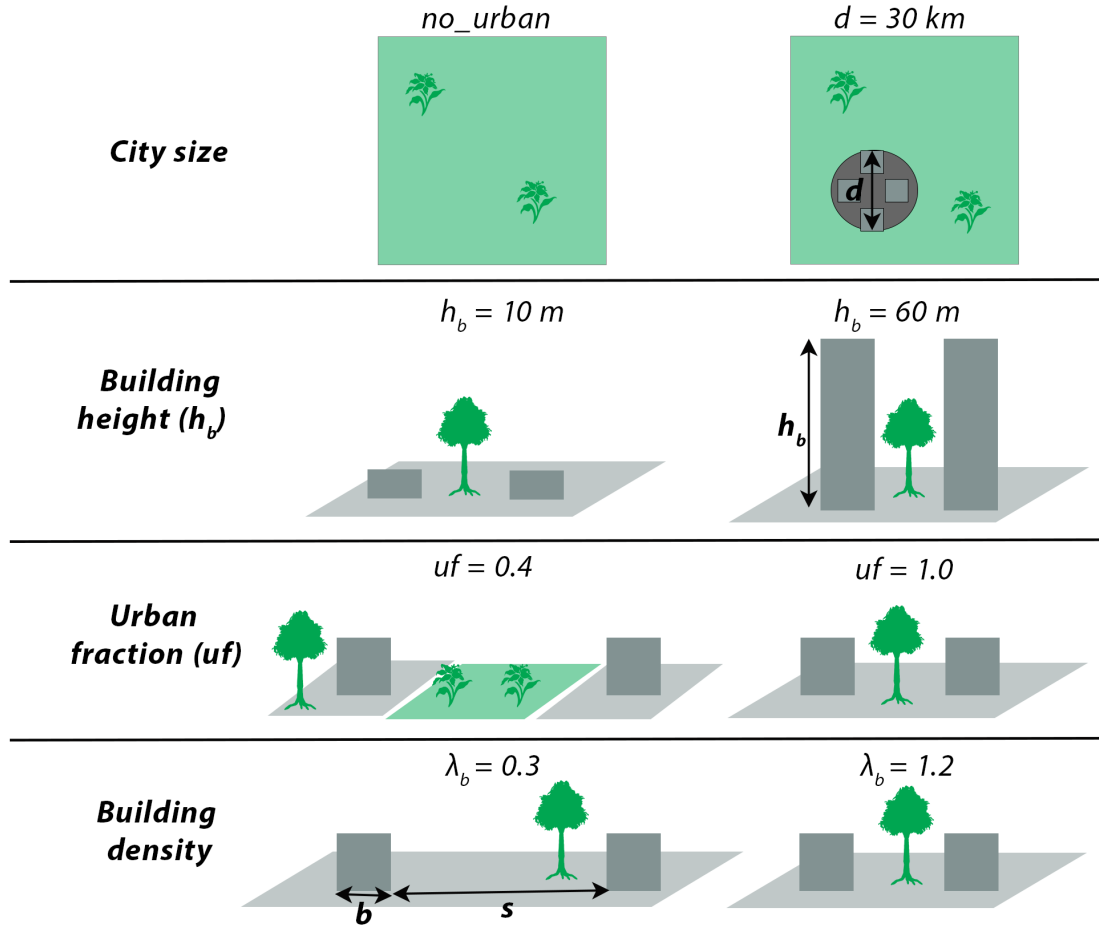


Figure 5.9: Schematic representation of some of the urban features explored in the idealized simulations. The *zero value* (left) and the *max value* (right) of the studied urban morphological parameters (city size, building height, urban fraction, and building density) are shown (see Table 5.6). In the first row, the simulation domain is depicted from above, showing croplands (green shading with plants) and the city (grey circle). In the remaining rows, the view is lateral. Grey rectangles represent buildings. d denotes the city diameter, h_b the building height, uf the urban fraction, s the street width, b the building width, and λ_b the ratio of building area to total area.

numerical experiments	zero value	intermediate values	max value
<i>city size 1</i> (km)	0 (no_urban)	5; 10; 15; 20; 25	30
<i>city size 2</i> (km)	0 (no_urban)	5; 10; 15; 20; 25	30
<i>urban fraction</i>	0.4	0.6; 0.8	1.0
<i>building height</i> (m)	10	20; 40	60
<i>building density</i> (λ_p, λ_b)	(0.1,0.3)	(0.2, 0.6) ; (0.3,0.9)	(0.4,1.2)
<i>green roofs</i>	no_urban	green roofs over all roofs	
<i>PV</i>	no_urban	PV panels over all roofs	
<i>NYC</i>	no_urban	New York City parameters	
<i>bubble</i>	no_urban	bubble shifted from (40,40) to (30,60)	

Table 5.6: Numerical experiments conducted and parameters explored.

	no_urban	5km_1	10km_1	15km_1	20km_1	25km_1	30km_1
city diameter (km)	0	5	10	15	20	25	30
urban fraction	0.63	0.63	0.63	0.63	0.63	0.63	0.63
building height (m)	20	20	20	20	20	20	20
λ_p	0.24	0.24	0.24	0.24	0.24	0.24	0.24
λ_b	0.78	0.78	0.78	0.78	0.78	0.78	0.78

Table 5.7: Urban morphological parameters used in the set of numerical experiments *city size 1*.

chosen as the reference city both because the study in Chapter 4 is based on it, and also because its morphological parameters are relatively typical of European cities (Table 5.2).

The same set of numerical experiments is also performed using an urban fraction of 1.0 instead of 0.63 (numerical experiments *city size 2*, Tables 5.6 and 5.8). An urban fraction of 1.0 means the entire urban surface is built-up (i.e., no croplands within the city radius, see third row in Figure 5.9). However, 20% of the street surface remains covered by trees in these experiments (Fig. 5.9). This condition is maintained in all simulations, as a city without any trees is unrealistic. Although a mean urban fraction of 1.0 is not observed in European or U.S. cities (Table 5.2), this experiment is conducted to explore the impact of city size under the extreme scenario of a fully urbanized city.

	no_urban	5km_2	10km_2	15km_2	20km_2	25km_2	30km_2
city diameter (km)	0	5	10	15	20	25	30
urban fraction	1.0	1.0	1.0	1.0	1.0	1.0	1.0
building height (m)	20	20	20	20	20	20	20
λ_p	0.24	0.24	0.24	0.24	0.24	0.24	0.24
λ_b	0.78	0.78	0.78	0.78	0.78	0.78	0.78

Table 5.8: Urban morphological parameters used in the set of numerical experiments *city size 2*.

	uf0.4	uf0.6	uf0.8	uf1.0
city diameter (km)	20	20	20	20
urban fraction	0.4	0.6	0.8	1.0
building height (m)	20	20	20	20
λ_p	0.24	0.24	0.24	0.24
λ_b	0.78	0.78	0.78	0.78

Table 5.9: Urban morphological parameters used in the set of numerical experiments *urban fraction* (uf).

	h10	h20	h40	h60
city diameter (km)	20	20	20	20
urban fraction	0.63	0.63	0.63	0.63
building height (m)	10	20	40	60
λ_p	0.24	0.24	0.24	0.24
λ_b	0.78	0.78	0.78	0.78

Table 5.10: Urban morphological parameters used in the set of numerical experiments *building heights* (h_b).

A third set of numerical experiments, called *urban fraction*, is conducted to study the sensitivity of the supercell to urban fraction. Values ranging from 0.4 to 1.0 are investigated for an idealized city with a diameter of 20 km and the other urban morphological parameters based on Milan (Table 5.9). This range is selected to encompass all the cities analyzed in Table 5.2, from Oklahoma City (urban fraction 0.42) to the extreme case of a fully urbanized city.

A fourth set of numerical experiments, called *building height*, is conducted to investigate the effect of building height on the simulated storm (Table 5.10). The building heights explored range from 10 to 60 m (sketch in Figure 5.9), representing cities from Chicago metro to downtown New York. Although the average building height in downtown Chicago reaches 114.5 m, this applies only to a small 2×2 km area (Table 5.2). As in the previous experiments, all other urban morphological parameters are kept consistent with those of Milan.

The fifth set of numerical experiments, called *building density*, explores the effect of

	λ_b 0.3	λ_b 0.6	λ_b 0.9	λ_b 1.2
city diameter (km)	20	20	20	20
urban fraction	0.63	0.63	0.63	0.63
building height (m)	20	20	20	20
λ_p	0.1	0.2	0.3	0.4
λ_b	0.3	0.6	0.9	1.2

Table 5.11: Urban morphological parameters used in the set of numerical experiments *building density*.

building density on the simulated storms (Table 5.11). In these experiments, building height is fixed at 20 m. Since the **BEP-BEM** scheme uses street and building widths as input parameters (rather than the λ parameters), it was chosen to keep the building width constant at 20 m, while varying the street width from 106 m to 11.5 m. This corresponds to a range of (λ_p, λ_b) values from (0.1, 0.3) to (0.4, 1.2), representing a transition from a very dispersed to a highly compact city (see the last row in Figure 5.9 for a schematic representation). This range covers the full spectrum of building densities observed in the cities analyzed using GLOBUS (Table 5.2).

Lastly, some additional experiments are conducted to investigate other specific questions. An experiment with green roofs (called *green roofs*, Table 5.6) and a second one with **photovoltaic** (PV) panels (called *PV*, Table 5.6) over all the roofs of the idealized city are carried out to understand the effect of these surfaces on storm dynamics. Both experiments are run with the settings of numerical experiment *30km_2* (described in Table 5.8). The parameterization used for green roofs and PVs are the ones developed in Zonato et al. (2021). Another experiment, called *NYC* (Table 5.6), is run to test the hypothesis of bifurcation of storms caused by the building barrier effect, proposed for the first time by Bornstein and Lin (2000) for **New York City** (NYC). This experiment is run using a city diameter of 8 km and the same parameters of downtown NYC shown in Table 5.2. An experiment, called *bubble* (Table 5.6), is also run to test the sensitivity of the urban-induced storm modification to the storm trajectory. In this experiment the thermal bubble is shifted from grid-point (40,40) to (30,60), while keeping the same parameter of experiment *30km_2* (Table 5.8).

5.1.9 Statistical method used to analyze the numerical experiments

For each urban morphological parameter studied, a number from 4 to 7 of experiments are conducted (Table 5.6). For each numerical experiment (described in section 5.1.8), 11 simulations are run. For each simulation, three values, one for each region (Fig. 5.8), of some diagnostic variables (Table 5.5) are extracted as explained in section 5.1.7.

Given a set of numerical experiments, a statistical analysis is conducted for each region and for each variable. For example, a statistical analysis is conducted for the *wmax* variable in the *upstream* region for the numerical experiments *city size 1*. The variables are normalized with respect to a reference zero value (defined case-by-case in Table 5.6). The normalized z_{ij} variable for the ensemble member j and numerical experiment i is given by:

$$z_{ij} = \frac{x_{ij} - x_{0j}}{x_{0j}} \quad (5.21)$$

where x_{ij} is the original variable and x_{0j} is the reference value for the ensemble member j . This approach is adopted because the interest is in understanding how storm diagnostics vary relative to a reference case (e.g., how updraft speed changes with increasing city size compared to the *no_urban* case). A linear fit is performed using the *curve_fit* function from SciPy (Virtanen et al., 2020) on the normalized variables. Any null values are excluded: specifically, if a storm dissipates, the corresponding ensemble member is omitted from the fit. This exclusion is necessary because, for some variables (e.g., Θ), including a null value in the linear fit lacks physical meaning. In general, excluding such values has been observed to increase the statistical significance of the fit.

We focus solely on the slope parameter of the linear fit, particularly its sign. The statistical significance of the slope is assessed using a 5-95% bootstrap confidence interval, obtained through 1000 iterations and computed with the *bootstrap* function of SciPy.

5.2 Results: supercell sensitivity to the urban morphological parameters

In this section, the results of the sensitivity analysis of the simulated supercell to urban morphological parameters are presented. Figure 5.7 shows the evolution of the storm in the control run of the *10km_1* experiment. At 04:20 pm (Fig. 5.7a), the first simulated radar echo appears in the numerical domain following the release of the thermal bubble at 04:00 pm. By 05:00 pm, the initial storm has split into right- and left-moving supercells (Fig. 5.7b). The well-known reason of the simulated storm splitting is provided in Appendix A.3. The upstream phase of the storm occurs between 05:00 pm and 05:40 pm (Fig. 5.7c), by the end of which the right-moving supercell has matured (as indicated by the relatively large area with UH values exceeding $100 \text{ m}^2 \text{ s}^{-2}$) and is approaching the city from the *WSW*. At 06:00 pm, the supercell is located over the city (Fig. 5.7d). By 06:20 pm (Fig. 5.7e), the downstream phase of the storm begins, ending at 07:00 pm (Fig. 5.7f), when the simulated supercell remains isolated and intense. Across the different ensemble members, there are variations in the storm tracks (e.g., Fig. 5.6) and intensity; however, all simulations of the *no_urban* reference experiment produce a long-lasting, right-moving supercell. The sensitivity of this storm to urban morphological parameters is discussed in the following sections.

5.2.1 City size 1

In the numerical experiments *city size 1*, the effect of increasing the diameter of the city with the same characteristics as Milan on the supercell is investigated. In Fig. 5.10a, each dot represents the normalized mean value of the storm's UH (calculated as described in Section 5.1.7) in the *upstream region* for each ensemble member as a function of city diameter. The mean linear fit and the associated confidence interval are also shown.

Overall, there is a clear decreasing trend of UH . However, a considerable ensemble spread is present: while most ensemble members exhibit a decreasing trend, some show an increasing one. A similar pattern is observed in the *city region* (Fig. 5.10c), whereas in the *downstream region* a more consistent decreasing trend is seen across all ensemble members (Fig. 5.10e).

Table 5.12 presents the mean slopes for each variable in the *city size 1* simulations in the three regions. A general negative trend is observed, though it is statistically significant only for some variables. The trend is particularly significant in the *city region* and for the vertical velocity and vorticity at 500 hPa, and for UH . This indicates that the supercell weakens as city diameter increases, with the weakening being more pronounced and statistically significant at the mid-levels and when the storm is over the city. The increasing trend showed by some ensemble members in Fig. 5.10a and 5.10b suggest that, even if rare, an intensification of the supercell is possible in the *upstream* and *city regions*.

Θ shows a positive trend. This is because the cold pools weaken as the storm weakens, leading to less pronounced negative Θ anomalies associated with downdrafts, i.e., a relative increase in Θ .

5.2.2 City size 2

In the *city size 2* experiments, a trend similar to the *city size 1* sample is observed, but the weakening of the storm is even more pronounced (Fig. 5.10). This weakening is particularly evident downstream of the city and, in this case, is statistically significant also for several low-level storm diagnostic variables (Table 5.13). A positive trend in vertical

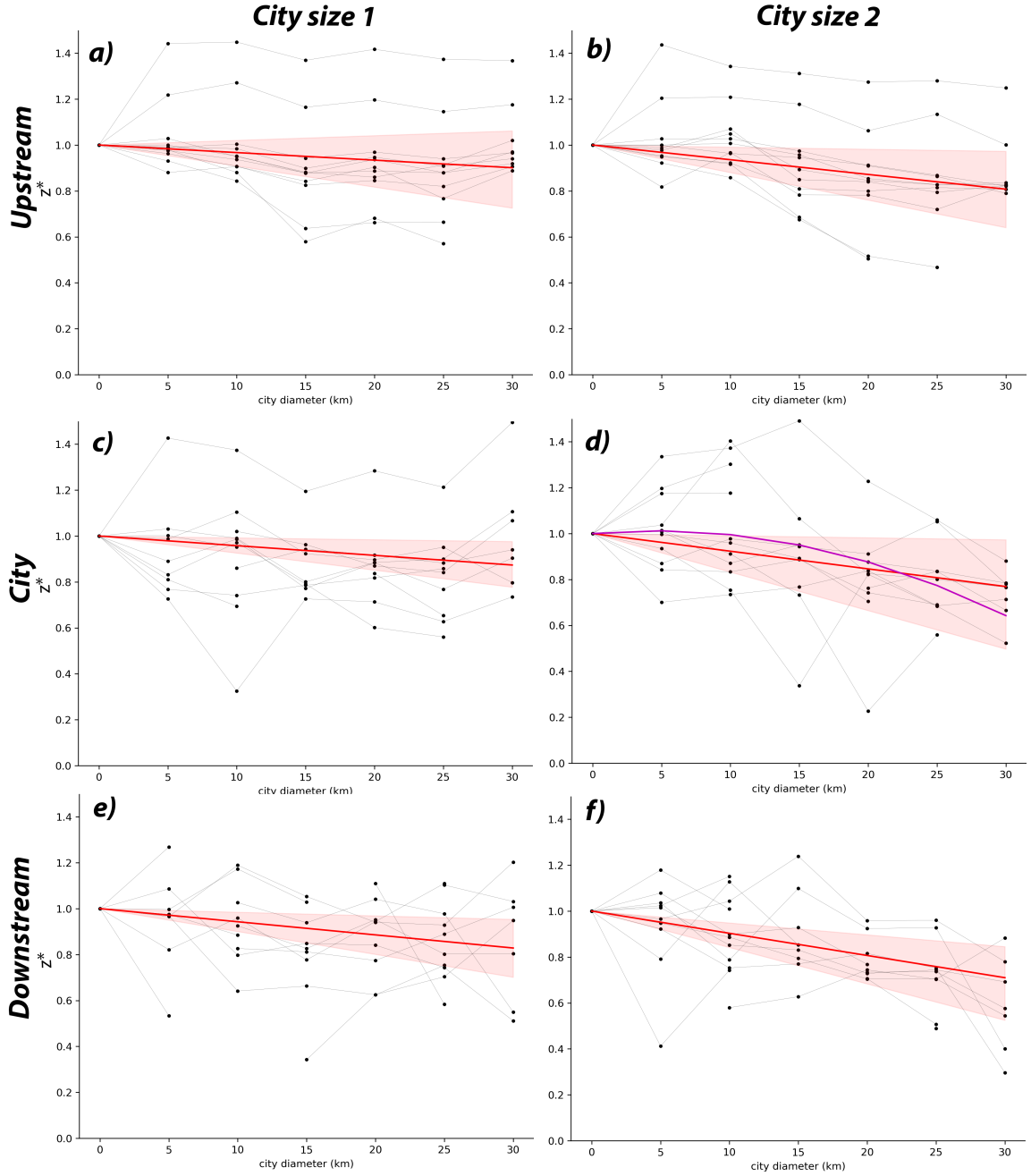


Figure 5.10: a) Normalized mean UH (z^*) in each ensemble member (black dots) of the *city size 1* set of experiments (Table 5.7) in function of the city diameter (km) in the *upstream* region. Ensemble members obtained with the same seed are connected. The mean linear fit is shown with a red line, while the 5-95% confidence interval is red shaded. c) and e) as in a) but for the *city* and *downstream* regions. b), d) and f) as in a), c) and e), respectively, but for the *city size 2* numerical experiments. In d) a quadratic fit is shown with a purple line.

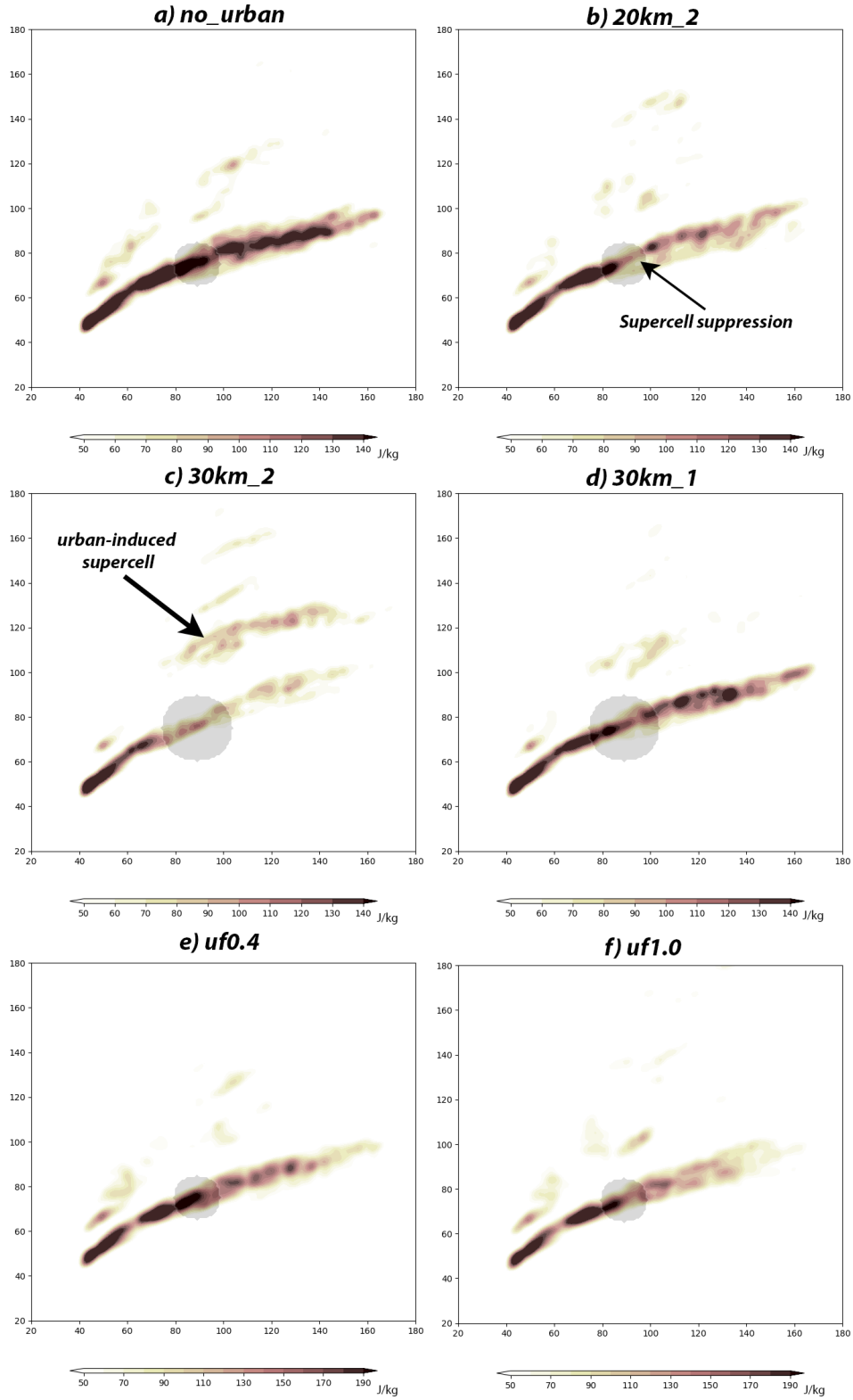


Figure 5.11: Cumulated UH ensemble mean in a) the *no_urban* experiment (Table 5.8), b) in the *20km_2* experiment (Table 5.8), c) in the *30km_2* experiment (Table 5.8), d) in the *30km_1* experiment (Table 5.7), e) in the *uf0.4* experiment (Table 5.9), f) in the *uf1.0* experiment (Table 5.9). The city is highlighted with a grey shading. In a) the city of the *20km* experiment is shown just for reference, in the experiment there are croplands everywhere.

variable	upstream	city	downstream
wmax	-2.6×10^{-3}	-2.1×10^{-3}	-1.1×10^{-3}
w500	-6.1×10^{-3}	-3.1×10^{-3}	-1.8×10^{-3}
ζ 500	-3.9×10^{-3}	-2.0×10^{-3}	-2.6×10^{-3}
uh max	-3.1×10^{-3}	-4.6×10^{-3}	-3.7×10^{-4}
uh mean	-3.3×10^{-3}	-4.2×10^{-3}	-5.7×10^{-3}
w700	-1.4×10^{-3}	-1.9×10^{-3}	-2.8×10^{-4}
ζ 700	$+3.8 \times 10^{-4}$	-8.7×10^{-4}	-6.5×10^{-4}
w850	-1.3×10^{-3}	-3.6×10^{-3}	-2.0×10^{-3}
ζ 850	-1.5×10^{-4}	$+2.1 \times 10^{-4}$	-5.7×10^{-4}
w950	$+2.1 \times 10^{-5}$	-5.4×10^{-3}	-3.7×10^{-3}
ζ 950	-7.8×10^{-4}	-2.6×10^{-3}	-5.1×10^{-3}
Θ_e (500 hPa)	-6.3×10^{-5}	-1.9×10^{-4}	-7.1×10^{-6}
Θ_e (500 hPa) mean	-3.7×10^{-5}	-3.4×10^{-5}	-1.2×10^{-5}
Θ (5 m) min	-2.9×10^{-5}	$+5.4 \times 10^{-5}$	$+8.1 \times 10^{-5}$
Θ (5 m) mean	-1.6×10^{-5}	$+4.9 \times 10^{-5}$	$+2.1 \times 10^{-5}$

Table 5.12: Slope (km^{-1}) of linear fitting lines between some diagnostic variables of the supercell simulated with *city size 1* numerical experiments (Table 5.7), and city diameter (km). Bolded values are statistically significant.

variable	upstream	city	downstream
wmax	-3.9×10^{-3}	-3.7×10^{-3}	-3.7×10^{-3}
w500	-8.1×10^{-3}	-8.7×10^{-3}	-4.1×10^{-3}
ζ 500	-4.8×10^{-3}	-4.3×10^{-3}	-4.3×10^{-2}
uh max	-4.1×10^{-3}	-6.8×10^{-3}	-3.3×10^{-3}
uh mean	-6.4×10^{-3}	-7.7×10^{-3}	-9.7×10^{-3}
w700	-1.8×10^{-3}	-1.8×10^{-3}	-3.1×10^{-3}
ζ 700	-7.8×10^{-4}	$+3.45 \times 10^{-5}$	-2.7×10^{-3}
w850	-2.4×10^{-3}	-4.3×10^{-3}	-3.4×10^{-3}
ζ 850	-8.1×10^{-4}	$+7.1 \times 10^{-4}$	-7.2×10^{-3}
w950	-9.1×10^{-4}	-6.1×10^{-3}	-5.6×10^{-3}
ζ 950	$+6.6 \times 10^{-4}$	-4.7×10^{-3}	-3.7×10^{-3}
Θ_e (500 hPa) max	-2.6×10^{-2}	-2.4×10^{-4}	-1.3×10^{-4}
Θ_e (500 hPa) mean	-9.6×10^{-5}	-5.5×10^{-5}	-7.1×10^{-5}
Θ (5 m) min	-8.9×10^{-7}	$+1.1 \times 10^{-4}$	$+1.7 \times 10^{-4}$
Θ (5 m) mean	-1.0×10^{-5}	$+7.5 \times 10^{-5}$	$+7.2 \times 10^{-5}$

Table 5.13: Slope (km^{-1}) of linear fitting lines between some diagnostic variables of the supercell simulated with *city size 2* set (Table 5.8), and city diameter (km). Bolded values are statistically significant.

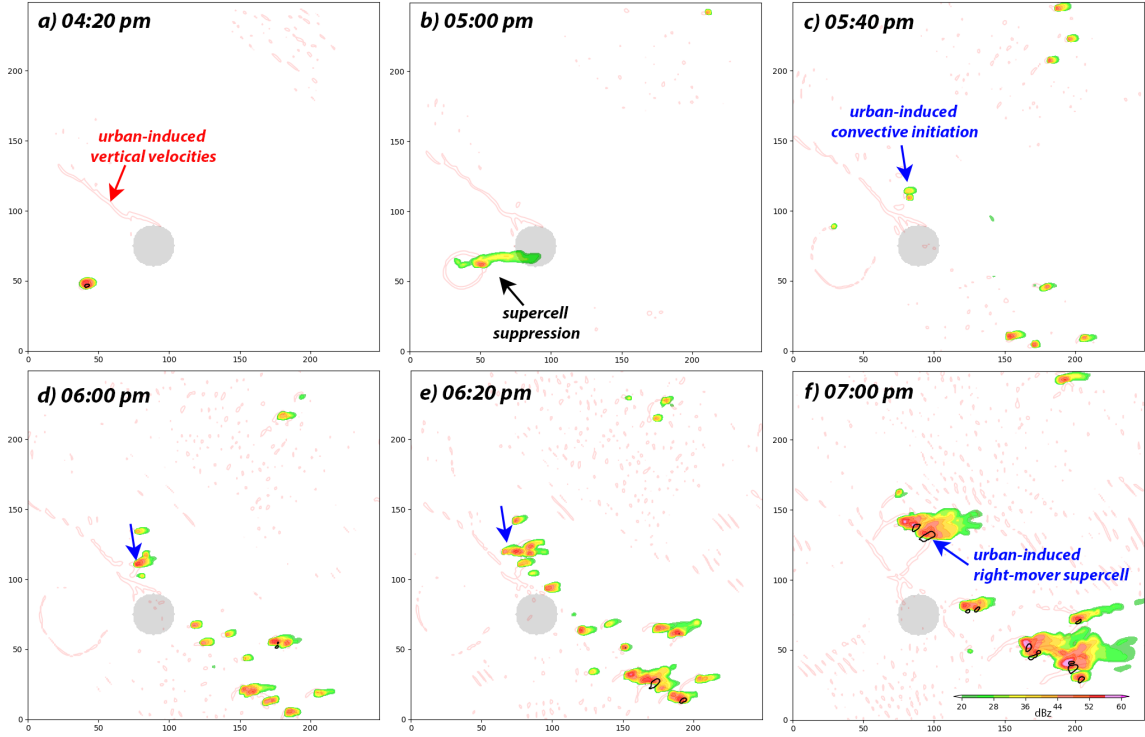


Figure 5.12: As in Fig. 5.7, but for the ensemble member 1 of 30km_2 numerical experiment. Vertical velocities at 950 hPa higher than 0.5 m s^{-1} are plotted with red contours. The blue arrows indicate the triggering of the urban-induced right-mover supercell.

vorticity at 850 and 700 hPa is observed in the *city region*, with statistical significance at 850 hPa.

In the *city region*, a quadratic fit of the form $ax^2 + bx + c = z^*$ is tested (purple line in Fig. 5.10d). For variables such as $wmax$, $w500$, and $mean uh$, the coefficient a is negative and statistically significant, while the coefficient b is also negative but not statistically significant. Examining Fig. 5.10d, it appears that for city diameters up to about 10 km, the city has little effect on the storm. For larger cities, the storm weakens with city size faster than a linear fit would suggest. A quadratic fit is tested also for the others sets of numerical experiments, but this is the only case where the quadratic fit is statistically significant for some variables.

The cumulative ensemble mean of UH illustrates even more clearly the effect of urban land use on the supercell. In the *no_urban* experiment, a strong UH swath is present (Fig. 5.11a). In the 20km_2 experiment, the supercell is considerably weakened as it passes over the city (Fig. 5.11b). The suppression becomes nearly complete in the 30km_2 experiment (Fig. 5.11c). Notably, the suppression is significantly more pronounced in the *city size 2* sample than in the *city size 1* sample (Fig. 5.11d), as also indicated by the steeper and more statistically significant slopes in Table 5.13 compared to Table 5.12.

A particularly interesting feature in the 30km_2 experiment is the emergence of a new UH swath north of the city (Fig. 5.11c). To better understand this behavior, the same time steps showing maximum reflectivity from ensemble member 1 of the 30km_2 experiment (Fig. 5.12) are compared with those from the control run of the 10km_1 experiment (Fig. 5.7). In the 30km_2 experiment, the initial supercell is suppressed very early, already by 05:00 pm (Fig. 5.12b). A stripe of vertical velocities exceeding 0.5 m s^{-1} extends from the city toward the northwest. Just east of this stripe, new convective cells begin to develop between 05:40 pm (Fig. 5.12c) and 06:20 pm (Fig. 5.12e). By 07:00 pm, a new

variable	upstream	city	downstream
wmax	-4.1×10^{-2}	-8.6×10^{-2}	-1.2×10^{-1}
w500	-4.8×10^{-2}	-1.6×10^{-1}	-1.8×10^{-1}
ζ 500	-3.6×10^{-2}	-7.6×10^{-2}	-7.5×10^{-2}
uh max	-2.7×10^{-2}	-1.0×10^{-1}	-1.9×10^{-1}
uh mean	-4.9×10^{-2}	-4.3×10^{-2}	-2.6×10^{-1}
w700	-1.8×10^{-2}	-1.2×10^{-1}	-1.1×10^{-1}
ζ 700	-5.8×10^{-3}	-3.4×10^{-3}	-1.6×10^{-3}
w850	$+5.0 \times 10^{-3}$	-7.2×10^{-2}	-8.7×10^{-2}
ζ 850	$+2.2 \times 10^{-2}$	-5.5×10^{-2}	$+1.1 \times 10^{-1}$
w950	-5.1×10^{-4}	$+1.4 \times 10^{-2}$	-1.1×10^{-1}
ζ 950	-1.6×10^{-3}	-2.8×10^{-2}	-1.8×10^{-1}
Θ_e (500 hPa) max	-1.0×10^{-3}	-6.3×10^{-3}	-5.9×10^{-3}
Θ_e (500 hPa) mean	-2.8×10^{-4}	-1.7×10^{-3}	-3.7×10^{-3}
Θ (5 m) min	-3.1×10^{-4}	$+4.0 \times 10^{-4}$	$+1.2 \times 10^{-3}$
Θ (5 m) mean	-3.7×10^{-5}	$+3.4 \times 10^{-4}$	$+5.0 \times 10^{-4}$

Table 5.14: Slope of linear fitting lines between some diagnostic variables of the supercell and urban fraction (experiments described in Table 5.9). Bolded values are statistically significant.

right-moving supercell forms from these earlier convective initiations (Fig. 5.12f). The evolution in the *10km_1* control run and in ensemble member 1 of the *30km_2* experiment are markedly different. In the latter, the original supercell is suppressed, while a new, urban-induced supercell develops. This behavior occurs in 8 out of 11 members of the *30km_2* experiment and is not observed in any of the other experiments (see Figs. 5.11 and 5.14). Additional evidence supporting the urban origin of the second supercell initiated at 07:00 pm will be presented in Section 5.3.

5.2.3 Urban fraction

In the *urban fraction* experiments, most storm diagnostic variables decrease as a function of the urban fraction (Table 5.14). However, only some of these trends are statistically significant in the *upstream region*, while no statistically significant trends are observed in the *city* and *downstream regions* (with the exception of Θ_e (500 hPa) in the *downstream regions*). In the *upstream region*, the ensemble spread is relatively small (Fig. 5.13a), which leads to statistically significant trends, even though the slope of the mean linear fit is smaller than in the other two regions. In contrast, in the *city* and *downstream regions*, the trends are not statistically significant due to the considerable ensemble spread (Fig. 5.13c,e). Nevertheless, examining the ensemble mean of the UH swath in the *uf0.4* and *uf1.0* experiments (Figure 5.11e,f) reveals a visible weakening of the storm. We conclude that the supercell weakens with increasing urban fraction, although this trend is not statistically significant.

5.2.4 Building height

An opposite trend is observed when analyzing the sensitivity of the supercell to building height. In all three regions, the mean UH increases with increasing building height (Fig. 5.13b,d,f). This trend is particularly strong in the *downstream region* (Fig. 5.13f). The positive trend in mean UH is statistically significant across all three regions studied (Table 5.15). Other mid-level storm diagnostic variables exhibit a similar positive trend (with the exception of ζ_{500} in the *downstream region*); but, these trends are statistically significant only in the *downstream* and *city regions* for some variables (e.g., w500 and w700). In agreement with this result, the ensemble mean of the UH swath in the *h60* experiment

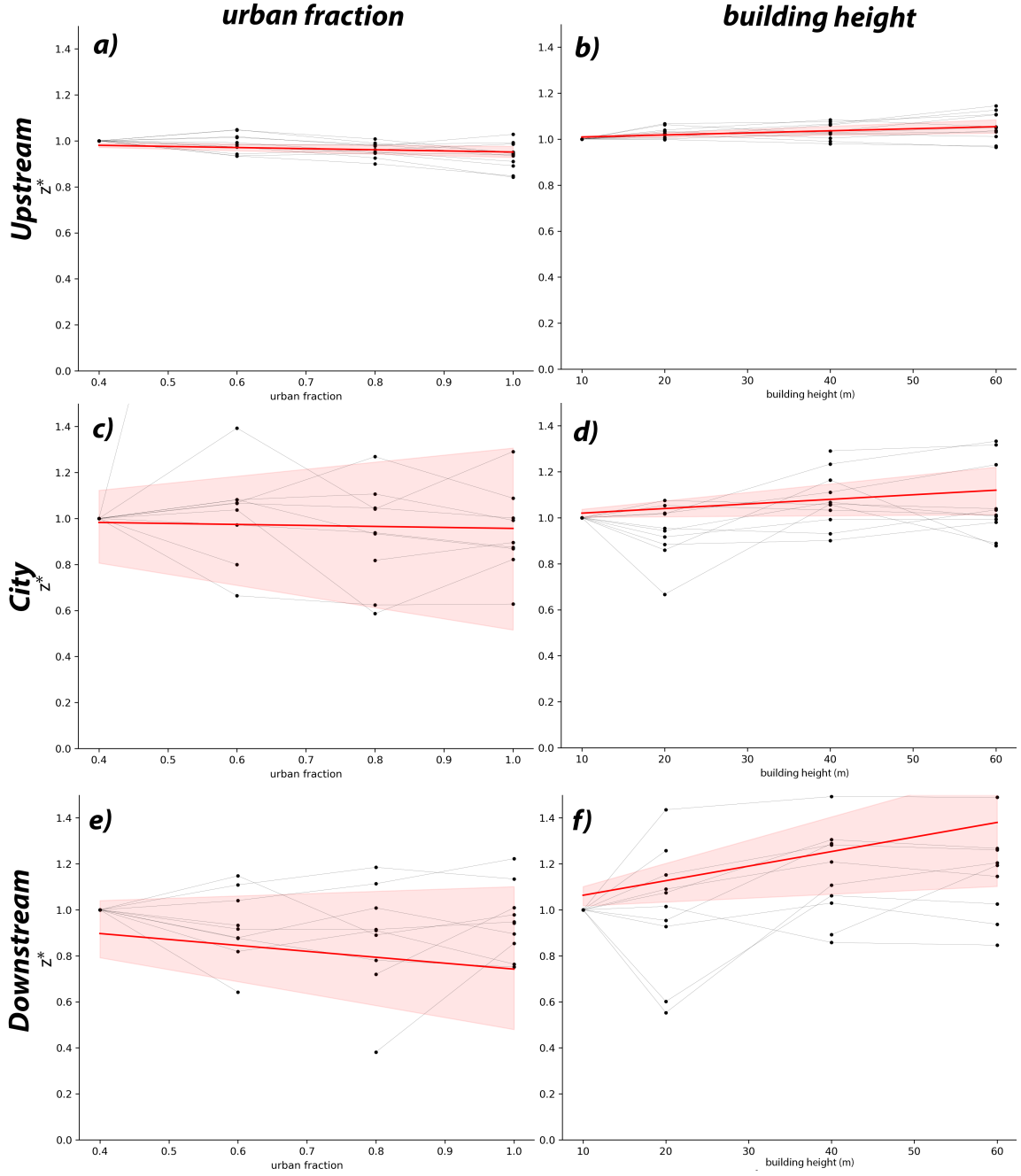


Figure 5.13: As in Fig. 5.10, but for *urban fraction* numerical experiments (Table 5.11) in a), c), and e), and for *building height* numerical experiments (Table 5.10) in b), d) and f).

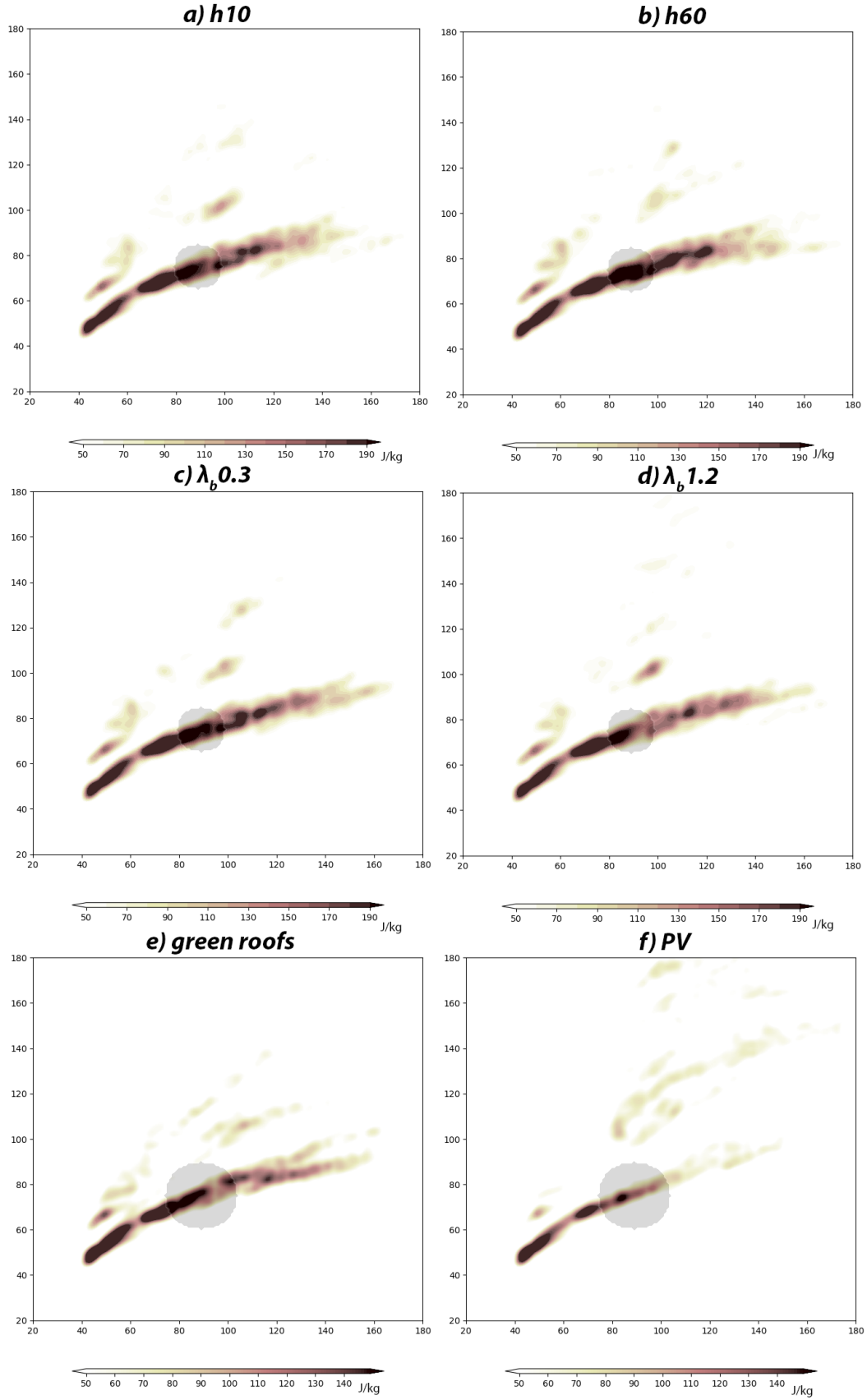


Figure 5.14: As in Fig. 5.11, but for a) h10 experiment (Table 5.10), b) h60 experiment (Table 5.10), c) $\lambda_b 0.3$ experiment (Table 5.10), d) $\lambda_b 1.2$ experiment (Table 5.10), e) green roofs experiment (Table 5.6) and f) PV experiment (Table 5.6).

variable	upstream	city	downstream
wmax	$+6.8 \times 10^{-4}$	$+9.4 \times 10^{-4}$	$+1.3 \times 10^{-3}$
w500	$+8.0 \times 10^{-4}$	$+8.6 \times 10^{-4}$	$+1.1 \times 10^{-3}$
ζ 500	$+8.1 \times 10^{-4}$	$+3.2 \times 10^{-4}$	-1.2×10^{-3}
uh max	$+5.2 \times 10^{-4}$	$+1.8 \times 10^{-3}$	$+7.7 \times 10^{-4}$
uh mean	$+5.1 \times 10^{-3}$	$+2.0 \times 10^{-3}$	$+6.3 \times 10^{-3}$
w700	$+3.2 \times 10^{-4}$	$+6.3 \times 10^{-4}$	$+3.7 \times 10^{-4}$
ζ 700	$+5.0 \times 10^{-5}$	-8.1×10^{-4}	-9.0×10^{-4}
w850	-4.6×10^{-5}	-2.7×10^{-4}	-4.1×10^{-4}
ζ 850	-5.6×10^{-4}	$+6.6 \times 10^{-5}$	-1.4×10^{-3}
w950	$+2.1 \times 10^{-4}$	-3.2×10^{-3}	-2.2×10^{-3}
ζ 950	-1.6×10^{-4}	-3.1×10^{-3}	-2.2×10^{-3}
Θ_e (500 hPa) max	$+1.1 \times 10^{-5}$	$+5.5 \times 10^{-5}$	$+9.5 \times 10^{-5}$
Θ_e (500 hPa) mean	-3.6×10^{-6}	$+3.4 \times 10^{-6}$	$+4.3 \times 10^{-5}$
Θ (5 m) min	$+3.3 \times 10^{-5}$	$+3.4 \times 10^{-5}$	-1.3×10^{-5}
Θ (5 m) mean	-1.6×10^{-7}	-8.0×10^{-6}	-5.1×10^{-6}

Table 5.15: Slope (m^{-1}) of fitting lines between some diagnostic variables of the supercell and building heights (experiments described in Table 5.10). Bolded values are statistically significant.

(Fig. 5.14a) is stronger than in the *h10* experiment (Fig. 5.14b).

In contrast, low-level storm diagnostic variables show a decreasing trend with increasing building height, and this trend is statistically significant for some variables (e.g., *w950* in the *city* region). This result is surprising and interesting, as it contrasts with the storm behavior aloft and also with the hypothesis that taller buildings should enhance vertical motion due to stronger surface wind convergence.

5.2.5 Building density

The mean **UH** does not exhibit a clear trend with increasing building density (Fig. 5.15). In the *downstream* region, there appears to be a slight positive trend, but the ensemble spread is considerable (Fig. 5.15c). Accordingly, the mean linear fits in all three regions are not statistically significant (Table 5.16). Other variables describing the storm's mid-level characteristics show a general weakening trend in the *upstream* and *city* regions, though this trend is statistically significant only for some variables (e.g., *w500* and ζ 500).

Low-level storm diagnostic variables do not show a consistent trend: some increase with increasing λ_b (e.g., *w850* in the *upstream* region), while others decrease (e.g., *w950* in the *city* region, which also shows a statistically significant trend). In the *downstream* region, no clear trend is observed either aloft or near the surface, and none of the studied variables exhibit statistically significant trends.

The ensemble mean of the **UH** swath shows slightly lower values *downstream* of the city in the $\lambda_b 1.2$ experiment (Fig. 5.14d) compared to the $\lambda_b 0.3$ experiment (Fig. 5.14c). Overall, we conclude that the storm does exhibit some sensitivity to building density; but, this parameter appears to be the least influential in determining storm dynamics than the previous studied.

5.2.6 Green roofs and **PV** panels

Green roofs have a remarkable impact compared to the original experiment with standard roofs. Comparing Fig. 5.11c with Fig. 5.14e, it is evident that the weakening of the storm as it passes over the city is much less pronounced when the roofs are covered with

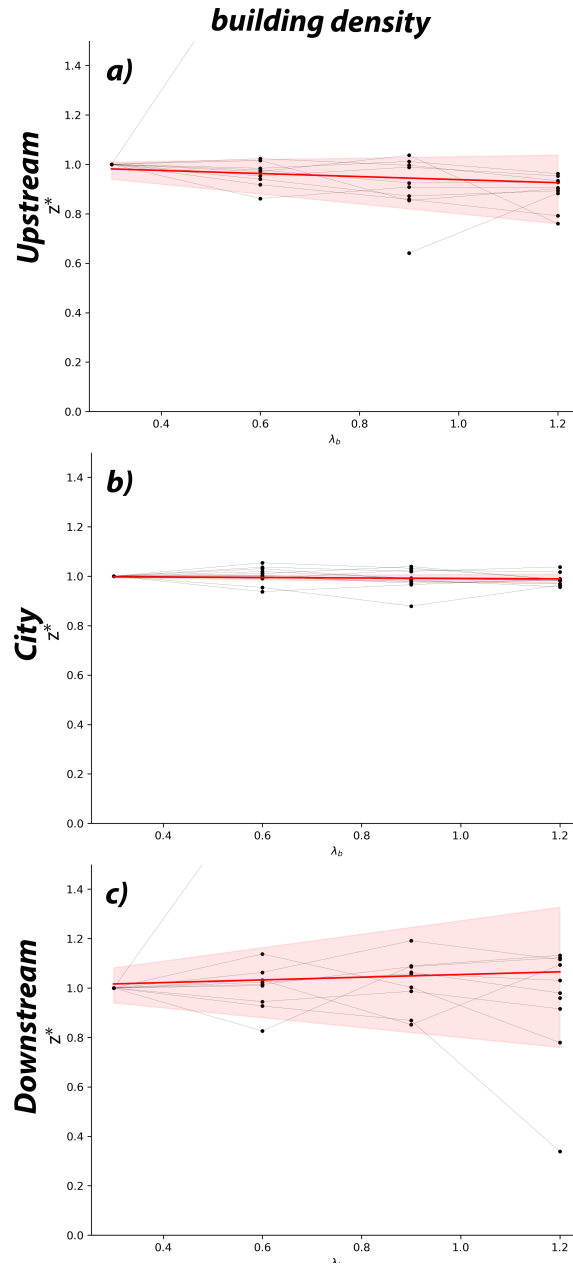


Figure 5.15: As in Fig. 5.10, but for numerical experiments *building density* (Table 5.11).

variable	upstream	city	downstream
wmax	-1.2×10^{-2}	-3.5×10^{-2}	$+2.1 \times 10^{-3}$
w500	-8.0×10^{-3}	-4.3×10^{-2}	$+7.5 \times 10^{-3}$
ζ 500	-1.6×10^{-2}	-2.2×10^{-2}	-3.8×10^{-2}
uh max	-1.1×10^{-2}	-3.8×10^{-2}	$+4.6 \times 10^{-2}$
uh mean	-9.2×10^{-3}	-6.2×10^{-2}	$+5.4 \times 10^{-2}$
w700	-9.2×10^{-4}	-3.4×10^{-2}	$+1.9 \times 10^{-2}$
ζ 700	$+4.3 \times 10^{-3}$	$+1.1 \times 10^{-2}$	$+7.4 \times 10^{-3}$
w850	$+2.9 \times 10^{-3}$	-1.8×10^{-2}	-2.1×10^{-3}
ζ 850	$+2.2 \times 10^{-2}$	-3.2×10^{-2}	$+1.0 \times 10^{-1}$
w950	-1.1×10^{-2}	-5.1×10^{-2}	-4.2×10^{-2}
ζ 950	$+1.3 \times 10^{-2}$	-5.9×10^{-2}	$+4.0 \times 10^{-2}$
Θ_e (500 hPa) max	$+1.2 \times 10^{-4}$	-2.6×10^{-3}	$+1.9 \times 10^{-3}$
Θ_e (500 hPa) mean	$+7.8 \times 10^{-5}$	$+1.5 \times 10^{-5}$	$+1.3 \times 10^{-4}$
Θ (5 m) min	$+5.6 \times 10^{-5}$	$+1.7 \times 10^{-3}$	-2.4×10^{-4}
Θ (5 m) mean	$+7.2 \times 10^{-6}$	$+6.9 \times 10^{-4}$	-8.8×10^{-4}

Table 5.16: Slope of the mean linear fit between some storm's diagnostic variables and λ_b (Table 5.11). Statistically significant values are bolded.

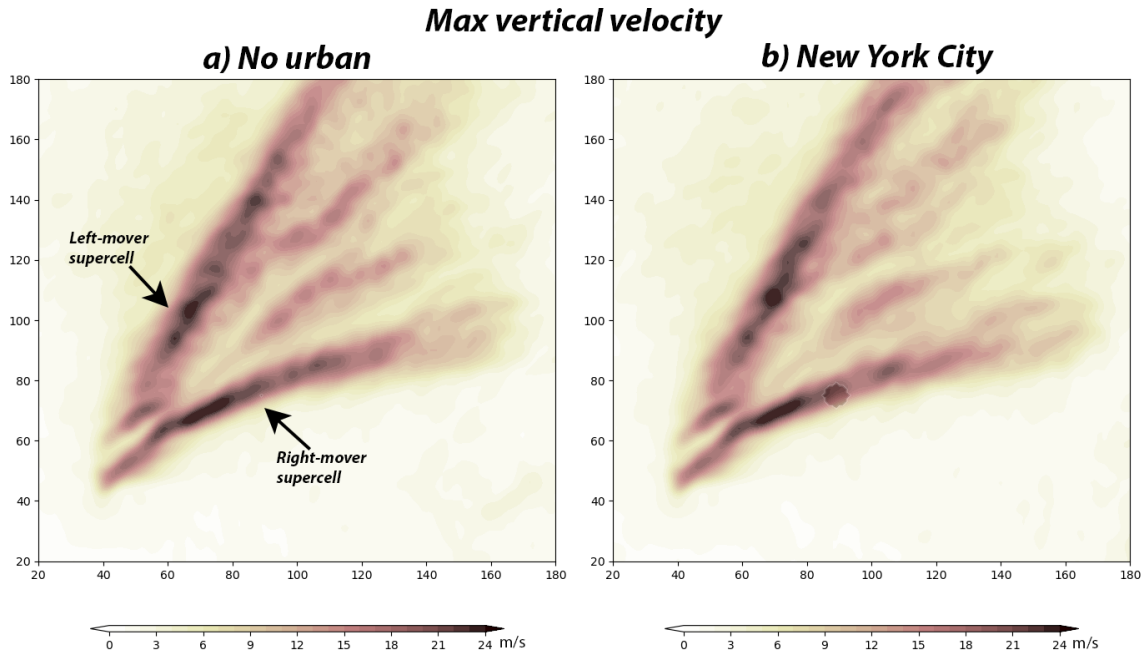


Figure 5.16: Ensemble mean of cumulated maximum vertical velocities in the a) *no_urban* and b) NYC experiment (Table 5.6). The city location is shown with a black circular shading in b).

vegetation. Moreover, the urban-induced supercell is absent. It appears that a city with green roofs exerts a significantly smaller influence on storm dynamics compared to a city with standard roofs.

On the other hand, *PV* panels do not affect the weakening of the supercell caused by a large city. The *UH* swath is similar in the experiment with *PV* panels (Fig. 5.14f) and in the original *30km_2* experiment (Fig. 5.11c).

5.2.7 New York City experiment

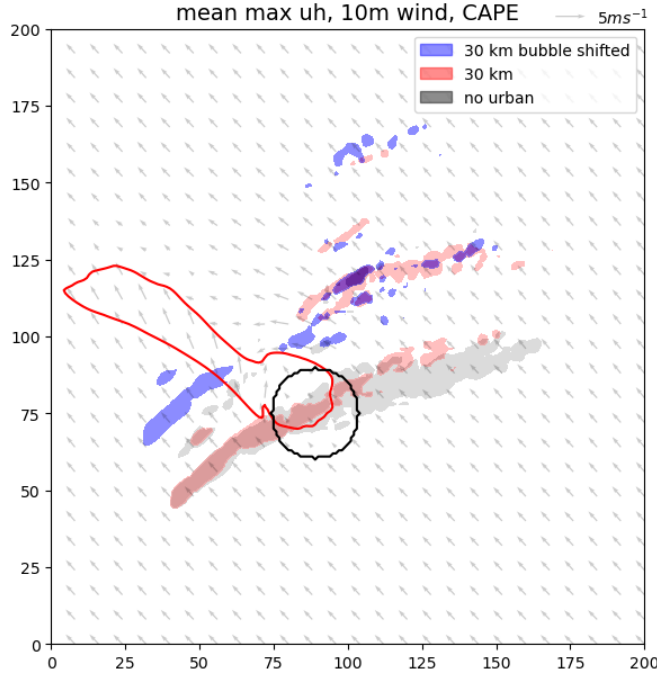


Figure 5.17: The ensemble mean of *UH* values exceeding 80 Jkg^{-1} is shown for the *no_urban* simulation (black shading), the *30km_2* simulations (red shading, Table 5.8), and the *bubble* experiment (blue shading). The difference in *CAPE* greater than 400 Jkg^{-1} between the ensemble mean in the *30km_2* experiment and in the *no_urban* experiment is indicated with a red contour. The ensemble mean of 10 m wind field from the *30km_2* experiment at 04:00 pm is shown with grey arrows. The city location is outlined with a black contour.

urban-induced supercell is less intense than in the original *30km_2* experiment (compare the blue and red shading north of the city in Fig. 5.17), likely because downdrafts associated with the initial storm create an environment less favorable for the development of the urban-induced supercell.

No remarkable differences are observed whether the right-moving supercell passes over a rural area or over downtown NYC (experiment NYC in Table 5.6). In Fig. 5.16, the ensemble mean of the cumulative maximum vertical velocities is shown for the *no_urban* experiment (Fig. 5.16a) and for the experiment with a city resembling downtown NYC with an 8 km diameter (Fig. 5.16b). The storm splitting attributed to NYC by Bornstein and Lin (2000) is not observed. Note the different trajectories of the left-mover and right-mover supercells: an explanation of this behavior is provided in Appendix A.3.

5.2.8 Bubble shifted

When the bubble is shifted northward in the *bubble* experiment (described in Table 5.6), the weakening of the supercell occurs even more rapidly than in the *30km_2* experiment. This is evident in Fig. 5.17, where the mean *UH* with the bubble shifted (blue shading) can be compared to the mean *UH* in the original *30km_2* experiment (red shading). Furthermore, the

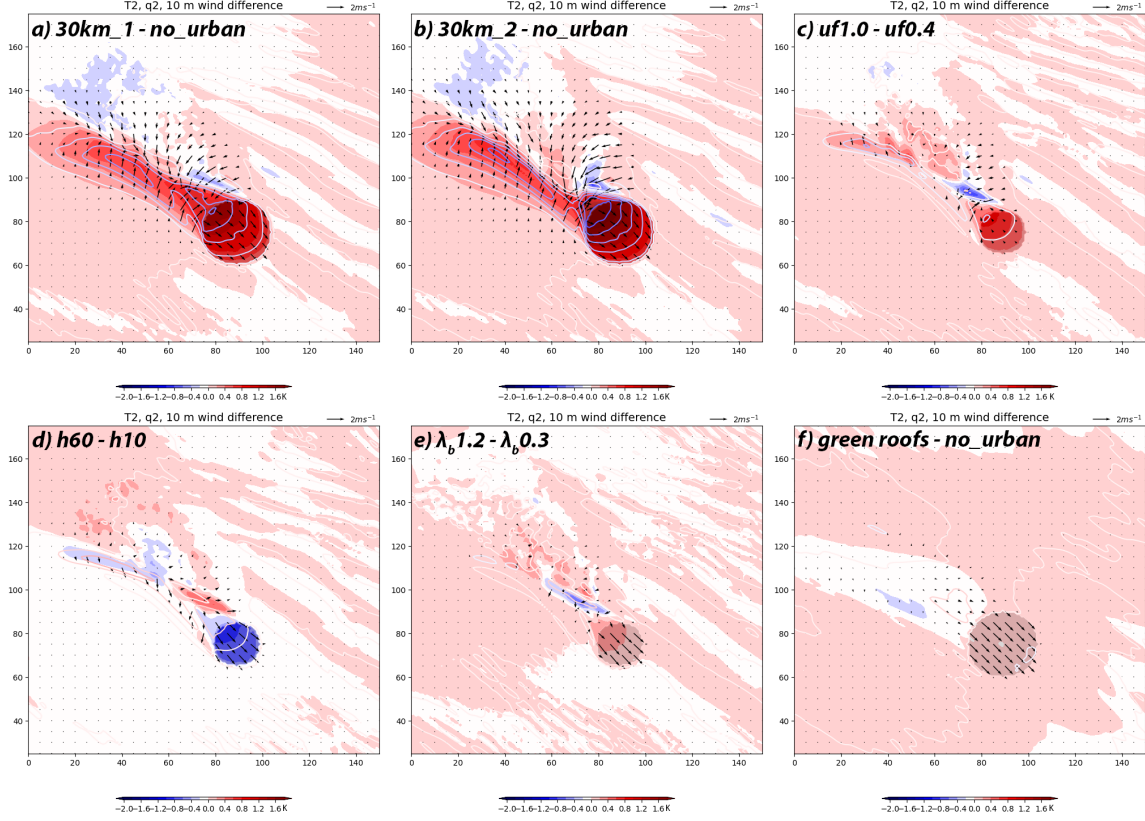


Figure 5.18: Ensemble mean of 2m Temperature (colour shading), 2m specific humidity (contours every 0.5 g kg^{-1} , blue negative, red positive) and 10m wind difference (vectors) at 04:00 pm between a) *30km_1* and *no_urban* experiments; b) *30km_2* and *no_urban* experiments; c) *uf1.0* and *uf0.4* experiments; d) *h60* and *h10* experiments; e) $\lambda_b 1.2$ and $\lambda_b 0.3$ experiments; f) *green roofs* and *no_urban* experiment. The city is shown with grey shading.

5.3 Results: sensitivity of the pre-storm environment to the urban morphological parameters.

To understand the urban-effects in the simulated storm discussed in the previous section, the effect of the urban land use on the pre-storm environment is analyzed hereafter.

5.3.1 Near-surface variables

The effect of morphological parameters on some near-surface variables at 04:00 pm (at convective initiation) is evaluated by computing the difference of the ensemble means between the simulations with the highest value ("max value" column in Table 5.6) and the lowest value ("zero value" column in Table 5.6) of the parameters explored in each set of numerical experiments. For example, in the *city size 1* set of experiments, simulations with a 30 km diameter city (*30km_1* simulations) are compared with the *no_urban* ones (Fig. 5.18a). The variables analyzed include the temperature at 2 m (T2), specific humidity at 2 m (q2), wind speed at 10 m, *mslp*. Temperature, specific humidity and wind at 850 hPa will also be analyzed.

The *30km_1* simulations produce a weak UHI of about 1.5 K over the city (Fig. 5.18a), in good agreement with typical daytime UHI magnitudes (Hidalgo et al., 2010). A pronounced UDI is also reproduced (peak of -2 g kg^{-1}), corresponding to a relative humidity

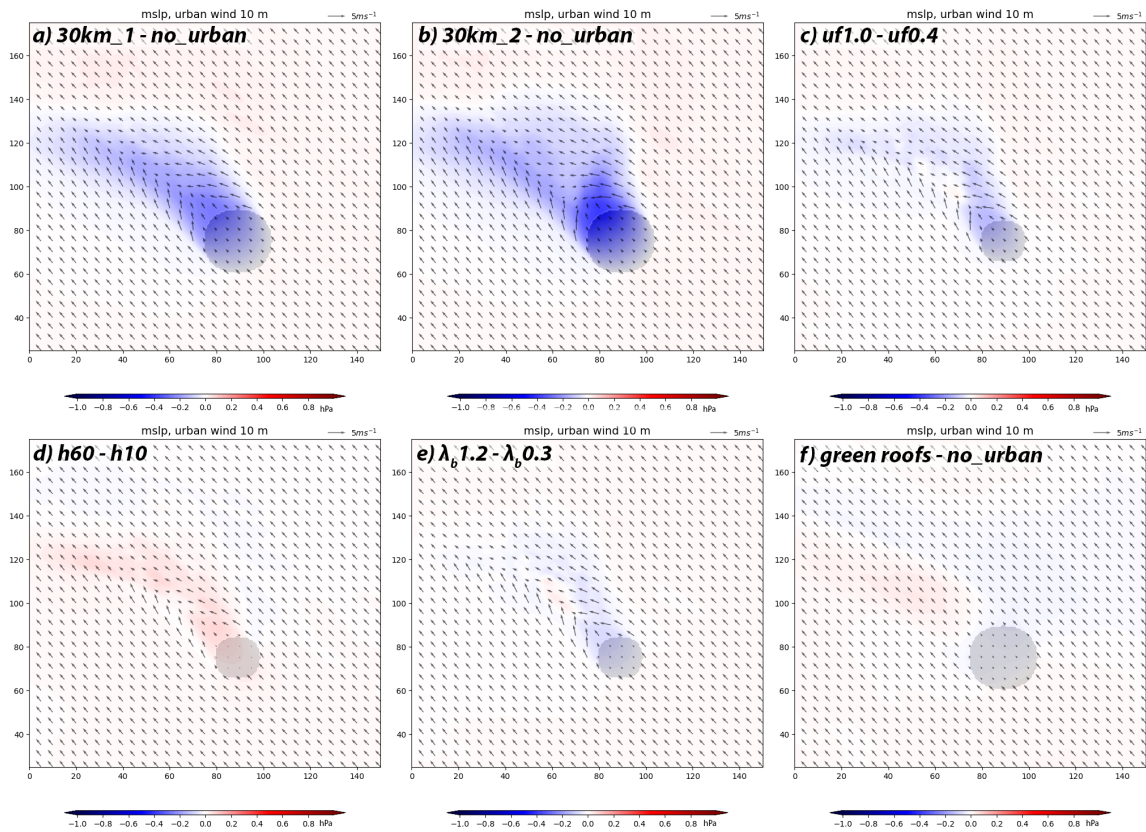


Figure 5.19: Ensemble mean of mean sea level pressure difference between a) *30km_1* and *no_urban* experiments; b) *30km_2* and *no_urban* experiments; c) *uf1.0* and *uf0.4* experiments; d) *h60* and *h10* experiments; e) $\lambda_b 1.2$ and $\lambda_b 0.3$ experiments; f) *green roofs* and *no_urban* experiment. The 10 m wind of the minuend simulation of the subtraction is shown with vectors. The city is shown with grey shading.

difference of about 13% between the city and the surrounding rural areas, consistent with global observational studies (Meili et al., 2022). Temperature and moisture perturbations form a gradient over the city along the surface wind direction. A wake of higher temperatures and lower humidity extends downwind (with respect to the southeasterly surface wind) up to 80 km from the city. Within this wake, a wind convergence is induced by the urban land use, while winds over the city are significantly reduced by increased drag. In the *30km_2* simulations, the **UHI** and **UDI** are slightly stronger, but the spatial pattern is similar (Fig. 5.18b).

Figure 5.18c shows the effect of urban fraction, obtained by subtracting the ensemble means of simulations with urban fraction 0.4 from those with urban fraction 1.0. The impact of urban fraction on **UHI** and **UDI** is much smaller than that of city size, about 0.5 K and -1.0 gkg^{-1} , respectively. The downwind wind convergence is also much weaker, and the warm, dry wake is not apparent.

An interesting result is that taller buildings reduce the **UHI** by about 1 K and increase moisture by 0.5 gkg^{-1} (Fig. 5.18d). Wind speeds are weaker in simulations with taller buildings, as well as the wind convergence downwind the city. The temperature reduction and moisture increase in the urban canopy are likely caused by reduced ventilation and consequent moisture stagnation (Wang et al., 2021). Our simulations begin with high moisture in the **PBL**, which decreases during the day due to mixing. In deeper urban canopies, drying is slower due to building shade, resulting in lower temperatures and higher moisture (Huang and Song, 2023).

A more compact city does not significantly alter the **UHI** or **UDI** but does reduce surface winds over the city (Fig. 5.18e). The weak influence of building density on the **UHI** contrasts with the findings of Li et al. (2020).

Finally, adding vegetation to all rooftops (*green roof* simulations) completely eliminates the **UHI** and **UDI**, as expected (Li et al., 2014), while maintaining the wind reduction over the city (Fig. 5.18f). This experiment can be interpreted as preserving the city's dynamic perturbation on the atmosphere (the weaker surface winds) while suppressing its thermodynamic effect (the **UHI** and **UDI**). Overall, city size is the factor with the strongest influence on the **UHI**, consistent with previous studies (Zhou et al., 2017).

A weak minimum in mean sea level pressure (about -0.5 hPa) is generated downwind of the city (with respect to the surface winds) by the urban land use (Fig. 5.19a). Over the city there is a pressure gradient, but not a minimum. The downwind pressure minimum coincides with the wind convergence. With a higher urban fraction, the pressure minimum is stronger (about -1.0 hPa ; Fig. 5.19b), as also suggested by the comparison between the *uf1.0* and *uf0.4* experiments (Fig. 5.19c). On the other hand, with taller buildings the pressure minimum is slightly weaker (by about 0.2 hPa ; Fig. 5.19d), while a more compact city plays only a marginal role in intensifying it (Fig. 5.19e). Lastly, in the *green roof* experiment the pressure minimum is absent.

A relationship between the **UHI** shown in Fig. 5.18 and the mean sea level pressure is evident: in simulations with a stronger **UHI** (e.g., *30km_2* experiment), the pressure minimum is also stronger, whereas when the **UHI** is weaker (*h60* experiment) or absent (*green roof* experiment), the pressure minimum is weaker or absent. Thus, it appears that the pressure minimum is thermodynamically induced by the **UHI** (i.e., it is a consequence of the buoyancy and not of the dynamical modification of the flow by the buildings). To further support this hypothesis, an additional analysis is carried out.

The *dynamic* p'_d (Eq. A.25) and *buoyancy* p'_b (Eq. A.26) pressure perturbations are computed for the ensemble mean of the *30km_2* simulations. The reference p_0 value is taken from the ensemble mean of the *no_urban* experiment. Equations A.26 and A.25 are elliptic partial differential equations, solved using a Gauss-Seidel iterative method. The

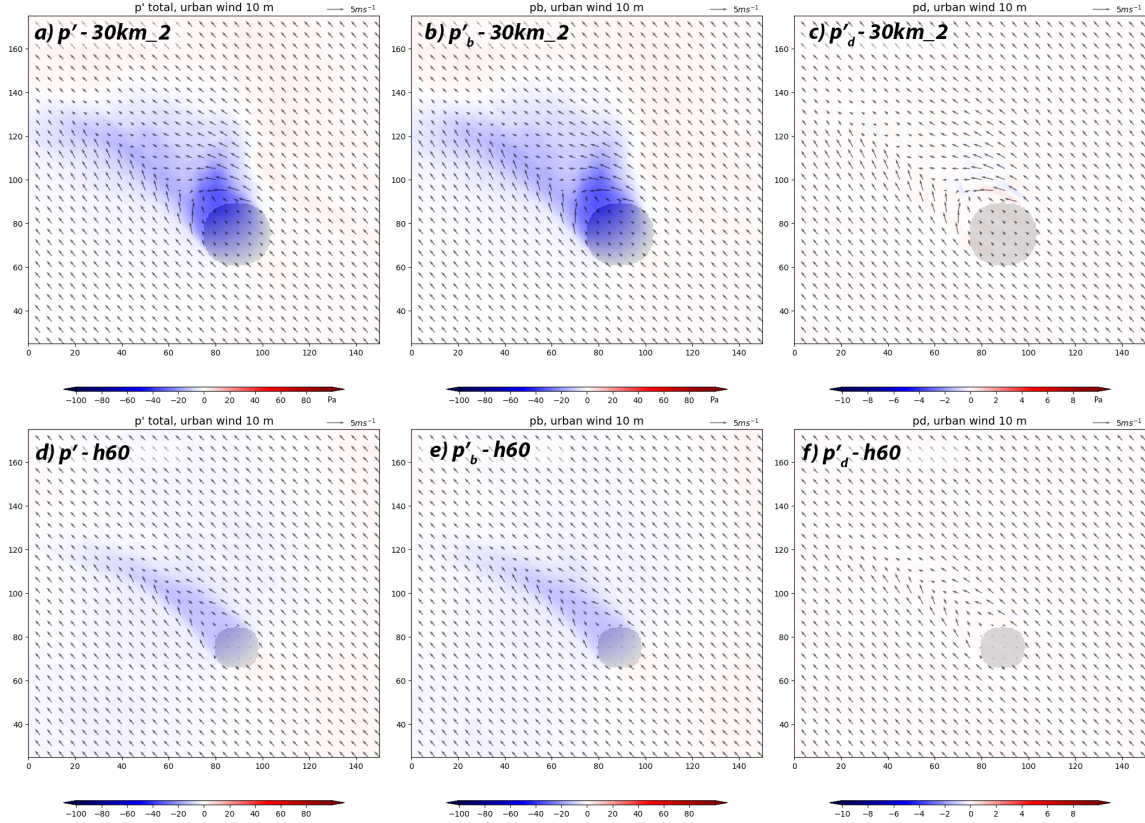


Figure 5.20: Ensemble mean of a) total pressure perturbations p' at 300m AGL in the *30km_2* experiment; b) as in a) but for buoyancy pressure perturbations p'_b ; c) as in b) but for dynamic pressure perturbations p'_d ; d), e) and f) as in a), b) and c), respectively, but for the *h60* experiment.

first guess is set to $p'_b = p'_d = 0$. A zero-gradient lateral boundary condition is imposed because the simulated pressure perturbations are away from the domain boundaries. A zero-gradient boundary condition for p'_d is also imposed at the bottom. Figure 5.20 shows the results at the 10th vertical level (about 300 m AGL). Buoyancy pressure perturbations p'_b clearly dominate the generation of the downstream pressure minimum, consistent with the previously noted sensitivity between the pressure minimum and the intensity of the UHI. The contribution of the dynamic pressure perturbations p'_d is negligible. This outcome is valid also for the *h60* experiment (Fig. 5.20d,e,f): tall buildings have a marginal effect on the generation of the downwind pressure minimum with respect to the thermodynamic modification of the PBL induced by urban land use.

The urban-induced atmospheric modifications extend up to the top of the PBL. For instance, in Figure 5.21 the same variables shown in Figure 5.18 are plotted at the 850 hPa isobaric level. Compared with the *no_urban* case, the wind field is markedly reduced over the city and in a wake downwind of it, where the temperature is reduced by about 0.3 K and the specific humidity is increased by 1.0 g kg^{-1} (Fig. 5.21a). There is also a region west of the city containing a warm bubble of about 1 K with reduced moisture. Additional cross-section analysis shows that this warm bubble originates from the displacement of the UHI dome by the wind profile. The same pattern is observed in both the *30km_1* and *30km_2* experiments, although in the latter the urban-induced perturbations are stronger. In contrast, urban fraction, building height, and building density have negligible effects at the 850 hPa level (Fig. 5.21c,d,e). Lastly, with green roofs, a warm bubble (thermal perturbation of about 0.5 K) is present downstream of the city (Fig. 5.21f).

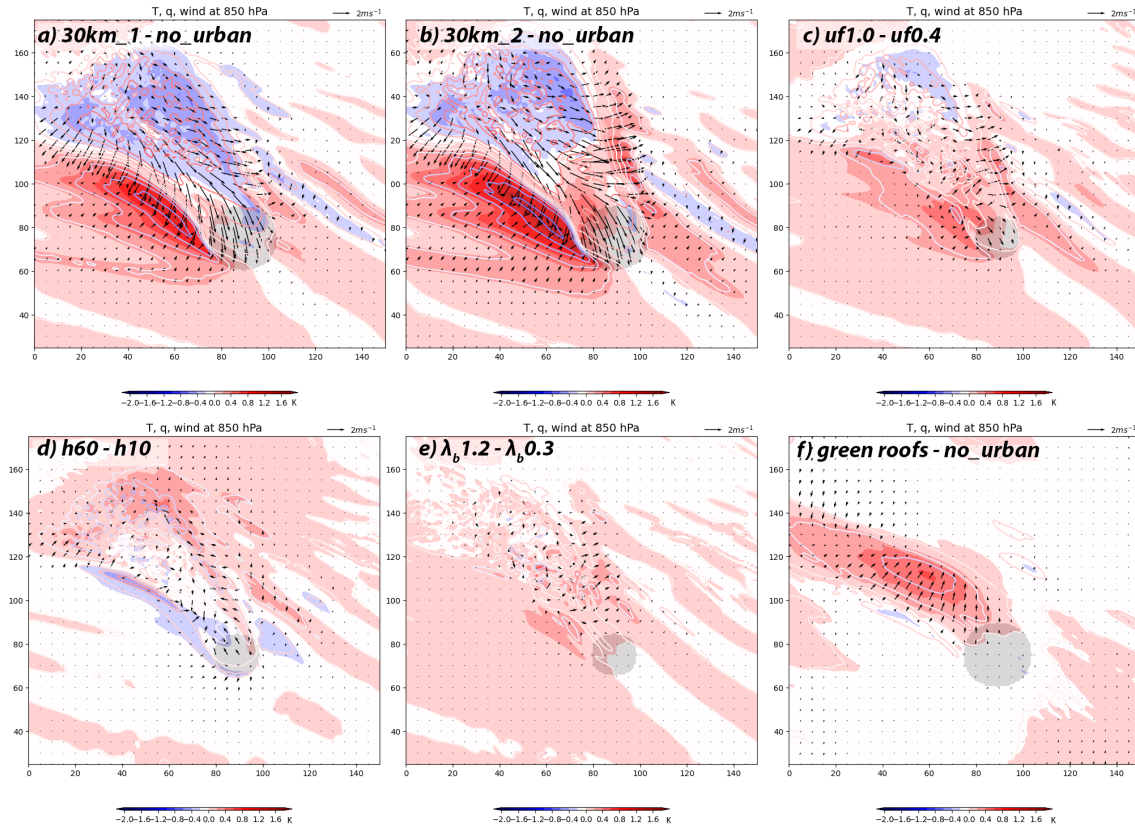


Figure 5.21: Ensemble mean of temperature (colour shading), specific humidity (contours every 0.5 g/kg, blue negative, red positive) and wind difference (vectors) at 850 hPa between a) *30km_1* and *no_urban* experiments; b) *30km_2* and *no_urban* experiments; c) *uf1.0* and *uf0.4* experiments; d) *h60* and *h10* experiments; e) $\lambda_b 1.2$ and $\lambda_b 0.3$ experiments; f) *green roofs* and *no_urban* experiments. The city is shown with grey shading.

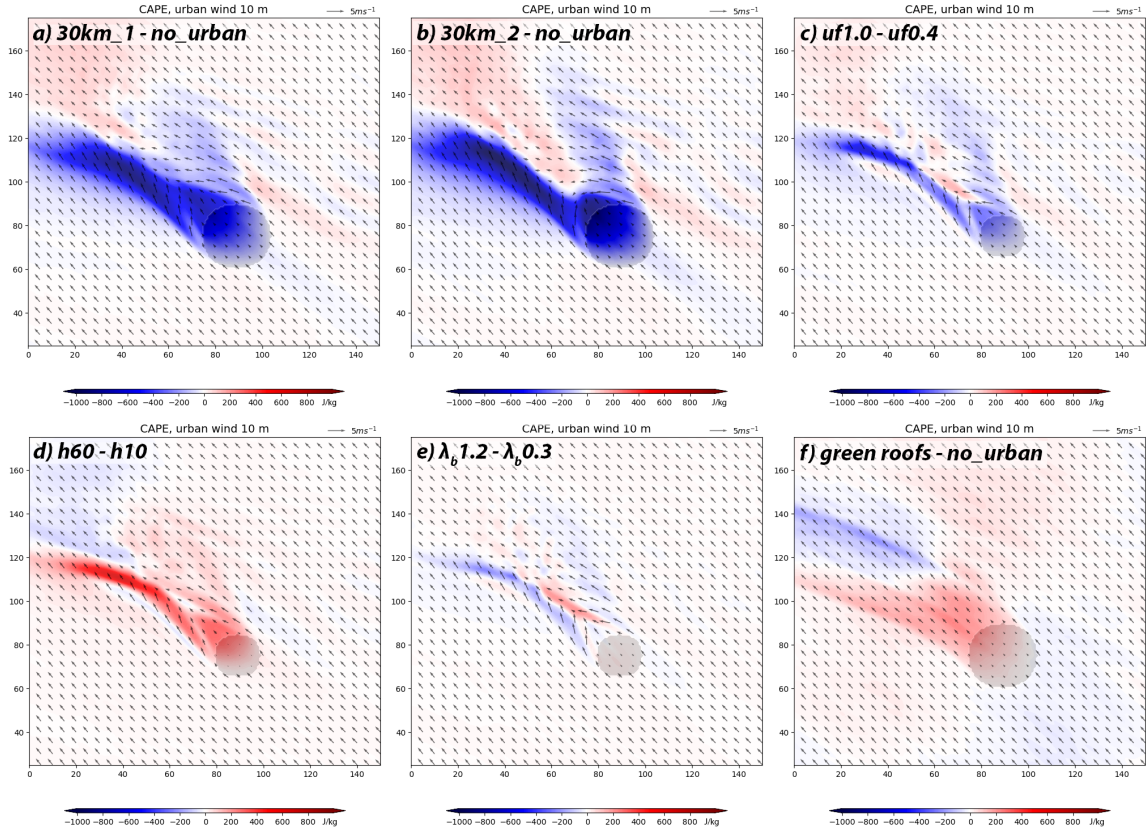


Figure 5.22: CAPE difference between a) *30km_1* and *no_urban* experiments; b) *30km_2* and *no_urban* experiments; c) *uf1.0* and *uf0.4* experiments; d) *h60* and *h10* experiments; e) $\lambda_b1.2$ and $\lambda_b0.3$ experiments; f) *green roofs* and *no_urban* experiment. The 10 m wind of the minuend simulation of the subtraction is shown with vectors. The city is shown with grey shading.

5.3.2 Latent Instability: CAPE

CAPE is markedly reduced by urban land use along a wake downstream of the city (Fig. 5.22a,b). The reduction in CAPE reaches up to 1000 Jkg^{-1} and coincides with the wake of higher temperature and lower moisture observed in Fig. 5.18. This reduction is consistent with the results presented in Chapter 4 and with the findings of Yang et al. (2021b), Reames and Stensrud (2018), and Yang et al. (2014). Urban fraction contributes to a decrease in CAPE of up to about 200 Jkg^{-1} (Fig. 5.22c), while taller buildings increase CAPE by up to 400 Jkg^{-1} compared with simulations with smaller buildings (Fig. 5.22d). Building density has no significant effect on CAPE (Fig. 5.22e), consistent with the results previously discussed for other near-surface variables. In the green roof experiments, CAPE increases slightly, by about 100 Jkg^{-1} , downstream the city (Fig. 5.22f).

The same sensitivity of CAPE to urban morphological parameters as observed for moisture suggests that the decrease in moisture more than offsets the temperature increase, which alone would cause an increase in CAPE. While the UHI drives the generation of the downwind pressure minimum, the UDI drives the formation of the wake with reduced CAPE.

5.3.3 Trigger: Vertical Velocities at 950 hPa

To assess the potential of the city to trigger convective initiation, the differences in ensemble means of vertical velocity at the 950 hPa isobaric level are analyzed. Upstream

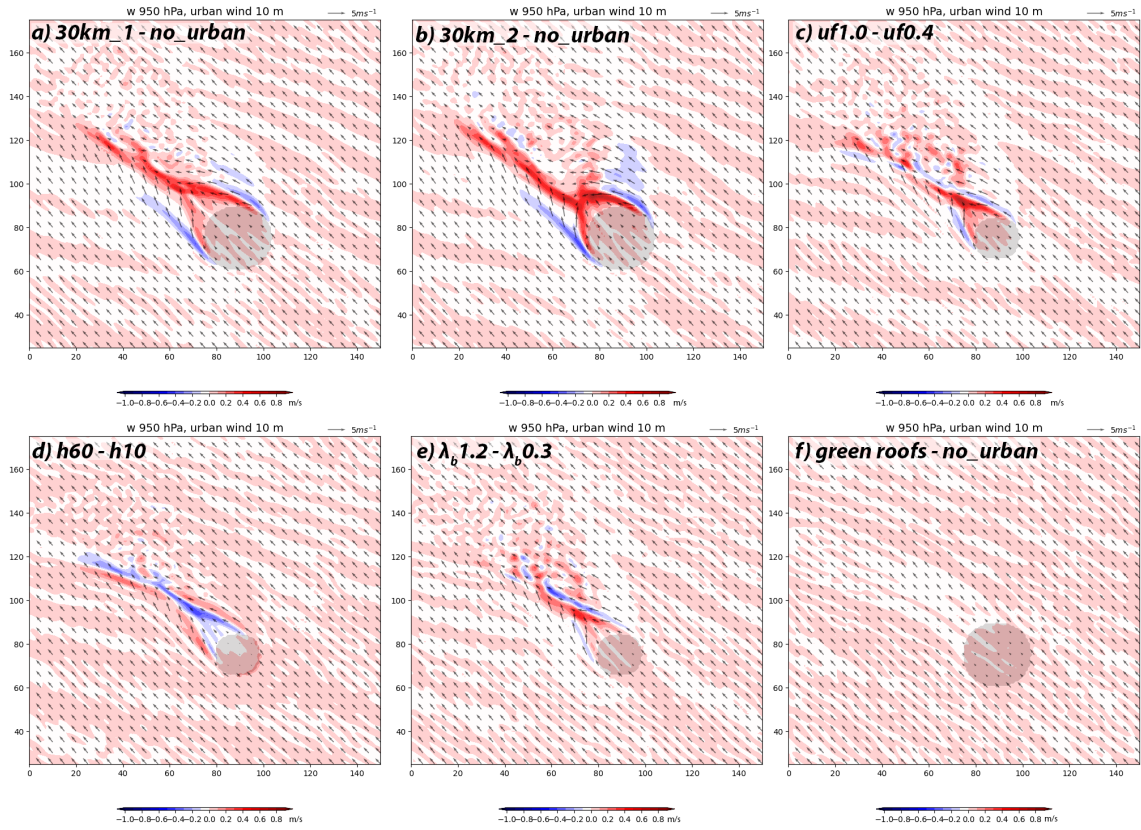


Figure 5.23: Difference of vertical velocities at 950 hPa between a) 30km_1 and no_urban experiments; b) 30km_2 and no_urban experiments; c) uf1.0 and uf0.4 experiments; d) h60 and h10 experiments; e) $\lambda_b 1.2$ and $\lambda_b 0.3$ experiments; f) green roofs and no_urban experiment. The 10 m wind of the minuend simulation of the subtraction is shown with vectors. The city is shown with grey shading.

and over the city, vertical velocities are not induced by urban land use (Fig. 5.23a,b). In contrast, downstream of the city, a line of intensified vertical velocity extends up to 60 km from the city. In the *30km_2* experiment, the intensified vertical velocity reaches locally up to 1 ms^{-1} . This line coincides with the surface wind convergence. A higher urban fraction produces stronger vertical velocities (Fig. 5.23c), while taller buildings reduce their intensity (Fig. 5.23d). A more compact city slightly intensifies vertical velocities downstream (Fig. 5.23e), whereas vertical velocities are unaffected in the *green roof* experiment compared with the *no_urban* scenario (Fig. 5.23f).

The intensification of vertical velocities is strictly related to the intensity of the UHI. A stronger UHI causes a stronger pressure minimum, a stronger wind convergence at the surface, and stronger vertical velocities. For example, in the simulation with taller buildings vertical velocities are weaker (Fig. 5.23d) because the pressure minimum is weaker (Fig. 5.19d) and the UHI is weaker (Fig. 5.18d). In the *green roof* experiment there are no vertical velocities (Fig. 5.23f) because there is no pressure minimum (Fig. 5.19f) and no UHI (Fig. 5.18f). The effect of building is negligible: even 60 m tall buildings does not cause an intensification of vertical velocities. This outcome explains why low-level vertical velocities in the supercell do not increase with increasing building height, as discussed in section 5.2.4. For this reason we believe that the *building-barrier effect* discussed in the literature for decades (see Shepherd 2013 for a review) is overestimated, as also supported by NYC experiment. The heights of the buildings is not enough to cause remarkable vertical velocities. The only exception could be in situation with strong surface flow and gust fronts, as discussed in Chapter 4.

5.3.4 Wind Shear: SRH 0-3 km

In the *30km_1* and *30km_2* experiments, SRH 0-3 km is reduced by up to 100 Jkg^{-1} (about 50 % less than in the *no_urban* simulations) south of the convergence line discussed in Section 5.3.3, and increased north of it with a similar magnitude (Fig. 5.24a). The increase (reduction) of SRH north (south) of the convergence line is due to the greater (smaller) veering of the wind caused by a stronger (weaker) surface easterly wind component (Fig. 5.18a), which in turn is associated with the pressure minimum discussed in Section 5.3.1. Thus, the modifications of SRH are induced by the pressure minimum generated by the UHI. The urban fraction amplifies this SRH pattern (Fig. 5.24c), whereas taller buildings cause the opposite pattern, i.e., higher SRH values south of the convergence line, due to the reduced intensity of the UHI. Building density and the presence of green roofs have little effect on SRH, except for slightly higher values over the city (Fig. 5.24e,f). The increase of SRH over the city results from reduced surface winds (Fig. 5.18e,f), consistent with the findings of Reames and Stensrud (2018). Notably, this result differs from what was found in Chapter 4.

5.4 Discussion

Ensembles of idealized numerical experiments carried out with the WRF model coupled with BEP-BEM consistently show a remarkable sensitivity of supercells to urban land use. In general, the incoming supercell is weakened by the city, especially downstream of it. This weakening occurs primarily for medium-large cities, with diameters greater than 10 km and a high urban fraction (Figs. 5.10 and 5.13). For smaller cities, the effect is less evident, and in some ensemble members the supercell is even intensified (e.g., Fig. 5.10d). The weakening of the supercell is caused by the UDI, which drives a substantial reduction in CAPE, up to 1000 Jkg^{-1} (Fig. 5.22), in the wake downstream of the city (relative

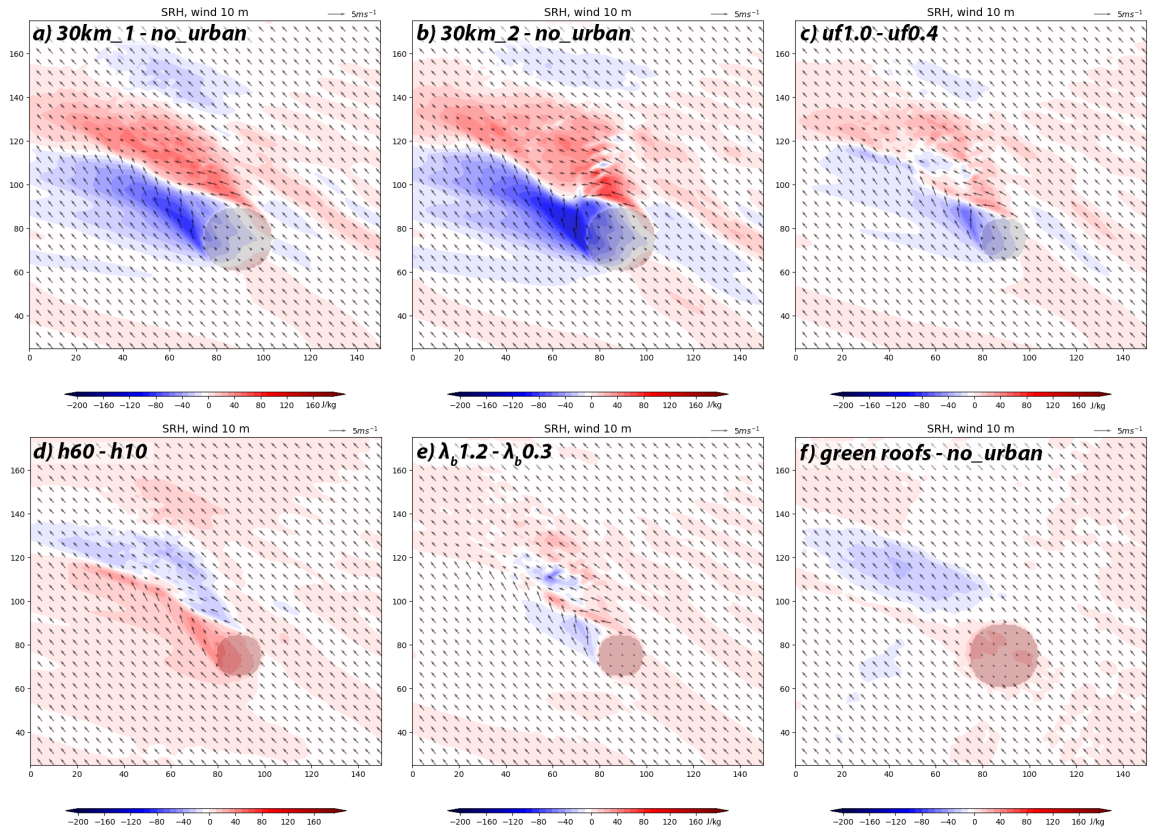


Figure 5.24: SRH 0-3 km difference between a) 30km_1 and no_urban experiments; b) 30km_2 and no_urban experiments; c) uf1.0 and uf0.4 experiments; d) h60 and h10 experiments; e) $\lambda_b1.2$ and $\lambda_b0.3$ experiments; f) green roofs and no_urban experiment. The 10 m wind of the minuend simulation of the subtraction is shown with vectors. The city is shown with grey shading.

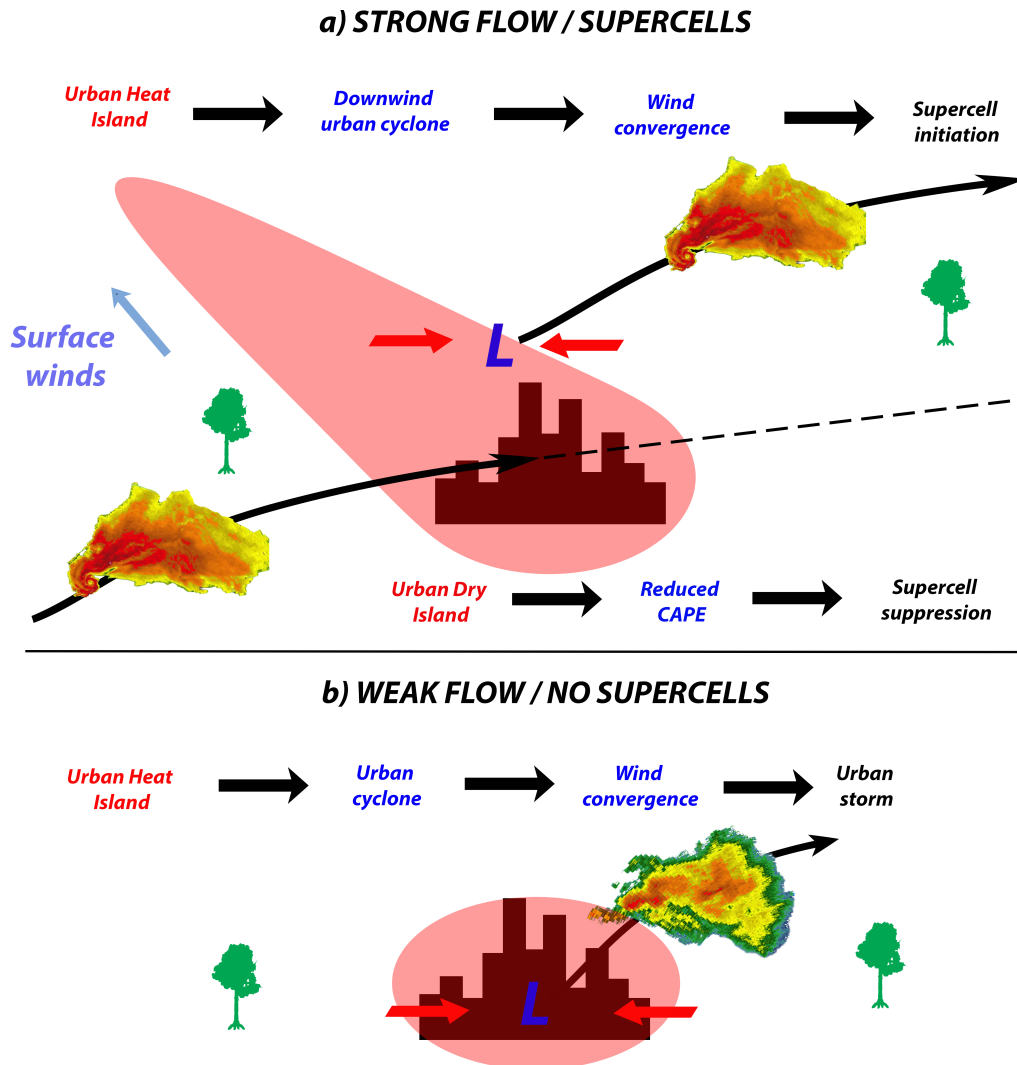


Figure 5.25: a) sketch of the results obtained: the UHI and UDI (red shading), city (black histogram), supercell (yellowish-reddish shading with hook-echo shape), *urban cyclone* (letter L) and wind convergence (red arrows) are shown. b) as in a), but for a weak flow situation without supercells.

to the surface winds, Fig. 5.25a). Previous studies have also found that urban land use can reduce CAPE (e.g., Reames and Stensrud 2018; Yang et al. 2014); but, to the authors knowledge, this is the first time that an urban-induced reduction of CAPE has been directly linked to an urban-induced weakening of storms. Our conclusions differ markedly from those of Naylor (2020), who did not consider the UDI in his idealized setup, a factor that proves to be crucial in influencing supercell dynamics in our simulations.

Furthermore, with a sufficiently large city and a high urban fraction, a new urban-induced supercell develops in 8 out of 11 ensemble members downstream of the city, while the initial incoming supercell is dissipated by the UDI (e.g., Fig. 5.12). This new urban-induced supercell is triggered by a convergence line downstream of the city, generated by a pressure minimum of about 1 hPa (hereafter referred to as the *urban cyclone*, Fig. 5.25a). The UHI thermodynamically induces the *urban cyclone*, which in turn produces the convergence line (Fig. 5.19). The connection between the *urban cyclone* and the UHI is strongly supported by a quantitative analysis of pressure perturbations (Fig. 5.20). Since the simulated environment is favorable for supercell storms, the triggered convective cells evolve into an urban-induced supercell. Although in our experiments this scenario occurs only with a 30 km wide city and an urban fraction of 1.0, we believe that it could also occur with smaller cities and lower urban fractions. The key requirement may simply be a sufficiently strong UHI combined with an environment conducive to supercells. In our numerical simulations, the threshold for the development of an urban-induced supercell corresponds to a UHI of 1.5 K, which is not particularly high.

Summing up, the city exerts a dual influence on supercells: it can weaken or dissipate an incoming supercell when the UDI is sufficiently strong (about 1 gkg^{-1} in our case), or it can generate a new urban-induced supercell if the UHI is strong enough. There is some observational evidence supporting these numerical results. The radar-based climatological study of Kingfield et al. (2018) found that, for four Great Plains cities in the United States, reflectivity values of approaching storms were higher upwind than downwind. The opposite pattern was observed when selecting an urban-favorable subset of nonsupercells occurring during summer under weak synoptic forcing. Similarly, the radar-based study of Lorenz et al. (2019) found that storms striking Berlin were more frequently weakened than intensified. Thus, in strongly forced environments, it is not uncommon for approaching storms to be dissipated or weakened by cities, consistent with our numerical simulations. This scenario has received relatively little attention in the literature, as most studies have focused on the potential of cities to generate urban storms and enhance precipitation (e.g., Yang et al. 2024; Shen and Yang 2023; Shepherd 2013; Shepherd et al. 2002; Bornstein and Lin 2000; Changnon Jr et al. 1976). The latter outcome is likely common in weakly forced synoptic environments and can also occur at times under strongly forced conditions. However, the weakening of severe storms by the urban land use is an equally important outcome that should be emphasized. Similarly, the capability of cities to generate their own supercell, as highlighted in this work, has not been emphasized in past studies, even though Lin et al. (2021) documented the intensification of a supercell downstream of Kansas City, and Reames and Stensrud (2018) reported a consistent supercell intensification in ensemble simulations north of Dallas-Forth Worth.

An additional relevant difference from previous works is that the urban-induced storm modifications in our simulations are exclusively the result of thermodynamic changes in the PBL, caused either by the UHI or the UDI. The direct effect of buildings on storms is negligible in our experiments, even with building heights of 60 m. In these cases, the *urban cyclone* is still induced by the UHI (Fig. 5.20), and the vertical velocities generated by 60 m tall buildings are much weaker than those associated with the UHI-

induced *urban cyclone* (Fig. 5.23d). Furthermore, building height is not correlated with higher values of low-levels vertical velocities in the simulated supercell (Table 5.15). Several studies have suggested that buildings may significantly influence urban rainfall, either through flow merging downstream of the city (Shen and Yang, 2023; Dou et al., 2015) or through dynamical blocking of the flow (Dou et al., 2024; Shen and Yang, 2023; Yang et al., 2021a; Zhang et al., 2017). Other authors have proposed that buildings can create a pressure minimum through their dynamic effect on surface flow (Yang et al., 2021b), or even induce storm splitting via the so-called *building-barrier effect* (Dou et al., 2015; Bornstein and Lin, 2000). None of these hypotheses are supported by our simulations. In particular, quantitative analyses of dynamic and buoyancy pressure perturbations indicate that the influence of buildings on downstream flow merging and on the formation of the *urban cyclone* is negligible. The building-barrier effect was explicitly tested with an ad-hoc experiment using an NYC-like city and was also rejected. To the authors knowledge, this is the first time that the dynamic effect of buildings on convective storms has been shown to be marginal in strongly forced environments. Nevertheless, the *building-barrier effect* could still play a role in other strongly forced situations not explored in these idealized simulations, such as QLCS events with intense outflows (see, for instance, Chapter 4).

The marginal effect of buildings on convective storms is likely valid for most atmospheric environments. If building height has little impact on storms under relatively strong surface flow, its influence should be even less relevant in weakly forced environments. For this reason, we argue that our conceptual model can be extended to such environments: in these cases, urban storms are triggered by wind convergence generated by the UHI-induced *urban cyclone*, while the dynamical effects of buildings play only a minor role (Fig. 5.25b). Under weak flow conditions, the *urban cyclone* is not advected downstream but remains centered over the urban area (e.g., Kawabata et al. 2025; Shen and Yang 2023). This interpretation is consistent with the pioneering work of Rozoff et al. (2003), who showed that the development of a convective storm over St. Louis, Missouri, was primarily driven by the UHI, while the urban momentum flux was negligible. More recently, similar results were obtained by Forster et al. (2024) and Kawabata et al. (2025) for a case study in Paris, and by Swain et al. (2025) for New York City.

An additional interesting outcome of this work is the sensitivity of the simulated supercell to surface conditions. Relatively small differences in temperature (1 K) and moisture (1 g kg^{-1}) are sufficient to markedly alter supercell behavior. A direct implication of this result is that operational weather forecasting models must be driven by up-to-date land use data. For instance, changes in vegetation caused by drought, if not assimilated into the models, can lead to significantly inaccurate forecasts. Similar conclusions were reached by Zhang et al. (2025), who studied the sensitivity of MCS to irrigation in the US Corn Belt, and by Reames and Stensrud (2018), who examined the effects of urban land use on a supercell.

Lastly, green roofs showed great potential for suppressing the UHI and the UDI, as already well established (e.g., Wang et al. 2022). The novelty of this study lies in showing that while a city with limited vegetation can weaken or dissipate an incoming storm, a city with widespread green roofs exerts only a marginal influence, allowing the original storm to maintain its intensity. We do not consider this a reason to discourage increasing vegetation in cities. In the worst-case scenario of a fully green city, given similar atmospheric environmental conditions over the city and the surrounding countryside, the incoming storm would simply maintain the same intensity it had over rural areas, without intensifying.

Although the analysis presented here is robust, several limitations remain that should

be addressed in future work. In the model employed, momentum and heat exchanges between urban land use and the atmosphere are parameterized using the [BEP-BEM](#) schemes. While reliable and state-of-the-art for urban mesoscale modeling, they do not explicitly resolve the underlying physics. Small-scale urban processes - which may, in some situations, strongly influence urban storms (e.g., under shallow surface-based inflow layers) - are not directly represented. [LES](#) simulations should be conducted to assess the importance of these processes for convective storms and to test whether the results of this study hold under that approach. Another aspect not investigated here but equally important for future research includes the effects of urban land use on nocturnal events. Furthermore, idealized simulations of a [QLCS](#), rather than a supercell, interacting with a city should also be carried out to evaluate whether building effects on the gust front may dominate over the [UHI](#) and [UDI](#) in such cases.

Chapter 6

Conclusions

The thesis analyzed the effect of complex topography and urban land use on high-impact weather events, with a special emphasis on local severe storms. Several benefits are associated with this line of research, ranging from improved early warning systems and more effective emergency plans to increased public preparedness for severe weather and strengthened socio-economic resilience. Considering the growing population living in urban areas, this topic is of compelling importance.

The understanding and prediction of high-impact weather in complex topography is challenging for several reasons, as reviewed in sections 1.2 and 2.3. Topography influences the dynamics of mesoscale phenomena, accurate and detailed observations are often insufficient to provide a complete picture of them, and numerical models struggle to produce reliable simulations (e.g., due to coarse resolution and limitations in physical parameterizations). Complex topography is interpreted in a broad sense, including not only mountains and water bodies but also urban land use. Urban areas are characterized by distinctive sensible and latent heat fluxes, and buildings modify the near-surface flow. Similar issues of mountains arise when studying the interactions between high-impact weather and urban areas. In urban environments an additional scale problem emerges (section 1.2): turbulence (~ 1 m) generated by buildings can influence mesoscale phenomena (~ 10 – 100 km), requiring the simulation of a broad scale of phenomena.

In the introduction, three scientific questions were posed, and the analysis carried out in this thesis provides answers to them. They are discussed point by point in the following paragraphs.

***Question 1:** How reliable are simulations of high-impact weather events in complex topographic regions?*

The performance of cutting-edge numerical models (ICON and WRF) was assessed in three case studies of windstorms: a downslope winter Bora event in Trieste (at the interplay between mountains and the Adriatic Sea, section 3.2), and two summer convective windstorms, one along the coast of Sicily (section 3.1) and one in the urban area of Milan, located in a plain surrounded by mountain ridges (section 4). The results showed that, overall, the models were able to reproduce the occurrence of severe windstorms with a good level of temporal and spatial accuracy. The physical dynamics of the downslope windstorm (the hydraulic jump) and of the convective windstorm (the downward propagation of a RIJ) were also successfully reproduced. However, some shortcomings were noted in the simulations of wind gust within the city of Milan (section 4.2.4) and in the Trieste case study (section 3.2.3). These discrepancies were attributed to the effects of local obstacles (e.g., trees and buildings) on the near-surface flow, which are not taken

into account in mesoscale simulations (section 3.2.3). Furthermore, mesoscale models are unable to resolve the very rapid and turbulent features of wind gusts close to the surface, systematically underestimating their intensity in inland regions (Cao and Fovell, 2018).

The predictability of windstorms was not specifically assessed. Nevertheless, it is worth mentioning that finding an initial condition capable of reproducing the storm that hit Milan was challenging (section 4.1.1). On the other hand, the convection-permitting ensemble analyzed for the *Bayesian* sinking case showed good probabilities of severe wind gust occurrence in the Southern Tyrrhenian Sea during the time window when the convective windstorm actually took place. Although recent years have seen improvements in the predictability of severe storms in regions with complex topography (De Martin et al., 2025c), this aspect remains a major challenge (Fischer et al., 2025).

Question 2: Can urban land use influence the development or behavior of local severe storms?

Yes, according to either a real case study analysis (Chapter 4) and idealized numerical simulations (Chapter 5), urban land use remarkably influences severe storm dynamics. In the Milan case study, the **QLCS** is shifted in the upwind region, its updrafts are intensified over the city while wind gusts are reduced, and the storm is split and delayed downstream of the city, producing a band of intensified precipitation (Figure 4.14). These results are consistent with previous studies that found an intensification of downwind precipitation (e.g., Yang et al. 2024) or downwind storm splitting (e.g., Bornstein and LeRoy 1990) and stronger updrafts over cities (e.g., Platonov et al. 2024). However, to the authors knowledge, this is the first time that the urban effect on a damaging convective windstorm generated by a **QLCS** has been assessed, whereas previous works focused on ordinary and less intense convective storms.

A different outcome is obtained when studying the effect of urban land use on a supercell using idealized simulations. A statistically significant weakening of the approaching supercell is observed, while in some ensemble members a new supercell is triggered downstream when a sufficiently large city is included in the domain.

Question 3: Which physical processes are responsible for these modifications?

A definitive answer to this question cannot be provided, as the Milan case study and the idealized simulations produce two different results. These are not necessarily conflicting and can be interpreted consistently.

In Milan, the observed storm modifications are attributed to the *barrier-building effect*, whereby buildings slow down the gust front and, due to mass conservation, intensify vertical motions over the city. This ultimately causes more intense precipitation downstream. By contrast, this mechanism is found to have a negligible impact in the idealized simulations of supercells. An *ad-hoc* numerical experiment is designed to test the *barrier-building effect* hypothesis for New York City, where it was originally proposed, and the hypothesis is rejected (section 5.2.7). Instead, most of the observed modifications are attributed to the thermodynamic modification of the **PBL** by urban surfaces. The **UDI** weakens an approaching storm by reducing **CAPE**, while the **UHI** generates an *urban cyclone*, capable of inducing downstream wind convergence that can trigger a new urban-induced supercell when the **UHI** is sufficiently intense. A higher fraction of vegetation within a city (e.g., more parks or green roofs) weakens both the **UHI** and the **UDI**, thereby decreasing the urban effect on the convective storm.

The discrepancies between the two studies can be explained by three factors. First, the storms under investigation are different: in a **QLCS**, cold pool propagation is an essential part of the storm dynamics (section 2.2.1), whereas a supercell is smaller and more

isolated, relying on the ingestion of a thermally buoyant rotating inflow (section 2.2.1). Thus, it is reasonable to expect that building-induced perturbations to the gust front play a greater role in [QLCSs](#) than in supercells. In supercells, by contrast, local thermodynamic conditions may have a stronger impact. Second, the first case study occurred in an environment favorable to elevated convection, i.e., the inflow was deep (Fig. 4.13) and not surface-based (Fig. 4.7). The idealized simulations, on the other hand, were intentionally designed with a surface-based inflow to maximize the urban effect (noting that surface-based convection is more common than elevated convection). Therefore, it is reasonable to expect that the observed modifications to low-level Θ_e in the Milan case study (Fig. 4.8) do not affect the approaching storm, because the air mass just above the city do not feed the storm, or if it does, it is mixed within a deep inflow layer not influenced by urban land use. In the idealized simulations, by contrast, the near-surface thermodynamic urban modifications (Fig. 5.18) do affect the storm. Lastly, it is worth emphasizing that the importance of the *barrier-building effect* in the Milan case study remains only a hypothesis that is not specifically tested¹. In contrast, the thermodynamic and dynamic effects of urban land use on the supercell were quantitatively investigated.

Summing up, it is believed that buildings may influence storm behavior in certain circumstances, such as those investigated in the Milan case study. But, in general, the thermodynamic alterations to the [PBL](#) induced by urban land use, namely the [UHI](#) and the [UDI](#), play a much more important role in weakening storms or initiating new convective cells via the *urban cyclone*. These results significantly advance our understanding of the interactions between urban land use and severe storms. The effect of buildings on convective cells is less pronounced than suggested by several previous works (e.g., Bornstein and LeRoy 1990; Niyogi et al. 2011; Yang et al. 2021a), while the role of thermodynamics is emphasized, as already proposed by some authors (e.g., Rozoff et al. 2003; Wang et al. 2025). However, those studies did not focus specifically on local severe storms. A further novelty is the highlighted importance of the [UDI](#) in creating a low- Θ_e island that reduces [CAPE](#), leading to the weakening of approaching storms. Although this outcome has been noted in a few observational studies (Kingfield et al., 2018; Lorenz et al., 2019), it has so far received limited attention from the scientific community. It is believed this is a relevant result that must be considered when discussing the interaction between local severe storms and urban areas.

This analysis sheds light on the understanding of severe storms in urban areas and, more generally, on the interactions between high-impact weather and complex topography. Several questions remain open, and new ones have arisen, requiring additional future work. Thus, this concluding paragraph is devoted to providing suggestions for future research directions that could further develop this line of study.

To better test the effect of buildings on linear convective storms and to more accurately quantify their impact on gust front dynamics, idealized simulations could be carried out using an [LES](#) approach similar to that employed by Rozoff et al. (2023). In their study, the effect of buildings on the intense winds associated with a landfalling tropical cyclone was investigated. A similar setup could be adopted here, but instead of forcing the domain with a hurricane wind profile, a convectively generated cold pool could be propagated across the domain. The effect of buildings on cold pool propagation and updraft enhancement could be evaluated more precisely, better resolving also the turbulence generated by buildings.

Another aspect that should be investigated in a future work is the effect of urban

¹It is not straightforward to test this in a real case study.

land use on nocturnal storms, using the same idealized setting adopted in Chapter 5. Typically, nighttime convection is elevated, and therefore near-surface atmospheric alterations induced by urban land use may have only a marginal effect on storms (Liu and Niyogi, 2019). On the other side, both the UHI and the UDI are much stronger at night than during the day. Hence, understanding the influence of urban land use on nocturnal storms is not straightforward and requires a dedicated study.

Furthermore, this thesis focused mainly on supercells and fast-moving QLCS. A future study employing idealized simulations could investigate the effect of urban land use on *parallel stratiform* QLCS, which are capable of producing flash floods. Moreover, instead of focusing exclusively on mid-latitude atmospheric conditions, one could adopt a tropical atmospheric profile (characterized by weak flow but extremely high moisture) in the idealized simulations to explore the impact of urban land use on tropical flash floods.

Lastly, the effect of aerosols on urban storms was not considered in this thesis. It is recognized that urban aerosols can significantly influence storm development (Liu and Niyogi, 2019; Fan et al., 2020). A future study employing a similar idealized simulation setup as in Chapter 5 could investigate the impact of urban aerosols on local severe storms.

Regarding the noted weaknesses in model performance for simulating surface wind gusts in complex terrain settings (sections 3.2.3 and 4.2.4), a future study could focus on running CFD simulations initialized by a numerical weather prediction model. A ML model could then be trained on the CFD simulations and used to provide a more accurate representation of surface wind gusts (compared to the original mesoscale weather model) with a faster computation time (compared to the CFD model). While CFD simulations are considered a valuable tool to improve our understanding of atmosphere-building interactions, ML methods may be more useful for forecasting applications that needs feasible computational times.

One of the key takeaways from Chapter 3 was the importance of observations in complex terrain for understanding high-impact weather events and for better evaluating numerical weather prediction models, ultimately improving forecasts and severe weather warnings. For instance, high-quality radar data are not available everywhere in Italy, which limits both the understanding of weather phenomena and the effectiveness of early warning systems, as discussed in section 3.1. Therefore, countries should allocate funding to strengthen their observational networks. In addition to fixed instruments, field campaigns should be conducted to collect as much data as possible during high-impact weather events. Historically, much of the understanding of mesoscale and microscale weather phenomena has come from analyses of data collected during such campaigns. The ongoing TEAMX, RISKADAPT, and WIND RISK campaigns, as well as the planned TIM campaign, could be extremely beneficial in advancing our understanding of the impacts of high-impact weather events and complex topography. Future work exploiting these datasets may enable a significant step forward in the understanding and prediction of severe weather in complex terrain.

Appendix A

Pressure perturbations

The equation of pressure perturbations derived from the Boussinesq momentum equations is a powerful tool to understand many phenomena at the atmospheric mesoscale. Since this equation is not typically derived in Atmospheric Science programs in Italy, its explicit derivation is presented in this Appendix. The common partition of pressure perturbations in dynamic and buoyancy pressure perturbations, employed in section 5.3.1, is introduced. Lastly, an applications of this equation to explain the storm splitting and the deviant motion of supercells observed in Figure 5.7 is presented.

A.1 A diagnostic equation for pressure perturbations

To obtain a diagnostic equation for pressure perturbations, it is first necessary to define the total pressure p and density ρ as the sum of a homogeneous base state (p_0 and ρ_0 , respectively) and a perturbation from it (p' and ρ' , respectively).

$$p = p_0 + p' \quad (\text{A.1})$$

$$\rho = \rho_0 + \rho' \quad (\text{A.2})$$

It's necessary to introduce the buoyancy B :

$$B = -\frac{\rho'}{\rho_0}g \quad (\text{A.3})$$

where g is the gravity acceleration. Recall that the momentum equation along the vertical axis can be written as (Eq. 7.4 in De Martin 2020):

$$\frac{dw}{dt} = -\frac{1}{\rho_0} \frac{\partial p}{\partial z} + g = -\frac{1}{\rho_0} \frac{\partial p'}{\partial z} + B. \quad (\text{A.4})$$

where w is the vertical component of the wind. Then, consider the Boussinesq momentum equation for the atmospheric fluid:

$$\frac{\partial \vec{v}}{\partial t} + \vec{v} \cdot \nabla \vec{v} = -\frac{1}{\rho_0} \vec{\nabla} p' + B \vec{k} - f \vec{k} \times \vec{v} \quad (\text{A.5})$$

where f is the Coriolis parameter, v the wind vector and $\vec{k}$ the unit vector along the vertical direction. The divergence of the previous equation gives:

$$\frac{\partial(\vec{\nabla} \cdot \vec{v})}{\partial t} + \vec{\nabla} \cdot (\vec{v} \cdot \nabla \vec{v}) = -\frac{1}{\rho_0} \nabla^2 p' + \frac{\partial B}{\partial z} - \vec{\nabla} \cdot (f \vec{k} \times \vec{v}) \quad (\text{A.6})$$

The second term on the [LHS](#) of [A.6](#) can be written with the Einstein notation as $\partial_j(v_i \partial_i v_j)$, where sums are implicit. Then, applying the product rule it yields:

$$\partial_j(v_i \partial_i v_j) = \partial_j v_i \partial_i v_j + v_i \partial_i \partial_j v_j \quad (\text{A.7})$$

The second term on the [RHS](#) of [A.7](#) is composed by second derivatives. Since we are studying first order pressure perturbations, this term can be neglected. Consequently:

$$\vec{\nabla} \cdot (\vec{v} \cdot \vec{\nabla}) \vec{v} = \partial_i v_j \partial_j v_i \quad (\text{A.8})$$

Since we are considering an incompressible flow (Boussinesq approximation), $\vec{\nabla} \cdot \vec{v} = 0$. Thus, considering Equation [A.8](#), Equation [A.6](#) becomes:

$$\partial_i v_j \partial_j v_i = -\frac{1}{\rho_0} \nabla^2 p' + \frac{\partial B}{\partial z} - \vec{\nabla} \cdot (f \vec{k} \times \vec{v}) \quad (\text{A.9})$$

Evaluating the third term on the [RHS](#) of equation [A.9](#) yields:

$$\vec{\nabla} \cdot (f \vec{k} \times \vec{v}) = -f\zeta + \beta u \quad (\text{A.10})$$

where $\zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$ is vertical component of the vorticity and $\beta = \frac{df}{dy}$ is the beta parameter. This last term is small, even at the synoptic scale and can be neglected. Thus, considering Eq. [A.10](#) and neglecting the beta effect, Equation [A.6](#) becomes:

$$\frac{1}{\rho_0} \nabla^2 p' = -\partial_i v_j \partial_j v_i + \frac{\partial B}{\partial z} + f\zeta \quad (\text{A.11})$$

Making the first term on the [RHS](#) explicit, gives:

$$\frac{1}{\rho_0} \nabla^2 p' = - \left[\left(\frac{\partial u}{\partial x} \right)^2 + \left(\frac{\partial v}{\partial y} \right)^2 + \left(\frac{\partial w}{\partial z} \right)^2 \right] - 2 \left(\frac{\partial v}{\partial x} \frac{\partial u}{\partial y} + \frac{\partial w}{\partial x} \frac{\partial u}{\partial z} + \frac{\partial w}{\partial y} \frac{\partial v}{\partial z} \right) + \frac{\partial B}{\partial z} + f\zeta \quad (\text{A.12})$$

Equation [A.12](#) is a diagnostic equation for first order pressure perturbations in the Earth's troposphere, assuming the Boussinesq approximation.

At the synoptic scale the Coriolis force is the dominant term on the [RHS](#), giving:

$$\frac{1}{\rho_0} \nabla^2 p' = f\zeta \quad (\text{A.13})$$

Considering p' as a wavelike function $p' = p'_0 e^{i\Omega t}$, where Ω is a frequency, t the time, and i the complex unit, the Laplacian of p' is proportional to $-p'$. Thus, $p' \propto -f\zeta$. This gives the same qualitative result of the gradient wind balance: anticyclonic (cyclonic) flow is associated with high (low) pressure in the Northern Hemisphere.

In the next part of the derivation, the Coriolis term $f\zeta$ will be neglected, since the focus will be in the mesoscale.

A.2 Buoyancy and dynamic pressure perturbations

To obtain the partition of Equation A.12 in dynamic and buoyancy pressure perturbations, it's useful to introduce the rate of strain tensor e_{ij} :

$$e_{ij}^2 = \frac{1}{4}(\partial_j v_i + \partial_i v_j)^2 \quad (\text{A.14})$$

That can be expanded as:

$$e_{ij}^2 = \frac{1}{4}((\partial_j v_i)^2 + (\partial_i v_j)^2 + 2\partial_j v_i \partial_i v_j) \quad (\text{A.15})$$

Vorticity is defined as:

$$\vec{\omega} = \vec{\nabla} \times \vec{v} \quad (\text{A.16})$$

In Einstein notation:

$$\omega_i = \epsilon_{ijk} \partial_j v_k \quad (\text{A.17})$$

The square of vorticity is then:

$$|\vec{\omega}|^2 = \omega_i \omega_i = (\epsilon_{ijk} \partial_j v_k)(\epsilon_{ilm} \partial_l v_m) \quad (\text{A.18})$$

Recalling that:

$$\epsilon_{ijk} \epsilon_{ilm} = \delta_{jl} \delta_{km} - \delta_{jm} \delta_{kl} \quad (\text{A.19})$$

Then:

$$\omega_i \omega_i = (\delta_{jl} \delta_{km} - \delta_{jm} \delta_{kl}) \partial_j v_k \partial_l v_m \quad (\text{A.20})$$

Finally:

$$\omega_i \omega_i = \partial_j v_k \partial_j v_k - \partial_j v_k \partial_k v_j \quad (\text{A.21})$$

Adding Eq. A.21 in both sides of A.15, it yields:

$$-e_{ij}^2 + \frac{1}{2}|\vec{\omega}|^2 = -\frac{1}{4}((\partial_j v_i)^2 + (\partial_i v_j)^2 + 2\partial_j v_i \partial_i v_j) + \frac{1}{2}(\partial_i v_j \partial_i v_j - \partial_i v_j \partial_j v_i) \quad (\text{A.22})$$

Noting that $\partial_i v_j \partial_i v_j = (\partial_i v_j)^2$ three terms can be eliminated, and at the end you have:

$$-e_{ij}^2 + \frac{1}{2}|\vec{\omega}|^2 = -\partial_i v_j \partial_j v_i \quad (\text{A.23})$$

Then, Equation A.11 can be written as:

$$\frac{1}{\rho_0} \nabla^2 p' = -e_{ij}^2 + \frac{1}{2}|\vec{\omega}|^2 + \frac{\partial B}{\partial z} \quad (\text{A.24})$$

Dynamic pressure perturbations p'_d are defined as:

$$\frac{1}{\rho_0} \nabla^2 p'_d = -e_{ij}^2 + \frac{1}{2}|\vec{\omega}|^2 \quad (\text{A.25})$$

while *buoyancy* pressure perturbations p'_b are defined as:

$$\frac{1}{\rho_0} \nabla^2 p'_b = \frac{\partial B}{\partial z}. \quad (\text{A.26})$$

A.2.1 Hydrostaticity in buoyancy pressure perturbations

An interesting exercise is to prove that buoyancy pressure perturbations arise from both hydrostatic and non-hydrostatic components.

Indeed, another common partition of pressure perturbations is between *hydrostatic* (p'_h) and *non-hydrostatic* (p'_{nh}) pressure perturbations:

$$p' = p'_h + p'_{nh}. \quad (\text{A.27})$$

Hydrostatic pressure perturbations are defined as:

$$\frac{\partial p'_h}{\partial z} = -\rho' g. \quad (\text{A.28})$$

Then, evaluate the Laplacian of the definition of buoyancy pressure perturbations p_b (Eq. A.26):

$$\frac{1}{\rho_0} \left(\frac{\partial^2 p'_b}{\partial x^2} + \frac{\partial^2 p'_b}{\partial y^2} + \frac{\partial^2 p'_b}{\partial z^2} \right) = \frac{\partial B}{\partial z}. \quad (\text{A.29})$$

Then, compute the derivative along the vertical direction of the buoyancy defined in Eq. A.3:

$$\frac{\partial B}{\partial z} = -\frac{1}{\rho} \frac{\partial \rho'}{\partial z} g - \frac{\rho'}{\rho^2} \frac{\partial \rho}{\partial z} g \quad (\text{A.30})$$

Consider the first term on the RHS of Eq. A.30 and the third term on the LHS of equation A.29:

$$\frac{\partial^2 p'_b}{\partial z^2} = -\frac{\partial \rho'}{\partial z} g \quad (\text{A.31})$$

Hydrostatic pressure perturbations (Eq. A.28) are obtained integrating along the vertical Equation A.31.

$$\frac{\partial p'_b}{\partial z} = \frac{\partial p'_h}{\partial z} = -\rho' g \quad (\text{A.32})$$

Thus, hydrostatic pressure perturbations are a portion of p_b , while nonhydrostatic pressure perturbations comprises the rest of p'_b and p'_d .

A.3 The effect of dynamic pressure perturbations on supercell propagation

In this section, the diagnostic equation for pressure perturbations (Eq. A.11) is used to explain the splitting of supercell storms and their deviant motion. To do that, it's useful to obtain a slightly different version of Eq. A.11. The wind can be partitioned between an horizontally uniform mean wind ($\bar{u}(z), \bar{v}(z), 0$), and a perturbation component ($u'(x, y, z), v'(x, y, z), w'(x, y, z)$):

$$u(x, y, z) = \bar{u}(z) + u'(x, y, z) \quad (\text{A.33})$$

$$v(x, y, z) = \bar{v}(z) + v'(x, y, z) \quad (\text{A.34})$$

$$w(x, y, z) = w'(x, y, z) \quad (\text{A.35})$$

With the aforementioned wind partition, the [RHS](#) of equation [A.23](#) can be written as follow:

$$\partial_j v_i \partial_i v_j = \partial_j (\bar{v}_i + v'_i) \partial_i (\bar{v}_j + v'_j). \quad (\text{A.36})$$

Evaluating the product:

$$\partial_j v_i \partial_i v_j = \partial_j \bar{v}_i \partial_i \bar{v}_j + \partial_j \bar{v}_i \partial_i v'_j + \partial_j v'_i \partial_i \bar{v}_j + \partial_j v'_i \partial_i v'_j. \quad (\text{A.37})$$

The first term on the [RHS](#) is zero, because $\bar{u}$ and $\bar{v}$ do not depend on x and y , and $\bar{w} = 0$. For the same reason, some derivatives in the second and third terms on the [RHS](#) are null. The only derivatives that are not null in the second and third terms on the [RHS](#) are:

$$\partial_j \bar{v}_i \partial_i v'_j + \partial_j v'_i \partial_i \bar{v}_j = 2 \left(\frac{\partial w'}{\partial x} \frac{\partial \bar{u}}{\partial z} + \frac{\partial w'}{\partial y} \frac{\partial \bar{v}}{\partial z} \right) \quad (\text{A.38})$$

Introducing the shear vector $\vec{S}$:

$$\vec{S} = \left(\frac{\partial \bar{u}}{\partial z}, \frac{\partial \bar{v}}{\partial z} \right) \quad (\text{A.39})$$

equation [A.38](#) can be written as:

$$\partial_j \bar{v}_i \partial_i v'_j + \partial_j v'_i \partial_i \bar{v}_j = 2(\vec{S} \cdot \nabla_h w'). \quad (\text{A.40})$$

Considering Eq. [A.23](#), the fourth term on the [RHS](#) of Eq. [A.37](#) can be written as:

$$\partial_j v'_i \partial_i v'_j = e'_{ij}{}^2 - \frac{1}{2} |\vec{\omega}'|^2 \quad (\text{A.41})$$

Consequently, with this partition of the wind, the equation of pressure perturbations [A.11](#) becomes:

$$\frac{1}{\rho_0} \nabla^2 p' = -e'_{ij} + \frac{1}{2} |\vec{\omega}'|^2 - 2\vec{S} \cdot \nabla_h w' + \frac{\partial B}{\partial z} \quad (\text{A.42})$$

Evaluating the first and second terms on the [RHS](#) in Equation [A.42](#) (similarly to Eq. [A.12](#)) gives:

$$\begin{aligned} \frac{1}{\rho_0} \nabla^2 p' = & \left[\left(\frac{\partial u'}{\partial x} \right)^2 + \left(\frac{\partial v'}{\partial y} \right)^2 + \left(\frac{\partial w'}{\partial z} \right)^2 \right] + 2 \left(\frac{\partial v'}{\partial x} \frac{\partial u'}{\partial y} + \frac{\partial w'}{\partial x} \frac{\partial u'}{\partial z} + \frac{\partial w'}{\partial y} \frac{\partial v'}{\partial z} \right) \\ & - 2\vec{S} \cdot \nabla_h w' + \frac{\partial B}{\partial z}. \end{aligned} \quad (\text{A.43})$$

$2\vec{S} \cdot \nabla_h w'$ is the linear dynamic pressure perturbation term, while the first six terms on the [RHS](#) are the non-linear dynamic pressure perturbations terms. The first three terms on the [RHS](#) are called fluid extensive term. These terms are maximized in the central axis of an updraft, and they do not influence the storm propagation and splitting, as proved by Rotunno and Klemp (1982). The last term on the [RHS](#) is the buoyancy term, and it is also not important for storm splitting and propagation because buoyancy pressure perturbations are located over the updrafts. Lastly, the fourth, fifth and sixth terms on the [RHS](#) are the fluid shear terms: these terms cause the storm splitting.

Consider for instance the fourth term:

$$2 \frac{\partial v'}{\partial x} \frac{\partial u'}{\partial y} = 2 \partial_x v' \partial_y u'. \quad (\text{A.44})$$

Rearranging:

$$2 \partial_x v' \partial_y u' = -\frac{1}{2} (-\partial_x v' \partial_y u' - 2 \partial_x v' \partial_y u' - \partial_x v' \partial_y u') \quad (\text{A.45})$$

A rotating updraft can be approximated as a cylindrical, rotating fluid in which the relation $\partial_x v' = -\partial_y u'$ holds. Then:

$$\begin{aligned} 2 \partial_x v' \partial_y u' &= -\frac{1}{2} (\partial_x v' \partial_x v' - 2 \partial_x v' \partial_y u' + \partial_y u' \partial_y u') \\ &= -\frac{1}{2} (\partial_x v' - \partial_y u')^2 \\ &= -\frac{1}{2} \zeta^2 \end{aligned} \quad (\text{A.46})$$

This term is responsible for the storm splitting. When an updraft develops in an environment with a straight hodograph, vertical velocities tilt the vorticity into the vertical (as described in section 1.4 of De Martin 2022). A couplet of vorticity maxima and minima is located on the flanks of the updraft, to the right and left of the mean wind¹. Equation A.46 shows that the displacement of the vorticity couplets at mid-levels is associated with a couplet of pressure minima on the two flanks of the updraft. These mid-level pressure minima produce two upward-directed non-hydrostatic pressure gradient forces, which drive the updraft to propagate toward both the right and the left (see Fig. 8.35 in Markowski and Richardson 2011). This leads to the splitting of the initial updraft: one updraft moves to the right of the mean wind, and the other to the left.

After the split, the RM storm ingests more streamwise vorticity because of its more southerly propagation, becoming a supercell. As long as there is some degree of cross-wise vorticity, ζ' and w' are not collocated. Consequently, non-linear dynamic pressure perturbations continue to force the storm to propagate to the right.

In a curved hodograph, the linear term $2\vec{S} \cdot \nabla_h \vec{w}'$ produces a mid-level pressure minimum on the right side of the updraft and a mid-level pressure maximum on the left side (see Fig. 8.40b in Markowski and Richardson (2011)). The resulting upward-directed dynamic vertical pressure perturbation forces drive the propagation of the right-mover supercell to the right, while suppressing the left-mover supercell. In any case, with a curved hodograph, the left-mover storm ingests less streamwise vorticity after the splitting, hindering its transition into a supercell. Weisman and Rotunno (2000) argued that even with a curved hodograph, the non-linear dynamic pressure perturbation term is the dominant factor in the rightward propagation of the supercell, because the tilting process produces an offset between the ζ' and w' in the lower levels (Fig. 15 in Weisman and Rotunno (2000)). Note also that linear dynamic pressure perturbations result from the interaction between an updraft and a wind profile with a curved hodograph, but they do not require the presence of vorticity in the updraft.

In conclusion, non-linear dynamic pressure perturbations are responsible for both storm splitting and the deviant motion of a supercell, as long as there is an offset between ζ' and w' . Linear dynamic pressure perturbations, on the other hand, lead to the dissipation of the left-mover supercell in an environment with a curved hodograph and promote the intensification of the right-mover supercell.

¹In a straight hodograph, vorticity is perpendicular to the mean wind.

Appendix B

Added value of an interdisciplinary Ph.D. program

This Ph.D. thesis is part of the *Future Earth, Climate Change and Societal Challenges* program, also known as *Frontiers*. *Frontiers* is an interdisciplinary program composed of five research themes and curricula: Curricula A, The Earth System; Curricula B, Impacts, adaptation and vulnerability; Curricula C, Technological innovations for a decarbonized society; Curricula D, Socio-economic and legal studies for mitigation of climate change; Curricula E, One Health. While these curricula share the same overarching goal, namely, helping society to face, adapt to, and mitigate the risks associated with climate change, they are linked to very different scientific fields, ranging from Physics to Social Sciences, and from Engineering or Chemistry to Law and Medicine.

Being part of such a program has provided several benefits. First, attending lectures from professors belonging to very different departments in the first year gave us the opportunity to explore new scientific challenges beyond our original background (for instance, the relationship between diseases and climate change) as well as novel approaches to scientific problems (such as the involvement of local communities in addressing the impacts of climate change). This cross-fertilization among disciplines was reinforced by discussions with fellow graduate students from other fields.

Second, at the end of each year we were required to present our results to a committee of the Ph.D. program, composed of professors from diverse disciplines. This compelled us to work hard on the dissemination aspect of our research, making it understandable to a broad audience. Such a skill cannot usually be practiced in a scientific conference, where the audience typically consists of experts in the same field. Developing this additional soft skill throughout the years can strongly benefit our future careers, both in convincing stakeholders and citizens of the importance of our research for society and in securing research funding.

Lastly, during the annual seminars we received highly constructive feedback from the committee professors. Very often, this feedback highlighted aspects we had not previously considered, since it came from scholars only marginally connected to our field. Such external criticism and suggestions helped reveal limitations and strengths of our work that we ourselves could not identify. This debate with academics from different departments greatly contributed to strengthening our research and to better emphasizing the key points with potential societal impact.

In conclusion, an interdisciplinary Ph.D. program presents additional challenges compared to more traditional ones. However, these challenges help graduate students enhance their research: they gain new ideas and methods from peers and professors in

other departments; they practice communicating their results to a broader, non-specialist audience; and they receive constructive criticism from academics outside their field, thereby improving the societal relevance of their research. Ultimately, a Ph.D. candidate trained in such a program is better prepared to address contemporary challenges, which are characterized by high levels of complexity and interactions among multiple domains, and which therefore require a holistic and systemic approach.

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