Numerical Investigation Of Full Scale Thunderstorm Downbursts: A Parametric Study And Comparison To Meteorological Model

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Graduate Program in Mechanical and Materials Engineering
A thesis submitted in partial fulfillment of the requirements for the degree in Master of Engineering Science
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Abstract

A series of simulations using the Large Eddy Simulation with an atmospheric meteorological cloud model have been carried out to investigate the important geometric and thermal parameters that influence a thunderstorm downburst outflow, as it pertains specifically to the idealized cooling source model. A separate set of Large Eddy Simulations make use of the same idealized cooling source model, in a realistic atmospheric base state using real field sounding data, in an attempt to make a quantifiable comparison to a downburst from a full cloud simulation. Randomness has been added to the cooling source forcing function to mimic the thermal variation in a real thunderstorm. It is found that the initial source parameters of the cooling source have a strong influence on the downburst outflow wind field metrics. It is also shown that scaling such events in size and height above the ground results in outflow wind field quantities that are comparable when using a scaling method for liquid drop release experiments. Additionally, it is found that using the cooling source model in a realistic atmospheric base state results in outflow wind characteristics that are more typical of sophisticated cloud models. Peak outflow wind speeds occur at a comparable height and overall magnitude, and the vertical profile of radial wind speed takes on a similar shape.

Keywords

Downburst, Thunderstorm, Outflow, Wind Loading, Localized High-intensity Wind, Microphysics, Cooling Source, Cloud Model, Scaling, Simulation
Co-Authorship Statement

This thesis has been prepared in a manner conforming to the specifications of an Integrated Article format thesis by the School of Graduate and Postdoctoral studies at the University of Western Ontario. This thesis contains previously published work as well as other co-authored articles that are in the process of being published. The following statements detail the type of involvement and amount of contribution of each of the co-authors in the individual chapters. Computational resources and data processing expertise for all simulations were made possible through L.G. Orf’s efforts.

Chapter 2:

The simulations conducted for this parameter study were designed and carried out by C. Oreskovic, with the assistance of L.G. Orf and E. Savory. L.G. Orf consulted on model parameter selection. The text was written by C. Oreskovic with some assistance from E. Savory and L.G. Orf. Data output would not have been made possible without L.G. Orf’s expertise and familiarity with CM1.

Chapter 3:

The simulations performed for this study were carried out by C. Oreskovic with assistance from L.G. Orf. Model parameters, including the recommendation to perform calculations using a 6th order explicit advection scheme, were selected with assistance from L. G. Orf. The cooling source random temperature perturbation code was written and incorporated into the model by L.G. Orf. The text was written primarily by C. Oreskovic with some assistance from E. Savory and L.G. Orf.

Appendix A:

The figures included in this appendix are taken from Orf et al. (2014) and Oreskovic et al. (2015). C. Oreskovic performed most of the analysis and construction of the quantitative plots, with assistance and guidance from E. Savory and L. G. Orf. L. G. Orf is responsible for the qualitative contour plots that show potential temperature of the downflow.
Acknowledgments

I would first like to thank my supervisor, Dr. Eric Savory, for the enormous help and support that he provided throughout the duration of this project. I would also like to thank him for giving me the opportunity to work part-time with the Advanced Fluid Mechanics Research Group during my undergraduate degree, where I was introduced to this project and more importantly academic research in general. I would never have been able to grow my mind in such a way without the opportunities I was given by him.

I would also like to thank Dr. Leigh Orf, who provided me with the meteorological insight into the project that I never would have gained without his guidance. I am also gracious for the computational resources that were provided to me with his assistance, and the time he spent teaching me how to run the necessary software for this work.

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Dedication

I would like to dedicate this work to my parents, for their ongoing love and support. Without them none of this would have been possible.
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<td>AGL</td>
<td>Above Ground Level</td>
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<tr>
<td>ARPS</td>
<td>Advanced Regional Prediction System</td>
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<td>CAPE</td>
<td>Convective Available Potential Energy</td>
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<td>CM1</td>
<td>Cloud Model 1</td>
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<td>DMC</td>
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<td>EF</td>
<td>Enhanced Fujita Scale</td>
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<td>RANS</td>
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<td>RAMS</td>
<td>Regional Atmospheric Modelling System</td>
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<td>SHARCNET</td>
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<td>TKE</td>
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<td>WRF</td>
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<td>XSEDE</td>
<td>Extreme Science and Engineering Development Environment</td>
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<td>[m]</td>
</tr>
<tr>
<td>ADV</td>
<td>Advection operator</td>
<td>[-]</td>
</tr>
<tr>
<td>B</td>
<td>Buoyancy</td>
<td>[N]</td>
</tr>
<tr>
<td>Dₛ</td>
<td>Optional diffusive tendencies in the scalar equations</td>
<td>[-]</td>
</tr>
<tr>
<td>Dᵤ</td>
<td>Optional diffusive tendencies in the u equation</td>
<td>[m/s²]</td>
</tr>
<tr>
<td>Dᵥ</td>
<td>Optional diffusive tendencies in the v equation</td>
<td>[m/s²]</td>
</tr>
<tr>
<td>D₴</td>
<td>Optional diffusive tendencies in the w equation</td>
<td>[m/s²]</td>
</tr>
<tr>
<td>Ek₀</td>
<td>Kinetic energy scaling parameter</td>
<td>[J]</td>
</tr>
<tr>
<td>Ep₀</td>
<td>Potential energy scaling parameter</td>
<td>[J]</td>
</tr>
<tr>
<td>Eₜ</td>
<td>Total energy scaling parameter</td>
<td>[J]</td>
</tr>
<tr>
<td>Kh</td>
<td>Eddy diffusivity</td>
<td>[m²/s]</td>
</tr>
<tr>
<td>Km</td>
<td>Eddy viscosity</td>
<td>[kg/s•m]</td>
</tr>
<tr>
<td>Nₘ</td>
<td>Brunt-Väisälä frequency</td>
<td>[Hz]</td>
</tr>
<tr>
<td>Nₛ</td>
<td>Newtonian relaxation terms in the scalar equations</td>
<td>[-]</td>
</tr>
<tr>
<td>Nᵤ</td>
<td>Newtonian relaxation terms in the u equation</td>
<td>[m/s²]</td>
</tr>
<tr>
<td>Nᵥ</td>
<td>Newtonian relaxation terms in the v equation</td>
<td>[m/s²]</td>
</tr>
<tr>
<td>Nₜ</td>
<td>Newtonian relaxation terms in the w equation</td>
<td>[m/s²]</td>
</tr>
<tr>
<td>Q</td>
<td>Volume of dense fluid</td>
<td>[m³]</td>
</tr>
<tr>
<td>R</td>
<td>Ideal gas constant</td>
<td>[J/kg•K]</td>
</tr>
<tr>
<td>Re</td>
<td>Reynolds number</td>
<td>[-]</td>
</tr>
<tr>
<td>R₀</td>
<td>Length scale, equivalent spherical radius</td>
<td>[m]</td>
</tr>
<tr>
<td>Sᵢⱼ</td>
<td>Mean strain tensor</td>
<td>[-]</td>
</tr>
<tr>
<td>S²</td>
<td>Deformation</td>
<td>[m²]</td>
</tr>
<tr>
<td>Tₛ</td>
<td>Tendency from sub-grid turbulence in the scalar equations</td>
<td>[-]</td>
</tr>
<tr>
<td>Tᵤ</td>
<td>Tendency from sub-grid turbulence in the u equation</td>
<td>[m/s²]</td>
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<tr>
<td>Tᵥ</td>
<td>Tendency from sub-grid turbulence in the v equation</td>
<td>[m/s²]</td>
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<tr>
<td>Tₜ</td>
<td>Tendency from sub-grid turbulence in the w equation</td>
<td>[m/s²]</td>
</tr>
<tr>
<td>T₀</td>
<td>Time scale</td>
<td>[s]</td>
</tr>
<tr>
<td>Symbol</td>
<td>Description</td>
<td>Unit</td>
</tr>
<tr>
<td>--------</td>
<td>-------------</td>
<td>------</td>
</tr>
<tr>
<td>$U_r$</td>
<td>Radial component of wind</td>
<td>[m/s]</td>
</tr>
<tr>
<td>$V_0$</td>
<td>Velocity scale</td>
<td>[m/s]</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>Generic variable in the advection equation</td>
<td>[-]</td>
</tr>
<tr>
<td>$\varepsilon$</td>
<td>Turbulent kinetic energy dissipation rate</td>
<td>[J kg$^{-1}$ s$^{-1}$]</td>
</tr>
<tr>
<td>$\pi$</td>
<td>Non-dimensional pressure</td>
<td>[-]</td>
</tr>
<tr>
<td>$\pi'$</td>
<td>Non-dimensional pressure perturbation</td>
<td>[-]</td>
</tr>
<tr>
<td>$\rho_0$</td>
<td>Base state atmospheric density</td>
<td>[kg m$^{-3}$]</td>
</tr>
<tr>
<td>$\tau_{ij}$</td>
<td>Turbulent shear stresses</td>
<td>[-]</td>
</tr>
<tr>
<td>$\Delta$</td>
<td>Spatial filter size</td>
<td>[m]</td>
</tr>
<tr>
<td>$\Delta x$</td>
<td>Grid spacing in the east-west direction</td>
<td>[m]</td>
</tr>
<tr>
<td>$\Delta y$</td>
<td>Grid spacing in the north-south direction</td>
<td>[m]</td>
</tr>
<tr>
<td>$\Delta z$</td>
<td>Grid spacing in the vertical direction</td>
<td>[m]</td>
</tr>
<tr>
<td>$\theta_p$</td>
<td>Potential temperature</td>
<td>[K]</td>
</tr>
<tr>
<td>$\theta_{p0}$</td>
<td>Base state potential temperature</td>
<td>[K]</td>
</tr>
<tr>
<td>$\theta'$</td>
<td>Potential temperature perturbation</td>
<td>[K]</td>
</tr>
</tbody>
</table>
Chapter 1

1 Introduction

1.1 Atmospheric phenomenon

1.1.1 Deep Moist Convection (DMC) and downdraft formation

Thunderstorms, otherwise known as deep moist convection (DMC), are relatively small meteorological phenomena that involve the formation of a cumulonimbus cloud and are most often associated with strong winds, intense precipitation of various types, as well as other phenomena such as damaging downburst winds (Doswell 2000). A number of components are required in the atmospheric in order for deep moist convection to mature. These components include a sufficient level of moisture, a low level of static stability in the atmosphere and the ascent of parcels of air to a level of their free convection by a number of different mechanisms. The thunderstorm forms as a result of warm moist air rising due to the effect of buoyancy. As this air rises it thermodynamically cools due to expansion at lower pressure, as a result, the moisture in the air condenses to form the cumulonimbus cloud, as well as various forms of precipitation (Doswell 2000). The mode by which warm moist air rises can be different and, as a result, there are three primary types of deep moist convection (i) the orographic thunderstorm, where moist air is forced upwards due to some type of topographic obstruction (ii) the air mass thunderstorm, where moist air rises due to more localized convection in an unstable (strong vertical movement of air) base state (iii) and the frontal thunderstorm, where thunderstorms form along the boundaries of opposing weather fronts. The thunderstorm can be further classified into four distinct types of events. These are (i) the single cell storm (ii) the multicellular cluster (iii) the multicellular line which is also known as the squall line (iv) and the supercell storm (Fujita 1955, Fovell and Dailey 1995). Deep moist convection is the parent event of the thunderstorm downburst phenomenon and it is the microphysical processes and strong thermal convection within the parent storm that leads to the evolution of the downburst.
1.1.2 The downburst

The thunderstorm downburst (also known as a microburst or macroburst depending on their geometric size) is a volume of air that rapidly descends out of a thunderstorm cloud eventually impinging upon the earth’s surface and causing a highly divergent outflow of locally intense wind that has the strong potential to damage surface structures (Abd-Elaal et al. 2013), as illustrated in Fig. 1.1 and Fig. 1.2. Downburst formation is a density driven event, which is primarily caused by thermodynamic cooling often associated with the formation of the thunderstorm cloud itself. The thermodynamic cooling is the direct result of the evaporation of precipitation such as rain, snow, hail and graupel (Wakimoto 1985). These processes, which occur inside the thunderstorm cloud, effectively create a volume of air that is at a lower temperature than ambient, resulting in a negatively buoyant parcel that has a potential to fall towards the earth. In addition to the thermodynamic cooling, the drag induced by the falling of precipitation aids in the formation and strength of the downflow (Fujita 1985).
A downburst can be described as a three part process, as can be seen in the diagram from Fujita (1985) of the evolution of a downburst in Fig. 1.3.

The downburst exits the cloud at a typical altitude on the order of 1-3 km and moves towards the ground in a spatially complex downflow. Roll vortices are evident in the downflow, primarily the result of baroclinic vorticity generation (vorticity caused by misaligned gradients of pressure with density) as the downflow interacts with ambient air (Vermeire et al. 2011a, Bluestein 2007). Baroclinic vorticity is defined by,
\[
\frac{D}{Dt}(\nabla \tilde{v}) = [(\nabla \tilde{v}) \nabla] \tilde{v} - \nabla (\alpha \nabla p') + \nabla (B \hat{k}) \quad (1.1)
\]

For a single horizontal component,

\[
\frac{D}{Dt} \left( \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x} \right) = -\frac{\partial B}{\partial x} \quad (1.2)
\]

where \( k \) is a unit vector pointing in the vertical direction, \( p' \) is the pressure perturbation, \( \alpha \) is the specific volume, \( u \), \( v \) and \( w \) are wind velocity vectors, \( x \) and \( z \) are Cartesian directions in the horizontal and vertical planes, \( B \) is buoyancy and \( t \) is time. The downflow then impinges upon the ground, resulting in a radial outflow of potentially damaging winds with a characteristic magnitude of peak velocity of \( >10 \text{ m/s} \) (Orville 1989). A roll vortex forms, as can be seen in photographs such as those depicted in Fig. 1.4 and illustrated in Fig. 1.1. The eddies found in the outflow can have wind velocity magnitudes that are on the order of F3, which is defined as winds with a magnitude between 70.5 m/s and 92 m/s (Fujita 1981), and have the strong potential to damage man-made surface structures.
Figure 1.4 - Sequence of Photographs of a downburst taken in the field (top), enhanced view of the roll vortex (bottom), (Fujita 1985)
Thunderstorm downburst research originated with the concern to the aviation industry while observing this unique locally intense surface wind. One of the first instances of the downburst appearing in literature was in the study by Veimester (1961), this study focused on the unique wind patterns encountered by a BOAC Argonaut aircraft taking off from Kano airport in Nigeria. What was encountered was a strong headwind, followed by a downdraft. The term downburst however was originally introduced in (Fujita and Byers 1977), a study investigating a 25 minute period of intense winds at John F. Kennedy international airport in New York City in 1975. During the 25 minute period, 4 to 5 intense downdrafts were observed within the spearhead echo (a radar echo associated with downburst winds with an ‘arrow’ pointed towards the direction of the echo), coined as ‘downdraft cells’. Early research of the phenomenon was met with some skepticism, as to whether or not a difference existed between a so called downburst and a typical thunderstorm precipitation downdraft. Fujita performed a number of studies in the late 1970s and early 1980s exploring downburst activity in more detail. These studies included JAWS (Joint Airport Weather Studies), NIMROD (Northern Illinois Meteorological Research on Downbursts) and MIST (Microbursts and Severe Thunderstorms) (Orville et al. 1989). During these studies many more downburst events were observed, which led to a more detailed explanation of the phenomenon.

Downburst damage is unique from that of a typical synoptic (everyday, straight line) type wind or that of a tornadic wind front. Surface damage caused by a downburst outflow results in a radial damage pattern, where surface objects (trees, grass, etc.) are flattened radially outwards. This type of damage is especially evident in Fujita’s aerial survey of surface damage caused by the April 3rd 1974 wind outbreak, as shown in Fig. 1.5 (Fujita 1985).
Figure 1.5 - Starburst damage pattern unique to downburst winds (Fujita 1985)

As can be seen, the damage is spread radially outwards from the centre, a pattern that is not consistent with damage from typical synoptic winds or tornado gusts.

The physical size of a thunderstorm downburst is variable, ranging on the order of a few hundred metres, to much larger events a few kilometres across (Fujita 1985). Due to this range in size, the thunderstorm downburst is mainly classified into two groups, the microburst and the macroburst. According the (Fujita 1985) the macroburst is considered to be a downdraft event that has an impingement diameter that is greater than 4km with outflow winds lasting between 5 and 20 minutes (Wakimoto 1985), and the microburst is classified as an event that has a diameter of 0.8 km to 4 km (Wilson et al. 1984) with peak winds lasting between 2 and 5 minutes (Wakimoto 1985). The 4km upper limit of the microburst has no particular basis in meteorology or fluid mechanics, it is simply an arbitrary value established historically (Wolfson 1988). However, the lower limit of 0.8 km is defined as the largest value of the diameter of an eddy that would be experienced in aviation that would be considered turbulence (Wolfson 1988). The parametric study that will follow in later chapters will focus on the phenomenon known as the microburst. The microburst can be further classified into two more groups, the dry and wet microburst. The dry microburst is defined as a downflow that is accompanied by little or no rain and is typically associated with virga (rain column that evaporates or sublimates before reaching the ground) from altocumuli or high based cumulonimbi (Wakimoto 1985). The wet downburst is defined as a downflow which is accompanied by heavy rain during the period
of high winds and is typically associated with the precipitation shafts of thunderstorms (Wakimoto 1985). The driving mechanisms behind the two types of downbursts (dry and wet) are, however, unique. The dry downburst is principally driven by the thermodynamic cooling effect of evaporating precipitation. The wet downburst is more complex, driven by multiple different forcing mechanisms including evaporative cooling and other microphysical effects such as the precipitation interaction (such as drag) with air (Wakimoto 1985).

Thunderstorm downbursts are extremely difficult to capture in field measurements as they are highly unpredictable and spatially highly variable. The root cause of this large spatial and temporal instability is the variation of temperature, pressure, wind velocity and direction, and moisture content that is associated with the vertical position in the earth’s atmosphere. There are three principal conditions that are required for a thunderstorm and ultimately a downburst to form, these are an inherently unstable atmospheric base condition, a very steep thermal lapse rate (the decrease in air temperature as elevation increases) and high atmospheric moisture content (the relative humidity of the air) (Czernecki et al. 2014). Once a thunderstorm forms, the primary thermodynamic cooling in the storm as a result of high moisture content will promote the formation of a downburst.

1.2 Motivation of the thesis

High Intensity Wind (HIW) events are known to cause upwards of 80% of all catastrophic power transmission line failures in the North and South America, Australia and South Africa (Dempsey 1996). Similarly, in Australia from 2008 to 2013 greater than 40 different power transmission tower failures have occurred due to high intensity wind events (Abd-Elaal et al. 2013). Wind loads play a large role in the design of power line transmission structures, and the codes which govern the design of transmission line structures are primarily intended for large magnitude synoptic type winds (Shehata et al. 2005). Fig. 1.6 shows a transmission tower structure failure most likely caused by a High Intensity Wind event (CEATI 2015).
High Intensity Wind events, such as downbursts, display a unique vertical profile of horizontal wind speed that is not consistent with typical synoptic winds. Downburst wind profiles show a peak wind magnitude that is always at a lower vertical position than the synoptic wind maximum (although synoptic wind profiles do not have a true peak, they have a max value at gradient height), making the design constraints different for downburst events. In order to properly design such infrastructure to withstand these unique wind types, a better understanding of these winds need to be established. This unique wind profile can be observed in Fig. 1.7, which compares a downburst outflow wind profile to a typical conventional atmospheric boundary layer (although it should be noted that an ABL rarely reaches a constant value). As can be seen, the peak outflow wind velocity occurs at an altitude lower than a boundary layer flow. However, the shape of the downburst wind profile is not the only cause of transmission line failures due to downbursts. When a downburst event occurs on one side of a tower, an unbalanced loading causes a torsional moment on the span that the structure is not designed for (Lin et al. 2012). Also, well-correlated gust fronts from a downburst outflow is usually of a scale at least as large as a typical conductor span (Shehata et al. 2005).
The thunderstorm downburst is differentiated from another high intensity wind event, the tornado, by the orientation of the primary vortex as well as the origin of the vortex itself. In a tornado the primary vortex is oriented vertically, driven by a central low pressure region. As a downburst impinges upon the ground, a roll up vortex forms and expands stretching outwards, the winds below the vortex ring can reach a magnitude typically around 50 m/s (Fujita 1985). This difference can be seen in Fig. 1.8.

Figure 1.8 - Downburst and tornado vortex comparison (Fujita 1985)
1.3 Established wind engineering models

Most established numerical and physical models of the thunderstorm downburst phenomenon rely on overly simplified physics. Observably, thunderstorm downbursts are extremely complex events which involve microphysical thermodynamic processes in their formation. In the area of wind engineering, physical studies and numerical simulations of downburst events have been mostly limited to axi-symmetric, impulsively-driven impinging circular jet (IJ) models (Letchford 2002) and (Holmes and Oliver 2000), see Fig. 1.9 (right). However, these downburst models ignore realistic effects that are encountered in the natural environment such as the drag-induced downflow due to the presence of precipitation and the complex thermodynamic cooling in the cloud. In addition, the IJ model is fundamentally flawed in that it relies entirely upon the unrealistic forcing of an impulsively started jet which is not present in nature (Mason et al. 2009, Anabor et al. 2011, Vermeire et al. 2011a,b, Zhang et al. 2013). However, the impulsive IJ is the most popular model in the wind engineering community for a number of reasons. First, the type of flow itself is relatively simple and well understood in the fluid mechanics community (Abramovich 1963, Gautner et al. 1970, Donaldson and Snedeker 1971, Rajaratnam 1976). Second, the IJ model flow field is very scalable, as the peak outflow wind velocity magnitude and position can be well estimated by the initial jet speed and the jet diameter, respectively. The IJ model makes use of an impulsively driven downward facing flow out of a nozzle placed at some altitude above the surface, in an attempt to model the downflow and subsequent outflow of a natural event. The impinging jet model was first hypothesized to be a suitable simplification of the physics of a real event by Fujita (1985). The first IJ models seemed to have been by Landreth and Adrian (1990) and Selvam and Holmes (1992). More recently Kim and Hangan (2007) performed an unsteady computational fluid dynamics (CFD) simulation using an impulsively started IJ. Unfortunately, the IJ model lacks the realistic physics present in a natural event as the primary driving mechanism of the jet flow is not a density gradient caused by thermodynamic cooling, but instead an impulsively driven mass of air.

Experimentally, the liquid drop release method has been explored in some detail (Lundgren et al. 1992, Alahyari and Longmire 1994, Yao and Lundgren 1996), an example is shown
in Fig. 1.9 (left). The liquid drop release method makes use of a dense fluid suspended within a less dense ambient liquid. The more dense fluid is allowed to fall to the ground, and the flow spreads out radially. This is an improvement over impulsively driven methods, as the primary means of flow is caused by a buoyancy driven effect. Although, there still exists some limitations of this approach, most notably the density gradient is not thermodynamically caused or temporally dependent as it is in a natural event.

The other, more recently introduced and less popular model, is the cooling source (CS) numerical model. The CS model makes use of a pre-defined cooling source forcing function that models the thermodynamic cooling caused by the evaporation of precipitation in the thunderstorm cloud, in order to create a cold mass of air analogous to the cool downdrafts observed in nature. This model was first introduced in Anderson et al. (1992) in a meteorological sub-cloud model, and has been subsequently studied in Anderson et al. (1996) and Orf and Anderson (1999). More recent adaptations of the model have been made in Lin et al. (2007), where the CS forcing function of Anderson et al. (1992) was applied to the Cloud Model 1 (CM1) (Bryan and Fritsch 2002). Mason et al. (2009) also introduced the same CS forcing function in an axi-symmetric two dimensional simulation, with promising results. Despite the increased realism, because density is the driver of flow, the CS model has not been studied in as much detail as the IJ model. The main reason that the CS model has not become more popular in the engineering community is due to its spatial and temporal complexity, and the lack of clear scaling criteria up to the present time.
Most recently, the IJ model has been compared in detail to the more complex CS type model in Vermeire et al. (2011a), both models being used inside of the CM1 framework. It was found that the CS model much better predicted the outflow wind fields, as well as the means by which vorticity is generated within the downburst initial column. From this study it was recommended that all future modelling of thunderstorm downbursts make use of the CS model over the IJ model, as the IJ model completely lacks the physics present in natural events. Thus, the present work will focus exclusively on the CS model.

1.4 Current modelling approach

1.4.1 Cloud Model 1 (CM1)

The numerical model that will be used for this study is the Cloud Model 1 (CM1), a three-dimensional, time-dependent, non-hydrostatic Large Eddy Simulation (LES) model which was developed by Dr. George Bryan at The Pennsylvania State University as well as at The National Center for Atmospheric Research (Bryan 2011). CM1 is primarily used to simulate extremely high resolution thunderstorms, although not limited to this use, with
little memory overhead requirements (inherent to the efficiency of the code), allowing for an extremely large number of grid points (on the order of $10^9$). CM1 is optimized to run in a massively parallel environment, making use of the Message Passing Interface (MPI) on such systems as the Shared Hierarchical Academic Research Computing Network (SHARCNET) and the Texas Advanced Computing Center (TACC) among others. CM1 is unique from other meteorological cloud models, such as the Advanced Regional Prediction System (ARPS) (Xue et al. 2003), Regional Atmospheric Modeling System (RAMS) (Pielke et al. 1992) and the Weather Research and Forecasting Model (WRF) (Michalakes et al. 1998), as CM1 has a better ability to conserve mass and energy (although some aspects of the CM1 code is borrowed). This is done by better retaining all terms in the internal energy equation and the pressure equation, terms which are ignored in other previously mentioned meteorological models (Bryan and Fritsch 2002, Bryan and Rotunno 2009). CM1 is an open source code written primarily in Fortran90, available for open use available at Bryan (2014). Downbursts are simulated in the CM1 environment using an addition originally written for the Wisconsin Model Engine (WME) (Anderson et al. 1992, Orf et al. 1996, Orf and Anderson 1999), adapted to work with CM1 (Lin et al. 2007). The downburst CM1 addition makes use of an ellipsoidal cooling source function placed within the computational domain made to mimic the realistic thermodynamic effects present in real thunderstorm downbursts. More details of the CM1 downburst model will be included in later chapters of this thesis.

### 1.4.2 Full cloud simulation of thunderstorm downbursts

Chapter 2 and, more notably, Chapter 3 of this thesis will make a detailed comparison to the results from a full cloud thunderstorm simulation of a downburst. For the reader to fully understand that comparison, a brief outline of the simulation will be presented in this subsection as well as relevant figures in Appendix A. The full cloud simulation that the comparison will be made to was developed and analyzed in Orf et al. (2012, 2014) as well as in Oreskovic et al. (2015). Key details of the simulation will be briefly explained here, although the reader can find more detailed information in these above-mentioned publications.
The primary driving mechanism of the formation of thunderstorm downbursts is the evaporative cooling caused by the change in phase of certain types of precipitation (Fujita 1985). Further, the downflow is enhanced by the drag that is induced by falling precipitation, most commonly hail or graupel (small hail). Meteorological cloud model simulations are used to numerically develop thunderstorms in atmospheric base states that are conducive to their formation. In order to capture that same thermodynamic and physical complexity present in natural downburst producing thunderstorms, Orf et al. (2012, 2014) proposed a simulation that resulted in a downburst-producing thunderstorm that captured the spatial complexity of more natural events, within the same meteorological cloud model (CM1) discussed in subsection 1.4. This is not the first instance of a meteorological cloud model being used to replicate near surface winds of interest to wind engineers (Nicholls et al. 1993, Yamada and Koike 2011), although this particular simulation was carried out on an extremely high resolution domain that focused on the thunderstorm downburst phenomenon and included five orders of magnitude more calculations than typical meteorological models that study deep moist convective storms. Orf et al. (2012) found that the peak outflow wind velocities occurred well behind the area of the primary roll vortex, and time histories of outflow wind found strong fluctuating wind components indicating a high degree of spatial variability and a strong turbulent outflow structure. The current author continued this work in Orf et al. (2014) by performing a complete spatial circumferential analysis of the downburst outflow wind field, see Fig. A.1 of Appendix A. Orf et al. (2014) made use of the same thunderstorm downburst simulation, although focus was shifted away from looking at single instantaneous profiles at peak times, rather the focus was on a complete circumferentially averaged \((r,z)\) vector plane at the same times during the event. It was found that peak outflow wind velocities can be reasonably approximated by a simple multiplier of the mean outflow wind velocities (plotted against radial position), see Fig. A.7 of Appendix A. More recently, the current author continued work on this same simulation in Oreskovic et al. (2015), completing the same circumferential spatial averaging approach, although focusing on the complete temporal analysis of the wind field. It was found that the same linear approximation of peak outflow wind speeds can be applied to the temporal history of the outflow.
Cloud Model 1 (CM1) was used in Orf et al. (2012, 2014), making use of a 3rd order Runge-Kutta time differencing scheme, and a 5th order scheme for advection terms. Most notably for this type of simulation, a parameterization scheme was employed to model the microphysical interaction, specifically the Morrison double-moment scheme (Morrison et al. 2009), of precipitation particulate and the atmosphere. It is this inclusion of microphysics that differentiates these types of simulations from the more idealized impinging jet (IJ) model of thunderstorm downbursts, as well as the simplified thermodynamic based cooling source (CS) model. These types of simulations offer the most realistic model of a thunderstorm downburst that can be achieved at this time because this model directly makes an effort to replicate the thermodynamic cooling that is the observed cause of thunderstorm downbursts in nature. The simulation of Orf et al. (2012) made use of a model time step of $\Delta t = 1/3 \text{ s}$ with 10 acoustic substeps, and a model domain of 92x92x14 km (x, y, z) using a horizontal grid spacing of $\Delta x = \Delta y = 20 \text{ m}$. The vertical mesh was stretched with $\Delta z = 5 \text{ m}$, at the ground up to 95 m at the top boundary. This resulted in a total number of grid points of 716,800,000, far greater than most computational wind models. The numerical model was initiated in a horizontal homogenous atmosphere, with humidity, pressure, temperature and horizontal winds varying only in the vertical direction. The sounding used was based on Brown et al. (1982), which was taken in the field in conditions that were observed to be conducive to the formation of downbursts. This same atmospheric base state will be employed in Chapter 3 of this present CS work.

What should be taken away from this brief overview is that this full cloud simulation of a downburst is computationally and spatially far more complex than any downburst model that has been explored in the past. The results from the literature shows that simulating a thunderstorm downburst using a full scale meteorological model results in outflow wind fields that are very physically realistic and reasonably reliable when compared to the natural event. It has only very recently been made possible that simulations of this order of magnitude (spatially and computationally) have been made possible and the size of this simulation (Orf et al. 2012, 2014) should be noted. Obviously, these full-scale simulations are not reasonable for use as an engineering model and, thus, this thesis will explore the CS model and an attempt will be made to increase the realism of such idealized studies.
The figures of Orf et al. (2014) and Oreskovic et al. (2015), and some supplementary text can be found in Appendix A of this thesis. In order to fully understand the comparison made in Chapter 3, it is recommended that the reader becomes familiar with the literature of Orf et al. (2012, 2014) and Oreskovic et al. (2015).

1.5 Purpose of the thesis

This thesis is in collaboration with industry, made possible by funding from Hydro One Inc. and The Institute of Catastrophic Loss Reduction. Hydro One Inc. is interested in high intensity wind events, such as thunderstorm downbursts, as they pose a particular threat to ground structures such as power transmission line towers and other important utility infrastructure. Due to a thunderstorm downburst’s non-typical vertical profile of horizontal winds, designing for wind loading for these events is a particular challenge that has been explored in some detail (Shehata et al. 2005, Lin et al. 2012, Ladubec et al. 2012, Aboshosha and El Damatty 2015, Aboshosha et al. 2016). This type of research is extremely important for industry design in order to prevent structural failures in the field. This thesis is part of a wider research project completed for Hydro One Inc. alongside our colleagues at the Department of Civil and Environmental Engineering at Western University.

The primary purpose of this thesis is to conduct a parametric study on the important physical aspects (peak wind speeds, locations of peak wind speeds, etc.) of a downburst and to make useful comparison to real field data as well as realistic high resolution full scale thunderstorm simulations of Orf et al. (2012, 2014). Current modelling of thunderstorm downburst phenomena (impinging jet models) appear to ignore the realistic physics present in real world thunderstorms. This thesis intends to explore multiple aspects (scalability, physical improvements) of a more realistic cooling source model, eventually leading to more reliable statistical model that can be used in industry. The primary outcome of this work is to develop a simple, yet more physically accurate, thunderstorm downburst model.

A number of simulations will be conducted in the following chapters to provide evidence in support of answering a number of relevant research questions;
• Does a cooling source model more closely resemble field data than the established engineering models?

• How is the outflow of the downburst affected by modifying various parameters in the initial cooling source condition (i.e., the shape, size, cooling intensity and temporal cooling function)?

• How does an idealized cooling source model compare to more sophisticated full cloud model simulations when the CS forcing function is placed in a realistic atmospheric base state?

• What additions can be made to a CS model to increase the physical realism, without increasing computational requirements?

1.6 Organization of the thesis

The following chapters will be organized as following, which will include two articles. The first chapter will investigate the outflow dynamics of a downburst in an idealized full scale cooling source model, by means of a parametric study. The second chapter will more closely examine an idealized cooling source model by comparison to a more realistic, full-scale, extremely high resolution meteorological thunderstorm simulation, in the same atmospheric base state. The conclusions chapter of this thesis examines the results established in the previous chapters, makes recommendations to future simulations and discusses the overall impact of this work.

1.7 Summary

Thunderstorm downbursts are a type of intense wind that are the direct result of thermodynamic cooling in a typical thunderstorm cloud resulting in negatively buoyant air descending from the cloud and impinging upon the ground. Thunderstorm downbursts are a well-documented phenomenon that poses a particular threat to ground infrastructure such as power transmission structures. Currently, our physical and numerical modelling techniques lack the realistic physics present in the real world events. The current research aims to improve upon numerical modelling techniques of downbursts, by performing a parametric study on a cooling source model, and comparing the results to field data as well as to data from sophisticated full-scale thunderstorm simulations.
1.8 References


Fujita, T.T., 1955. Results of detailed synoptic studies of squall lines. Tellus. 4, 405-436.


Chapter 2

Preface

The following text is a version of a manuscript that is being prepared for submission to The Journal of Wind Engineering and Industrial Aerodynamics, as part of a more broad study of full scale thunderstorm downburst simulations that make use of Cloud Model 1 (CM1). This chapter contains supplementary text and equations that will not be submitted as part of the journal publication but which, instead, will be used specifically for this thesis.

Keywords: downburst, CM1, cloud model, cooling source model, parametric study, Lundgren scaling

2 A Full Scale Parametric Study of an Idealized Cooling Source Downburst Model

2.1 Introduction

Downbursts are downdrafts of air which descend out of a thunderstorm cloud, impinging upon the ground causing a radial outflow of wind (Fujita 1985). They are the result of thermodynamic processes in the thunderstorm, such as the formation of rain, snow, hail and other types of precipitation (Fujita 1985). The formation of precipitation results in thermodynamic cooling, where heat is removed from the entrained air, creating a large body of cooler more dense air within the cloud which descends to the earth’s surface due to negative buoyancy. Additionally, the drag which is induced by the falling of this precipitation also aids in the evolution and strength of the downburst (Orf et al. 2012). The winds which result from this have enormous potential to damage man-made structures on the ground, such as buildings and power transmission line structures (Kim and Hangan 2007, Aboshosha and El Damatty 2015, Aboshosha et al. 2016), and follow a wind speed profile which does not conform to those of the well-studied synoptic winds. The difference from synoptic winds makes designing structures for this type of wind loading a particular challenge. A characteristic downburst descends out of the cloud producing a primary roll
vortex due to baroclinically generated vorticity (Bluestein 2007, Vermeire et al. 2011a), the event then impinges upon the ground creating a secondary stronger roll vortex along the surface, travelling radially outward. A downburst has characteristically strong radial peak winds, as well as large positive and negative vertical winds within the roll vortex. Peak outflow winds within a downburst profile also occur at elevations much closer to the ground than the typical synoptic wind profile (Fujita 1985).

Proctor (1988) simulated a downburst by replicating the thermodynamic cooling in the atmosphere using environmental conditions observed during the Joint Airport Weather Studies (JAWS) project (Hjemfelt 1986, 1988) and initiated the initial downdraft by specifying precipitation at the top of the domain and allowing it to descend. Early attempts to replicate downburst outflow winds by means of a simplified approach more analogous to modern engineering models were made by Selvam and Holmes (1992), where an impinging jet model was employed. The impulsively driven impinging jet (IJ) model of a downburst seems to originate from Fujita (1985), even though it lacks the realistic physics present in natural events because the primary mechanism driving the jet flow is not negative buoyancy but, rather, an artificial impulse of momentum. Additionally, the IJ model does not accurately capture the formation and evolution of the primary roll vortex, as the down flows of natural events are not steady state but transient. Kim and Hangan (2007) performed an Unsteady Reynolds-Averaged Navier-Stokes (URANS) simulation of an IJ and found that, although the outflow roll vortex formed, it was initiated by a Kelvin-Helmholtz instability at the shearing interface between the nozzle and ambient fluid, an artifact not observed in natural events.

As established in Anderson et al. (1992) and Vermeire et al. (2011a), any suitable numerical model that aims to accurately capture the outflow dynamics of a natural event should take into account the primary driving mechanism of the flow, buoyancy. The cooling source (CS) model, an idealized numerical approach, attempts to replicate the thermodynamic processes in the thunderstorm cloud by introducing a spatially and temporally dependent CS which ‘grows’ within the atmosphere. This approach appears to better replicate the primary means of vorticity generation and, as a result, more accurately replicates the peak outflow wind velocities (magnitude and shape of vertical profiles)
(Vermeire et al. 2011a). This present work focuses on the CS model, as previous studies suggest that the IJ model does not accurately represent the downburst evolution observed in nature. The more sophisticated numerical CS model that is investigated in this study was first introduced in Anderson et al. (1992), where a spatial and temporal CS was placed into a dry adiabatic atmosphere. The imposed thermal forcing functions in that study were originally estimated from the set of ice-phase cloud model simulations of Straka and Anderson (1992). The physical dimensions of the ellipsoidal CS function were approximated to represent the thermodynamic cooling region within the full cloud model thunderstorm simulation of Anderson et al. (1992). The more idealized CS sub-cloud model was run using the Wisconsin Model Engine (WME) (Anderson et al. 1992, Orf et al. 1996, Orf and Anderson 1999). WME is a reduced sound speed system (Anderson et al. 1986, Droegemeier and Wilhelmson 1987) introduced for the parallelization of computationally expensive meteorological simulations. Anderson et al. (1992) made use of a CS with a horizontal half width of 1200 m, vertical half width of 1800 m, a peak cooling forcing rate of -0.052 K/s and a vertical height of the centre of the CS of 2000 m. The cooling ramp-up function consisted of a 2 min cooling rate ramp-up period followed by a 10 min steady state cooling period, and then a 2 min ramp-down period to zero cooling rate. This simulation was run on a domain with 50 m grid spacing in all directions, using a 0.25 s advection equation time step. The Anderson et al. (1992) simulations consisted of two CS functions placed next to each other, with varying separation distance to represent the paired downdrafts observed in the full cloud simulation runs. The results from these simulations were promising as they closely represented the wind fields in the more realistic simulations, whilst the source of the flow was generated in a physically realistic way that matched the thermodynamic cooling present in natural events; a density perturbation in the cloud region. It was also found that agreement was reasonable when compared to the axisymmetric isolated downburst simulation of Proctor (1988). The \( \cos^2 \) spatially and temporally dependent CS model has since been employed in other studies. Citing the CS realism, Orf et al. (1996) performed another numerical study using the WME to examine in more detail colliding microburst outflows, by performing a parametric study investigating the effect of the spatial separation of two CS. Orf and Anderson (1999) studied the effects of horizontal translation of the CS function in a unidirectional sheared
environment. This was investigated because the thermodynamic cooling in a natural thunderstorm event is a function of the evaporation of local precipitation that translates with the environmental winds at the height of the cloud base.

More recently in Lin et al. (2007) the same CS forcing function code was carried over to Cloud Model 1 (CM1) (Bryan and Fritsch 2002), a more sophisticated cloud model specialized for simulations of deep moist convection (DMC) which can easily be modified for more simple sub-cloud model simulations like thunderstorm downburst winds. A modified CS ramp-up function peak value was used in that study to approximate the higher outflow wind speeds observed in some more intense natural thunderstorm downbursts, a magnitude of temperature perturbation four times greater than that presented in Orf and Anderson (1999). Lin et al. (2007) concluded that an idealized CS model is a practical simplification of the thermodynamic cooling in a natural thunderstorm as the various parameters of the source itself can be modified.

Vermeire et al. (2011a,b) made use of a similar CS model from the WME within CM1. Vermeire et al. (2011a) compared the model to an impulsively driven IJ model adapted to run within CM1, finding that the IJ model cannot capture the realistic buoyancy driven effects that are found in natural events, concluding that all further study of simplified downburst models should be conducted using the CS approach. Vermeire et al. (2011b) conducted a colliding downburst line outflow study, again using CM1, finding that colliding outflows result in unique wind fields with larger damage footprints and peak outflow velocities greater than those of a single event, particularly in the region of the colliding outflows caused by a burst swath. Vermeire et al. (2011b) recorded a 70% increase in the area where a surface structure would encounter damage due to the increased surface footprint of a downburst line event, and 55% increase in peak outflow radial wind speeds when compared to an isolated event. It was also found that the LES approach of CM1 resulted in more reliable data, when compared to the scale adaptive simulation (SAS) URANS simulations of Mason et al. (2009). The same CS function presented in Anderson et al. (1992) has also been used in other numerical studies including Anabor et al. (2011), which concluded that the sub-cloud LES CS model is capable of replicating the characteristic length and time scales present in full cloud simulations.
In downburst events there is large spatial variability within the thermodynamic cooling present in the thunderstorm. Cooling rate and the size and shape of the CS are all subject to atmospheric conditions, such as variation in temperature and wind shear. The CS model was investigated in Mason et al. (2009) in a parametric study that employed a similar CS approach to that of Anderson et al. (1992), but using a commercial software package. Various physical attributes of the CS were modified including the CS diameter, shape, forcing intensity, temporal downdraft characteristics, environmental lapse rate and surface roughness. It was found that the normalized peak outflow velocities were not greatly affected by changing the various parameters of the CS. However it was noted that the relationship between outflow velocities and the downdraft diameter are not linearly related as they are in the IJ model. Notably, the shape of the CS had a significant effect on outflow vortex development and so it was concluded that any future study should carefully consider the shape of the CS. Similar effects were observed for variations in other parameters such as the temporal characteristics of the ramp-up function and the elevation of the CS.

According to Pryor (2005) the atmospheric conditions specifically associated with thunderstorms that are favorable for downburst formation are a strong instability in the atmosphere, large amounts of Convective Available Potential Energy (CAPE) and the presence of a mid-tropospheric layer of dry air. CAPE is a measure of the energy per unit mass in the air that is available for free vertical convection, essentially, representing the positive buoyancy acting on a parcel of air. CAPE is defined as,

\[ \text{CAPE} = \int_{z_f}^{z_n} g \left( \frac{T_{v,\text{parcel}} - T_{v,\text{env}}}{T_{v,\text{env}}} \right) dz \]  

(2.1)

where \( z_f \) and \( z_n \) are the boundaries of the vertical domain, \( T_{v,\text{parcel}} \) is the temperature of the convecting parcel of air and \( T_{v,\text{env}} \) is the environmental air temperature. Typically, the value of CAPE is a strong indicator of a thunderstorm’s potential severity. Some attempts at predicting the intensity of a thunderstorm downburst have been proposed such as Wind Index (WINDEX), Dry Microburst Index (DMI) and \( \theta_e \) Deficit as well as the Wet Microburst Specific Index (WMSI), which incorporates \( \theta_e \) and CAPE measurements (Pryor
and Ellrod, 2004). In Pryor (2005) the Hybrid Microburst Index (HMI), intended to predict the magnitude of downburst convective winds, was developed, incorporating the sub-cloud temperature lapse rate and dew point depression. All of these models aim to predict the downburst peak outflow wind speeds and, effectively, the size and intensity of the thermodynamic cooling region, by considering the atmospheric conditions that cause their formation. This begs the questions: what do these models suggest for these relationships and can they be applied to simplified downburst models? These types of forecasting models aim to predict the severity of a downburst event by making a connection to the gross potential energy of the parent storm. Although CAPE has no quantifiable meaning in the CS model of a downburst, since a dry CS run has no moisture available for condensation and, thus, convection, this work aims to make a parallel between the meteorological HMI or WMSI and the relationship between CS initial potential energy and peak outflow wind speeds. Since it remains unclear in the downburst modelling community whether such a relationship exists, the present work investigates that connection.

Mason et al. (2009) investigated the outflow velocities, among other metrics, in a parametric study involving the CS model, but they did not offer a scaling approach for quantifying the effects of the physical changes of the source and their strong temporal dependence. Comparing the transient features of the downburst outflow can be particularly challenging. The purpose of the present parametric study is to determine if a relationship exists between the physical characteristics of the initial CS (size, shape, cooling rate, etc.) and the outflow wind fields. There exists substantial variation in the physical dimensions of the initial cold source of air within a downburst-producing thunderstorm and, thus far, very few (Lin et al. 2007, Mason et al. 2009, 2010, Vermeire et al. 2011b) CS studies have attempted to investigate this variation.

Currently, a reliable scaling method exists for the IJ model which linearly relates peak outflow wind speed and the spatial locations in the wind velocity field to the magnitude of the initial velocity and nozzle diameter of the impinging jet, respectively (Letchford and Chay 2002, Kim and Hangan 2007). Due to this simplicity of scaling, the IJ model has become enormously popular among the wind engineering community, despite its complete lack of physical realism. The present study seeks to investigate if a suitable scaling method
also exists for the CS model, relating the CS size, shape and cooling rate to outflow properties such as peak wind speed and bulk flow kinetic energy. This is of particular interest to wind engineers, since predicting the peak outflow wind velocity from the strength of the thunderstorm could aid in designing more suitable structures in areas where downburst producing thunderstorms occur. The scaling approach introduced in Lundgren et al. (1992) and Yao and Lundgren (1996) will be investigated, as well as a number of newly introduced unique scaling parameters. Lundgren et al. (1992) performed a physical downburst model study that involved the release of dense liquid parcels, into a less dense ambient fluid environment, which then impinged on a smooth horizontal surface. An inviscid scaling law was proposed for comparison between the transient features obtained from multiple experiments. It was found that this scaling law worked fairly well for simplified experiments where the fluid density changes abruptly between the source and ambient environment, as was the case in other liquid release experiments (Alahyari and Longmire 1994). The present work investigates whether the same scaling laws are applicable to the more spatially-complex CS model, where the density change between the source and the environment is more gradual.

The overall goal of this numerical study is to investigate the relative importance of the initial parameters of the CS, in terms of the general outflow characteristics of the downburst, in order to answer the following questions:

- Does the overall size of the CS have any impact on peak outflow wind velocity magnitudes, and their locations (both radial and vertical)?
- What is the effect of modifying the CS peak cooling rate at the geometric centre of the CS?
- Can a CS type downburst model be scaled in a similar way to the IJ model (i.e. a source diameter and height)?
- Can the Lundgren et al. (1992) scaling approach be applied to the more complicated spatially and temporally dependent CS model, or is it limited to more simplified constant density sources?
• How do the sizes of the near surface horizontal areas that experience Enhanced Fujita Scale EF0 and EF1 winds compare between the different simulations in this study?
• Does there exist a relationship between CS initial potential energy and peak outflow wind speeds, similar to the connection between CAPE and peak wind speeds in natural events?

Although there is significant spatial and temporal variability in the cold mass of air forming a thunderstorm downburst, the present study will quantify some cases that may be considered physically reasonable. The next section presents details of the numerical modelling approach, followed by the computational set-up, discussion of results and, finally, some concluding remarks and recommendations for future research.

2.2 Details of the numerical model

2.2.1 Model background and advection scheme

The numerical model used for this study is Cloud Model 1 (CM1) release 18 (cm1r18) (Bryan and Fritsch 2002), with some minor custom modifications made for surface roughness selection and data acquisition. CM1 is a model developed for atmospheric studies specifically involving the investigation of deep moist convection (DMC). The governing equations are discretized on an Arakawa C grid, where spatial (advection) derivatives are solved using a 5th or 6th order scheme and temporal terms are solved using a 3rd order Runge-Kutta (RK3) scheme. A 5th order (dependent on Courant number) spatial advection scheme was used for vertical scalars and velocities, whilst a 6th order explicit (kdiff,6=0.05) advection scheme (Wicker and Skamarock 2002) was selected for horizontal terms to maintain numerical stability, which showed an improvement over Vermeire et al. (2011a,b). It is recommended that all future studies using CM1, specifically CS simulations of downbursts, make use of a full numerical domain with four open radiative boundary conditions and a 6th or higher-order even explicit scheme for horizontal advection terms. The advantages of the explicit advection scheme is discussed in greater detail in sec. 4.3. A more detailed description of the governing equations can be found in the full model description of Bryan (2015).
2.2.2 Governing equations of the model

CM1 makes use of the spatially filtered momentum equations;

\[
\frac{\partial u}{\partial t} + c_p \theta_p \frac{\partial \pi'}{\partial x} = ADV(u) + f v + T_u + D_u + N_u \tag{2.2}
\]

\[
\frac{\partial v}{\partial t} + c_p \theta_p \frac{\partial \pi'}{\partial y} = ADV(v) + f u + T_v + D_v + N_v \tag{2.3}
\]

\[
\frac{\partial w}{\partial t} + c_p \theta_p \frac{\partial \pi'}{\partial z} = ADV(w) + B + T_w + D_w + N_w \tag{2.4}
\]

where ADV is the advection operator, T are the tendencies from sub-grid scale turbulence, D are tendencies from other diffusive processes that do not include sub-grid scale turbulence, N are the Newtonian relaxation terms and f terms are only included if Coriolis acceleration is considered. Finally, B represents the buoyancy force term defined below.

The terms u, v and w are the spatially filtered orthogonal velocity vector components in the directions x (east-west), y (north-south) and z (vertically), respectively (the spatial filtering will be described in greater detail later). \(C_p\) is the specific heat of dry air at constant pressure, \(\theta_p\) is defined as the potential temperature and \(\pi'\) is defined as the non-dimensional pressure perturbation from the base state.

The advection operator is given as the following, where \(\alpha\) is any generic variable,

\[
ADV(\alpha) = \frac{1}{\rho_0} \left[ \frac{\partial (\rho_0 u \alpha)}{\partial x} - \frac{\partial (\rho_0 v \alpha)}{\partial y} - \frac{\partial (\rho_0 w \alpha)}{\partial z} \right. \\
\left. + \alpha \left( \frac{\partial (\rho_0 u)}{\partial x} + \frac{\partial (\rho_0 v)}{\partial y} + \frac{\partial (\rho_0 w)}{\partial z} \right) \right] \tag{2.5}
\]

The governing equation for potential temperature perturbation, \(\theta'\), is given by,
\[
\frac{\partial \theta'}{\partial t} = ADV(\theta) + T_\theta + D_\theta + N_\theta + \frac{1}{c_p \pi} \varepsilon + q(x, y, z, t) \tag{2.6}
\]

where \(q(x, y, z, t)\) is the spatial and temporal CS forcing function introduced in a later section of the present work, and the potential pressure perturbation is determined by the following relationship,

\[
\frac{\partial \pi'}{\partial t} = -\frac{g}{c_p \theta_0} w + \frac{R}{c_v \pi} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) = ADV(\pi') + \frac{R \pi}{c_v \theta_0} (T_\theta + D_\theta + N_\theta) \tag{2.7}
\]

where \(\rho_0\) is the base density. Buoyancy, \(B\) in the above equations, is defined as the following,

\[B = g \frac{\theta - \theta_{\rho_0}}{\theta_{\rho_0}} \tag{2.8}\]

where \(\theta_{\rho_0}\) is the base potential temperature, and \(g\) is the gravitational constant.

### 2.2.3 Sub-grid turbulence model

CM1 offers two sub-grid scale (SGS) turbulence closures, a turbulent kinetic energy (TKE) model based on Deardorff (1980) and simpler Smagorinsky (1963) type model. The turbulent kinetic energy model of Deardorff (1980) was selected as it offers three critical advantages over the later, first sub-grid turbulence is only active in locally sheared conditions for the Smagorinsky (1963) model, second Smagorinsky (1963) makes use of the assumption that turbulence is steady and isotropic which is not ideal for lower resolution numerical grids, and finally the Smagorinsky (1963) scheme includes no stability dependence to the inherent length scales in the flow. The turbulence tendencies of the momentum equation \((T_i)\) are solved using the following;

\[T_i = \frac{1}{\rho} \left[ \frac{\partial \tau_{i1}}{\partial x} + \frac{\partial \tau_{i2}}{\partial y} + \frac{\partial \tau_{i3}}{\partial z} \right] \tag{2.9} \]
\[ T_v = \frac{1}{\rho} \left[ \frac{\partial \tau_{12}}{\partial x} + \frac{\partial \tau_{22}}{\partial y} + \frac{\partial \tau_{23}}{\partial z} \right] \] (2.10)

\[ T_w = \frac{1}{\rho} \left[ \frac{\partial \tau_{13}}{\partial x} + \frac{\partial \tau_{23}}{\partial y} + \frac{\partial \tau_{33}}{\partial z} \right] \] (2.11)

\[ T_s = \frac{1}{\rho} \left[ \frac{\partial \tau_{11}^s}{\partial x} + \frac{\partial \tau_{22}^s}{\partial y} + \frac{\partial \tau_{33}^s}{\partial z} \right] \] (2.12)

In the above, u, v and w represent the orthogonal component of wind, and s represents any scalar term (temperature or moisture). The stress terms in the sub-grid domain are solved using the following relation;

\[ \tau_{ij} = \rho u_i' u_j' = 2\rho K_m S_{ij} \] (2.13)

where \( \rho \) is the air density and \( K_m \) is the eddy viscosity, which is determined for this study using a turbulent kinetic energy closure. \( S_{ij} \) is the mean strain tensor;

\[ S_{ij} = \frac{1}{2} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) \] (2.14)

The turbulent fluxes for temperature within the computational domain, for the \( T_\theta \) equation are solved using;

\[ \tau_i^\theta = \rho u_i' \theta' = -K_h \rho \frac{\partial \theta}{\partial x_i} \] (2.15)

where \( K_h \) is the eddy diffusivity and \( \rho \) is the air density. The scheme for eddy viscosity and eddy diffusivity are similar to that of Deardorff (1980). \( K_m \) (viscosity) and \( K_h \) (diffusivity) are solved using the relationship;

\[ K_m = c_m l e^{1/2} \] (2.16)
\[ K_h = c_h l e^{1/2} \]  
(2.17)

where \( e \) in these equations is the sub-grid scale turbulent kinetic energy;

\[ e = \frac{1}{2} \overline{u_i' u_i'} \]  
(2.18)

\[ \frac{\partial e}{\partial t} = \text{ADV}(e) + K_m S^2 - K_h N_m^2 + \frac{1}{\rho} \frac{\partial}{\partial x_i} \left( 2 \rho K_m \frac{\partial e}{\partial x_i} \right) - \varepsilon \]  
(2.19)

\( \varepsilon \) is the dissipation, \( S^2 \) is the deformation, both given by;

\[ \varepsilon = c_\varepsilon e^{3/2} / l \]  
(2.20)

\[ S^2 = 2 S_{ij} S_{ij} \]  
(2.21)

\( l \), the characteristic length scale, is determined by the following relation;

\[ l = 0.8165 \left( \frac{e}{N_m^2} \right)^{1/2} \]  
(2.22)

\( N_m^2 \) is the Brunt- Väisälä frequency (also known as the buoyancy frequency), which is the frequency at which a parcel of air has a tendency to oscillate in a statically stable environment. In CM1 for sub-saturated air, the air in this parametric study, the frequency is formulated by the following;

\[ N_m^2 = \frac{g}{\theta_p} \frac{\partial \theta_p}{\partial z} \]  
(2.23)

And \( c_\varepsilon \) is determined to close the equations;

\[ c_\varepsilon = 0.2 + 0.787 \frac{l}{\Delta} \]  
(2.24)
The spatial filter size $\Delta$ for the computational domain is a function of the smallest three-dimensional sub grid volume;

$$\Delta = (\Delta x \Delta y \Delta z)^{1/3}$$  \hspace{1cm} (2.25)

In the meteorology community, it is common to define the surface roughness, $z_0$, as a function of the surface drag coefficient;

$$z_0 = \frac{10}{k} \frac{1}{e^{\sqrt{c_d}} - 1}$$  \hspace{1cm} (2.26)

where $k$ is the Von Karman constant (set to be 0.40 in CM1) and $c_d$ is the surface drag coefficient.

The version of the model used here is revision 18 (cm1r18). A number of changes and improvements have been made since Vermeire et al. (2011a,b) which used revision 13 (cm1r13). Most changes are minor, although some major improvements have been made to the surface treatments for simulations considering friction.

### 2.2.4 Computational details

The governing equations are discretized on a three-dimensional (non-uniform in the vertical direction) computational grid. Although Vermeire et al. (2011a) made use of a simplified mirrored one quarter domain, increased computational resources allowed a full 3-D domain to be used in the present work. Improvements to the model used in Vermeire et al. (2011a,b) were made, the distance in the north-south and east-west directions from the CS centre was increased from 3.5 km to 4.8 km and mirrored boundary conditions were not included.

CM1 makes use of an Arakawa C type Cartesian grid, which solves velocity variables on a staggered grid. For this parametric study all simulations used an identical numerical domain. The model domain encompassed approximately 9.6 x 9.6 x 4.0 km with a constant horizontal grid spacing of $\Delta x = \Delta y = 10$ m. In the vertical direction a stretching was employed to reduce computational load and focus the resolution in the region of peak outflow winds.
and the primary roll vortex. The vertical grid stretching is based on Wilhelmson and Chen (1982). The first grid point above the ground is at Δz=1 m stretching to Δz=50 m at the top of the domain, making use of 160 horizontal planes. This results in a total of 147,456,000 grid points for each simulation. The north, south, east and west (see Fig. 2.1) lateral boundaries for this domain are treated as open-radiative surfaces, allowing for flow to enter and exit the domain. The bottom or surface boundary is set to be a partial slip (semi-slip) with surface roughness set to z₀=0.10 m to correspond with Vermeire et al. (2011a,b), Mason et al. (2009) and Orf et al. (2012, 2014). The ground roughness length of z₀=0.10 m is applied to Monin-Obukhov theory which specifies the logarithmic law of the wall. Surface fluxes are not included between the ground and atmosphere. The top boundary condition for winds is set as a free-slip condition.

For the present work, as in Vermeire et al. (2011a,b), Orf et al. (1997) and Mason et al. (2009), a dry adiabatic lapse rate is employed in a quiescent atmosphere. A uniform atmospheric temperature is set to 300 K. No heat flux is included between the bottom boundary and the air, as no temperature difference exists.

A model time step of Δt=0.0500 s with 10 acoustic substeps is used for all simulations in this study. This minor change from the Δt=0.0625 s of Vermeire et al. (2011a,b) was required to maintain computational stability (Courant–Friedrichs–Lewy number>1) for some of the simulations where increased peak surface outflow winds were observed. All simulations were run for 10 min (600 s), as this period of time encompassed both the period of peak winds as well as the decay of the primary vortex. Due to the volume of data produced by the high resolution computational domain, data acquisition was focused at the period of peak outflow winds. Full three dimensional data for important variables was recorded every 30 s from 0 s to 250 s, every 5 s from 250 s to 450 s, and again every 30 s from 450 s to 600 s. Statistical data (peak quantities) were recorded every 0.5 s for the duration of the simulations.
2.2.5 Cooling source model

This parametric study makes use of a CS model to model the atmospheric conditions within a thunderstorm during the evolution and duration of a real downburst event. A region of cooler negatively-buoyant air is grown within the numerical domain and allowed to descend towards the ground. This type of numerical downburst model is based on the CS model established in Anderson et al. (1992, 1996), which replicates the cooling rate found in a downburst-producing thunderstorm. The growth of the CS is governed by the following \( \cos^2 \) ramp-up spatial relationship:

\[
q(x, y, z, t) = \begin{cases} 
g(t) \cos^2 \pi R & \text{for } R < 1 \\
0 & \text{for } R > 1 
\end{cases}
\]  

(2.27)

where \( g(t) \) is the growth of the cooling rate, and \( R \) is the scaled radial position in the ellipsoidal region, bounded by 1. The subdomain interval was erroneously reported in previous work (Vermeire et al. 2011a,b, Orf and Anderson 1999, Mason et al. 2009), and has since been corrected here. The subdomain interval for the ramp up function is limited
to 1 and not 0.5 as previously published, although this previous interval was only an error in the text, and all model equations in Anderson et al. (1992) and Vermeire et al. (2011a,b) are correct. For all simulations here, including the reference simulation, \( g(t) \) reaches its peak value after a 120 s \( \cos^2 \) ramp-up, remains constant from 120 s to 720 s and ramps down after 720 s;

\[
g(t) = \begin{cases} 
-\cos^2 \left( \frac{\pi [t - 120]}{2(120)} \right) & t < 120 \\
-1 & 120 < t < 720 \\
-\cos^2 \left( \frac{\pi [t - 720]}{2(120)} \right) & 720 < t < 840 \\
0 & t > 840 
\end{cases}
\]

(2.28)

The CS ellipsoid is physically defined using a non-dimensional scaled radial relation \( R \);

\[
R = \sqrt{\left( \frac{x - x_0}{h_x} \right)^2 + \left( \frac{y - y_0}{h_y} \right)^2 + \left( \frac{z - z_0}{h_z} \right)^2}
\]

(2.29)

where \( h_x, h_y \) are the horizontal half widths and \( h_z \) is the vertical half height of the CS ellipse (Fig. 2.2). The shape of the ellipse is one of the primary parameters investigated in the parametric study reported here. \((x_0, y_0, z_0)\) shown by \( h_c \) is the location of the spatial centre of the CS ellipse.

![Figure 2.2 – Cross sectional view of a generic cooling source in full domain with important dimensions](image)
2.3 Methodology

2.3.1 Details of the parametric study

To model the realistic CS variation found in nature, the following parametric study was conducted, focusing on a number of physical variables that influence the size, shape, cooling intensity and height above the ground of the CS (Tab. 2.1, Fig. 2.3, Fig. 2.4). These values were selected to remain within the reasonable bounds of those encountered within a simulated thunderstorm event (Straka and Anderson 1992, Anderson et al. 1992). The responsiveness of the wind field to a change in any of the following parameters is discussed and analyzed.

Table 2.1 – Study parameters

<table>
<thead>
<tr>
<th>Type of Parameter Change</th>
<th>( h_x ) (m)</th>
<th>( h_y ) (m)</th>
<th>( h_z ) (m)</th>
<th>( q ) (K/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reference Cooling Source</td>
<td>1200</td>
<td>1800</td>
<td>2000</td>
<td>-0.08</td>
</tr>
<tr>
<td>(Vermeire et al. 2011)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Aspect Ratio</td>
<td>1200</td>
<td>1200</td>
<td>2000</td>
<td>-0.08</td>
</tr>
<tr>
<td></td>
<td>1800</td>
<td>1200</td>
<td>2000</td>
<td>-0.08</td>
</tr>
<tr>
<td>Size</td>
<td>1000</td>
<td>1500</td>
<td>2000</td>
<td>-0.08</td>
</tr>
<tr>
<td></td>
<td>1400</td>
<td>2100</td>
<td>2000</td>
<td>-0.08</td>
</tr>
<tr>
<td></td>
<td>1000</td>
<td>1500</td>
<td>1666</td>
<td>-0.08</td>
</tr>
<tr>
<td></td>
<td>800</td>
<td>1200</td>
<td>1333</td>
<td>-0.08</td>
</tr>
<tr>
<td>Cooling Source Intensity</td>
<td>1200</td>
<td>1800</td>
<td>2000</td>
<td>-0.04</td>
</tr>
<tr>
<td></td>
<td>1200</td>
<td>1800</td>
<td>2000</td>
<td>-0.06</td>
</tr>
<tr>
<td></td>
<td>1200</td>
<td>1800</td>
<td>2000</td>
<td>-0.10</td>
</tr>
</tbody>
</table>
2.3.2 Reference simulation

The CS geometry was based on the original CS of Anderson et al. (1992), whilst the cooling rate peak intensity of Vermeire et al. (2011a,b) was used, giving an ellipsoid half-width of
1200 m, a half-height of 1800 m, a centroid height above the ground of 2000 m and a CS cooling rate of -0.08 K/s.

2.3.3 Naming convention

The naming convention used here in tables and figures takes the following form, CoolingRate_HalfWidth_HalfHeight_HeightAboveGround. Where Cooling Rate is the peak cooling rate of the CS (K/s). HalfWidth, HalfHeight and HeightAboveGround are all physical dimensions of the ellipsoidal CS (m).

2.3.4 Post-processing of data

Due to the need to compare the present data to other axisymmetric or simplified two-dimensional downburst data, circumferential averaging was employed to effectively simplify the full three-dimensional data into a (r,z) plane. A similar process was used in Orf et al. (2012, 2014) as well as Aboshosha et al. (2015). The centre of the downbursts on the horizontal plane was located at x=4.8 km, y=4.8 km and, then, at each radial distance from the centre of the event all data were spatially averaged around the circumference of the imposed circle of radius (r). The process was repeated for all heights (z) and all radial positions (r). This circumferential averaging process was performed for potential temperature, potential pressure, air density, three components of wind velocity and the three components of vorticity. For this study, the (r,z) plane encompasses 480 radial positions from (r=0 m to 4800 m) and 160 vertical positions from (z=0 m to 4000 m).

2.3.5 Scaling method

Although the IJ peak outflow wind speeds scale linearly with the jet nozzle velocity (Shehata et al. 2005), no such jet nozzle exists in nature and so scaling the CS model data are more challenging. Due to the inherent sensitivity of the outflow characteristics to the CS geometric properties of the CS, a non-dimensional scaling method should be adopted for comparison of dimensional quantities. A scaling method incorporating potential energy was developed by Lundgren et al. (1992) and further investigated in Yao and Lundgren (1996), based upon the fundamental driving mechanism of thunderstorm downbursts, which is a density difference between the source volume and the ambient surrounding
atmosphere. This scaling method allows for a somewhat reasonable comparison between CS type simulations, as well as comparison to full cloud (Oreskovic et al. 2015) and experimental data (Lin et al. 2007, Roberto et al. 2015). The characteristic length scale, $R_0$, is the equivalent radius of a sphere with volume, $Q$, defined by Lundgren et al. (1992) as the volume of an ellipsoid with the physical dimensions $h_x$, $h_y$ and $h_z$, the two half widths and half height. The characteristic time scale, $T_0$, is a function of the equivalent radius, the density of the ambient fluid ($\rho$ in this case the average density of the atmosphere in the region of the outflow), the density difference between the ambient and the downburst fluid and the gravitational acceleration constant. The characteristic velocity scale, $V_0$, is determined as a ratio of the characteristic length scale and time scale. Finally, a Reynolds number can be calculated using the average kinematic viscosity, $\nu$, of the surrounding atmosphere in the region of the CS and the velocity and length scales.

$$Q = \frac{4}{3} \pi h_x h_y h_z \quad R_0 = \left(\frac{3Q}{4\pi}\right)^{\frac{1}{3}} \quad T_0 = \left(\frac{R_0 \rho}{g \Delta \rho}\right)^{\frac{1}{2}} \quad V_0 = \frac{R_0}{T_0}$$

(2.30)

Whilst this scaling method was developed for dense liquid parcel release laboratory experiments (Lundgren et al. 1992, Alahyari and Longmire 1994), the CS model introduces a new level of complexity, most notably the temporally and spatially dependent CS ramp-up function that this scaling procedure was not specifically designed for. Hence, the present research investigates whether these laws are still applicable. Additionally, this scaling method does not take into consideration the vertical variation of density present in the real atmosphere. However, since the fundamental driving mechanism is the same for both CS type simulations and simple liquid drop releases, it was postulated that this approach may also be suitable for the former.

Other scaling parameters can be constructed from the Lundgren et al. (1992) parameters. Using these scales an expression for the total kinetic energy and potential energy can be written. The mass of the dense fluid is defined as the product of the volume of the ellipsoid and its mean density difference from ambient. The mean density was determined by
considering only the region up until the temperature perturbation reaches zero, and only once the CS had reached the peak cooling rate (120 s). The ambient density is taken as the spatial mean density of the environment not included in the CS region. The scale factor for the kinetic energy of the flow can be written as a product of the mass and the square of the Lundgren velocity scale factor. Similarly, the gravitational potential energy of the CS is written as the product of the volume of CS region, the air density difference, the gravitational constant and the height of the centre of mass of the ellipsoid above the ground. Additionally, the total energy of the flow can be scaled using a summation of the total kinetic and total energy scaling factors.

\[ \Delta \rho = \bar{\rho} - \rho_{amb} \]

\[ m_0 = Q \Delta \rho \quad E_{k0} = m_0 V_0^2 \quad E_{p0} = m_0 g h_c \quad E_T = E_k + E_p \]  

(2.31)

2.4 Results and discussion

2.4.1 Grid independence

To check for grid independence, three computational grids were considered. The horizontal grid spacing (\( \Delta x = \Delta y \)) was set to be uniform at 15 m, 10 m and 6 m, respectively, for each grid. The vertical spacing remained the same for all three cases, using a vertical stretching, that gave a 1m grid spacing close to the ground and 50 m at the top boundary. This gave a total number of grid points within the computational domain of approximately 65 million, 147 million and 409 million, respectively. For this analysis a few important quantities were examined, namely the three components of wind velocity and potential temperature. It is typically suggested that a grid refinement factor greater than 1.5 be selected between the coarsest and finest mesh (Roache 1993a,b, Franke et al. 2007, Tominaga et al. 2008). The representative cell (h) and grid refinement (r_g) factors are defined as,

\[ h = \frac{1}{N} \sum_{i=1}^{N} \Delta V_i \]  

(2.32)
\[ r_g = \frac{h_{\text{coarse}}}{h_{\text{fine}}} \] (2.33)

where \( \Delta V_i \) is the volume occupied by the cell in the computational domain and \( N \) is the total number of cells. The grid refinement factor for this study, between the coarsest and finest mesh is approximately 6.3.

**Table 2.2 – Summary of the meshes and corresponding maximum difference in peak outflow winds**

<table>
<thead>
<tr>
<th>Mesh</th>
<th>( \Delta x ) (m)</th>
<th>( \Delta y ) (m)</th>
<th>( \Delta z ) (m)</th>
<th># of Grid Points</th>
<th>Difference (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fine</td>
<td>6</td>
<td>6</td>
<td>1-50 (Wilhelmson and Chen 1982)</td>
<td>409 million</td>
<td>0.09</td>
</tr>
<tr>
<td>Baseline</td>
<td>10</td>
<td>10</td>
<td>1-50 (Wilhelmson and Chen 1982)</td>
<td>147 million</td>
<td>1.87</td>
</tr>
<tr>
<td>Coarse</td>
<td>15</td>
<td>15</td>
<td>1-50 (Wilhelmson and Chen 1982)</td>
<td>65 million</td>
<td>-</td>
</tr>
</tbody>
</table>

Grid independence was based on a 1% difference between the peak values on a vertical profile. The percent difference is defined as,

\[ \text{Percent Difference} = \left| \frac{A_{\text{mesh1}} - A_{\text{mesh2}}}{A_{\text{mesh2}}} \right| \times 100\% \] (2.34)

where \( A \) is any generic variable, in this case the radial component of wind velocity (Fig. 2.5). The grid independence test was conducted for all three grids at a model time of \( t=360 \) s and a horizontal position of approximately \( x=8.925 \) km and \( y=4.785 \) km. This model time and horizontal position corresponded to the temporal and physical condition close to the peak outflow wind velocity in the east-west direction. A difference in peak radial wind speed between the 15 m and 10 m grids was approximately 1.87\%, whilst the 10m and 6m grids gave a difference of 0.09\%. The percent difference for either case is even lower for
heights greater than 100 m, showing the highest percent difference in the region of peak velocity. Therefore, the 10 m resolution grid was used for this parametric study. A similar result was obtained by Vermeire et al. (2011a,b), where it was determined that a 10 m horizontal grid spacing was sufficient to produce grid independent results for this particular model.

![Figure 2.5 - Vertical profile of the radial wind speed at a region near the peak outflow wind velocity (Left). Magnified version of the same profile (Right)](image)

2.4.2 Observed improvement using a 6\textsuperscript{th} order scheme in full domain

When using the 5\textsuperscript{th} order advection scheme for both horizontal and vertical terms, it was observed that a cleft occurred in the temperature fields at locations in the outflow of 0\textdegree, 90\textdegree. This instability can be characterized by the premature development of fluctuations in the temperature field which have the appearance of a turbulent anomaly (Fig. 2.6, 2.7) and are also visible in Vermeire et al. (2011b). It was concluded that before proceeding with any further simulations, the instability needed to be corrected as it was observed that the anomaly did have a small, but noticeable impact on the outflow wind field. Outflow wind speeds were found to be different from those observed at the centre of the outflow (45\textdegree). Further simulations were completed to determine if the origin of the instability was caused by the imposed symmetry boundary condition. A simulation using 5\textsuperscript{th} order advection for
both horizontal and vertical terms was conducted without the symmetry boundary conditions. Instead, the CS was placed in the centre of the domain and the north south and east west walls were set to outflow boundary conditions. From this it was determined that the symmetry conditions were not the entire cause of the instability, as the anomaly was recorded again at 0°, 90°, 180° and 270°. It was concluded that using the 5th order scheme combined with unrealistic symmetry boundary conditions was forcing the flow to be influenced by the numerical grid. It is hypothesized that simulating a purely axi-symmetric flow that is primarily in the radial direction on a Cartesian grid space is not ideal. This type of simulation should be discretized on a radial grid, an option not available in CM1. However, the instability was resolved by employing a 6th order explicit (k_{diff,6}=0.05) advection scheme for the horizontal terms in a full computational domain without boundary conditions whereupon the outflow appeared to contain no numerical grid-influenced instability. This can be observed in Fig. 2.6 where two isosurface plots are shown for the same model time (t=370 s) for two simulations with different advection schemes and domains. Fig. 2.6 (right) contains a temperature isosurface plot of the replicated Vermeire et al. (2011a) simulation showing the observed instability associated with the 5th order advection. (left) contains the same isosurface plot with a 6th order explicit advection scheme in the full computational domain, showing no cleft.

Using odd-ordered techniques of this nature are implicitly diffusive, with a diffusive coefficient proportional to the Courant number, whilst the even-ordered techniques require explicit diffusion which, for this application, gave more control over the kinetic energy of the flow at small scales. By carefully selecting the diffusion coefficient (k_{diff,6}) of the higher order explicit scheme, the minor cleft in the temperature fields that was observed when using the 5th order scheme in a quarter domain with mirrored east and south boundary conditions, as used in Vermeire et al. (2011a,b) and in early test runs in the present work, was largely reduced.
Figure 2.6 - Temperature isosurface plots of the downburst outflow. 5th order advection in a symmetry domain (right), 6th order advection in a full domain (left) for t=370 s. Note: the full domain data are cropped to a single quarter for comparison.

Figure 2.7 – Close up view of the turbulent anomaly. 5th order advection in a symmetry domain (right), 6th order advection in a full domain (left) for t=370 s. Note: the full domain data are cropped to a single quarter for comparison.

2.4.3 Parametric study results

Ten simulations were carried out in order to investigate the relationship between the outflow and the initial characteristics of the CS. This includes 4 sets of simulations; 3 which scale both the size of the initial CS and the height of the source above the ground; 4 that investigate the magnitude of the cooling rate of the CS; 3 that modify the CS size, while
all other parameters remain constant and 2 which modify the aspect ratio of the CS, to represent the variation observed in nature. There exists some overlap between these groupings, as the reference simulation based on Vermeire et al. (2011a) is used as the reference comparison for multiple groups. Fig. 2.8 plots the time history of the overall peak radial wind speeds observed over the entire domain. It can be seen that there exists a large variation in the temporal evolution of each of the events, which is a major challenge in this type of parametric study. Peak outflow wind speeds range from nearly 30 m/s to 65 m/s over the range of all of the simulations and, temporally, the peaks occur over a range of 150 s. There also exist variation in other metrics, including the time at which the max outwards acceleration occurs ($\partial u_r/\partial t$max, which occurs at approximately 225 s – 300 s) as well as the magnitude of residual wind speeds ($t>500$ s). These quantities that show trends are summarized in Sec. 4.3.1. Additionally, the time histories of these peak wind speeds fall into two categories, those with ‘rounded’ peak regions and those with ‘sharp’ peak regions.

Figure 2.8 - Time history of the maximum radial wind velocity observed in the entire computational domain for each simulation in the study

This discussion focuses on the effect of CS parameters on quantitative aspects of the downburst outflow wind field, specifically; the radial position and height where the peak radial wind speed occurs, and the magnitude of that wind speed in terms of Enhanced Fujita scale area swaths. It is also of specific interest to this study to see if it is possible to
geometrically scale the CS and collapse the data by use of the idealized Lundgren et al. (1992) scaling approach.

### 2.4.3.1 Investigation of important outflow quantities

The simulations where the size of the CS event is scaled up and down from the baseline Vermeire et al. (2011a) case is considered below. The events are scaled in ellipsoidal half width ($h_x = h_y$) up from the baseline 1200 m to 1400 m and down to 1000 m. The height at which the CS is placed above the ground (centre of the event) is 2000 m for all three cases. The aspect ratio of the cooling sources remains constant at a 1:1.5 ratio for all three cases. In effect, only the half widths and half heights of the CS is changed, all other parameters remain constant, including the rate at which the air is cooled. Fig. 2.9 shows the $r_{\text{max}}$ and $z_{\text{max}}$, which is defined in this study to be the radial location where the peak outflow radial wind speed ($u_{r,\text{max}}$) occurs, and the height at which that peak wind is observed. Fig. 2.9 (left) show that for the radial location where $u_{r,\text{max}}$ occurs there is a distinct linear relationship between the CS size (horizontal half width of CS) and the corresponding radial position. Seen in Fig. 2.9 (right), the height at which the maximum wind speed occurs however shows a less obvious relationship that seems to suggest that $z_{\text{max}}$ is insensitive to the CS size. Additionally, the data appear to be staggered in levels that correspond to the grid locations at those heights, indicating that the true locations of the peak wind speeds may lie somewhere in between the grid levels.

![Figure 2.9](image_url)  
**Figure 2.9** – Radial location (left) and vertical position (right) where the maximum radial wind speed occurs versus CS ellipsoidal volume
Four simulations were performed where the peak cooling rate was modified to reflect variation in the thermodynamic cooling in a natural event. Peak cooling rates of -0.04 K/s, -0.06 K/s, -0.08 K/s (baseline) and -0.10 K/s were considered. The shape, size and height above the ground remained the same as that presented in Vermeire et al. (2011a). Fig. 2.10 (bottom left) shows a strong relationship between the CS cooling rate and peak outflow radial wind speed. Assuming that a cooling rate of 0 K/s would result in 0 m/s outflow wind speed, the data fits along an exponential trend (approximately second order polynomial) line that suggests that peak outflow wind speeds are exponentially related to the peak rate at which the air is cooled in the source. This is not entirely surprising as the peak outflow wind speeds are proportional to the bulk kinetic energy of the flow, which is shown to be related to the initial potential energy of the source caused by the temperature perturbation (Sec. 2.4.3) i.e. $\Delta \rho \propto u^2$. A first order differencing method was used to calculate the maximum radial acceleration of the flow, showing a nearly linear trend (Fig. 2.10 bottom right). Fig. 2.10 (top) shows that a less clear relationship is present for the locations (radial and vertical) of that same peak outflow wind velocity. Although it can be seen that a stronger cooling source (larger cooling rate) results in both radial and vertical locations that are closer to the region of impingement, suggesting that a colder downburst is spatially more compact. Although it should be noted that the radial locations of the peak fall within a range of approximately 50 m, and the vertical locations within a range of 20 m in a physical domain that extends 4.8 km in either direction, perhaps showing that the variation is not particularly significant.
Figure 2.10 – The radial (top left) and vertical (top right) position where the maximum radial wind speed occurs and the maximum radial wind speed (bottom left) and maximum radial acceleration (bottom right) versus CS cooling rate.

Three simulations were performed where the size of the event was scaled from the baseline Vermiere et al. (2011a) simulation and the height of the CS centre was also scaled by the size of the CS. This set of simulations forms the basis of the Lundgren et al. (1992) scaling approach section of this paper, as it scales all geometric parameters of the event (in a similar manner to the IJ model scaling). In Fig. 2.11 the radial location of the peak outflow wind appear to take on a linear relationship, whereby the event is scaled up the locations of the winds also seem to increase out away from the impingement. This can be seen in Fig. 2.12, where equivalent temporal isosurface temperature fields for the same simulations are plotted and the primary vortex is located where one would expect (a radial location that would scale with the size of the CS) for geometrically scaled downbursts. The overall magnitude of the peak outflow wind speed that corresponds to Fig. 2.11 (bottom left) also
appears to show a linear relationship related to the height at which the CS is located above the ground. It should be noted that the entire CS is scaled in all aspects (size and height above the ground) and $h_c$ is only one metric that differentiates each CS. Fig. 2.11 (bottom right) also shows a linear relationship between the CS ellipsoidal half width and peak outflow wind speed. This suggests that by scaling all dimensional aspects of the CS, the exponential relationship between total initial potential energy and bulk outflow kinetic energy ($\Delta \rho \propto u^2$) remains, but the size of the CS reduces the kinetic energy of the flow by a proportional amount.
Figure 2.11 – The radial (top left) and vertical (top right) position where the maximum radial wind speed occurs and the maximum radial wind speed versus height at which the scaled CS is placed (middle left) and half width of CS (middle right) and maximum outward radial acceleration versus CS height (bottom)

2.4.3.2 The Lundgren scaling approach

Due to the variation in outflow wind fields between different downburst simulation methods, a scaling approach is necessary to compare the flow evolution. Here, the scaling
approach of Lundgren et al. (1992) has been adopted. However, the CS approach poses a unique problem. The source is a spatial and temporal ramp up function and, therefore, the density field is not uniform in space or time. In the liquid release model of the downburst (Lundgren et al. 1992, Yao and Lundgren 1996, Alahyari and Longmire 1995), there exists a change in density at the interface between the downburst source and the ambient surrounding environment. Since the temperature of the air in the CS is proportional to the density and pressure of the local air, this poses a minor problem that needs to be considered by defining an equivalent average density value. Further, the meteorological cloud model makes use of a stratified atmosphere, where the density of the air is also proportional to the pressure of the air as defined by the ideal gas law. As a result $\rho_{amb}$ is not uniform everywhere, unlike the liquid release experiments upon which the Lundgren et al. (1992) scaling is based. As a result, some approximation of the density of the ambient atmosphere needs to be made. Here, the ambient density has been taken to be the spatial mean of that portion of the atmosphere into which the CS is placed, up to where there exists no temperature perturbation. Temporally, the density is computed once the region of cold air reaches its maximum size, which corresponds to the time at which the cooling rate peaks (120 s). The density of the downburst cold air volume ($\rho$), is taken to be the spatial average of the CS region (ellipsoidal volume). This resulted in a value of $\Delta \rho / \rho$ of 0.040, 0.038 and 0.035 (Tab. 2.4) for the scaled simulations that correspond to 008_1200_1800_2000 008_1000_1500_1666 and 008_800_1200_1333. This value is a computed mean, and it is recommended that future studies that make use of the CS approach, investigate different methods in determine the appropriate densities. This is an improvement over the approach to estimating the relevant densities that was made in Oreskovic et al. (2015) from the full cloud simulation data of Orf et al. (2012), where an estimate with no spatial averaging was performed for that study. The density difference for the full cloud simulations (1.3%) is less than those computed in the present CS simulations. However, present agreement with the experimental study of Roberto et al. (2015) is good, where a URANS model of a downburst showed density differences of approximately 3% and 4% which replicated an experimental set up with a density difference of 3.3%.
Table 2.3 – A summary of the Lundgren et al. (1992) scaling parameters

<table>
<thead>
<tr>
<th>Parameter</th>
<th>008_1200_1800_2000</th>
<th>008_1000_1500_1666</th>
<th>008_800_1200_1333</th>
</tr>
</thead>
<tbody>
<tr>
<td>R₀ (m)</td>
<td>1373</td>
<td>1144</td>
<td>916</td>
</tr>
<tr>
<td>T₀ (s)</td>
<td>59.19</td>
<td>55.67</td>
<td>51.53</td>
</tr>
<tr>
<td>V₀ (m/s)</td>
<td>23.21</td>
<td>20.56</td>
<td>17.77</td>
</tr>
<tr>
<td>Δρ₀/ρ</td>
<td>0.040</td>
<td>0.038</td>
<td>0.035</td>
</tr>
</tbody>
</table>

The mean density of the source volume of air for the current study is computed by first defining the interface at which the source boundary exists. For this study, the interface is considered to be where the temperature perturbation of the air is greater than 299 K (the quiescent atmosphere is defined to be a uniform temperature 300 K). This is taken at a model time of 120 s, which corresponds to the time at which the peak cooling rate is first reached. All spatial densities values are then averaged and the corresponding downburst density is computed (ρ\textsubscript{spatial}). Similarly, the ambient fluid density is computed by taking the spatial mean of all density information that falls outside of the downburst volume (>299 K). Future studies may seek to investigate other methods by which the source density could be defined, such as a volume integral of the source. For this study however, the spatial mean is considered to be approximately equivalent to the volume integral of the same region, summarized in the following equation, where Q is the integration region and x, y and z are the Cartesian locations within the region, and Q is the cooling source volume.

\[
\bar{\rho}_{\text{spatial}} \approx \frac{\iiint_{Q} \rho(x, y, z) \, dx \, dy \, dz}{Q} \tag{2.35}
\]
Figure 2.12 – Potential temperature perturbation contour of the circumferentially averaged r,z temperature field, and potential temperature perturbation isosurface at $T=299$ K for the three scaled simulations. 008_800_1200 (top) 008_1000_1500 (middle) 008_1200_1800 (bottom), for three times which are considered approximately equivalent for each simulation.
Figure 2.13- Potential temperature perturbation contour of the circumferentially averaged r,z potential temperature perturbation field for the three scaled simulations. 008_800_1200 (top left) 008_1000_1500 (middle left) 008_1200_1800 (bottom left), for t=330 s, and temperature isosurface plots at t=300 s and T=299 K (right).
The temporal evolution of the downflow and the outflow front is an important metric when comparing cases. Here, the height above the ground of the downflow is defined as the distance from the ground surface to where the wind magnitudes begin to increase from zero in the downward direction (which also corresponds to a temperature of approximately 4 K below ambient). This height above ground is represented by $z_m$. A similar quantity is the radial position from the centre of impingement to where the downburst outflow front is located, represented by $r_m$ (Fig. 2.14). This, too, is defined as being where the temperature interface reaches 296 K. The 296 K interface is selected as it is this temperature perturbation that results in a noticeable boundary in downward and outward wind speeds from the ambient stationary atmosphere. Fig. 2.14 shows this definition for both quantities, for two times in the simulation where the downflow makes up the majority of the downburst (left) and after the downburst impinges on the ground and the radial outflow is visible (right).

![Figure 2.14 – Definition of the location of $z_m$ and $r_m$ of the downburst outflow front](image)

Fig. 2.15 plots the evolution of the downburst outflow front as a function of time, normalized by the Lundgren et al. (1992) scaling parameters, together with data from other studies (Lundgren et al. 1992, Alahyari and Longmire 1995, Mason et al. 2009, Vermeire et al. 2011, Roberto et al. 2015). It can be seen that the agreement with other numerical studies is good for regions after impingent, which occurs at $t/T_0=5.5$. However, agreement is not as good before impingement. This is not surprising, since Alahyari and Longmire
(1995) and Lundgren et al. (1992) make use of liquid release experiments that do not include the realistic spatial and temporal growth of the source that occurs with the CS. For simplified liquid drop release experiments, the radial position of the outflow begins at $r=R_{cylinder}$ and, thus, cannot begin at $r/R_0=0$. Liquid drop release experiments in that instance lack the realistic physics that a real thunderstorm cloud would experience as the cold mass of air develops. However, agreement after impingement seems to show that the outflow evolution is relatively similar.

Figure 2.15 – Time history of the radial position of the downburst outflow front normalized by the Lundgren et al. (1992) equivalent spherical radius ($R_0$) and time scale ($T_0$)

Fig. 2.16 tracks the height of the 296 K interface above the surface of the ground as a function of time, non-dimensionalized by the Lundgren et al. (1992) parameters. It can be seen that in CS driven flows, the evolution of the down flow front is not similar to those of the liquid parcel releases. For liquid parcel experimental studies, the drop cylinder is typically released at a height of around 3$R_0$ and, as such, the downburst downflow front is considerably higher above the ground than the event is wide. For the CS studies presented
here, the centre of the CS is placed at a height ranging from 1333 m to 2000 m, while the width of the event ranged from 800 m to 1200 m. Also, the CS intensity varies spatially such that the downburst front ‘grows’ out from the centre and once the CS fluid becomes heavy enough for buoyant forces to cause it to fall, it is already near the ground. As a result, the downburst front is typically far closer to the ground at the beginning of the simulation (approximately \(z_m=800\) m for the largest simulation) than the liquid drop release studies. The Lundgren et al. (1992) scaling approach does seem to collapse the downflow and outflow fronts quite well, making comparison in this form promising.

Figure 2.16 – Time history of the vertical position of the downburst outflow front, measured from the ground to the surface of the outflow front, normalized by the Lundgren et al. (1992) time scale (\(T_0\)) and length scale (\(R_0\))

Peak outflow radial wind speeds also appear to collapse reasonably well when normalized by the Lundgren et al. (1992) parameters, as is shown in Fig. 2.17. The peak wind velocities in the east-west direction (which are reasonably approximated as the radial direction, since the flow is largely radial) over the entire computational domain are plotted as a function of time (the same data set from Fig. 2.6). It can be seen that, temporally, the data between the simulations matches fairly well, although an offset of scaled wind speeds can be seen between \(t/T_0=4\) and \(t/T_0=6\) with a shift of approximately \(t/T_0=0.2\). Agreement is especially
good in the regions up until impingement, as the bulk of the flow remains extremely similar until after impingement and loss of symmetry around the outflow circumference.

![Figure 2.17](image)

**Figure 2.17 – Time history of the maximum radial velocity in the outflow (right) and normalized by the Lundgren et al. (1992) velocity scale ($V_0$) time scale ($T_0$) (left)**

Due to the differences in the way the outflow evolves between the three scaled events, shown in Fig. 2.12, it is somewhat difficult to compare the temporal history of the outflow at a single location. Fig. 2.18 shows the history of peak outflow radial wind speeds at approximately equivalent radial positions. Selecting an equivalent radial position is somewhat challenging and this equivalence is based on the observed shape of the primary outflow vortex, as well as similar $r/R_0$ values. Radial positions of 1350 m, 1110 m and 930 m are selected, and the peak wind velocity at that radial location, for all heights, is plotted against time, normalized by the Lundgren et al. (1992) approach. It can be seen that the data collapse fairly well, with a minor temporal offset. These results are encouraging, as the general shape of the time history is preserved, showing that the wind velocities do scale with size and height of the CS above the ground.
Another comparison of interest is the scaling of the bulk kinetic energy (KE) in the outflow (Fig. 2.19). CAPE is correlated to the peak outflow wind speeds (and, thus, the bulk kinetic energy of the flow) in natural events (Pryor and Ellrod, 2004) and any suitable model of a downburst should make an attempt to quantify the relationship between source potential energy (PE) and outflow KE. CM1 produces statistical data that is representative of the total kinetic energy in the computational domain, specified by the raw statistic output variable ‘ek’ (Bryan 2015). This statistic will be used as the basis to approximately quantify the evolution of the actual total kinetic energy in the flow. An approximation of the initial potential energy of the CS is made whereby $E_{p0}$ is the gravitational potential scaling factor calculated by approximating the mean density difference of the air contained within the CS from ambient (from the Lundgren et al. (1992) scaling approach), together with the height of the CS above the ground. Non-dimensionalized total kinetic energy differs between each simulation by approximately 500% at peak. The data collapse extremely well when the scaling approach is applied (<1% difference), indicating that the bulk KE energy in the flow evolves consistently no matter how large the initial CS is (when height is scaled with it). It is clear however, that the introduced scaling term for potential energy, overestimates the initial potential energy of the cold air contained in the downburst.

Figure 2.18 – Time history of peak radial wind velocity for three scaled simulations at three radial positions that are considered to be equivalent and correspond approximately to the radial position where the peak outflow wind speeds occur (left), made non-dimensional by the Lundgren et al. (1992) velocity and time scales (right).
Figure 2.19 – Time history of the total kinetic energy within the computational domain (left) and then normalized by the initial potential energy scale term and the Lundgren et al. (1992) time scale term (right)

Assuming that the peak wind velocities in the outflow and the bulk potential energy of the event scale fairly well for these events, the location within the outflow of the peak wind speeds should also compare fairly well. Fig. 2.20 plots the radial location where the peak outflow wind velocity occurs as a function of time and Fig. 2.21 plots the corresponding height AGL of the same wind. A scatter plot is used for these data sets, as it is clear that different parts of the flow structure are plotted depending on the time in the event and, therefore, a continuous line should not be used here.

Figure 2.20 – Time history of the radial position of peak radial wind speed, normalized by the Lundgren et al. (1992) equivalent spherical radius ($R_0$) and time scale ($T_0$)
It can be seen that the vertical location of the peak wind speeds in the outflow evolve consistently between all three events when compared non-dimensionally (Fig. 2.21). Earlier, Fig. 2.10 showed the flow structure of the circumferentially averaged temperature field, as well as the full domain temperature data set, for a qualitative comparison. Despite the size of the cooling sources and the corresponding outflow, it is clear that they have an extremely similar structure, as the data for the most part collapses.

Figure 2.21 – Time history of the height to the peak radial wind speed, non-dimensionalized by the Lundgren et al. (1992) equivalent spherical radius (R₀) and the time scale (T₀)

Vertical profiles of the circumferentially averaged radial winds at the radial locations where the peak winds are recorded, are plotted in Fig. 2.22, for the simulations where the size and height above ground of the events are scaled accordingly. Fig. 2.22 (top left) shows the profiles fully dimensionalized to show a better sense of scale of the actual wind speeds between the three events. It is clear that the profiles follow nearly identical trends, with the only difference being the overall magnitude of the peak wind speed and the corresponding height. When scaled by the maximum wind speed and the height at which that maximum occurs, the data collapse for regions below the peak height, and compares well up to a height of twice the height of the maximum. Agreement is not as close for heights above 3 times the maximum (~300 m). In Fig. 2.22 (bottom), the Lundgren et al. (1992) scaling
approach has been applied to the same vertical profiles, collapsing the data fairly well, especially near the height at which the peak occurs.

Figure 2.22 – Vertical profiles of radial wind speeds for the three simulations which geometrically scale both the size of the CS and the height above the ground (top left) at approximate locations where the peak outflow velocities are recorded. Radial wind velocity vertical profiles scaled by the peak outflow wind velocity, and the height which the peak wind occurs at (top right). Radial wind velocity vertical profiles scaled by the Lundgren et al. (1992) velocity scale ($V_0$), and the equivalent spherical radius ($R_0$) (bottom).
2.4.3.3 Enhanced Fujita Scale wind

Vermeire et al. (2011b) quantified the horizontal wind swaths where near ground EF0 and EF1 winds were predicted for a CS outflow. The present work focuses on the near ground region, from 0m to 50m AGL, investigating the maximum winds in this region in both space and time. The maximum surface wind swath (Vermeire et al. 2011b, Bryan 2015) is a two dimensional wind plane where the wind at each point in space is recorded over all time, and only the maximum is recorded. Up to now, as far as the authors are aware, no attempt has been made to investigate the instantaneous horizontal area where EF0 and EF1 winds are recorded for a specific height above ground level for a downburst. However, it is noted that the EF scale is designed specifically for tornadic wind damage, the EF scale in this study will specifically be used to only indicate wind speed, as presumably the damage caused by downbursts and tornados are fundamentally different.

Of particular interest to wind engineers is the near ground area that experiences EF0 or greater wind speeds. High Intensity Wind (HIW) events are of interest specifically to the power transmission line industry, as it is these classes of winds that cause the most notable damage to such structures (Shehata et al. 2005, Ladubec et al. 2012, Aboshosha and El Damatty 2015, Aboshosha et al. 2016). The Enhanced Fujita scale is a modified version of the Fujita scale (Fujita 1971, 1973, 1981), used to quantify the damage caused by intense tornadic events (McDonald and Mehta 2006, Hamada and El Damatty 2015). EF0 winds are characterized by winds with a magnitude between 29 m/s and 37 m/s. EF1 winds are winds in the range of 38-49 m/s and are synonymous with moderate damage including overturned mobile homes and severely stripped roofs (McDonald and Mehta 2006). In an effort to quantify the near ground winds that ground based structures may experience during a downburst event, the horizontal area at 50 m AGL that experiences EF0 and EF1 winds has been tracked in time for a number of the downburst events in this study. Although 10 m AGL is typically taken to be the reference height for near ground winds, 50m is also studied in this case as this is the height above ground that corresponds to the highest peak outflow wind velocities for the present CS simulations and the full cloud results of Orf et al. (2012, 2014). Typically a peak height of between 50 m and 150 m is observed, as can be seen in Fig. 2.22. Fig. 2.23 tracks the temporal history of the area that
experiences EF0 winds for the simulations where both the size and height above the ground has been scaled accordingly (008_1200_1800_2000, 008_1000_1500_1666, 008_800_1200_1333) for both 10 m AGL and 50 m AGL. Not surprisingly, the largest of the three events shows a larger wind swath of EF0 winds during the region of peak radial outflow wind velocities. It can be seen that the largest area where these winds occur seems to be just after the time of downburst impingement (near 300 s model time) for all three events. A dip following the initial peak in the 50m AGL plot (Fig. 2.23 right) is, presumably, caused by the primary roll vortex passing through the 50m AGL horizontal plane as it is lifted off of the ground by the stationary near surface air, a phenomenon observed in the current study, as well as by Vermeire et al. (2011a). Fig. 2.23 shows that an area of up to 4.5 km² experiences EF0 winds at 50 m AGL and up to 14 km² at 10 m AGL for the largest CS. However, it can be seen that the smallest of the scaled events experiences almost no EF0 winds during this time period, less than a quarter of that which is experienced during the largest event, even though the size and height of the initial CS has only been scaled by a factor of 1.5.

**Figure 2.23 – Temporal history of horizontal area at 10 m (right) and 50 m (left) AGL that is experiencing EF0 winds for simulations where both the size and height AGL is scaled**

Fig. 2.24 captures the horizontal area at which EF1 winds (38-49 m/s) occur at 50 m AGL for three different variations of CS parameter modifications, and Fig. 2.25 shows EF1 winds at 10 m AGL. The top left plot corresponds to the same parameter variation of Fig. 2.23, the modification to both size and height above ground of the initial CS. Top right corresponds to the simulations where a modification to the peak cooling rate at the
geometric centre of the source has been made (in this case -0.04 K/s, -0.06 K/s and -0.10 K/s). Fig. 2.24 (bottom left) corresponds to the simulations where the geometric size of the CS has been modified, but all other parameters including aspect ratio remain the same. For simulations where the size and height of the CS has been scaled, it can be seen that the largest CS results in the largest horizontal area with EF1 winds and the smallest CS has no appreciable horizontal area where EF1 winds occur. The medium sized CS has an area of approximately 0.5 km² where EF1 winds occur. It is noted that the time at which the largest area where these types of winds are observed is approximately 325 s for both simulations. The simulations where the peak cooling rate is modified shows a trend where the size of the area where the EF1 winds occur are not only larger for the more intense CS but also staggered in time. The most intense CS shows the largest EF1 wind area at a time approximately 175 s before the least intense CS that shows a non-zero value. The time at which the peak area occurs also appears to be proportional to the CS intensity, as does the peak area. The least intense CS in this set of simulations (-0.04 K/s) shows no substantial area of EF1 winds, not surprising since this specific simulation shows a maximum outward radial wind velocity of only 40 m/s, which is on the lower end of the EF1 scale. The simulations where the size of the CS is changed, but all other parameters remain the same, shows an interesting trend. The area that experiences EF1 winds appears to be linearly proportional to the size of the CS. However, the time at which the peak area occurs appears to remain nearly constant for all three cases, with only an offset of approximately 10 s from peak to peak.
Figure 2.24 - Temporal history of horizontal area at 50 m AGL that is experiencing EF1 winds for simulations where both the CS size and height AGL is scaled (top left), simulations where the peak CS intensity has been modified (top right) and simulations where the CS size has been changed (bottom)
Figure 2.25 - Temporal history of horizontal area at 10 m AGL that is experiencing EF1 winds for simulations where both the size and height AGL is scaled (top left), simulations where the peak CS intensity has been modified (top right), and simulations where the size of the CS has been changed (bottom)

Some clear relationships can be observed between these three types of parameter modifications and the area over which EF0 and EF1 winds are observed. Changing the peak cooling rate of the CS seems to have a strong effect on the time at which the largest area of EF1 winds is observed, not surprising since the cooler air falls faster and spreads outward radially sooner. The size of the CS seems to have a strong impact on the size of the area of the winds, but not on their temporal position, indicating that the height of the CS above the ground has little impact on this specific metric.

2.5 Conclusions

This study consisted of 10 CS simulations of thunderstorm downbursts making use of the meteorological cloud model, Cloud Model 1 (CM1). The CS forcing function of Anderson et al. (1992) and Vermiere et al. (2011a) was used as the baseline simulation case, while a
number of important geometric and thermal parameters were modified in order to investigate their impact on the downburst outflow wind fields. The primary objective of this work was to determine if a scaling procedure, such as that used with the more common IJ downburst model, can be used for the more sophisticated CS model. The scaling procedure developed in Lundgren et al. (1992) was applied to both complete time histories and instantaneous profiles of the outflow wind field. The present research provided some encouraging results and a number of conclusions and recommendations are made that are detailed here, answering the initial questions presented in the introduction chapter of this work.

- Using a 6th order explicit advection scheme (and carefully selecting the diffusion coefficient) in a full computational domain resulted in improved numerically stability, removing grid a dependent numerical anomaly that was present in Vermiere et al. (2011a). Also, a slightly larger computational domain was able to be used due to increased resources, allowing the outflow wind field to flow uninterrupted by boundary conditions. Future studies involving CM1 should make use of at least 6th order explicit horizontal advection terms in a full domain.

- The size of the CS (while keeping all other geometric parameters the same) has a linear effect on the radial location and height above the ground where the peak outflow wind speeds occur.

- There exists an exponential relationship between the peak cooling rate of the CS and the peak outflow wind speeds that suggests that the density of the source is proportional to the square of the outflow wind speed \( \Delta \rho \propto u^2 \). Total potential energy of the source term also appears to be tied to the bulk kinetic energy of the outflow when considering Lundgren et al. (1992) scaling methods.

- The height of the CS (while scaling the other geometric parameters equivalently) has a linear effect on the peak outflow wind speeds and its corresponding radial position.

- The Lundgren et al. (1992) scaling approach appears to work very well for CS type simulations of downbursts, and does not appear to be limited to only liquid release models of thunderstorm downbursts. Selecting the source density, however, can be challenging, due to the spatial and temporal dependence of the cooling source.
forcing function. Future work should aim to develop a better method to select the source density.

- The evolution of the downburst front of any CS model of a downburst compares reasonably well to other models during the time after impingement of the source fluid. However, the front develops differently in the CS function, as is expected since the cooling source size is temporally and spatially dependent.

- The downflow front of the cooling source model does not agree at all with more idealized liquid drop release experiments. Again, due to the spatial and temporal growth of the CS forcing function, and the height at which the front of the downflow is above the ground at the beginning of the simulation.

- The height and radial position where peak velocities in the outflow occur scale extremely well using the Lundgren et al. (1992) parameters, showing that the evolution of the outflow front is temporally similar between the scaled events.

- The total kinetic energy of the downburst scales extremely well when using a total potential energy scaling term, estimated using the Lundgren et al. (1992) parameters. The bulk energy data collapses, furthering the hypothesis that the outflow wind field bulk flow energy scales with CS size.

- The total horizontal area that experiences EF0 and EF1 winds at 50 m (the height where peak outflow winds are observed) and 10 m AGL is proportional to the initial geometric parameters of the CS. Cooling rate modification adds a temporal influence on the EF areas that is not observed in simulations when the cooling rate is kept the same.

This parametric study shows promising results that indicate that the CS model is scalable in a similar way to the IJ model. It is clear that there exists a quantifiable relationship between the CS initial geometric parameters and the outflow wind fields. Lundgren et al. (1992) scaling parameters can be used for CS studies to collapse data and make a quantitative comparison between physically different downbursts, although some caution is needed when defining the source density. Future studies should perform a larger number of simulations, in an effort to establish an even more clear understanding of the relationship between source size and the impact on the outflow.
2.6 Acknowledgements

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2.7 References


Chapter 3

Preface

The following text is a draft manuscript that is being prepared to be submitted to The Journal of Wind Engineering and Industrial Aerodynamics, as part of a more broad study of full scale thunderstorm downburst simulations that make use of Cloud Model 1 (CM1). This draft contains supplementary text and equations that will not be submitted as part of the publication, but, instead will be used specifically for the thesis submission.

Keywords: downburst, CM1 cloud model, thunderstorm, numerical modeling

3 A Thunderstorm Downburst in a More Realistic Atmosphere: Comparison of Idealized Cooling Source to Meteorological Cloud Model

3.1 Introduction

Thunderstorm downbursts are air masses that descend rapidly from the lower atmosphere and impact the ground, resulting in radial outflows of intense, potentially damaging winds (Wakimoto 1985). Downbursts are density driven events, resulting from the thermodynamic cooling of the earth’s atmosphere, associated with deep moist convection cloud development. Thermodynamic processes, including the formation of rain, snow and graupel (small hail) create pockets of negatively buoyant air which, eventually, descend towards the ground (Fujita 1985). Capturing field data is extremely difficult (Gunter and Shroeder 2015), due to the unpredictable nature and large spatial variability of these events, and so research in this field has primarily focused on physical and computer modelling.

The most common engineering physical and numerical model of downbursts is the axi-symmetric impulsively-driven impinging jet (IJ) model which was studied initially by Landreth and Adrian (1990), Letchford and Chay (2002), Chay and Letchford (2002) and, more recently, by Kim and Hangan (2007), Sengupta and Sarkar (2008) and Xu and Hangan (2008), which introduces some observable limitations as the key physics present in nature are ignored. The IJ model will not be considered in this study, as it was shown in
Vermeire et al. (2011a) that there are notable deficiencies in its physical realism. In meteorological research and in this study, an improved model has been introduced, known as the cooling source (CS) model (Anderson et al. 1992, Orf et al. 1996, Orf and Anderson 1999, Vermeire et al. 2011a,b). The CS model is created by specifying an axi-symmetric cooling forcing function that mimics the thermodynamic cooling that occurs within a thunderstorm during formation. However, the CS does have some limitations, such as the highly idealized method by which the thermodynamic cooling is introduced and, most notably, the quiescent (uniform and stationary) atmospheric base state that is most often used in these studies (e.g. Anderson et al. 1992, Orf et al. 1996, Mason et al. 2009, Anabor et al. 2011, Vermeire et al. 2011a,b, Oreskovic et al. 2016a). The thermodynamic cooling which causes a thunderstorm downburst to form, inherently does not evolve in such simplified or idealized ways. More complex models do exist that attempt to simulate the thermodynamic cooling more realistically, the most sophisticated being the meteorological cloud model which includes microphysical modelling. These types of models introduce a realistic atmospheric base state and initiate the formation of a thunderstorm by including a warm bubble of air, to mimic the heating from the earth’s surface (Orf et al. 2012). Realistic microphysics (the interaction of air with different forms of precipitation) is included, which adds to the realism of the downburst by including the drag induced by falling precipitation. These types of simulations do, however, have limitations, most notably the increased computational resources required to perform the calculations demanded by the wind engineering community, five orders of magnitude higher than other three-dimensional cloud model studies (Orf et al. 2012). In fact, the computational resources available have only recently been capable of performing these types of simulations at the resolution required by the engineering community. Full cloud models solve additional equations beyond those of the simplified CS, introducing a number of variables related to the microphysical interactions and formation of precipitation. Additionally, a significantly large computational domain is required (120x120x14 km is typical), when compared to the CS model, since not only does the downburst outflow have to be captured, but the entire region that encompasses the thunderstorm cloud must be included. This computational requirement makes full cloud meteorological downburst models impractical at the present time for engineering applications. The full cloud model, however, is important when
validating the results of more simplified models, ensuring the IJ and CS wind fields can be reasonably compared to field data.

Full scale meteorological thunderstorm models have been used to investigate deep moist convection phenomenon for some time. Indeed, sophisticated three dimensional numerical models made an early appearance in Lilly (1979), while Klemp et al. (1981) simulated the Del City thunderstorm, a storm which occurred in Oklahoma on May 20 1977 by using sounding data recorded in the field and a microphysical parameterization from Kessler (1969). The storm was particularly long lasting, reaching a tornadic phase, and data recorded from Doppler radar were compared to the simulated results. Rotunno (1981) investigated the velocity data of early cloud development and storm rotation from the LES of Wilhelmsen and Klemp (1978). Rotunno and Klemp (1981) used a numerical model to simulate a thunderstorm in a veering environmental wind shear, to investigate why a symmetric updraft typically acquires cyclonic rotation. Farley and Orville (1987) simulated a strong hailstorm in a two dimensional time dependent simulation with microphysical parameterization. The simulation was initiated using a sounding taken from the field during the observed thunderstorm in Red Deer Alberta on 26 July 1983. The simulated thunderstorm produced a very accurate replication of the thunderstorm that was observed in the field, as a consistent comparison was made to aircraft measurements of the thermodynamic structure of the sub-cloud and internal region of the storm.

Meteorological models are versatile in the way that they can simulate any type of deep moist convection, including thunderstorms that can lead to the formation of strong downdrafts and downbursts. Full cloud models were initially used to investigate thunderstorm downbursts in Proctor et al. (1988) where a simplified downburst was initiated in an atmospheric base state adapted from field data taken from the 1982 JAWS study (Hjelmfelt 1987). The downdraft was initiated by drag induced by falling precipitation from the top of the domain. Tuttle et al. (1989) made use of a meteorological cloud model to simulate a strong thunderstorm that was observed during the MIST study (Wakimoto and Bringi 1988). This model included realistic microphysics and captured the evolution of the thunderstorm, as well as a microburst that descended to the surface under the effects of mass loading, a notably thermodynamic cooling as a result of the evaporation
of rain and sublimation of hail. Knupp (1989), performed a study where two downbursts were simulated to determine the characteristics of the low-level downdraft initiation within the parent storm. For each simulation the environment was initiated differently, using similar temperature profiles but modified moisture profiles. It was found that the downburst characteristics are closely controlled by the way in which precipitation formed and arrived at lower levels in the domain. It was found that the downdrafts from each simulation had similar outflow wind speeds (on the order of 12m/s), but their spatially variability differed greatly as a result of different initial atmospheric conditions. It was also determined that different initial atmospheric conditions (wind shear near the ground and initial moisture in the atmosphere) resulted in differences in the microphysical processes that lead to the formation of the downburst. That study concluded that downburst initiation and evolution (outflow) is closely related to the initial atmospheric base state which is of critical importance to the present CS study. Proctor and Bowles (1992) numerically investigated a natural thunderstorm downburst that occurred in Denver Colorado in 1998 and which a number of aircraft encountered on their final approach at Stapleton Airport. The parent conditions were simulated on the Terminal Area Simulation System (TASS), and it was found that a strong downdraft formed in the simulation downwind from the primary rain shaft. It was concluded from this study that the evaporation and sublimation of snow and ice participles were the primary driving mechanism of the downdraft. Straka and Anderson (1992) performed a number full cloud thunderstorm simulations using sounding data from the Cooperative Huntsville Meteorological Experiment (COHMEX). These simulations were performed at 500 m and 250 m grid resolution, and were able to produce hail shafts that were 1 km - 2 km wide. In the second part of the study the influence of microphysical effects were investigated. It was found that storms which included the sublimation of snow and ice resulted in downdraft wind speeds that were the same or greater in magnitude than simulated storms that did not include such ice phase precipitations, suggesting that the evaporation of snow and ice leads to an additional thermodynamic cooling forcing that can drive stronger downdrafts. A more recent downburst producing thunderstorm simulation was completed by Orf et al. (2012), where a thunderstorm was initiated in the realistic atmospheric base state of Brown et al. (1982). A primary downdraft was observed, accompanied by a number of adjacent secondary
downdrafts, with peak outflow winds in excess of 35 m/s. Orf et al. (2012) focused their analysis on the near ground region of the downburst, the area of interest to wind engineers. They concluded that full cloud simulations capture the realistic spatial variability present in natural downburst events, but noted the lack of repeatability for the wind engineering community. A statistical approach to the full cloud simulation revealed that the peak radial outflow wind speeds can be linearly approximated from the circumferentially averaged mean radial wind speed.

The CS model is a relatively new tool in the study of thunderstorm downbursts (Anderson et al. 1992). Vermeire et al. (2011a) established that the spatial and temporal dependence of the CS function more accurately models the primary means by which vorticity is generated in the outflow. The CS model is also the only simplified engineering model that seeks to model the thermodynamic cooling which is the primary driving mechanism by which downbursts are formed. However, most CS studies are completed in a quiescent atmospheric base state that does not incorporate realistic variations in the vertical profiles in the atmosphere. Mason et al. (2009) performed a parametric study using the CS forcing function of Anderson et al. (1992), examining among other variables, the effect of using a simple atmospheric lapse rate, based on the Proctor (1989) ‘worst case scenario’ lapse rate. A steep lapse rate with a high melting level was employed, and it was found that the outward radial component of wind velocity was increased in magnitude slightly, although no large impact on the instantaneous velocity profile shapes was observed. However, it should be noted that this lapse rate was employed with no wind shear. Similarly, Mason et al. (2010) conducted a CS study that made use of an atmospheric base state with simple wind shear profiles, which showed good agreement to full scale simulations. As far as the present authors are aware, no CS studies have employed the CS forcing function inside a realistic atmospheric base state and attempted to make a comparison to full cloud thunderstorm simulations which make use of the same atmospheric base state. In essence, this involves performing a full cloud simulation without the computational complexity of the microphysics scheme and emulating the thermodynamic cooling using an idealized CS. Nearly all simplified engineering numerical models insist on estimating the peak wind speeds (those of interest to wind engineers) in a stationary quiescent environment that is
simply not observed in nature. A more realistic and reasonable model should be completed in an atmosphere that is based upon a composite field sounding.

The present study investigates the effect of placing an idealized CS forcing function into the more realistic atmosphere used in the more realistic full scale meteorological cloud model of Orf et al. (2012), Orf et al. (2014). The shape, size and CS intensity is set to approximately match the thermodynamic cooling of the full cloud simulation of Orf et al. (2012) and a comparison of the wind fields is made. It will be shown that the geometric properties of the CS of Anderson et al. (1992) and the cooling rate of Vermiere et al. (2011a) were fairly consistent with the attributes of the downdraft observed in Orf et al. (2012) and, as a result those geometric properties were selected for the current work. The purpose of this numerical study is to investigate if it is possible (or reasonable) to more-realistically estimate peak outflow winds when a simplified model is placed in a more realistic atmospheric base state. From a wind engineering standpoint, this method could offer a way to increase the realism of more simplified engineering IQI and CS models without increasing the complexity or computational requirements to the levels demanded by meteorological cloud models.

Hence, the overall objective of this work is to determine, using a numerical modelling approach based on Large Eddy Simulations, whether a simple CS model can replicate the structure of a downburst produced by a more sophisticated, full-physics, large-scale thunderstorm model, for the same horizontal, homogeneous ground (represented by the roughness length, $z_0=0.1$ m) and the same, realistic, atmospheric base state. Details of the numerical model and atmospheric base state will be explored, the scaling approach will be discussed, followed by the results of the simulations and, finally, the conclusions of the study will be presented.

### 3.2 Details of the numerical model

#### 3.2.1 Background information

The numerical model that is used for this study is Cloud Model 1 (CM1) release 18, henceforth referred to as cm1r18. CM1 is a numerical meteorological cloud model that was developed specifically for the idealized simulation of deep moist convection. Developed
by George Bryan at the National Centre for Atmospheric Research (NCAR). CM1 makes use of an Arakawa C type numerical grid for discretizing the Navier Stokes equations. For this study a 5th order scheme for vertical advection terms is employed and a 6th order explicit scheme for horizontal terms.

3.2.2 Governing equations of the model

More details of the governing equations can be found in Oreskovic et al. (2016a), and a full set of equation descriptions can be found in Bryan (2015). The primary governing equations will be summarized here.

CM1 iteratively solves the inviscid momentum equations,

\[
\frac{\partial u}{\partial t} + c_p \theta_p \frac{\partial \pi'}{\partial x} = ADV(u) + f v + T_u + D_u + N_u
\]  

(3.1)

\[
\frac{\partial v}{\partial t} + c_p \theta_p \frac{\partial \pi'}{\partial y} = ADV(v) + f u + T_v + D_v + N_v
\]  

(3.2)

\[
\frac{\partial w}{\partial t} + c_p \theta_p \frac{\partial \pi'}{\partial z} = ADV(w) + B + T_w + D_w + N_w
\]  

(3.3)

ADV is the advection operator of the variable u, v and w. f terms are the Coriolis acceleration effects, not considered in this study due to the relatively small scale of the flow, and B is the tendency of buoyancy. T is the tendency of sub grid scale turbulence, D is the tendency of diffusivity and N are the Newtonian relaxation terms.

The advection operator is described by the following equation, for any generic variable

\[
ADV(\alpha) = \frac{1}{\rho_0} \left[ - \frac{\partial (\rho_0 u \alpha)}{\partial x} - \frac{\partial (\rho_0 v \alpha)}{\partial y} - \frac{\partial (\rho_0 w \alpha)}{\partial z} \\
+ \alpha \left( \frac{\partial (\rho_0 u)}{\partial x} + \frac{\partial (\rho_0 v)}{\partial y} + \frac{\partial (\rho_0 w)}{\partial z} \right) \right]
\]  

(3.4)
The potential temperature perturbation is solved using the following relation,
\[
\frac{\partial \theta'}{\partial t} = ADV(\theta) + T_\theta + D_\theta + N_\theta + \frac{1}{c_p \pi} \varepsilon + q(x, y, z, t) \tag{3.5}
\]

where \( q \) is the spatially and temporally dependent CS forcing function. The potential pressure perturbation is solved using the following,
\[
\frac{\partial \pi'}{\partial t} = -\frac{g}{c_p \theta_0} w + \frac{R}{c_v} \pi \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) = ADV(\pi') + \frac{R \pi}{c_v \theta} (T_\theta + D_\theta + N_\theta) \tag{3.6}
\]

### 3.2.3 Sub-grid turbulence model

For this study, a turbulent kinetic energy (TKE) sub grid scale turbulence closure was selected, similar to the TKE scheme described in Deardorf (1980). The sub-grid turbulence scheme of Deardorf (1980) has been selected over the standard Smagorinsky (1963) option, as it offers some improvements in performance (Oreskovic et al. 2016a). The sub-grid turbulence tendencies are formulated in the following way;

\[
T_u = \frac{1}{\rho} \left[ \frac{\partial \tau_{11}}{\partial x} + \frac{\partial \tau_{12}}{\partial y} + \frac{\partial \tau_{13}}{\partial z} \right] \tag{3.7}
\]

\[
T_v = \frac{1}{\rho} \left[ \frac{\partial \tau_{12}}{\partial x} + \frac{\partial \tau_{22}}{\partial y} + \frac{\partial \tau_{23}}{\partial z} \right] \tag{3.8}
\]

\[
T_w = \frac{1}{\rho} \left[ \frac{\partial \tau_{13}}{\partial x} + \frac{\partial \tau_{23}}{\partial y} + \frac{\partial \tau_{33}}{\partial z} \right] \tag{3.9}
\]

\[
T_s = \frac{1}{\rho} \left[ \frac{\partial \tau^s_1}{\partial x} + \frac{\partial \tau^s_2}{\partial y} + \frac{\partial \tau^s_3}{\partial z} \right] \tag{3.10}
\]

where \( u, v \) and \( w \) subscripts represent the filtered velocity terms, and \( s \) is any scalar spatially filtered term. The shear stress terms are solved using the following approximation,
\[
\tau_{ij} = -\rho u'_i u'_j = 2\rho K_m S_{ij} \tag{3.11}
\]
A more detailed presentation of the turbulence closure can be found in Oreskovic et al. (2016a).

### 3.2.4 Numerical details

#### 3.2.4.1 Grid independence

A grid independence test was performed in Oreskovic et al. (2016a), where the velocity profile of the radial component of wind was plotted for three different computational grids. The grids considered in that test were 15 m, 10 m and 6 m horizontally homogeneous grids, with identical vertical grid spacings. It was found that peak wind velocities in the profiles corresponding to the two finest meshes resulted in a difference of less than 0.1% and, therefore, the 10 m homogeneous grid was selected for that study and for the present work. A similar grid sensitivity test was performed in Vermeire et al. (2011a), where it was also confirmed that the 10 m grid spacing resulted in a grid independent solution.

#### 3.2.4.2 Computational details

Although the numerical model is described in more detail in Oreskovic et al. (2016a), a summary will be given here. All simulations in this study make use of the same computational grid and domain. The computational domain consists of four lateral boundary conditions that are treated as outflows (open radiative surfaces). The ground boundary consists of a semi-slip wall and the top boundary is considered a free-slip surface. The ground is treated with a surface roughness of $z_0=0.10$ m to match that used in Vermiere et al. (2011a) and Orf et al. (2012). The domain is discretized using an Arakawa C type numerical scheme (Arakawa and Lamb 1977), with horizontal homogeneous grid spacing, and a grid stretch is employed in the vertical direction using the Wilhelmson and Chen (1982) stretching scheme. The horizontal grid points are spaced at $\Delta x=\Delta y=10$ m, and the vertical domain points are spaced at $\Delta z=1$ m at the surface and up to $\Delta z=50$ m at the top of the domain. The vertical stretching was used to focus the resolution near the surface of the ground, where the peak outflow wind velocities are typically recorded for thunderstorm downbursts. The computational domain extends to approximately 9.6 km in the horizontal directions and 4 km in the vertical direction and, as a result, the computational domain
occupies approximately 369 km$^3$. The total number of nodes in the computational domain is approximately 147 million. An improved time step of $\Delta t=0.05$ s with 10 acoustic substeps was used, compared to the $\Delta t=0.0625$ s of Vermeire et al. (2011a). This change was made as some numerical instabilities occurred when peak velocities greater than those recorded in previous studies were encountered. This was made possible through the use of XSEDE’s (Extreme Science and Engineering Development Environment) stampede clusters’ increased resources. All simulations in this study were run for a period of 600 s (10 min) model time. Full domain data were recorded every 30 s for the first 240 s, 5 s from a period of 245 s to 450 s and, again, every 30 s from 450 s until the end of the simulation. This was done to reduce the computational load and large memory overhead caused by saving data too frequently. Statistical data were saved every 0.5 s for the duration of the simulation.

### 3.2.5 Cooling source function

The same CS function defined in Anderson et al. (1992) and Vermeire et al. (2011a) is used for this study. The CS is ellipsoidal in shape and is both spatially and temporally dependent, mimicking the natural thermodynamic cooling that occurs in downburst producing thunderstorms. The CS ramp up function uses a $\cos^2$ spatial ramp-up,

$$q(x, y, z, t) = \begin{cases} g(t)\cos^2 \pi R & \text{for } R < 1 \\ 0 & \text{for } R > 1 \end{cases}$$

where $R$ is the normalized non-dimensional position within the ellipsoid,

$$R = \sqrt{\left(\frac{x-x_0}{h_x}\right)^2 + \left(\frac{y-y_0}{h_y}\right)^2 + \left(\frac{z-z_0}{h_z}\right)^2}$$

$x_0, y_0$ and $z_0$ are the locations of the centre of the ellipse, $x, y$ and $z$ are the location anywhere within the ellipsoid, and $h_x, h_y$ and $h_z$ are the horizontal half widths and half height of the ellipsoid, respectively.
The function $g(t)$ is the temporal cooling rate ramp-up, a piecewise function which includes an initial $\cos^2$ ramp up function, followed by a steady cooling rate period, and then a $\cos^2$ ramp down function to zero, defined as,

$$g(t) = \begin{cases} 
-\cos^2 \left[ \pi \frac{t - 120}{2(120)} \right] & \text{if } t < 120 \\
-1 & \text{if } 120 < t < t_2 \\
-\cos^2 \left[ \pi \frac{t - t_2}{2(120)} \right] & \text{if } t_2 < t < t_3 \\
0 & \text{if } t > t_3 
\end{cases} \quad (3.14)$$

The present study investigates the effect of modifying the values of $t_2$ and $t_3$, to match the cooling ramp down to that which is observed in nature and in more sophisticated full cloud models. The values that will be used for $t_2$ and $t_3$ are 300 s and 420 s, and 200 s and 320 s, for those simulations which will investigate the effect of reducing the duration of the thermodynamic cooling.

Figure 3.1 – Diagram of the ellipsoidal CS with axis naming convention

3.2.6 Random temperature perturbation addition

Although an improvement over simple IJ models of downbursts, the smooth temperature and wind profiles found in the CS model of Vermeire et al. (2011a,b) and Oreskovic et al.
do not exist in nature. The thermodynamic cooling caused by the evaporation of precipitation present in a realistic atmosphere does not take on the perfectly uniform shape seen in the idealized CS model. In order to mimic the more realistic spatial variation that would be present in natural thermodynamic cooling, randomness has been added to the CS ramp up function. Random perturbations in temperature have been added to the CS function at every model domain grid point.

\[ \theta_{\text{rand}} = a(n_{\text{rand}} - 0.5) \]  

(3.15)

where \( \theta_{\text{rand}} \) is the random temperature perturbation at any given grid point within the CS ellipsoid and \( a \) is the specified amplitude (fixed for a given simulation) of the perturbation. The amplitude of the perturbation can be controlled from 0 (no perturbation) to 1. \( n_{\text{rand}} \) is a random number (between the value of 0 and 1). This script makes use of the Fortran mpi (message passing interface) random number generator of random_seed, which initializes a pseudo random number seed from the mpi rank. In order to have random spatial variability in the perturbations that are added to the CS, it is necessary to ensure that the random seed is unique for each mpi rank in the simulation.

3.3 Imposed atmospheric base state

For this study, a more realistic atmospheric base state is imposed on the model, in an effort to provide atmospheric conditions more representative of those encountered by a downburst in the natural environment. The atmospheric base state is the same as that in the full cloud simulation of Orf et al. (2012), where a downburst producing thunderstorm was simulated in a fairly high resolution numerical environment. This base state is constructed using an atmospheric sounding which has been found to be conducive to the formation of downburst producing thunderstorms on the high plains of the United States (Brown et al. 1982). The Brown et al. (1982) sounding was recorded in the field on July 6th 1980 in Boulder, Co. where a downburst was observed with peak outflow wind speeds of 25 m/s lasting for a period of 1 min – 2 min accompanied by light rain and hail and a second period of strong winds of approximately 20 m/s lasting for another 2 min. The base state used for this simulation is horizontally homogenous, with variation only in the vertical direction. The present study was run in a dry environment and so only information regarding the
temperature and pressure of the air and wind direction (and magnitude) are considered here. The sounding contains a deep (up to an elevation of 4 km, the top of the model domain) and well mixed boundary layer, with wind shear near the surface. This atmospheric sounding indicated a lifted condensation level of 3779 m above ground level, a level of free convection 3858 m above ground level and a calculated convective potential energy of 277 J/kg, and exhibits a strong potential for thunderstorms that produce dry downbursts. The data shows a classic ‘inverted-V’ sounding, with a deep, well-mixed boundary layer with very low relative humidity values that is conducive to the very efficient evaporation of liquid, suitable for a dry downburst. The sounding is shown as a Skew-T Log-P plot in Fig. 3.2. The red line represents temperature as a function of height, and the wind barbs on the right contain information regarding the wind magnitude and direction as does the hodograph in the upper right. The green line represents the dew point temperature, but is not considered in this simulation as this study contains a dry atmosphere (the moisture related equations in the model are not solved).

Figure 3.2 - Horizontally homogeneous base state sounding used for this study, adapted from Brown et al. (1982)
The wind barbs on the far right show wind velocity and magnitude by the orientation and number of barbs. Each long barb represents 10 kts and a short barb represents 5 kts. The orientation is given by the typical North-South-East-West arrangement, where a barb is vertical and pointing up when the wind comes from the North.

### 3.4 Methodology

#### 3.4.1 Details of the study

This study consists of 5 separate simulations. The first replicated the baseline simulation of Anderson et al. (1992) with a modified CS intensity of -0.08 K/s to match Vermiere et al. (2011a). The second simulation is similar to the first, but with random perturbations added to the region of the CS to investigate the effect on the outflow wind fields. The other three simulations investigated the effect of placing the baseline CS into a non-quiescent environment, with a non-zero atmospheric lapse rate and modest wind shear near the surface. The duration of the CS was modified for each of these three simulations, the first of which maintained the same CS temporal evolution of Anderson et al. (1992), namely a ramp up period from 0 s to 120 s, followed by a steady state period from 120 s to 720 s, and a ramp down period from 720 s to 840 s. The other two simulations modified this evolution in order to better replicate the time history when compared to more realistic simulations and field data. The ramp down occurred at 200 s and 300 s and lasted for the same 120 s period. All three simulations placed in the atmospheric base state of Brown et al. (1982) also subjected the CS to a horizontal translation of 5.2 m/s in the north direction ($v_{\text{move}}$) and 2.2 m/s in the east direction ($u_{\text{move}}$). This translation corresponded to the mean wind velocity at an elevation of the centre of the CS (2 km). This was done since the thermodynamic cooling that causes downbursts is the result of the evaporation of precipitation, which translates at the same speed as the parent storm.
### Table 3.1 – Details of the simulations of the current study

<table>
<thead>
<tr>
<th>Reference Cooling Source (Vermeire et al. (2011))</th>
<th>Temperature Perturbation Addition</th>
<th>$u_{move}$ (m/s)</th>
<th>$v_{move}$ (m/s)</th>
<th>Base State</th>
<th>$t_{ramp down}$ (s)</th>
<th>$t_{CSoff}$ (s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0%</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>Quiescent, Vermeire et al. (2011a)</td>
<td>720</td>
<td>840</td>
</tr>
<tr>
<td>Random Temperature Perturbation Addition Reference Simulation</td>
<td>100%</td>
<td>0</td>
<td>0</td>
<td>Quiescent, Vermeire et al. (2011a)</td>
<td>720</td>
<td>840</td>
</tr>
<tr>
<td>Atmospheric Base State</td>
<td>100%</td>
<td>2.2</td>
<td>5.2</td>
<td>Brown et al. (1982)</td>
<td>720</td>
<td>840</td>
</tr>
<tr>
<td></td>
<td>100%</td>
<td>2.2</td>
<td>5.2</td>
<td>Brown et al. (1982)</td>
<td>300</td>
<td>420</td>
</tr>
<tr>
<td></td>
<td>100%</td>
<td>2.2</td>
<td>5.2</td>
<td>Brown et al. (1982)</td>
<td>200</td>
<td>320</td>
</tr>
</tbody>
</table>

Fig. 3.3 shows the temporal dependence of the CS forcing function for the three different scenarios used in this study. The cooling of the air ramps up as a $\cos^2$ function from zero, reaches a peak of unity, and ramps back down.
3.4.2 Naming convention

From here on in the simulations will be referred to in a shortened form to easily state the names of the simulations with as little confusion as possible. The reference simulation will be known as Vermeire et al. (2011a). The random temperature perturbation addition reference simulation will be known as ‘random temperature perturbation reference’. The three simulations which include an imposed realistic atmospheric base state will be known as ‘$t_{g(t),\text{max}}=\infty$’, ‘$t_{g(t),\text{max}}=300$’ and ‘$t_{g(t),\text{max}}=200$’, for the simulations that use the Vermeire et al. (2011a) temporal ramp up/down, the simulation that uses a thermodynamic cooling cut off at 300 s, and the simulation that uses a ramp up cut off at 200 s, respectively.

3.4.3 Circumferential averaging of raw data

Due to the large volume of data produced by the simulations, some post processing is required for two reasons. First, to reduce the amount of raw data to be analyzed, and second to simplify the data to a two dimensional (r-z) plane, as is most common in other work (axi-symmetric). Also, data for this study are compared to the full cloud simulation of Orf et al. (2012, 2014) which made use of a circumferential averaging processing method to convert the highly spatially variable full cloud data into a single (r-z) plane. The circumferential averaging process used for this study is identical to Oreskovic et al. (2016a) which is based on a similar process outlined in Orf et al. (2014), by which the centre of the event is identified (directly beneath the centre of the CS) and the data are averaged along the circumference of an imposed circle of radius r (measured from the impingement
This procedure is then repeated for all radii and heights. A more detailed explanation can be found in Oreskovic et al. (2016a). It should be noted, however, that for the non-scalar quantities of wind velocity in the u and v directions there exists a horizontal translation that must be added to (or subtracted from) the radial wind velocities in order to obtain the ground referenced wind speeds. For the leading edge of the CS outflow the horizontal components of translation are added to the radial wind speeds and for the trailing edge of the outflow the translation vector is removed from the outflow radial winds. For the two sides of the outflow that are not parallel to the flow, no such addition is needed. This can be summarized in the following equations.

\[ U_{translation} = \sqrt{u_{move}^2 + v_{move}^2} \]  

\[ U_{r, leading edge} = U_r + U_{translation} \]  

\[ U_{r, trailing edge} = U_r - U_{translation} \]

For non-directional scalar terms, such as temperature, no translation needs to be considered and circumferential averaging can be directly applied, since all quantities can be referenced to the approximate (since the wind shear of Brown et al. (1982) causes symmetry to be lost) centre of the down flow, not the ground. This circumferential spatial averaging process is indicated in the Fig. 3.4, including the horizontal translation of the CS represented as the vector addition of the translation of the domain.
Figure 3.4 - Computational domain and circumferential averaging coordinates with vector diagram of translation wind

The north-south direction will be defined here on in as the compass standard north-south and will correspond to the ‘v’ components of velocity. The east-west direction will be similarly defined here on in as the standard east-west direction and will correspond to the ‘u’ component of wind velocity, where the origin (0 km, 0 km) is located at the South-West corner. This convention is used for all subsequent diagrams that show spatial data.

3.4.4 Establishing spatial and temporal scales

The present work will focus on making a comparison between the idealized CS model of a thunderstorm downburst and the more sophisticated meteorological cloud model approach of Orf et al. (2012) when performed in the same atmospheric base state. The CS is not an exact replicate of the full cloud set up, and gross differences between the two downburst winds are evident in Fig. 3.10 and Fig. 3.12 of the present work. Some simplifications of the physics are made in the CS model and, as a result, there are some obvious differences in the spatial structure of the two events. In order to make the comparison between the two events more clear, it is necessary to give a qualitative comparison of the spatial scales for both of the events. The full cloud simulation downburst of Orf et al. (2012) is initiated by
a cool mass of air with a main downflow column with a radius of approximately 650 m, while the CS cold air mass has a radius of approximately 1200 m. The radial position \( r_{\text{max}} \) where the peak outflow wind speeds are observed for the CS simulation occur at \( \sim 1100 \) m, while the full cloud downburst shows an \( r_{\text{max}} \) of approximately 1500 m. The height at which the peak outflow wind speed is observed \( (z_{\text{max}}) \) is approximately 40 m for both simulations. The base of the cloud, or the centre of the cold mass of air for the thunderstorm simulation is around \( z=2000 \) m, and the CS forcing function is also centred at \( z=2000 \) m. In the temporal domain, the peak outflow wind speeds are observed to occur at \( \sim 100 \) s after impingement, while for the full cloud simulation these winds are observed to occur similarly at approximately 120 s after impingement.

Additionally, the full cloud simulation downburst is accompanied by a number of other adjacent but weaker downburst-like masses of air, evident in Fig. 3.12. In the full cloud simulation, the main downburst is preceded by weaker downward air that ‘preconditions’ the environment in which the main downburst occurs. This ‘preconditioning’ and interaction with other adjacent events is most likely responsible for the bulk of the differences between the full cloud simulation results and the simplified CS. The CS model in its current form does not allow for adjacent events or a ‘preconditioning’ in the atmospheric base state, as the CS models the thermodynamic cooling of a single event. It is this fundamental difference that illustrates the realism of the full cloud model approach to simulating downbursts that is very challenging to replicate in simplified models. Additionally, applying a unified scaling approach (as is introduced in sec. 4.5 and discussed in sec 5.6) is challenging due to the spatially ill-defined cooling region in the full cloud model.

### 3.4.5 Scaling approach

As outlined in Oreskovic et al. (2016a), the non-dimensional scaling approach of Lundgren et al. (1992) and Yao and Lundgren (1996) can be applied to the CS simulations with some success. The Lundgren et al. (1992) scaling approach is designed for use with liquid drop release experiments, where the initial volume of dense liquid is clearly defined and the interface between the dense fluid and the ambient fluid is also very clear. For a CS, however, the source of dense fluid varies spatially and temporally. As a result, selecting
the correct initial volume of cold air, and the density difference between the source and the surrounding air is more difficult. This problem is even more difficult to overcome for the full scale meteorological simulations, as the cold volume of air is even more poorly-defined due to locally variable environmental conditions. An attempt to apply the Lundgren et al. (1992) scaling to these more complex simulations is also attempted here, because developing a non-dimensional scaling method is important for comparing the results from different types of models. The equations for the scaling variables are found below.

\[
Q = \frac{4}{3} \pi h_x h_y h_z \quad R_0 = \left( \frac{3Q}{4\pi} \right)^{\frac{1}{3}} \quad T_0 = \left( \frac{R_0 \rho}{g \Delta \rho} \right)^{\frac{1}{2}} \quad V_0 = \frac{R_0}{T_0} \quad Re = \frac{V_0 R_0}{v}
\]

A more detailed description of the equations can be found in Oreskovic et al. (2016a) as well as Lundgren et al. (1992). Here, Q is the volume of dense liquid, \( R_0 \) is the equivalent spherical radius of the dense liquid, \( T_0 \) is the dimensional time scale, \( V_0 \) is the dimensional velocity scale, and \( Re \) is the Reynolds number. The values \( h_x, h_y \) and \( h_z \) are the two CS horizontal half widths and the half height, respectively.

### 3.5 Results and discussion

#### 3.5.1 Circumferential averaging of CS simulations

Both the idealized CS simulations introduced in Anderson et al. (1992) and those used in this study produce outflow wind fields that are extremely smooth when compared to those arising from more sophisticated numerical studies such as Orf et al. (2012). When comparing vertical profiles of wind magnitude from full cloud simulations, there exists a large spatial variation with a fairly constant standard deviation of values around the imposed circumference where the spatial average is taken. The difference between the circumferential mean and the corresponding peak is significant, as discussed in Orf et al. (2014). However, for the CS simulations the ratio between mean and peak is essentially unity in the outflow before the onset of turbulence or before symmetry breaks. This spatial
variability can be observed in Fig. 3.5, where the circumferential mean and peak are plotted for the full cloud simulation.

![Figure 3.5 - Circumferentially averaged radial wind velocity for the full cloud simulation at t=3606s and r=1500m, and the corresponding peak value around the circumference](image)

For the CS simulation in a quiescent environment with no random temperature perturbation addition, for an approximately equivalent radial profile (this location is selected to be in the region near the peak velocity and also in a location that is visually similar) in Fig. 3.6 right, there exists almost no spatial variability around the imposed circumference (before the onset of turbulence), owing to the extreme smoothness of such idealized simulations. Any type of variability only begins to exist once the symmetry is broken, as turbulence sets in. This is not unexpected, as the CS forcing function is symmetrical by definition. Interestingly, even with the addition of random temperature perturbations (as seen in Fig. 3.6 left), the variability present in the full cloud simulation cannot be matched, presumably without increasing the randomness unrealistically.
Figure 3.6 - Circumferential mean and corresponding peak wind velocities for r=1100m at t=370s (before symmetry is broken) with random temperature perturbation addition (left) and without addition (right) in a quiescent atmosphere

The more realistic vertical profiles of circumferentially-averaged wind speeds that can be observed in the full cloud simulation (Fig. 3.5), where the mean and peak radial wind speeds are very different, are also found when the CS forcing function is placed into the same conditions that includes the effects of atmospheric stratification and wind shear found in the real atmosphere. The non-uniform atmospheric base state introduces considerable circumferential variation in wind speed, as can be seen in Fig. 3.7. The variation in wind speed and direction as a function of height introduced by the sounding, breaks the symmetry that would be present in a quiescent atmospheric base state. This is an encouraging result, as thunderstorm outflows do not show the symmetry present in idealized simulations, such as the CS or IJ in a quiescent environment.
Figure 3.7 - Circumferential mean and corresponding peak wind velocities for r=1100m at t=370s (before symmetry is broken) with random temperature perturbations in Brown et al. (1982) base state

It would appear that the addition of randomness into the cooling forcing function does not change the circumferential spatial variability of the outflow, as there is almost no observable difference between the circumferential mean and peak values for the vertical profile of the instantaneous radial component of the wind vector. It seems that the addition of random temperature perturbations assists in achieving a more numerically stable simulation and, importantly, a realistic outflow that has no preference for the numerical grid orientation, as is discussed in the next section.

3.5.2 Observed improvement using random temperature perturbations

As outlined in Oreskovic et al. (2016a), using a 5th order advection scheme for both horizontal and vertical terms, as well as placing the CS in an imposed symmetry domain (Vermiere et al. 2011a), produced a cleft in the outflow temperature fields at locations of 0° and 90°. This anomaly was corrected in the present work by using a 6th order explicit scheme for horizontal advection terms (while carefully selecting the diffusion coefficient) and removing the imposed symmetry by placing the CS in a full four quarter domain. Making this change resulted in a flow that appears to not have any preference for the orientation of the numerical grid (the X-Y direction). The 6th order scheme resulted in a corrected outflow that had no imposed ‘ripples’ at the symmetry interface. However,
despite the major improvements in the computationally stability that were made over Vermeire et al. (2011a), some minor, seemingly grid imposed, anomalies still occurred at some positions around the circumference of the outflow even in a 6th order explicit scheme. With the 6th order advection scheme the outflow reached a maximum radial velocity at 0°, 45° and 90° a few seconds (model time) before the rest of the outflow. These anomalies were not observed in isosurface plots of the temperature field (as they were when using the 5th order advection scheme), but the subtle variation in wind speed around the circumference can be seen in the surface wind swath plots. As far as the authors are aware, the limitations of simulating a radial downburst flow on a Cartesian grid for these idealized CS forcing functions has not been investigated previously. The largest criticism of the idealized CS model in Oreskovic et al. (2016a) is that it produces outflows that are far too smooth when compared to observations in nature or to a full cloud meteorological simulation. This ‘over symmetry’ causes turbulence to arise earlier in certain regions in the outflow before others, presumably due to rounding errors in the solution. The instability that occurs in the 6th order explicit scheme is revealed when observing the surface wind swath at the surface (the maximum recorded surface wind speed). This numerical anomaly is reduced in this study by introducing random temperature perturbations to the CS as discussed in Sec. 3.2.6. The improvement can be observed in Fig. 3.8, where on the left is the surface wind swath of the baseline CS of Vermeire et al. (2011a) in a full domain with 6th order explicit advection (k_{diff,6}=0.05) and on the right is the same simulation with added randomness to the CS forcing function. Turbulence arises earlier and in a more realistic way, visually closer to the outflows observed in the full cloud simulation of Orf et al. (2014). Peak wind speeds seem to develop earlier at angular positions that are a multiple of 45° for the simulation with the 5th order advection scheme, a phenomenon that would not be observed in nature.
Figure 3.8 - Comparison of the surface wind swath between a simulation with no random temperature perturbations (left) and with random temperature perturbations to the CS (right) in a quiescent atmospheric base state

3.5.3 Temporal relationship between CS and Outflow Strength

Other studies that make use of the CS of Anderson et al. (1992), such as Vermiere et al. (2011a), neglected the impact of the temporal variation of the CS ramp up function on the decay of the outflow wind field. Mason et al. (2009), however, did investigate some aspects of the temporal ramp up function, completing a total of six unique simulations where the CS ramp up time was modified. Mason et al. (2009) selected ramp up periods of 30 s, 60 s, 120 s, 240 s, 360 s, as well as a ‘short burst’ ramp up where the CS was shut off immediately once the CS reached peak cooling. They determined that the length of the ramp up function had little impact on the normalized (by peak value) $u$ (x-direction or East-West direction) velocity profiles as a function of height. It was noted, however, that the peak velocity did appreciably change for simulations where the ramp up function lasted longer than the baseline event. Mason et al. (2009) did not, however, investigate the impact of the ramp up function duration on the decay of the outflow wind velocities. The present
study investigated three simulations where the CS ramp up function is cut short when compared to Vermeire et al. (2011a). Orf et al. (2014) found that the full scale cloud model simulations differed from more idealized IJ and longer duration CS simulations in the way that the peak outflow wind velocities decay with time. This temporal decay of outflow winds is typically observed in field measurements (Hjelmfelt 1987, Gunter and Shroeder 2015). Shown in Fig. 3.9 is the circumferentially-averaged maximum radial wind speed in the computational domain, for each of the three simulations for different steady state cooling periods. It would seem that the $t_{g(t),\text{max}}=200$ simulation produces a peak wind decay that is most analogous to the full cloud simulation of Orf et al. (2014), as can be seen in App. A Fig. A.7 and, hence, most of the analysis here on in will focus on this simulation.

![Figure 3.9 - Time history of peak radial wind velocity for three simulated CS events](image)

### 3.5.4 Influence of wind shear on outflow

As is shown in the vertical profiles of circumferentially averaged radial wind speed, the wind shear that is imposed on the CS has a significant impact on the downburst outflow shape. The symmetry imposed by the symmetrical CS forcing function is broken and the outflow resembles that observed in the full cloud simulations. The effect of a very simple wind shear on a downburst outflow was also studied in Orf and Anderson (1999), showing similar realism, although the wind shear found in Brown et al (1982) is more complex. Fig. 3.10 plots the temperature surface of $T=323$ K for the entire computational domain. It can
be seen that the CS deforms as it descends from the central cooling location and interacts with the higher altitude (~2 km) winds. Upon impingement, the outflow is affected by the near ground wind shear, so that the leading edge lifts up off the ground faster than the trailing edge, which is forced downwards. The imposed temperature base state results in a lower energy potential between the cool downburst air and the surrounding ambient atmosphere, as the overall temperature of the downburst column is warmer by approximately 15 K compared to Oreskovic et al. (2016a) and Vermeire et al. (2011a,b). This resulted in a slower moving outflow (peak outflow winds are slightly less than those observed in Vermeire et al. (2011a)), as expected. Fig. 3.11 shows the atmospheric air temperature at a height of 1 m AGL for the entire domain. It can be seen that symmetry is lost in this simulation, as the downburst outflow interacts with the environmental winds at this altitude. The circular shape of the CS forcing function is most notably lost at the most eastward flank of the outflow. Fig. 3.12 illustrates the spatial complexity present in full cloud simulations of downbursts, as well as the ‘pre-conditioning’ of the lower atmospheric base state.
Figure 3.10 – Isosurface (T=323K) temperature field of the $t_{g(t),\text{max}}=200$ simulation at model time of $t=370$ s (top) and $t=445$ s (bottom)
Figure 3.11 - Potential temperature at 1 m AGL for $t_{g(t),max}=200$ at $t=445$ s

Figure 3.12 – Volume render of the potential temperature field for the full cloud simulation of Orf et al. (2012) for $t=3426$ s (top) and $t=3614$ s (bottom). Temperature perturbation ranges from 1 K (yellow) to 4 K (dark blue), with annotations indicating the primary and adjacent (secondary) event.
3.5.5 Downflow and outflow characteristics

Fig. 3.13 shows the vertical profiles of the circumferentially averaged data from Orf et al. (2014), which correspond to the region of the outflow where the highest radial wind speeds are observed, as well as the circumferentially averaged profiles from the current study for the region (and time) near the corresponding highest radial outflow winds for the three cases ($t_{g(t),\text{max}}=\infty$, $t_{g(t),\text{max}}=300$, $t_{g(t),\text{max}}=200$). Specifically, these profiles come from $t=310$ s and a radial position of approximately 1100 m (visually comparable). The profiles have been normalized by the peak outflow wind speeds in those regions, as well as the height at which that peak value occurs. It is seen that agreement near the surface, at altitudes below the peak wind speed location, is very good, presumably due to the consistent surface roughness of $z_0=0.1$ m. Agreement is fairly reasonable for regions above the peak velocity location, most notably for $t=3588$ s in the full cloud simulation. Some divergence occurs at $z/z_{ur,max}$ greater than 8. As is stated in Orf et al. (2014), these types of normalized vertical profile plots used to compare downburst outflows of various studies should be read with some caution, as the normalization method causes all of the curves to collapse at $(1,1)$, signifying good agreement even if a good agreement is not present in the non-normalized data. However, this figure does show that the CS outflow results are contained within the envelope of those from the more sophisticated model. This is somewhat encouraging, as in Appendix A of this thesis, a similar plot is constructed comparing the full cloud work to the vertical profiles from other simplified CS studies such as Vermeire et al. (2011) and Lin et al. (2007), which fall outside of the boundary of the full cloud simulation profiles. It is also found that the present profile results match fairly closely to the CS results of Mason et al. (2009).
A better comparison of the two events can be made in a fully dimensional plot of the vertical profiles of wind at a given location in the outflow. Fig. 3.14 plots the circumferential mean vertical profile, along with its corresponding peak value, for $t=3606$ s and $r=1500$ m for the full cloud simulation of Orf et al. (2012), and the circumferential mean for the current study at $t=310$ s and $r=1100$ m for the $tg(t)_{max}=200$ simulation. These times and radial locations were selected as representing the time and radial location at which the maximum radial wind speed occurred for both cases. It can be seen in the plot that the agreement between the circumferential mean profiles is rather poor, although this is expected. The spatial variability present in the full cloud simulation results in a large standard deviation between the circumferentially averaged mean and corresponding peak values, as there is a large discrepancy between peak and minimum wind velocities around the circumference. However, the agreement between the circumferentially averaged mean profile for the CS results, and the peak winds of the full cloud simulation is extremely good (Fig. 3.14). Both the height of the peak circumferential of the CS, and the height of the peak of the peak of the full cloud simulations are around 40 m AGL, and the magnitude of the winds are in the 35 m/s range. The peak of these profiles are also in reasonable
agreement with field data from (Gunter and Shroeder 2015) that show peaks in the 25 m/s to 45 m/s range. It should be noted that this CS profile does not incorporate the horizontal translation speed increase. The unfilled triangle marks represent the circumferentially averaged radial velocity including the horizontal translating velocity of the CS, and the diamond markers represent the trailing edge of the downburst subtracting the translating component. It is encouraging that the boundary of the maximum and minimum wind velocities at this radial and temporal position envelope the peak wind speeds of the full cloud simulation. However, in the CS simulations, a maximum temperature deficit of approximately 12 K is recorded, whereas in the full cloud simulations the deficit reaches a maximum of approximately 4 K. It is clear that a far larger temperature deficit is required in the CS model to reach similar outflow wind speeds. Orf et al. (2012) speculated that the drag induced by falling precipitation aids in the intensity of the outflow wind speeds, a result that seems to be supported in the present work. Since microphysics is not modelled in the CS, there exists no falling precipitation that can create a stronger downflow.

Figure 3.14 - Circumferentially Averaged mean value and corresponding peak for full cloud simulations at t=3606 s and r=1500 m, and the circumferentially averaged mean for t=310 s and r=1100 m for the $t_{g(t),\text{max}}=200$ simulation
Fig. 3.15 shows the radial position at which the maximum circumferentially averaged radial wind speed occurs as a function of time for \( t_{g(t),\text{max}=\infty} \), \( t_{g(t),\text{max}=300} \), \( t_{g(t),\text{max}=200} \). This type of plot is particularly important as it illustrates the growth of the outflow in terms of its peak wind speed position, rather than the outflow front as in Oreskovic et al. (2016a). The time of impingement can be observed to take place at approximately 300 s model time.

![Graph showing radial position vs time](image)

**Figure 3.15 - The radial position where the maximum circumferentially averaged radial wind speed occurs in the outflow**

Fig. 3.16 investigates the evolution of the height at which the maximum radial wind velocity occurs as a function of radial location in the downburst outflow, for three different times centred around the time at which the peak outflow velocity occurs for the \( t_{g(t),\text{max}=200} \) simulation, as well as the same metric for the full cloud simulation. It can be seen that the height grows almost linearly as the roll vortex travels along the ground outwards. The vortex climbs in height and grows in size further along in time. A different pattern is observed for the full cloud data set, where no peak occurs presumably due to a lack of roll vortex (any roll vortex present in the full 3-D data is ‘washed away’ during the circumferentially averaging process). The growth in the height of the location of the peak radial wind speed is far less structured and more gradual.
Figure 3.16 - Height to the maximum radial velocity for each radial position in the outflow for $t_{g(t),\text{max}}=200$ simulation at $t=280\text{s}$, $t=310\text{s}$ and $t=340\text{s}$(left), and for the full cloud simulation of Orf et al. (2014) at $t=3588\text{s}$, $t=3606\text{s}$ and $t=3628\text{s}$ (right), near the time at which peak outflow wind speed occurs.

Fig. 3.17 shows the variation of the maximum radial velocity for three different times in the present simulation (left) and Orf et al. (2014) (right) centred on the peak outflow radial velocity, as a function of the radial position. It can be seen that the peak outflow radial velocity for the $t_{g(t),\text{max}}=200$ simulation occurs somewhere in the region of $r=1000 \text{ m}$ from impingement. This is slightly less consistent when compared to the full cloud simulation, as the peak outflow radial velocity occurs closer to $r=1500 \text{ m}$. However, this is not completely unexpected as the spatial scales of the down flow, in particular, are not equivalent, as the full cloud simulation, despite the efforts to match the down flow column closely, has a wider outflow presumably due to adjacent downburst events. It is clear that both outflows follow a similar trend, where the peak outflow wind speeds reach an overall peak and decay with time.
Figure 3.17 - The circumferentially averaged radial velocity at the height at which the maximum occurs for each radial position in the outflow for the $t_g(t),_{\text{max}=200}$ simulation at $t=280s$, $t=310s$ and $t=340s$ (left), and for the full cloud simulation of Orf et al. (2014) at $t=3588s$, $t=3606s$ and $t=3628s$ (right).

Fig. 3.18 plots the same quantities as Fig. 3.17 but normalized by the radial position at which the peak outflow wind occurs, as well as by the peak velocity. Plotted along with these sets of simulations is the idealized model of Holmes and Oliver (2000). The ambient winds, which are not present in simplified models such as Holmes and Oliver (2000), can easily be seen in this normalized plot as the agreement between the simulations is quite good from $r/r_{\text{max}}=0$ to $r/r_{\text{max}}=1$. This seems to suggest that including environmental winds, at a minimum, significantly increases the realism of simplified simulations. Additionally, for radial positions far away from the region where the peak occurs, an even more realistic pattern occurs, as the winds do not decay away to zero, rather a residual value of the environmental winds remains. As is observed in nature, such as during the July 14 1982 downburst event at Stapleton Airport, and at Andrews Airforce base on August 1 1983 (Fujita, 1985).
Figure 3.18 - The variation of normalized circumferentially averaged radial velocity at the height of maximum velocity with the corresponding normalized radial position

Fig. 3.19 plots the maximum downwards flow velocity (solid line), and the maximum outward radial velocity (dashed) against time for the duration of the simulation for the entire computational domain. This plot contains data only for the simulation with a ramp-up period of 200 s, as it was found that this simulation results in a wind speed decay that is most typical of the full cloud simulation. It is observed that the outflow velocity occurs approximately 60 s - 100 s after the peak downwards velocity in the initial downdraft column. This is temporally consistent with the full cloud simulation of Orf et al. (2012), as the peak outflow and downflow velocities are separated by a period of approximately 1 min - 2 min (Orf et al. 2014). The transient nature of the CS descending cold mass causes a gap between peak downflow and peak outflow winds, no such gap exists in the IJ model as the source downflow wind is always active (Vermiere et al. 2011a). It is hypothesized that a density driven model will more accurately capture this gap, as all of the kinetic energy of the downburst arises from potential energy that exists because of gravity. For an impinging jet model, the peak downwards velocities occur at the nozzle exit (at the simulated base of the cloud), a phenomenon that is not observed in nature. Interestingly, the downwards and radial speeds appear to decay towards zero and the ambient background
horizontal winds of the atmosphere respectively, which is something that is consistent with what is observed in field measurements (Fujita, 1983). Further, these types of CS simulations show a peak downwards wind velocity that reaches a maximum and then also decays. For the IJ model data, the downwards velocity is defined as the jet nozzle velocity, and no decay as a function of time occurs at all (Vermiere et al. 2011a, Kim and Hangan 2007).

**Figure 3.19 - Peak down flow velocity component and peak outflow velocity for the simulation corresponding to a ramp up period of 200s**

Fig. 3.20 shows circumferentially averaged mean radial and vertical wind speeds at two radial locations in the computational domain. What is observed is a very similar trend to Fig. 3.17, where the peak down flow wind velocities are separated in time period of roughly 60 s. This gap can also be quantified not only by the temporal separation of the peak velocities, but also by the beginning of impingement, and the beginning of outflow winds observed at these two points. The beginning of impingement (defined where the outflow wind speed makes a rapid increase from the baseline ~7 m/s) can be seen to occur at approximately 100 s model time, and the time at which the outflow reaches the radial position of 1000 m (selected as this is approximately the radial position where peak outflow wind speeds are observed) occurs at approximately 220 s. This represents a period of
roughly 120 s or 2 min, which is comparable to what is observed in the full cloud simulation of Orf et al. (2014) (right). This is marked on the plot as red and blue lines.

Figure 3.20 – Circumferential peak down flow and outflow wind speeds for two radial positions for the current study, r=10m corresponding to the area directly under the CS and, r=1000m (left), and for the full cloud simulation of Orf et al. (2012) at r=0m and r=1500m (right), corresponding to the location where peak outflow and downflow wind speeds are typically observed

3.5.6 Lundgren scaling approach

Fig. 3.21 plots the maximum radial velocity observed for both the full cloud simulation of Orf et al. (2014) at r=1500 m as well as the present results at r=1100 m, normalized by the Lundgren et al. (1992) scaling velocity, against time normalized by the Lundgren et al. (1992) time parameter. The radial positions selected for both studies (r=1500 m, and r=1100 m) were chosen as they are considered to be relatively spatially equivalent. What is observed is that the Lundgren et al. (1992) time parameter does appear to match fairly consistently the two types of simulations, as the time at which the peak occurs seems to match fairly closely, with t/T₀=8 for both. However, the scaled radial velocities appear to match less closely, and do not exactly collapse. An offset of U/r/V₀=0.25 has been removed from the full cloud study data, in order to remove the more complex background wind that is present in the full cloud study as a result of adjacent downburst events not present in the simplified CS study. A similar subtraction was made (U/r/V₀=0.2) to the CS data, as the
sounding data introduced background winds. The time history for the full cloud simulation was also forced to start at 0 by removing simulation time prior to the onset of the downburst activity, which is at approximately 3000 s (50 min). This forces both plots to begin at the origin, for a more reasonable comparison. Ideally, both the temporal and wind velocity scales would collapse to the same peak, however having the temporal scale collapse is an encouraging result considering the vast differences between the two simulations. The Lundgren et al. (1992) scaling procedure for the full cloud simulations predicts a $V_0$ which is smaller than that for the CS simulations, resulting in a higher scaled radial velocity, perhaps effected by the adjacent downburst structures. For the full cloud simulation the downburst column diameter was taken to be approximately $D_0=1300$ m, and the source base height was described to be approximately $z=2500$ m above the surface (Oreskovic et al. 2015). An estimate of the density difference between the downdraft column and the ambient surrounding atmosphere was taken to be $\Delta \rho/\rho=0.0133$. Using these values, and applying the Lundgren et al. (1992) scaling equations, the dimensional scaling parameters for time and velocity were established to be approximately 70.5 s and 9.2 m/s respectively. For the CS simulations, the approximate $R_0$ was calculated to be 1373 m, somewhat wider than the full cloud simulations. Also the density difference between the CS downdraft column and the ambient air of the atmosphere was calculated to roughly be $\Delta \rho/\rho=0.095$. This resulted in a time and velocity scale to be 38.6 s and 35.6 m/s respectively. It should also be noted that the non-dimensional circumferentially averaged radial wind speeds for the current study decay most realistically back to the ambient winds for the $t_{g(t),\text{max}}=200$ simulation. The $t_{g(t),\text{max}}=\infty$ shows no appreciable decay back to ambient winds, more analogous to the impinging jet results of Vermeire et al. (2011a). The $t_{g(t),\text{max}}=300$ simulation shows some decay, although the best agreement seems to occur the sooner the CS forcing function is eliminated from the domain.

**Table 3.2 – Lundgren et al. (1992) scaling parameter estimates from the current study and the full cloud simulation of Orf et al. (2012) made in Oreskovic et al. (2015)**

<table>
<thead>
<tr>
<th></th>
<th>$\Delta \rho/\rho$</th>
<th>$R_0$ (m)</th>
<th>$T_0$ (s)</th>
<th>$V_0$ (m/s)</th>
<th>$Re$</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Present study</strong></td>
<td>0.095</td>
<td>1373</td>
<td>38.6</td>
<td>35.6</td>
<td>31199</td>
</tr>
</tbody>
</table>
Figure 3.21 - Non-dimensional maximum radial circumferentially averaged wind speed plotted against non-dimensional time using the Lundgren et al. (1992) time scale ($T_0$) and velocity scale ($V_0$)

What is encouraging is that shown in the temporal history in Fig. 3.21, the CS type simulations that involve a shorter steady state cooling period show a more realistic velocity ramp up and ramp down characteristic than simpler IJ models, and even the CS results of Vermeire et al. (2011a). The radial wind speeds ramp up, reach a short peak, and then fall eventually back to the background wind values over a period of approximately $t/T_0=8$.

The downflow and outflow column between the CS events of the present study and the full cloud simulation of Orf et al. (2012) take on noticeably different structure as is evident in Fig. 3.10 and Fig. 3.12 of the previous section. Plotted in Fig. 3.22 is the evolution of the height of the primary column of cool air for both types of simulations. Due to the spatial complexity, identifying a single primary downflow column for the full cloud simulation
results is challenging. Here, circumferentially averaging was performed, and the largest area near the centre of the event was identified, and the downdraft was tracked where a temperature difference of 4 K existed. It is observed that the leading edge of the downflow for the full cloud simulation falls to the surface faster than the CS, and also from a much higher height. It is established in Oreskovic et al. (2016) that the CS type simulations have a leading edge that forms much closer to the surface than any other type of modelling technique. The outflow evolution for the full cloud simulation also appears to rapidly move outwards, more so than the CS studies, and as is illustrated in Fig. 3.12, this is most likely caused by many adjacent events falling within the primary outflow causing an exaggerated growth.

Figure 3.22 – The temporal evolution of the downflow (left) and the outflow (right) of both the current study and the full cloud simulation of Orf et al. (2012)

3.6 Conclusions

This study made use of CM1, a large eddy simulation meteorological cloud model, to simulate the idealized CS simulation proposed in Anderson et al. (1992). A comparison was made to the full cloud simulation of Orf et al. (2012), Orf et al. (2014) and Oreskovic et al. (2015), to investigate whether an idealized CS simulation can more accurately capture the realism of a cloud model simulation, if the same atmospheric base state conditions are employed. Randomness was added into the CS forcing function in order to better represent the spatial and temporal variability that is present in the thermodynamic cooling in a natural thunderstorm cloud. A number of simulations were performed, one which investigated the
effect of the randomness addition and three others which investigated the temporal
dependence of the cooling on the downburst outflow. To the best of the author’s knowledge
this study represents the most sophisticated ‘low level’ downburst model that has been
utilized to date, with some encouraging results. The following conclusions have been
reached:

- The spatial variability around the circumference of the downburst impingement for
  a CS, which is observed in nature and full cloud simulations, is captured when the
  imposed atmospheric base state of Brown et al. (1982) is used. Peak circumferential
  wind speeds are approximately 1.3 that of the circumferential mean values,
  consistent with that (~2) observed for the full cloud simulation of Orf et al. (2012).

- The addition of randomness into the CS forcing function has little impact on the
  spatial variability of the downburst outflow, although it does result in a more
  realistic transition into the turbulent region and aids in achieving computational
  stability and a grid independent solution.

- The temporal dependence of the CS ramp up function does appear to have a
  noticeable effect on the decay of the radial wind velocities. Previous CS studies
  such as Vermeire et al. (2011a), Mason et al. (2009) and Anabor et al. (2011) and
  IJ studied such as Kim and Hangan (2007) showed an unnatural decay of wind
  velocities since the source remained ‘on’ throughout the simulation. The
  $t_{fg(t),max}=200$ simulation of this study showed best agreement to the full cloud
  simulation data since the radial winds decayed back to the ambient values similarly
  to how they did in Orf et al. (2014).

- The mean outflow wind velocities of the CS simulation appear to replicate the peak
  outflow wind velocities of the full cloud simulation. Both vertical circumferentially
  averaged radial wind speeds fall within 5 m/s of each other, and the heights to the
  peak winds are located at approximately the same height of 40 m. The later result
  being unsurprising since both simulations make use of the same numerical model
  with the same surface treatment options ($z_0=0.1$ m).

- Outflow wind speeds fall within the same range for the current study and the full
  cloud simulation, although temperature deficits are far larger in the CS simulations
  (~12 K) than that of the full cloud simulation (~4 K), supporting the concept that
the drag induced by falling precipitation is a large contributor to the magnitude of outflow wind speeds in real events.

- The peak downflow wind velocities in the region around impingement are followed by the peak outflow radial velocities at a location of $r=1$ km at approximately 1 min - 2 min (depending on how this temporal gap is measured), which is consistent with the full cloud simulation of Orf et al. (2014).

- The scaling procedure of Lundgren et al. (1992) can be applied to non-dimensionalize the temporal and spatial development of the downburst outflow between the CS simulations and the full cloud meteorological simulation. However, this scaling approach appears to be less effective at collapsing peak radial wind speeds, and future studies should investigate the limitations of this scaling method.

- Peak wind speed values of the CS study appear to somewhat overestimate the peak wind speed values observed for the full thunderstorm model. It is suggested for future studies than the cooling rate is lowered from -0.8 K/s to perhaps -0.6 K/s or lower.

The present work shows promising results that indicate that utilizing the simplified CS model in a more sophisticated atmospheric base state results in an outflow wind field that is more comparable to those from a sophisticated meteorological model. The overall goal of this work is to eventually create a computationally less expensive approach, than the full cloud simulations, that effectively captures the complexity of natural downburst events. Future work will involve performing these idealized CS simulations in more atmospheric base state conditions using sounding data from the field in conditions where downbursts have formed. Additionally, including moisture and microphysical effects may be important, as the drag-induced winds contribute a large part to the strength of the downflow from the thunderstorm (Orf et al. 2012). It is also recommended that future studies further investigate the addition of randomness into the cooling forcing function, as the meteorological cloud model shows that variation in the thermodynamic cooling is significant.
3.7 Acknowledgements

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3.8 References


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Chapter 4

4 Discussion Conclusions and Recommendations

4.1 Discussion

This chapter is a discussion of the findings that were reached in the previous two chapters (Chapters 2 and 3), as well as the significance of the current research and its impact on the numerical modelling of thunderstorm downbursts. Also, some recommendations for future studies are made, that should aid in the quality of future numerical models of the cooling source approach. Finally, an overall conclusion of the work is discussed highlighting what is accomplished in this work, and what still needs to be done for future studies.

4.1.1 The influence of the variation of CS parameters

A parameter study was performed for Chapter 2 of the current work that involved a total of 10 different Large Eddy Simulations that made use of the Cloud Model 1 (CM1) (Bryan and Fritsch 2002) with varying geometric and thermal properties of the cooling source (CS), in order to investigate the effect on the downburst outflow. It was found that an exponential relationship exists between the peak outflow radial wind speeds and the CS initial cooling rate. A linear relationship was also found to exist between peak outflow wind speeds and their locations, and the CS size. It was also found that the scaling laws presented in Lundgren et al. (1992) and Yao and Lundgren (1996), worked fairly well for the CS simulations of this study. The Lundgren et al. (1992) laws are based on liquid drop release experiments, and thus the CS more complicated spatial and temporal distribution of density made estimating the initial parameters of the scaling laws more difficult. However, when all aspects of the CS are scaled (shape, size and height above ground), a collapse in data was observed. Some comparison to other studies of the evolution of the downburst downflow and outflow were made (Lundgren et al. 1992, Alahyari et al. 1994, 1995, Yao and Lundgren 1996, Mason et al. 2009, Roberto et al. 2015). It was found that the outflow growth matches liquid drop release experiments fairly consistently after impingement, but regions before this and the growth of the downflow is not consistent. The use of scaling laws for this type of downburst model at this point can be considered encouraging, although more work needs to be done, particularly a larger set of data needs
to be collected (i.e., more simulations that investigate other parameters). Finally, a clear relationship between source parameters and the horizontal areas at 10m and 50m AGL that experience EF0 and EF1 winds were observed.

4.1.2 Cooling source in a more realistic atmosphere with random temperature perturbations

A set of 3 simulations were performed for Chapter 3 of the present work, that involved placing the idealized cooling source forcing function of Anderson et al. (1992) and Vermeire et al. (2011a) into the realistic atmospheric base state of Orf et al. (2012, 2014) which is based on the field measurements found in Brown et al. (1982). The cooling source ramp down region was modified to occur at three different times, a modification of Vermeire et al. (2011a,b) in an effort to more closely match the wind speed decay of Orf et al. (2012) and field studies such as Gunter and Shroeder (2015) and Fujita (1983, 1985). Results show that ramping down the cooling source earlier showed radial wind velocity decay more consistent with full cloud simulations. It was found that the circumferential spatial variability present in the full cloud simulation was reasonably replicated, as the overall peak circumferential velocities differed by the mean at time of peak by around 30%. This result is presumably due to the introduced wind shear of the atmospheric sounding of Brown et al. (1982). It was also determined that the peak wind velocity profiles matched in shape fairly consistently with the full cloud simulations, showing the height at which the peak wind velocity occurred in relatively similar positions (a result not entirely surprising due to the consistent computer model that was used, and similar wind magnitudes and surface roughness modelling between the two simulations). Finally, the growth of the downburst outflow appeared to be fairly consistent between the current study and the full cloud simulation of Orf et al. (2014) and Oreskovic et al. (2015), as the lag between peak outflow wind velocities and peak downflow wind velocities were consistent at around 120 s. Additionally, Lundgren et al. (1992) scaling was also applied to the time history of peak outflow wind velocity, showing a reasonable collapse of data in time.
4.1.3 Numerical considerations

It was observed in the data of Vermeire et al. (2011a,b), and confirmed in preliminary simulations of the current study, that using a 5th order advection scheme in a quarter computational domain (mirrored boundary conditions) resulted in a cleft in the temperature fields near the boundary walls. This anomaly was characterized by a premature transition into turbulence at 0° and 90° angular positions in the downburst outflow. In order to combat this issue in Chapter 2, a 6th order explicit horizontal advection scheme was selected with a careful selection of the diffusion coefficient to maintain better control of the energy at small scales in the flow. Subsequent simulations were performed on a full computational domain, with larger dimensions than that proposed in Vermeire et al. (2011a,b). This mitigated the issue to the point where the instability was not visually detectable in the downburst outflow three dimensional temperature fields.

In Chapter 3 of the current study, it was determined that despite the observed improvement by using a 6th order explicit advection scheme in a full computational domain, there still existed a minor grid influenced unnatural development of turbulence at factors of 45° in the outflow. This instability revealed itself only in the sws (surface wind swath) temperature fields. It was hypothesized that modelling the random disturbances of temperature variation (that exists in a natural thunderstorm), would result in a less axi-symmetric outflow, and reduce the turbulent anomaly. A novel way of introducing random temperature perturbation into the cooling source forcing function was developed, by randomizing the magnitude of the temperature cooling rate experiences within the cooling source region. This modification to the cooling source resulted in far more realistic looking outflows, that better matched the near surface temperature fields of Orf et al. (2012, 2014) and Oreskovic et al. (2015).

4.2 Significance and applications

This work focuses on connecting the meteorological concepts of Convective Available Potential Energy (CAPE) and real atmospheric sounding data to simplified engineering models of thunderstorm downbursts. The overall goal of this research is to be able to create more reliable engineering models, which encompass the physics that are present in real
field events and are not computationally cumbersome. Additionally, this research should be the starting point of developing reliable scaling methods (similar to those of the popular IJ model) for the CS model.

4.3 Recommendations

From the work presented in this thesis, a number of useful recommendations for future work can be made. First, the Lundgren et al. (1992) scaling method appears to be an extremely useful non-dimensional approach to examining results for even the more sophisticated temporal and spatially depended cooling source downburst model. However, it should be noted that the source density is very depended on the atmospheric pressure and temperature, and any future study should perhaps aim to further investigate how to more accurately estimate the density of the source and more importantly the resulting down flow. Also, Lundgren et al. (1992) scaling for this type of modelling does appear to be less accurate (and useful) in the regions well after the period of peak outflow wind velocities (best agreement appears to be in the early down flow of the cooling source). Any future study that considers Lundgren et al. (1992) scaling should perhaps investigate this discrepancy.

Making use of a more realistic atmospheric base state (from real field sounding data), appears to better model the outflow winds when compared to the more sophisticated full cloud simulations of Orf et al. (2012, 2014). The more complex lapse rate and wind shear included in these soundings when compared to the quiescent environment of previous studies (Mason et al. 2009, 2010, Anabor et al. 2011, Vermeire et al. 2011a,b, Oreskovic et al. 2016a), appears to incorporate more complicated environmental conditions present in the field. However, agreement between the full cloud simulation and the simplified CS with the sounding is not perfect, there still exists some obvious differences between the two models (most notably a temperature deficit of 12 K is required in the CS simulations to produce similar outflow wind speeds, while the full cloud simulations only have a temperature deficit of approximately 4 K), as a result it is clear that the microphysics that is included in the full cloud models add a further sense of realism that any idealized study cannot capture. It is recommended that future studies consider a more wide variety of base state conditions, to make comparison to both field data and other full cloud simulations.
Finally, it is recommended that for future studies, especially those that make use of Cloud Model 1 (CM1), be done using a 6th order explicit horizontal advection scheme (with careful selection of the diffusion coefficient) in a full computational domain (with no mirror symmetry boundary conditions) with added randomness to the cooling source forcing function. From this study it is observed that numerical instabilities occur as a result of using a quarter domain Cartesian (Arakawa C) grid space with lower order odd-numbered advection schemes in the solver. Perhaps future studies will investigate the effect of using different computational meshes or diffusion coefficients. Although a daunting task, a recommendation can be made to the authors of CM1 to incorporate other mesh options into the model. Simulating this cooling source in the full computational domain required enormous computational resources, perhaps in the future access to more computational resources will be available making these simulations even easier to complete.

4.4 Conclusions

The author hopes that the current studies presented in this thesis are a valuable contribution to the cooling source numerical model of the thunderstorm downburst. A number of topics have been investigated in depth, including the thermal and spatial variation that may exist in nature regarding the thermodynamic cooling of a thunderstorm cloud, and the effect of imposing a more realistic atmosphere on the simplified cooling source approach. During the length of the study, other equally important contributions were made/discovered including the necessity (in CM1) to simulate purely radial flows using a higher order even-numbered explicit advection scheme in full computational domains, and the advantages of seeding randomness into the cooling source. The author hopes that completing even more physically realistic cooling source simulations in the future will be made possible by the work done in this thesis.

4.5 References


Appendices

Appendix A - Additional figures

A.1 - Circumferential analysis of a simulated three-dimensional downburst-producing thunderstorm outflow

The following figures are taken directly from Orf et al. (2014), as it pertains to the text included in the Introduction chapter of the present work.

A detailed comparison to the full cloud thunderstorm simulation of Orf et al. (2012, 2014) is made in Chapter 3 of the current work. It is important for the reader to get a sense of scale of the full cloud type simulations of thunderstorm downbursts. The spatial complexity of such events is revealed in the following figures, in both the qualitative vector and contour plots but also in the quantitative plots which show the strong variability in wind data.

Figure A.1 - Sketch of the model domain showing dimensions and co-ordinate systems
Figure A.2 - A snapshot of the subcloud region of the thunderstorm focused on the downburst. The grey volume rendered field is rain mixing ratio and the coloured/greyscale horizontal (x,y) plane represents horizontal wind speed at \( z = 19 \text{m AGL} \) (above ground level) and 3606 s into the simulation.
Figure A.3 - Snapshot of horizontal wind velocity vector magnitudes in (x,y) plane at $z = 31.5$ m and 3606 s into the simulation.
Figure A.4 - Vector field in vertical plane through the centre of the main downburst region aligned (a) in the East-West direction (x,z) and (b) in the North-South direction (y,z)
Figure A.5 - Circumferentially-averaged velocity vectors in vertical (r,z) plane and velocity magnitude contours at time (a) $t = 3588$ s, (b) $t = 3606$ s and (c) $t = 3628$ s
Figure A.6 - Variation with radial distance from impingement of height to maximum circumferentially-averaged near-ground radial wind velocity, for the three different times.
Figure A.7 - Variation of circumferentially-averaged radial velocity (at height of maximum velocity) with radial position, together with maximum value around that circumference (a) $t = 3588$ s, (b) $t = 3606$ s and (c) $t = 3628$ s
Figure A.8 - Variation of normalized circumferentially-averaged radial velocity (at height of maximum velocity) with normalized radial position

Figure A.9 - Normalized outflow radial velocity versus normalized height for observed and cooling source CFD simulated events
Figure A.10 - Time histories of downwards vertical velocities (-Uz) at centre of column (r = 0) at different heights (blue/green or greyscale lines) and horizontal outflow velocities at two peak velocity locations A (orange line/grey line with symbol) and B (red line/black line with symbol)
A.2 - Temporal Evolution of a Simulated Downburst-Producing Thunderstorm Outflow

The following figures are taken directly from Oreskovic et al. (2015), as it pertains to the text included in the Introduction chapter of the present work.

Figure A.11 - Circumferentially-averaged velocity vectors in the vertical plane (r,z) and velocity magnitude contours, and snapshots of the subcloud region of the thunderstorm focused on the downburst. The grey volume in the frame is the rain mixing ratio, and the coloured horizontal (x,y) plane represents the horizontal wind speeds 19 m above ground level at time: (a) $t = 3472\text{ s}$ (b) $t = 3588\text{ s}$ (c) $t = 3606\text{ s}$ (d) $t = 3628\text{ s}$ (e) $t = 3744\text{ s}$. 
Figure A.12 - Variation of vertical wind speed with time within the downdraft vertical “cylindrical” column, spatially averaged over that column cross-sectional area (of diameter = 1300 m)

Figure A.13 - Variation with time of the maximum circumferentially-averaged radial wind speed at the radial location corresponding to its maximum value (r = 1500 m), together with the peak value around the circumference at each time.
Figure A.14 - Contours of potential temperature perturbation and vertical wind velocity component (with vectors) in vertical East-West (x-direction) plane through downburst centre at (a) $t = 3400$, (b) 3472, (c) 3588, (d) 3606, (e) 3626 and (f) 3744 s.
References


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