Momentum Transport by Organised Deep Convection

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Abstract

Deep convection is an important process that influences the vertical redistribution of heat, moisture, and momentum. Convective momentum transport (CMT) is composed of multiscale dynamical processes, including convective-scale CMT associated with updrafts and downdrafts, and mesoscale CMT associated with the quasi-steady circulation from organised systems. This multiscale nature of CMT is an active area of research especially regarding the influence of momentum transport on the mesoscale organisation of convective systems and vice-versa; these mesoscale processes and their associated CMT are the focus of this study.

Firstly, the CMT budgets of simulated idealised mesoscale convective systems are examined, and their sensitivity to horizontal resolution, domain size, and the boundary conditions, investigated. This is followed by an examination of a case study during the Tropical Warm Pool - International Cloud Experiment (TWP-ICE). The effect of horizontal resolution, time evolution, and cloud regime on the CMT were evaluated to determine the momentum profiles on both scales. These results are then compared to observational data derived from Doppler radar.

The idealised simulations reveal that for relatively large domains, horizontal gradient terms are still important, including the mesoscale pressure gradients; such terms are neglected in all CMT parameterisations. Like most convection parameterisations, current CMT parameterisations only represent convective-scale processes through relatively simple plume models. Thus, they do not properly represent the transports associated with organised systems, even though the tilted circulation associated with organised convection is a fundamental aspect of CMT. The horizontal pressure gradient and the sign of the momentum flux change sign as the system develops. Small domain calculations, which have become common for radiative-convective equilibrium experiments, are shown to suppress organisation through artificially large compensating subsidence and hence provide unrealistic representations of the processes involved. Finally, examination of the cross-updraft/downdraft pressure gradients demonstrates significant errors in their rep-
representation in current schemes.

The convective and mesoscale momentum transport profiles from the real cases reveal how these scales change at various stages of their evolution. Grid spacing > 4 km does not capture the convective scale CMT and the coarsest resolution (∼ 30 km) produces the wrong sign of CMT. Mesoscale organisation is highly influential when evaluating the CMT, as these tilted circulations are responsible for much of the momentum transport. These findings indicate that rather than just parameterising the convective or small-scale dynamics, these schemes must take into account the larger, mesoscale processes.

The observational data reveals differences in the dynamical structures to those identified using the model data. Analysing momentum transport profiles identifies a downshear-tilted system, suggesting that the model is not representing the full diversity of systems that actually occur. Another simulation using a different microphysics scheme and analysis of the model’s thermodynamics shows that both microphysics schemes create overly strong cold pools which preferentially generate upshear-tilted systems. The usefulness of this radar dataset has also allowed insights into the variety of systems and may be used to reproduce these systems in models.
Declaration

This is to certify that:

(i) the thesis comprises only my original work towards the PhD except where indicated;
(ii) due acknowledgement has been made in the text to all other materials used;
and
(iii) the thesis is less than 100,000 words in length, exclusive of tables, maps, bibliographies, appendices and footnotes.

________________________________________
Rachel L. Badlan
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“I love deadlines. I love the whooshing noise they make as they go by.” (The Salmon of Doubt, 2002)

So long, and thanks for all the fish.....


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1 Introduction

1.1 Opening Remarks

Moist convection is an immensely important process that is responsible for the redistribution of heat, moisture and momentum across the globe; of which momentum transport is the least understood. The vertical transport of horizontal momentum from convection has been shown by LeMone (1983) and LeMone et al. (1984) to be considerable across the globe. Convective momentum transport (CMT) is known to have a significant effect on the both the zonal and meridional mean flow (Zhang and McFarlane (1995)) which can contribute to the Hadley circulation (Schneider (1975), Helfand (1979)) and also has a similar magnitude to the rest of the terms comprising the atmospheric angular momentum budget (Houze (1973)). Momentum transport due to organised convection also affects the precipitation, the speed of its propagation (Mahoney et al. (2009)), its lifespan, and the regime of the convective system and so plays an important role in convective development.

Moist convection can organise into shallow or deep cloud systems in most regions across the globe and are strongly associated with the global atmosphere circulation; there are mid-latitude baroclinic systems, subtropical convective complexes and tropical systems. This connection between the convective organisation and the large-scale circulation suggests that the convective organisation may be represented by resolvable variables (i.e., parameterised (Moncrieff (2010))). Momentum transport is vital to the organisation of such systems and investigations have confirmed the importance of representing CMT in simulations, which typically results in improved forecasting of surface winds, precipitation, etc. (Richter and Rasch (2008)). Large-scale circulations are more accurately resolved when parameterised momentum transport is included in Global Circulation Models (GCMs) or regional forecasting. The momentum budgets of deep convection were also studied in the 1980’s by researchers such as Shapiro and Stevens (1980) and
Sui and Yanai (1986), and the importance of the vertical transport of horizontal momentum highlighted. On a more local level, the MCS momentum field has implications for potentially harmful surface winds which can affect the motion of the storm (Mahoney et al. 2009) making accurate forecasting of such systems difficult. The nature of these transports depends on the individual flow regime of the system and its evolution. Several factors such as the pressure gradients across updrafts and downdrafts, the strength of the cold pool, and the environmental wind shear can all affect the type of convective system that is generated. These organised convective systems also exert an influence over the environmental shear due to the momentum transport, in turn influencing the larger-scale flow and potentially feeding back on the convection itself, thereby influencing its convective regime (Lane and Moncrieff (2010)).

However, CMT is not comprised solely of the relatively simple transports as has been assumed to-date. It is composed of multiscale dynamical processes, including convective-scale momentum transport, which are associated with updrafts and downdrafts, and mesoscale CMT associated with the quasi-steady circulation associated with organised systems. Moncrieff (1992) argues for the separation of the transports at the convective-scale and mesoscale as these have different characteristics with the mesoscale momentum transport inherently linked to the tilt of the mesoscale convective system. It follows that these two scales need to be investigated separately, to better understand the contributions from mesoscale and convective-scale processes by investigating the momentum transports within convective systems at both those scales. The processes found in mesoscale convective systems (MCS) range from the microscale (turbulence and mixing) through the mesoscale (viz. the organised circulations) to the global scale, and it is this multiscale nature of momentum transport that requires further research.

Current CMT parameterisations only represent simple momentum transport associated with convective-scale processes through relatively basic plume models and therefore they do not properly represent the momentum transport associated with mesoscale organisation of the MCS. As the mesoscale component of CMT is specifically related to the tilted circulations associated with organised convection, so the ability of the parameterisation schemes to represent CMT properly, is limited. Therefore the overarching goal of this investigation is to examine the relative contributions of the convective-scale and mesoscale transports in organised convective systems, as this will reveal how well mesoscale models represent the mesoscale processes.
1.2. Aim and Scope

Momentum transport has not been researched as fully as the thermodynamic variables of heat and moisture, and both its effects on the convective systems and how the system affects the development of the CMT, are still questioned. CMT parameterisation schemes have only recently been incorporated routinely into convection schemes in numerical models. However, improvement in both our knowledge of how momentum transport operates at different scales and how to incorporate this information into parameterisation schemes is needed. Although inclusion of CMT into parameterisation schemes has improved outcomes (i.e., simulated variables and precipitation match the observational data more closely), there is still a need to obtain more accurate representation of convective systems, especially for deep convection within the maritime continent (R. Stratton, pers. comm.).

1.2 Aim and Scope

The main aim is to investigate how momentum transport is related to the mesoscale structures and circulations found in deep convection. This is to be achieved initially by using idealised simulations of a three-dimensional (3D) squall line to understand the composition of the momentum budget and its sensitivity to various spatial and temporal numerical model resolutions. This is then extended to a real-case simulation. Finally a comparison with radar observations is conducted.

The following are my intended research questions:

**A.** Which terms are most influential to the momentum transport profile and do these findings affect the way that momentum transport is represented in a parameterisation scheme?

**B.** What are the relative contributions of mesoscale and convective transports in organised convective systems?

**C.** Does observational data from radar reveal differences in convective momentum transport characteristics in real systems to those identified in models?
This research project is divided into three distinct phases, as detailed below.

The first phase specifically addresses Question A and uses an idealised simulation which provides a simple model of convection. The main aim is to examine the roles of different components of the momentum budget and their sensitivity to aspects of the simulation. This reveals whether the mesoscale organisation of the system contributes to the momentum budget. The hypothesis for Question A is that the mesoscale transports are influential.

The second phase addresses Question B and will be applied to a real case from Tropical Warm Pool - International Cloud Experiment (TWP-ICE) over Darwin, Australia. A scale-separation technique is used to identify the convective and mesoscale transports associated with different types of MCS and their sensitivity to resolution, which also reveals how different models simulate convection. It is hypothesised that the convective transports will be underrepresented in the coarsest domain simulations and that this resolution will give the wrong sign for the mesoscale CMT - following the results of Moncrieff and Liu (2006) - see Section 2.2.1 for further discussion).

The third stage will answer Question C using a dataset derived from radar observations during the same period and location as described in stage two. This is exciting as it will be a challenge to estimate $u'w'$ from this new radar dataset. These results will allow a direct comparison to be made between the systems from the model and the radar, and will reveal if the fluxes are of a similar magnitude. The hypothesis is that the systems will be similar in both, with potentially larger fluxes from the model due to stronger vertical winds.

1.3 Significance of the Study

The significance of this study is that it takes a multiscalar approach to the problem of how momentum transport behaves. Mesoscale processes, to this point, have been neglected in this representation of momentum and this study examines whether the different transports act downgradient or countergradient. This separation of scales has not been applied to a real case before, and this should reveal if those two scales act in a similar manner to each other. Another point of signif-
icance is the application of a derived 3D wind dataset from observational data to the analysis of momentum transport. Until now, observations have been spatially incomplete and this has restricted analysis of convective dynamics. This dataset should enable a comprehensive comparison of model and real MCS dynamics and identify and differences between the systems reproduced by the model and those observed by the Doppler radar.

1.4 Overview

In Chapter 2, the background literature is discussed and an overview of the relevant research to date is described in order; from convection and types of MCS, followed by a summary of research thus far into momentum transport and a brief description of how it is parameterised - this sets the current study in context. Chapter 3 comprises a detailed description of the terms studied in the momentum budget and includes a full decomposition of the equations of motion.

Chapters 4, 5 and 6 describe the results of the three main areas of study. In Chapter 4, the momentum budget is examined using idealised model runs as well as a discussion on how these results affect our understanding of current parameterisation schemes. The real case of TWP-ICE is analysed using scale-separation of CMT to identify contributions from mesoscale and convective-scale in Chapter 6. Chapter 6 uses a unique observational dataset derived from Doppler radar to examine the momentum transport and a comparison of the systems identified with those from the previous chapters. The final chapter (Chapter 7) summarises the main results of this research, and discusses their implications for further work.
1. Introduction
2 Background

This chapter describes the background information relevant to this study and serves to expand upon the information relating to the research, by providing details about both convective systems and momentum transport. The research project, its design, method and results are described in later chapters.

The aim of this chapter is to set the study in the appropriate context, in order to enable a deeper understanding of the subject of momentum transport due to deep or organised convection (CMT). This section considers the processes that give rise to convection and determines the characteristics of the momentum transport. It addresses whether CMT acts in a upgradient or downgradient manner, and how momentum transport can feedback and affect the environmental flow. This understanding of the relationships between momentum transport and deep, organised convection delivers an appreciation of the importance of CMT and describes why it is vital in the accurate representation of organised convection and its effects on the environment.

To gain an understanding of this topic, it is necessary to discuss and evaluate some processes related to convection, and the types of environmental variables such as wind shear, which set up the conditions for convective systems to grow and develop. Organised convection is responsible for the redistribution of thermodynamic, dynamic and chemical quantities throughout the globe, however this research focuses directly on the dynamic variable, momentum. This chapter then examines more specifically the types of MCS that can evolve in areas of deep convection and the structures that are associated with each type of convective regime. How these systems are directly linked to the nature of the momentum transport is then described with an explanation of the relationship between the tilt of the system and CMT.
2.1 Convection

This section introduces the background to the research and summarises the relevant research to date by introducing theory describing how convective systems form, some of their structural components and underlying dynamics, concluding with an overview of the main types of linear MCS that occur in both the mid-latitudes and the tropical regions of the globe.

Convection is important for the global circulation as it vertically transports the thermodynamic quantities of heat and moisture as well as those of a dynamic nature such as momentum, along with various chemicals vertically throughout the globe (Tung and Yanai (2002a), Moncrieff (2010)). Convective systems demonstrate considerable diversity in the way they form, their structural characteristics and, as a consequence, also determine the weather on a local scale. Storms can bring severe weather with associated potentially dangerous conditions and MCSs, which are the largest of the convective storms (Houze et al. (1990), Houze (2004), Bluestein (2013)) often produce severe weather and can lead to flash flooding, damaging hail and other devastating outcomes. MCSs have a number of physical processes that occur on various temporal and spatial scales which can affect the motion of these systems and present a challenge to forecasting and numerical weather prediction (NWP) (Mahoney et al. (2009)).

Clouds typically form when the air becomes saturated as it cools adiabatically normally via a lifting process to the lifting condensation level (LCL). Mechanisms of this lifting can be dynamic, and processes for this include one or a combination of: quasi-geostrophic lift, uplift along a mesoscale boundary or forced up a slope in the terrain (Bluestein (2013)). Alternatively, air can be heated resulting in a superadiabatic lapse rate with dry convection mixing moist air vertically to the saturation level; the convective condensation level. In order to be able to freely convect, the air mass must be displaced upward beyond its level of free convection (LFC), where it becomes positively buoyant or unstable (Emmanuel (1994)). Once a parcel of warm, moist, air is lifted above the environmental LFC, it is eventually decelerated beyond a level of neutral buoyancy by entraining ambient air as it rises (Riehl and Malkus (1958)). Convection can be separated into two groups: ordinary convection and organised. Ordinary convection (Fig. 2.1 (a)) does not grow upscale and has much in common with a thermal plume. An example of an ordinary convective cloud is *cumulus humilus* which is non-precipitating. This type of convection is small-scale and has local effects only; it does not prop-
2.1. Convection

agate (i.e., is stationary, being only advected by the mean wind), is short-lived and occurs in the absence of wind shear (Moncrieff and Klinker (1997), Moncrieff (2004), Moncrieff and Liu (2006)). It is quite different to organised convection which is described in more detail in the following section. The differences in these convective regimes have implications for the parameterisation for convection as will be discussed later in Section 2.2.3

2.1.1 Organised Convection

As convection develops, the systems can become organised, evolving into mesoscale systems with coherent flows. Some of the characteristics of organised convection include coherent circulations, propagation, and interaction with environmental flow. They live longer than the lifecycle of a convective cell, and are strongly influenced by vertical wind shear. This longevity and ability to propagate demonstrates that organised convection has an affect on the atmosphere (and ocean) that spans a range of scales, such as affecting the environmental shear and the generation of gravity waves.

This research is concerned with organised convection and specifically analyses squall-lines which are found in the mid-latitudes and the tropics. These convective systems have tilted structures and are long-lived, deeper, and more intense compared to ordinary convection which typically occurs in a low-shear environment, has flow vertically up and down, is short-lived and has no defined circulation or organised flow. These differences are summarised in Fig. 2.1. As Moncrieff and Liu (2006) state, the organised convection found in MCSs is primarily a result of the interaction of the dynamics due to the environmental wind shear, the latent heating and the evaporative cooling. The systems studied in this research are also mesoscale systems and the momentum transports are inherently linked to the type of MCS in which they are found.

2.1.2 Mesoscale Convective Systems

A general definition of an MCS is a cumulonimbus cloud with a contiguous area of precipitation of approximately 100 km or greater in one or more directions with Houze (1993) extending this to include mesoscale circulations which are induced by the convective and stratiform areas of the cloud system. They are responsi-
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Fig. 2.1: Distinction between (a) ordinary convection and (b) organised convection with attention to aspects pertinent to convective parameterisation. From Moncrieff and Liu (2006).

MCSs are attributable to and embedded within large-scale wave motions (Carbone et al. (2002), Jakob and Tselioudis (2003)). Intraseasonal (e.g. the Madden-Julian Oscillation (MJO)/Intraseasonal oscillation (ISO)) and interannual climate variations such as El Niño are also associated with the formation of MCSs. Nakazawa (1988) describes synoptic-scale superclusters within the MJO, whilst Chen et al. (1996) note the occurrence of cloud clusters in particular regions due to the influence of the ISO. Convectively couple equatorial waves can also influence the development and lifecycle of MCSs (Haertel and Kiladis (2004)).

Most of current knowledge of MCSs began around 40 years ago by field campaigns and projects, as well as by modelling MCSs using various numerical weather models, and some of these studies identified large MCSs in the tropics over the oceans (Zipser (1969), LeMone et al. (1984)). They have a range of structures in both the cloud form and the precipitation, however they all possess some common features.

A convective system’s precipitation may be divided into two distinct regions: convective and stratiform areas with different characteristics as shown by precip-
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itation from radar echoes (i.e., McAnelly and Cotton (1989)). This approach has been used by Gallus and Johnson (1992), Yang and Houze (1996), Mechem et al. (2006) and others, where these regions have been identified using some criteria (i.e., vertical wind speed, microphysical separation, or radar characteristics to define them). MCSs are composed of distinct regions of convective and stratiform precipitation which have fundamentally different properties both in terms of kinematics (Houze (1982), Houze et al. (1989)) and microphysics (Houze et al. (1989), Braun and Houze (1994a), Braun and Houze (1994b)). These systems develop mesoscale and convective circulations which influence the evolution of the system and its duration, and have defining structures reflecting the dynamics within.

Generally the convective region is characterised by intense, vertical cores and cover a smaller region in contrast to the stratiform areas. These stratiform regions which although larger, produce lighter precipitation and weaker vertical winds in general (Parker and Johnson (2000)). The stratiform component of the cloud is formed by the dissipating convective cells and mesoscale ascent (Yuter and Houze (1995a)) and also incorporates areas of cooling in the lower troposphere due to melting and evaporation. The two regions have different heating profiles which affects the behaviour of the large-scale flow (Mapes (1993), Mapes and Houze (1995)). These heating profiles are due to the dynamics of the MCS; the stratiform region has net heating above from the updrafts and the release of latent heat from condensation with a region of net descent below the cloud base due to cooling from the evaporation and melting of the precipitation from the stratiform region above, resulting in net cooling in the lower levels. Convection however, shows net heating throughout the troposphere as the convective updrafts and resultant heating have a greater affect than the small amount of cooling from the convective downdrafts (Houze (2004)).

One way of considering updrafts is as a layer which is lifted due to an assumed drop in the hydrostatic air pressure in the mid-levels, occurring across the updrafts. This method requires strong low-level shear and produces mesoscale circulations such as Moncrieff’s triple branch model, which is used throughout this project to interpret the mesoscale dynamics of each system. These flows are represented in Fig. 2.2 with the arrows indicating the slantwise layer overturning in the vertical plane described more fully in Moncrieff’s nonlinear theory of steady convective overturning (Moncrieff (1992), Moncrieff (2010)). This is driven by descent due to evaporative cooling. The triple branch model has the upward jump-like branch (indicated by the ascending front-to-rear flow), an overturning
branch travelling upwards (on the upper right of the figure) and a descending overturning branch or rear inflow jet (RIJ), described by various authors such as Kingsmill and Houze (1999) and Skamarock et al. (1994).

The slantwise layer overturning is fundamentally influenced by the convective available potential energy (CAPE) as well as the kinetic energy and the work done by the horizontal pressure gradient ($\Delta p/\rho$) (Moncrieff (2010)). This mechanism was initially formulated to account for the organisation of MCSs of $\sim 100$ km horizontal length but could also account for systems of $\sim 1000$ km i.e. superclusters (Moncrieff and Klinker (1997)).

The rear-to-front downward flow is accelerated by the pressure gradient force. As the strongest winds occur in this location, it is often termed the RIJ (Smull and Houze (1987)) and along with the strength of the downdrafts and negative buoyancy, can alter the downward transport of momentum and wind fields caused by the storm (Mahoney et al. 2009). This typical structure of trailing-stratiform precipitation can be explained by a tilted dipole heating structure in the system, with a narrow tilted band of heating preceding a wider area of cooling (Pandya and Durran (1996)).

![Fig. 2.2:](image)

**Fig. 2.2:** Conceptual model of the kinematic, microphysical, and radar echo structure of a convective line with trailing-stratiform precipitation perpendicular to the convective line. Areas of strong reflectivity are indicated by shaded areas. H and L indicate areas of positive and negative pressure perturbations, respectively. Dashed-lines indicate trajectories of ice particles passing through the melting layer. From Houze (2004).

Another key component of multicell storms are the various convective cells (represented in Fig. 2.2 by shaded regions) which are located at the front of most typical convective systems and last $\sim 30$ min, even though the storm may last for
many hours. These cells are composed of small-scale dynamical flows, and mass moves across these updrafts and downdrafts. These cells are embedded within the mesoscale flow, yet they act at the short temporal and spatial scale characteristic of updrafts i.e. they are short-lived (they regenerate with a periodicity of around 10 minutes (Lin et al. (1998))) and occupy an area from a few hundred metres to a few kilometres. At any time, the individual convective cells are at different stages of their life-cycles, which consist of the three phases: developing (forming ahead of the convective line), mature (becoming part of the convective line) and dissipating (when they are incorporated into the trailing stratiform precipitation). As the environmental shear is stronger, they do not interact with each other due to this vertical wind profile (Cotton and Anthes (1989)). These cells may be generated along the gust front as the cold pool lifts this conditionally unstable surface air, which flows up and over the cold pool and the downdraft produced by the previous cell (Thorpe and Miller (1978), Wilhelmson and Chen (1982)).

2.1.3 Types of MCS

The three distinct types of linear MCS identified and described by Parker and Johnson (2000) are the leading convection/trailing stratiform (TS), leading stratiform/trailing convection (LS) and parallel stratiform (PS) (Fig. 2.3). The most common of these is the TS system. Parker and Johnson (2004) found that this archetype accounted for approximately 60% of the systems in the central United States studied over a 2 month period, with the LS and PS archetypes contributing approximately 20% each to the total occurrences (Parker and Johnson (2000)). The main system that is identified from the model data in this research project is the front-fed TS (FFTS) archetype (an example of which is shown in Fig. 2.2), where the system is initiated and maintained by front-to-rear storm-relative inflow (corresponding to the jump-like structure discussed earlier). This system is described in detail by a number of significant papers, notably Newton (1950), Rutledge et al. (1988), Houze et al. (1989) and Houze (2004). However, as a summary, the main flow structures are the front-to-rear flow that ascends in the region of the convective updraft, weakly overturning at the front, with most of the circulation exiting the rear of the system. Melting, evaporation, and sublimation of precipitation enters the RIJ and this cooling is the reason for the descent of air from rear-to-front (Parker and Johnson (2004)).
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![Fig. 2.3: The three main types of linear MCSs. a) trailing stratiform with leading convection, b) leading stratiform and c) parallel stratiform. The small dark blue ovals represent the convective cells/updrafts, the mid blue oval represents the convective line and the light blue is the stratiform area of the cloud. The arrows indicate the direction of propagation.](image)

Other types of linear MCS are the front-fed LS (FFLS) and rear-fed LS (RFLS), which are maintained by front-to-rear and rear-to-front storm-relative flows respectively (Parker and Johnson (2004)). The RFLS is in most respects a mirror image of the FFLS and Pettet and Johnson (2003) reinforces this premise stating that, if reversed, the mesoscale circulations of the FFLS are kinematically almost the same as that of the RFLS. However, the caveat is that there are also differences between the two systems as well, such as the inflow jet reversing in the TS, but not with the LS example, which also has an elevated rear inflow, whereas the TS has a boundary layer frontal inflow. However, both these systems share the jump updraft as the dominant dynamic feature. All the systems have these typical characteristics when fully-developed (i.e. when mature) and change their dynamical structure as they move through the various stages of evolution.

Rotunno et al. (1988) demonstrated the change in tilt of a squall-line as it transitions from its initial stages to a mature system, and this implies that the momentum transport can change from upgradient to downgradient throughout the lifecycle. Upgradient (or ‘countergradient’) transport occurs when the sign of the momentum transport is the same as that of the environmental shear and, as it acts in the same direction, it can reinforce and strengthen the wind shear, thereby increasing the longevity of the MCS. Conversely, downgradient transport occurs when the momentum sign is of the opposite sign to that of the low-level shear and works against this shear, weakening the tendency and potentially reducing the system’s lifetime. Downgradient transport is the equivalent of diffusion, whereas
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countergradient can be thought of anti-diffusion.

Weisman and Rotunno (2004) further provided an explanation for the longevity and propagation of an MCS via cold pool dynamics. The evolution is comprised of three stages, which can be expressed in terms of the strength of the cold pool and the magnitude of the low-level ambient wind shear. The interaction between the wind shear and the cold pool is important in producing the initial lift which leads to the regeneration of convective cells, creating organised convection, and this has implications for the longevity of the system through its effect on momentum transport.

The low-level positive shear ($\Delta U$) in the initial stages (Fig. 2.4(a)) creates a downshear tilt to the system as the strength/speed of the cold pool (denoted by $C$) has not sufficiently developed. In this case, the positive shear is a positive horizontal vorticity, shown by the clockwise arrow to the right of the system ($\frac{D\eta}{Dt}$, where $\eta$ is the vorticity in the $y$-direction). As the cold pool forms, its positive temperature gradient develops negative vorticity due to the buoyancy gradient ($\frac{\pa b}{\pa x}$, where $b$ is the buoyancy and the vorticity is related to the buoyancy gradient by $\frac{D\eta}{Dt} = -\frac{\pa b}{\pa x}$). This buoyancy-driven vorticity balances the positive shear vorticity. This relationship between the buoyancy and the wind shear can be expressed as $-f\frac{\pa b}{\pa x} dt \sim \frac{\pa U}{\pa z}$, and the vorticity that is generated by the developing cold pool creates deeper uplift by 'pulling' the system upright (Fig. 2.4(b)). The vorticities that develop due to the temperature/buoyancy gradient are located on the cloud boundaries and at each end of the cold pool beneath the cloud.

Eventually the strength of the cold pool can dominate (i.e., $C > \Delta U$) and the system develops an upshear tilt, with a RIJ forming (Fig. 2.4(c)). This process is important to this research, because the tilt of the structure and its evolution affects both the pressure gradients across the system (i.e., mesoscale) and the sign of the momentum transport. The mesolow is also important (see results in Chapter 4). The reason for the diversity in the range of types of MCS, is therefore due to the different shear profiles, and also the thermodynamic profile of the atmosphere, as this affects the contributions from the two opposing vorticities. This work seeks to identify any such changes by examining the transports at different stages in the lifetime of organised convection.
Fig. 2.4: Three stages in the evolution of a convective system. From Weisman and Rotunno 2004.
2.2 Momentum Transport

Moncrieff (1992) developed an archetypal model describing how the sign of momentum transport is inherently linked to the tilt of convective systems, which provides a method of analysing the evolution of a MCS via its momentum transport profile. Figure 2.5 presents two eddies; the right hand eddy is tilted in the positive direction, the other, in the negative direction. The upwards component of the left hand eddy has $\rho u' < 0$ and $\rho w' > 0$, resulting in $\rho u'w' < 0$ (the same applies to the downward component) with the horizontal and vertical velocity perturbations being anti-correlated; therefore this tilt has a negative sign. When the $u'$ and $w'$ are positively correlated the eddy has a positive tilt, and $\rho u'w' > 0$, as in the right hand image. For a positive shear environment, this figure would represent upshear (downgradient) and downshear (upgradient) transports respectively. Momentum transport therefore modifies the shear which can enhance or suppress the organisation, thereby affecting the lifecycle of the system.

The multiscale nature of atmospheric processes serves to complicate the analysis of the dynamics. As shown in Fig. 2.6, the weather and climate is composed of processes which act at different temporal and spatial scales, such as momentum transport which is composed of mesoscale momentum transports, convective-scale momentum transports and, gravity wave momentum transports. This study is concerned with convective-scale, which includes turbulence and mixing and mesoscale, which are coherent circulations.

Mesoscale momentum transport (organised convection) can be counter-gradient or downgradient (Moncrieff (1981)), whilst convective-scale momentum transport is predominantly downgradient. Therefore it is the feedback from the mesoscale momentum transport that can produce stronger, longer-lived systems. It may also produce shear that is too strong, such that $\Delta U \gg C$, which can dissipate the system. As a consequence of these two scales which act in a different manner to each other (regarding CMT), the aim of this project is to separate the momentum transports thereby evaluating their relative contributions to the total momentum tendency using data from a real case.

Before discussing the importance of CMT, it is necessary to clarify what is meant by the term. Majda and Stechmann (2009) defines CMT as the “conversion of moist convective available potential energy to horizontal kinetic energy in the flow field”. It is an important distinction to make that CMT refers to the total transport due to a convective system - not only the momentum transport
at the *convective* scale. It has long been known that CMT has an influence on the environmental flow (Tung and Yanai (2002b)).

### 2.2.1 Influence of CMT

Momentum transport may affect the development of a convective system as it can change from downgradient to countergradient during its evolution and may thereby affect its development. Studies have demonstrated its influence on the motion (Mahoney et al. 2009), surface winds (Mechem et al. (2006), Mahoney and Lackmann (2011)), precipitation (Yuter and Houze (1995a), Yuter and Houze (1995b)), and the longevity of the system through enhanced shear (Moncrieff (1981)) as the convective system evolves. The importance of shear is also highlighted by Grubišić and Moncrieff (2000), with Lane and Moncrieff (2010) demonstrating how it affects the convective regime.

Studies across the maritime continent during TOGA-COARE by Moncrieff and Klinker (1997), and Tung and Yanai (2002b) have shown an increase in the westerly surface winds from the downward transport of westerly momentum, increasing the low-level convergence in the Madden-Julian Oscillation (MJO). Other work by Helfand (1979), Zhang and McFarlane (1995) and Gregory et al. (1997)
2.2. Momentum Transport

Fig. 2.6: The main scales referred to in the atmosphere and some of associated systems. The gridsize of two model types is suggested. Adapted from Cotton et al. (2011).

found an intensification in the upper branch of the Hadley circulation when CMT was included in the parameterisation. Richter and Rasch (2008) also found that inclusion of CMT improves representation of the Hadley circulation due to changes in the Coriolis torque, while Moncrieff (2004) demonstrated notable vertical fluxes of zonal momentum in the MJO.

Various observational studies by LeMone (1983), LeMone et al. (1984), and Gao et al. (1990) investigated the effects of the momentum budget on the large-scale mean flow and found that the greatest contribution to the overall momentum budget was from cross-line momentum generation (where the 'line' refers to the convective line or squall line). LeMone (1983) demonstrated that momentum fluxes normal to the line acted to increase the environmental wind shear, and the pressure gradients accelerated the updrafts and downdraffs. Observations by Hogan et al. (2008) and modelling of organised convection by Mechem et al. (2006) indicate that in certain circumstances, the convective organisation can reinforce and strengthen the wind shear (Stratton et al. (2009)). It is also suggested that momentum transport is dependent on the organisation with downgradient transport perpendicular to the convective line and countergradient possibly occurring, normal to the line (LeMone (1983), Wu and Yanai (1994), Tung and
Yanai (2002a)). Modelling of an MCS with a cloud resolving model (CRM) by Gray (2000) identified that momentum transport by mesoscale downdrafts in precipitating regions of the cloud has a similar contribution as that from convective updrafts and is countergradient. Zhang and Wu (2003) used a 2D version of the Clark-Hall anelastic cloud model (Clark et al. (1996)) and also found that westerly wind bursts transport momentum in a countergradient fashion. Therefore these studies suggest that CMT can be dependent on the convective regime.

2.2.2 Importance of CMT in mesoscale models

The importance of CMT in mesoscale models has been the focus of various studies. Moncrieff and Liu (2006) using a regional-scale model showed the effect of horizontal resolution on CMT. They ran a 10 km and a 30 km simulation using a convective parameterisation scheme that did not parameterise CMT and compared it to a 3 km simulation that explicitly resolved convection, and hence, CMT. The results found that the 10 km grid simulation (Fig. 2.7(b)) compared favourably to the 3 km simulation (Fig. 2.7(a)), however not only did the 30 km (Fig. 2.7(c)) simulation show weaker momentum transport, it also has the wrong sign. This system was also found to propagate too slowly with unrealistic mesoscale momentum transport (MMT). This work provides a motivation for part of this study, as the authors argued that a 10 km resolution model should reproduce mesoscale transports, and this resolution is used in Chapter 5 of this research as part of the real case study. Thus, we test these assertions here. Lackmann et al. (2011) found that simulations that included CMT parameterisation yielded results that had a closer agreement in the zonal wind with explicitly resolved simulations, than those that neglected CMT. This leads to the need for further research into the effect of momentum transport in mesoscale simulations.

The size and strength of the convective and stratiform areas of the convective system can also have a profound impact on the momentum transports. Yang and Houze (1996) found that the momentum budgets of these regions produce different tendencies depending on the structure of squall lines, so storms with weak stratiform precipitation would produce fundamentally different effects on the large-scale tendency than one with stronger stratiform precipitation. Figure 2.8 shows that the convective regions dominate at the lower levels, both contribute at the mid-levels and stratiform has the biggest effect at the higher levels.
2.2. Momentum Transport

Fig. 2.7: Zonal momentum transport ($\rho u'w'$): (a) 3 km grid-spacing explicit simulation, (b) 10 km grid-spacing simulation applying convective parameterisation, and (c) 30 km grid-spacing simulation applying convective parameterisation. Contour interval is 2 kg m$^{-1}$ s$^{-2}$. From Moncrieff and Liu (2006).
2. Background

**Fig. 2.8:** (a) Block diagram showing the balance of area-weighed momentum budget terms (VMF and VEF - vertical flux convergence by mean flow and standing eddies, HMF - horizontal mean-flow flux convergence and PGF - pressure gradient force). over 4 subregions of a large-scale area during the mature stage of a squall line. (b) as for (a) but net momentum tendency. From Yang and Houze (1996).

Further numerical studies and observational work have indicated that the horizontal pressure gradient force across a convective system may modify the momentum flux in both tropical and mid-latitude squall lines. Wu and Yanai (1994) developed a parameterisation scheme which, when evaluated by Han and Pan (2006) reduced the forecast track error of hurricane prediction and improved the intensity. Han and Pan (2006) included a CMT parameterisation of convectively-induced pressure gradient force (PGF), which produced a realistic organisation of precipitation and intensity of hurricanes.

NWP models may be run with high enough resolution to explicitly resolve convection and although superparameterisation (i.e. use of a CRM to simulate moist convection, see Grabowski and Smolarkiewicz (1999) and Grabowski (2001), for further information) may be used to resolve convective systems, for the foreseeable future global and climate models require convective parameterisation (CP) schemes. The use of CMT in CP schemes however, ranges from total neglect, to inclusion via pressure gradient forces. Several studies have highlighted the importance of the inclusion of CMT in model simulations (e.g. Moncrieff and Liu (2006), Mahoney et al. (2009), Lackmann et al. (2011)). The need to represent CMT in CP schemes was made evident by Wu and Moncrieff (1996) when they demonstrated that the amount of kinetic energy that was produced by shear
from CMT was of similar magnitude to that produced from buoyancy. Therefore the inclusion of such parameterisation is extremely important in coarse resolution models (such as general circulation and global models), which rely on parameterisation to represent these smaller-scale processes.

As stated, it is accepted that models with large gridlengths (e.g. $> 30$ km) should use parameterisation, however it is common to use nested domains at different resolutions to run simulations. For models with gridlength of less than 5 km, convection is normally not parameterised (e.g. for CRMs with gridlengths of $\sim 1$ km, the mesoscale dynamics are explicitly resolved). This is why convective parameterisation is usually not used for gridlengths of $< 5$ km in high resolution inner domains. However for mesoscale models which have horizontal resolutions of 5 - 25 km, as Moncrieff and Liu (2006) noted, organised convection is conspicuously absent from contemporary parameterisations and this is a concern, as the ability to simulate the mesoscale dynamics leads to realistic organisation in climate models (Moncrieff (2010)). Therefore, in some circumstances such as coarse parent domains, or when running ensembles with gridlengths around 12 km, (and larger e.g. global and climate models), it is still necessary to use CP schemes to realistically allow the development of convection.

2.2.3 Parameterisation Schemes

Various attempts have been made to parameterise CMT in general circulation models (GCMs). Schneider and Lindzen (1976), accounted for momentum transport in their cumulus parameterisation scheme using the assumption that in-cloud velocities depend only on the entrainment rate and the conservation of in-cloud horizontal momentum. This study was based upon the work by Ooyama (1971) who used a mass-flux approach to cumulus parameterisation and assumed that in-cloud horizontal momentum is conserved. However, this conservation only occurs if a negligible horizontal pressure gradient is present. This work was followed by LeMone (1983) and LeMone et al. (1984) who demonstrated that convection can cause changes to the pressure perturbation field around the convective area which in turn can affect the CMT, as in-cloud velocities are influenced greatly by the in-cloud pressure gradients. This point is important as the largest uncertainty in parameterisation of CMT in GCMs comes from estimating those velocities (Richter and Rasch (2008)) and these pressure gradients are also affected by the
tilt of the system. However, the large-scale horizontal pressure gradient also has a great influence on an MCS. Lafore and Moncrieff (1989) quantified the effects of the work done by the horizontal pressure gradient on the generation and the maintenance of mesoscale downdrafts in numerical simulations of tropical squall lines and argued that this should also be taken into account when representing CMT.

As momentum does not mix in the same way as thermodynamic scalar quantities such as heat and moisture, consideration must be taken of the accelerations due to the storm scale pressure fields. In the 1980’s the Fritsch-Chappell convection parameterisation scheme (Fritsch and Chappell (1980)) included convective momentum adjustments for mesoscale gridlengths and the Kain-Fritsch scheme (Kain (2004), Kain and Fritsch (1990), Kain and Fritsch (1993)), which was an extension of this scheme, is also well-suited for CMT.

Observational studies by Shapiro and Stevens (1980) and Flatau and Stevens (1987) included this pressure gradient in their parameterisation schemes and König and Ruprecht (1989) also used a mass-flux approach using vorticity. Recently, several parameterisation schemes of momentum transports by convection have been formulated using pressure gradients to represent in-cloud velocities and the CMT. Zhang and Cho (1991a) and Zhang and Cho (1991b) developed a scheme that parameterises the perturbation pressure gradients and Zhang and Cho (1991a) also included in-cloud pressure gradients by employing a simple model of flow around a cloud. This study was shown to improve the representation of the Hadley circulation acting downgradient and reducing the vertical wind shear (Zhang and McFarlane (1995)) although earlier work by Tiedtke (1989) found that CMT had little, if any, impact. Wu and Yanai (1994) employed a linearised approximation of the diagnostic pressure equation, while Kershaw and Gregory (1997) and Gregory et al. (1997) (hereafter GKI) used a CRM to derive an empirical relationship for the pressure gradient. Both these schemes use the large-scale vertical wind shear and convective mass flux to parameterise the cloud pressure gradients. There is some uncertainty in these parameterisations as the pressure distributions within the convective systems are derived from only a few simulations. However it is this latter scheme that is considered further in Chapter 4 as it has become the most common scheme in global models.
2.2.4 The Gregory parameterisation scheme

The Gregory parameterisation scheme was developed from the Gregory and Rowntree (1990) mass flux convection scheme and a modelling study by Kershaw and Gregory (1997). This section gives some relevant background on how the scheme parameterises momentum transport and the assumptions behind it. To begin, the momentum equations when decomposed, show that the total tendency on the mean flow is influenced by various accelerations, such as the horizontal advection terms, the pressure gradient force and the vertical transport of horizontal momentum. Figure 2.8 shows the terms that are involved in the generation of total momentum transport and the different regions in the cloud where these flows occur. For example, the lowest level shows positive net u-momentum tendency (as with the highest levels) due to increasing rear-to-front flow, however the regions within the levels show different terms dominating and influencing the flow. The equations governing momentum transport are fully described in Chapter 3. As a summary however, starting with the equations of motion:

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial x},$$  \hspace{1cm} (2.1)$$

where \(u, v\) and \(w\) are the zonal, meridional, and vertical wind components respectively, \(\rho\) is the density, and \(p\) is the pressure. Reynolds averaging is applied to each of the variables to separate into the temporally varying domain mean wind and corresponding perturbation. The variables are now represented as \(u = \bar{u}(t) + u', \ v = \bar{v}(t) + v', \ w = \bar{w}(t) + w', \) and \(p = \bar{p}(t) + p',\) where the overbar represents the domain average and the primes denote perturbations. Applying the mass continuity equation produces:

$$\frac{\partial \bar{u}}{\partial t} + 2\frac{\partial \bar{v}u'}{\partial x} + \frac{\partial \bar{v}v'}{\partial y} + \frac{\partial \bar{w}u'}{\partial z} + \frac{1}{\rho} \frac{\partial \rho \bar{u}w'}{\partial x} - \frac{1}{\rho} \frac{\partial \rho u'}{\partial x} - \frac{1}{\rho} \frac{\partial \rho u v'}{\partial y} - \frac{1}{\rho} \frac{\partial \rho u w'}{\partial z} = 0,$$ \hspace{1cm} (2.2)$$

If periodic flow is assumed, the equations of motion can be simplified by neglecting a few small terms, as the horizontal gradients are assumed to approach zero, and it can be shown that for the zonal wind component:

$$\frac{\partial \bar{u}}{\partial t} = -\frac{1}{\rho} \frac{\partial \rho u' w'}{\partial z},$$ \hspace{1cm} (2.3)$$

These variables are separated into the mean and perturbation components across
the domain and it is this scale separation assumption that underlies all parameterisations. This term is also called the momentum flux divergence term and it is this term that therefore determines whether the transport is downgradient or countergradient.

Kershaw and Gregory (1997) used a CRM to simulate ensembles of two different types of convective regime: deep convection for mid-latitude cold air outbreaks and tropical convection. The momentum transports from these simulations were then estimated and used to validate the GKI parameterisation. This parameterisation scheme includes a simple representation of the across-cloud pressure gradients and its effect on the in-cloud flow. It finds that the effects of these across-cloud pressure gradients are mostly balanced by the in-cloud vertical advection.

This GKI scheme is based on the Gregory and Rowntree (1990) mass flux convection scheme and represents the effect of the cloud-pressure gradients on the mass flow within the cloud. It seeks to calculate the vertical eddy flux of horizontal momentum due to convection (Eqn. 2.3 (RHS)) by estimating the mass flux in the updrafts and downdrafts and the in-cloud horizontal velocities. Determining the updrafts and downdrafts is done by modelling them as entraining plumes. The various terms that this scheme seeks to approximate are: a) the vertical velocity, which is calculated from the mass flux, b) the estimated horizontal velocities are based on the small-scale pressure gradients (across updrafts and downdrafts), and c) this scheme approximates the pressure gradient across the updrafts to be proportional to the mass flux and the environmental shear.

The relationship in b) is based on work by Rotunno and Klemp (1982) who used linear analysis to show there is a relationship between the mean shear of the cloud environment and the updraft vertical velocity. LeMone et al. (1988) demonstrated that this also holds true for deep convective clouds, whilst Wu and Yanai (1994) agreed that this relationship is also consistent with the analysis of Rotunno and Klemp (1982). The method that this scheme employs, is described more fully and evaluated further in Section 4.3.5. A derivation of the momentum equations follows, in order to understand the terms involved in the transport of momentum; this will also serve as the basis for investigating the momentum budget.
3 Derivation of the Momentum Budget

In this chapter the derivation of the momentum budget is provided in terms of the Reynold’s averaged momentum equations for a compressible atmosphere, here restricted to the x-component.

The anelastic approximation is usually employed for analyses of CMT as it assumes that density does not vary in space or time - only in the vertical. This assumption is used to remove acoustic waves and is adequate for most situations. However, for this project, the numerical modelling was carried out using the WRF model which uses fully compressible equations, and consequently the derivation of the momentum budget also utilises the continuity equation for compressible flow.

The aim of this chapter is to identify all the terms that contribute to the momentum budget. To achieve this, the equations of motion and the continuity equation are defined and combined to give a full set of terms that contribute to the momentum budget, which is used in the analysis of the momentum budget in Chapter 4.

Starting with the flux form of the equations of motion:

$$\frac{\partial \rho u}{\partial t} + \frac{\partial \rho u^2}{\partial x} + \frac{\partial \rho vu}{\partial y} + \frac{\partial \rho uw}{\partial z} = -\frac{\partial p}{\partial x}$$

(3.1)

where \(\rho\) is the density, \(u, v, w\) are the zonal, meridional and vertical wind components respectively, and \(p\) is the pressure. These variables are then separated into the mean and perturbation components. For example, the domain-averaged horizontal mean of \(u\) is represented by an overbar \(\bar{u}\) and the perturbation by \(u'\). \(\bar{u}\) is a 2D function of height and time and \(u'\) is a function of \((x, y, z, t)\).

Substituting in: \(u = \bar{u} + u', v = \bar{v} + v'\) and \(w = \bar{w} + w'\), gives:

$$\frac{\partial}{\partial t} \rho(\bar{u} + u') + \frac{\partial}{\partial x} \rho(\bar{u} + u')^2 + \frac{\partial}{\partial y} \rho(\bar{u} + u')(\bar{v} + v') + \frac{\partial}{\partial z} \rho(\bar{u} + u')(\bar{w} + w') = -\frac{\partial p}{\partial x}$$

(3.2)
Expanding out the terms:

\[
\rho \frac{\partial \bar{u}}{\partial t} + \rho \frac{\partial u'}{\partial t} + \bar{u} \frac{\partial \rho}{\partial t} + u' \frac{\partial \rho}{\partial t} + \frac{\partial \rho \bar{u}^2}{\partial x} + 2 \frac{\partial \rho \bar{u}'}{\partial x} + \frac{\partial \rho u'}{\partial y} + \frac{\partial \rho u v'}{\partial y} + \frac{\partial \rho u' v'}{\partial y} + \frac{\partial \rho u''}{\partial z} + \frac{\partial \rho u' v'}{\partial z} + \frac{\partial \rho u' v'}{\partial z} + \frac{\partial \rho u''}{\partial z} = -\frac{\partial p}{\partial x} \tag{3.3}
\]

Dividing by \(\rho\):

\[
\frac{\partial \bar{u}}{\partial t} + \frac{\partial u'}{\partial t} + \bar{u} \frac{\partial \rho}{\partial t} + \bar{u} \frac{\partial \rho}{\partial t} + \bar{u} \frac{\partial \rho^2}{\partial x} + \bar{u} \frac{\partial \rho u'}{\partial x} + \bar{u} \frac{\partial \rho u'}{\partial x} + \frac{\partial \rho u v'}{\partial y} + \frac{\partial \rho u' v'}{\partial y} + \frac{\partial \rho u' v'}{\partial y} + \frac{\partial \rho u''}{\partial z} + \frac{\partial \rho u' v'}{\partial z} + \frac{\partial \rho u' v'}{\partial z} + \frac{\partial \rho u''}{\partial z} = -1 \frac{\partial p}{\partial x} \tag{3.4}
\]

Expanding and cancelling terms that are identically zero i.e. horizontal derivatives of mean variables, gives:

\[
\frac{\partial \bar{u}}{\partial t} + \frac{\partial u'}{\partial t} + \frac{\partial \rho}{\partial t} + \frac{\partial \rho}{\partial t} + \frac{\partial \rho^2}{\partial x} + \frac{\partial \rho u'}{\partial x} + \frac{\partial \rho u'}{\partial x} + \frac{\partial \rho u v'}{\partial y} + \frac{\partial \rho u' v'}{\partial y} + \frac{\partial \rho u' v'}{\partial y} + \frac{\partial \rho u''}{\partial z} + \frac{\partial \rho u' v'}{\partial z} + \frac{\partial \rho u' v'}{\partial z} + \frac{\partial \rho u''}{\partial z} = -1 \frac{\partial p}{\partial x} \tag{3.5}
\]

leaves:

\[
\frac{\partial \bar{u}}{\partial t} + \frac{\partial u'}{\partial t} + \frac{\partial \rho}{\partial t} + \frac{\partial \rho}{\partial t} + \frac{\partial \rho^2}{\partial x} + \frac{\partial \rho u'}{\partial x} + \frac{\partial \rho u'}{\partial x} + \frac{\partial \rho u v'}{\partial y} + \frac{\partial \rho u' v'}{\partial y} + \frac{\partial \rho u' v'}{\partial y} + \frac{\partial \rho u''}{\partial z} + \frac{\partial \rho u' v'}{\partial z} + \frac{\partial \rho u' v'}{\partial z} + \frac{\partial \rho u''}{\partial z} = -1 \frac{\partial p}{\partial x} \tag{3.6}
\]

The continuity equation is:

\[
\frac{D\rho}{Dt} + \rho \nabla \cdot u = 0
\]

\[
\Rightarrow \frac{\partial \rho}{\partial t} + \nabla \cdot \rho u = 0
\]
\[ \Rightarrow \frac{\partial \rho}{\partial t} + \frac{\partial}{\partial x} \rho u + \frac{\partial}{\partial y} \rho v + \frac{\partial}{\partial z} \rho w = 0 \]

\[ \Rightarrow \frac{\partial \rho}{\partial t} + \frac{\partial}{\partial x} (\rho (\bar{u} + u')) + \frac{\partial}{\partial y} (\rho (\bar{v} + v')) + \frac{\partial}{\partial z} (\rho (\bar{w} + w')) = 0 \]

\[ \Rightarrow \frac{\partial \rho}{\partial t} + \frac{\partial \rho \bar{u}}{\partial x} + \frac{\partial \rho u'}{\partial x} + \frac{\partial \rho \bar{v}}{\partial y} + \frac{\partial \rho v'}{\partial y} + \frac{\partial \rho \bar{w}}{\partial z} + \frac{\partial \rho w'}{\partial z} = 0 \]

Expanding:

\[ \Rightarrow \frac{\partial \rho}{\partial t} + \bar{u} \frac{\partial \rho}{\partial x} + \bar{v} \frac{\partial \rho}{\partial y} + \bar{w} \frac{\partial \rho}{\partial z} + \nabla \cdot \rho \mathbf{u'} = 0 \]

Removing terms that are identically zero, i.e., horizontal derivatives of mean variables, gives:

\[ \Rightarrow \frac{\partial \rho}{\partial t} + \bar{u} \frac{\partial \rho}{\partial x} + \bar{v} \frac{\partial \rho}{\partial y} + \bar{w} \frac{\partial \rho}{\partial z} + \nabla \cdot \rho \mathbf{u'} = 0 \]

Rearranging:

\[ \Rightarrow \frac{\partial \rho}{\partial t} + \bar{u} \frac{\partial \rho}{\partial x} + \bar{v} \frac{\partial \rho}{\partial y} + \bar{w} \frac{\partial \rho}{\partial z} + \nabla \cdot \rho \mathbf{u'} = - \frac{\partial \bar{w}}{\partial z} \quad (3.7) \]

Replacing the bracketed terms in 3.6 with the term on the RHS of Eq. 3.7, leaves:

\[ \frac{\partial \bar{u}}{\partial t} + \frac{\partial u'}{\partial t} + \frac{\partial}{\rho \frac{\partial \rho}{\partial t}} + \bar{u} \frac{\partial \rho u'}{\partial x} + \bar{v} \frac{\partial \rho u'}{\partial y} + \bar{w} \frac{\partial \rho u'}{\partial z} - \bar{u} \frac{\partial \bar{w}}{\partial z} + w' \frac{\partial \bar{u}}{\partial z} \]

\[ + \frac{\partial \bar{w}}{\partial z} + \bar{u} \frac{\partial w'}{\partial z} + \frac{1}{\rho} \frac{\partial p u^2}{\partial x} + \frac{1}{\rho} \frac{\partial p u v'}{\partial y} + \frac{1}{\rho} \frac{\partial p u w'}{\partial z} = - \frac{1}{\rho} \frac{\partial p}{\partial x} \quad (3.8) \]

Taking the horizontal average of Eq. 3.8 and noting the assumption that \( \bar{v} = 0 \), which makes \( \bar{v} \frac{\partial \rho u'}{\partial y} = 0 \) and

\[ \left\langle \frac{\partial u'}{\partial t} \right\rangle = 0, \left\langle u' \frac{\partial \bar{w}}{\partial z} \right\rangle = 0, \left\langle w' \frac{\partial \bar{u}}{\partial z} \right\rangle = 0 \]

leaving:

\[ \left\langle \frac{\partial \bar{u}}{\partial t} \right\rangle + \left\langle \frac{u' \frac{\partial \rho}{\partial x}}{\rho} \right\rangle + \left\langle \bar{u} \frac{\partial \rho u'}{\partial x} \right\rangle + \left\langle \bar{w} \frac{\partial \rho u'}{\partial z} \right\rangle + \left\langle \frac{\partial \bar{u} \bar{w}}{\partial z} \right\rangle - \left\langle u' \frac{\partial \bar{w}}{\partial z} \right\rangle + \left\langle \frac{1}{\rho} \frac{\partial p u^2}{\partial x} \right\rangle \]

\[ + \left\langle \frac{1}{\rho} \frac{\partial p u v'}{\partial y} \right\rangle + \left\langle \frac{1}{\rho} \frac{\partial p u w'}{\partial z} \right\rangle = - \left\langle \frac{1}{\rho} \frac{\partial p}{\partial x} \right\rangle \]
The brackets $\langle \rangle$ are removed for the remainder of this derivation. As $\frac{\partial \bar{\omega}\bar{u}}{\partial z} = \bar{\omega}\frac{\partial \bar{u}}{\partial z} + \bar{u}\frac{\partial \bar{\omega}}{\partial z}$, so $\bar{\omega}\frac{\partial \bar{u}}{\partial z}$ can be substituted in to replace $\frac{\partial \bar{\omega}\bar{u}}{\partial z} - \bar{u}\frac{\partial \bar{\omega}}{\partial z}$, and therefore the equation of motion for fully compressible flow can be represented by:

$$\frac{\partial \bar{u}}{\partial t} + \frac{u'}{p} \frac{\partial \rho}{\partial t} + \frac{\bar{u}}{p} \frac{\partial u'}{\partial x} + \frac{\bar{w}}{p} \frac{\partial u'}{\partial z} + \frac{w}{p} \frac{\partial u^2}{\partial x} + \frac{1}{p} \frac{\partial p u'}{\partial y} + \frac{1}{p} \frac{\partial p u' w'}{\partial z} = -\frac{1}{p} \frac{\partial p}{\partial x}$$ (3.10)

Removing the $\frac{u'}{p} \frac{\partial \rho}{\partial t}$ and $\frac{\bar{w}}{p} \frac{\partial u'}{\partial z}$ terms which are close to zero and rearranging to separate the terms that affect the tendency to the RHS, leaves:

$$\frac{\partial \bar{u}}{\partial t} = -\frac{\bar{u}}{p} \frac{\partial u'}{\partial x} - \frac{w}{p} \frac{\partial \bar{u}}{\partial z} - \frac{1}{p} \frac{\partial u^2}{\partial x} - \frac{1}{p} \frac{\partial p u'}{\partial y} - \frac{1}{p} \frac{\partial p u' w'}{\partial z} - \frac{1}{p} \frac{\partial p}{\partial x}$$ (3.11)

This final equation retains all the terms to close the momentum budget and is used in Section 4.3.1. It is valid for no background meridional flow i.e. $\bar{v} = 0$. 
4 Momentum transport in idealised simulations

4.1 Introduction

As detailed in Chapter 2, CMT is known to have a considerable effect on both the zonal and meridional mean flow, including the Hadley circulation as well as surface winds, precipitation and the motion of MCS’s. CMT is normally parameterised in global models, but uncertainties remain about the dynamics of CMT, its effects, and its representation in current schemes. This chapter addresses the problem of CMT using idealised simulations to assess some of the assumptions underlying its parameterisation. This stage of the research begins by using the derivations from the previous chapter.

It is the aim of this chapter to build on previous work using a variety of idealised model simulations to test some of the assumptions inherent in the CMT parameterisation schemes. This part of the thesis serves to determine what terms are influential in the momentum budget, thereby addressing the first question posed in Chapter 1 (i.e., Which terms are most influential to the momentum transport profile?). It also serves as a preliminary investigation into how momentum transport is influenced by different scales - both spatially and temporally.

Specifically, a range of model domain sizes are used to assess the contributions of terms involving horizontal gradients, which are normally neglected in parameterisations. Their sensitivity to model resolution is also considered. The across-updraft and downdraft pressure gradients are also examined in the context of the relationships used in the GKI parameterisation and also to assess the relative contributions of the convective-scale transports to the domain-mean tendency. The remainder of this chapter is ordered as follows: the numerical model and its configuration is described in Section 4.2.1, with a description of the three-dimensional convective system studied in Section 4.2.2. The results of the idealised simulations are presented and analysed in Section 4.3. This is followed by an evaluation of the GKI parameterisation scheme; this chapter is then
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summarised in Section 4.4.

4.2 Idealised model simulations

4.2.1 Model description and configuration

All simulations use the Weather Research and Forecasting Model (WRF) - Advanced Research core (ARW) version 3.3 (Skamarock et al. (2008)). WRF-ARW is a mesoscale modelling system, developed principally by the National Center for Atmospheric Research and is used for real-time and operational forecasting as well as research applications. The model uses the fully compressible Eulerian non-hydrostatic equations, which are solved on a finite-difference mesh with Arakawa C-grid staggering and formulated using a mass-based vertical coordinate (Laprise (1992)). In this study moist processes are treated explicitly (no cumulus parameterisation) and represented by the WRF single-moment 6-class microphysics (WSM6) scheme (Hong and Lim (2006)). The effects of subgrid turbulence are parameterised using a predictive 1.5-order turbulence kinetic energy closure. For this study Coriolis effects and other physical processes (viz. radiation, surface friction, surface fluxes) are neglected.

The height of the computational domain is 35 km and a 10 km Rayleigh damping layer is imposed to prevent gravity wave reflection from the upper boundary. Three computational domains are used: 100 km x 50 km (denoted small, Sm), 400 km x 200 km (denoted medium, Med) and 800 km x 400 km (denoted large, Lg). Another domain - 50 km x 50 km (denoted square, Sq) was also run with 1 km horizontal grid spacing (Sq_1km) only. For each of the main three domain sizes, three different horizontal grid spacings were used to explore resolution sensitivities (500 m, 1 km and 3 km). There are 71 vertical levels with an approximate vertical grid spacing of 500 m for all domains. The timesteps are 3 s, 6 s and 12 s for 500 m, 1 km and 3 km gridlengths, respectively. All model simulations are run for a duration of 6 hours - the domain configurations are summarised in Table 4.1. All domain configurations are run for both open and cyclic boundary conditions, except the Sq domain which uses cyclic boundary conditions only.
4.2. Idealised model simulations

The initial model thermodynamic environment for all simulations is defined using the Weisman and Klemp (1982) analytic sounding (Fig 4.1(b)-(c)). The initial wind profile is shown in Fig. 4.1(a) and contains constant vertical shear near the surface such that

\[
U_0(z) = \begin{cases} 
0, & z > h; \\
U_0 - \frac{U_0}{h}z, & z \leq h.
\end{cases}
\]  

(4.1)

where \( U_0 = -18.0 \text{ m s}^{-1} \) and \( h = 3.25 \text{ km} \). This wind profile is used because previous studies have shown that this low-level linear shear produces a strong, long-lived MCS (Thorpe et al. (1982), Weisman and Rotunno (2004)). The meridional wind component \( v \) is initially zero. Convection is initiated by a temperature perturbation (‘warm bubble’) of 3 K in the horizontal centre of the domain with a radius of 10 km and a height of 1.5 km. The mean wind profile is free to evolve with the simulation; an example evolution of the mean zonal wind \( u \) is shown in Fig. 4.1(a) (from Med_1km, See Table 4.1, with cyclic boundary conditions). This evolution represents the effect of the mean momentum tendency, which leads to an increase in surface \( u \), a decrease in upper-level \( u \) and an eventual reduction in the low-level shear.

<table>
<thead>
<tr>
<th>Model run</th>
<th>Domain size</th>
<th>( \Delta x, \Delta y )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sq_1km</td>
<td>50 km x 50 km</td>
<td>1 km</td>
</tr>
<tr>
<td>Sm_500m</td>
<td>100 km x 50 km</td>
<td>500 m</td>
</tr>
<tr>
<td>Sm_1km</td>
<td>100 km x 50 km</td>
<td>1 km</td>
</tr>
<tr>
<td>Sm_3km</td>
<td>100 km x 50 km</td>
<td>3 km</td>
</tr>
<tr>
<td>Med_500m</td>
<td>400 km x 200 km</td>
<td>500 m</td>
</tr>
<tr>
<td>Med_1km</td>
<td>400 km x 200 km</td>
<td>1 km</td>
</tr>
<tr>
<td>Med_3km</td>
<td>400 km x 200 km</td>
<td>3 km</td>
</tr>
<tr>
<td>Lg_500m</td>
<td>800 km x 400 km</td>
<td>500 m</td>
</tr>
<tr>
<td>Lg_1km</td>
<td>800 km x 400 km</td>
<td>1 km</td>
</tr>
<tr>
<td>Lg_3km</td>
<td>800 km x 400 km</td>
<td>3 km</td>
</tr>
</tbody>
</table>

**Tab. 4.1:** Simulation domain configurations.
4. Momentum transport in idealised simulations

Fig. 4.1: Profiles of (a) zonal wind $u$ (m s$^{-1}$), (b) mixing ratio $q$ (g kg$^{-1}$) and (c) potential temperature $\theta$ (K) used in the initial sounding for all simulations. Also shown in (a) is the mean wind $\bar{u}$ every two hours for the medium 1 km domain with cyclic boundary conditions.

4.2.2 Convective system

The result of all simulations is a three-dimensional mesoscale convective system of the leading convective-line trailing-stratiform type (e.g. Houze et al. (1989), Parker and Johnson (2000)), though there are structural differences in spatial extent and strength between the convection in the various size and resolution domains. As an example, the evolution and mature structure of a simulated storm (Med_1km, see Table 4.1) is shown in Figs. 4.3 & 4.2. The system covers about 200 km and 100 km in the x- and y-directions respectively (Fig. 4.2). The vertical cross-section of the zonal wind in Fig. 4.3(c) implies the triple-branch model from Moncrieff (1992) (his Fig. 1(b)) with front-to-rear ascending flow, an overturning updraft branch, and a descending mesoscale downdraft at the rear of the system. A stagnation point is present at the surface, near x=170 km, identifying the leading edge of the cold pool that helps maintain the system by
triggering new convection. The contours in Fig. 4.3(d) are perturbation pressure and the red contours represent the area of greatest negative perturbation pressure, which indicates the position of the mesolow, located behind the updraft. This is reinforced by the presence of a region of higher pressure at the front of the system.

Fig. 4.2: Horizontal cross-section of cloud at 8 km height at (a) 0145 hours and (b) 0300 hours. Cloud outline is cloud mixing ratio 0.1 g kg$^{-1}$. Black contours represent areas of convective updrafts (defined as in-cloud w $>$ 4 m s$^{-1}$ and 15-min surface rain $>$ 1 mm h$^{-1}$). The blue shading represents the cloud mixing ratio, the darkest shades represent values of 1.1 g kg$^{-1}$ and the lightest shading is 0.05 g kg$^{-1}$.

The mean wind profile has positive low-level wind shear (Fig. 4.1) and Fig. 4.3 demonstrates that the main convective region of the mature system is upshear-tilted. The simulated systems evolve in a similar way to that depicted by Weisman and Rotunno (2004), with the convective region being downshear-tilted in its earliest stages and becoming upright as the cold pool develops (Fig. 4.3(b)). This evolution will be considered later.
Fig. 4.3: Cross-section of cloud (cloud mixing ratio contour outline 0.1 g kg\(^{-1}\) is shown by solid black line) averaged in the y-direction and time-averaged during the following time periods: (a) 0000-0045 hours, (b) 0100-0145 hours and (c) & (d) 0300-0345 hours when the system is fully developed. (a)-(c) shows contours of \(u\) (every 2.0 m s\(^{-1}\), with zero removed) - positive values shown by solid lines and negative by dashed) for medium 1 km resolution cyclic domain. The cold pool is shown by blue shading, using an outer contour of \(-1^\circ\)C temperature perturbation with the darkest shading approximately \(-5^\circ\)C. (d) The black contours are pressure perturbations every 30 Pa, with negative values shown by dashed lines and positive values shown by solid lines. The red contours shows the pressure perturbations every 30 Pa, less than 150 Pa i.e. the region of lowest pressure.
4.3 Results

To calculate the momentum budgets and tendencies, the model output was first interpolated from the WRF native mass-based coordinate system to height coordinates. This interpolation was necessary to close the momentum budget as written in Eq. 3.11. The sum of all the terms comprising the momentum budget (Eq. 3.11) along with the mean tendency of $\bar{u}$, were calculated for all simulations with the difference between the sum of the terms and the mean tendency defining a residual. Budgets were initially compared for calculations using model output every 3, 15, 30 and 60 minutes for the Med_1km_open simulation (Fig. 4.4). As expected, the budgets with the 3-minute output showed the smallest residual compared to the other time resolutions; 15-minute data provided similar results, with a notable increase in the residual for 30- and 60-minute data. The small amount of added accuracy for the 3-minute output was not worth the considerably larger data storage, and 15-minute model output is used for all budgets found herein.

4.3.1 Analysis of momentum budgets and tendencies

In order to compare the domain-mean vertical flux of horizontal momentum with the other terms comprising the momentum budget (Eqn. 3.11), Fig. 4.5(a) shows all terms that contribute to the mean tendency for the open boundary conditions, medium domain. It is obvious that although the momentum flux ($\rho u'w'$) term dominates the other terms and accounts for the overall sign of the tendency across the domain in the upper-levels, there are other terms that significantly contribute to the tendency on the mean flow. The terms that are most important within the momentum budget are $\frac{1}{\rho} \frac{\partial \rho u'w'}{\partial z}$, $\frac{1}{\rho} \frac{\partial \rho u'}{\partial x}$ and $\frac{1}{\rho} \frac{\partial p}{\partial x}$. As stated, the sign of the total tendency (indicated by the solid red line) at upper levels is consistent with the divergence of the vertical flux of horizontal momentum but this is offset by $\frac{1}{\rho} \frac{\partial p}{\partial z}$ and $\frac{1}{\rho} \frac{\partial \rho u'}{\partial z}$, which comprise the static and dynamic Bernoulli pressure terms, respectively.

The sum of all the budget terms (Fig. 4.5(b) - red line) is very similar to the actual tendency, i.e., there is a very small residual. The largest residual is found near the tropopause at approximately 10.5 km altitude. This larger residual at these heights is related to transient behaviour, presumably linked to overshooting convective updrafts, because calculations with 3-minute data showed a smaller
**Fig. 4.4:** Residuals for model output every 3, 15, 30 and 60 minutes for the medium 1 km domain with open boundary conditions.
4.3. Results

Fig. 4.5: Profiles of (a) the domain mean momentum tendency (red line) and all contributing terms (m s\(^{-1}\) h\(^{-1}\)) averaged over the 6 h simulation, (b) comparison of the momentum tendency (red line) and the sum of all the terms (black line) (m s\(^{-1}\) h\(^{-1}\)), (c) momentum transport \(\rho u'w'\) (N m\(^{-2}\)) and (d) the tendency in m s\(^{-1}\) h\(^{-1}\) for the 1 km horizontal resolution, medium domain with open boundary conditions. (c) and (d) show various stages throughout the 6 h simulation.

In the first two hours of the simulation the mean momentum flux and tendency is much smaller than at other times (and over the whole 6-hour simulation) (Fig. 4.5(c)). The tendency has opposite sign at early times near the tropopause compared to later times, but is very small (Fig. 4.5(d)). Fig. 4.3(b) shows a slight downshear tilt suggesting that any transports due to mesoscale organisation, albeit weak, are potentially offset by those from convective-scale transports as the domain-mean momentum flux is near zero.

The three dominant terms \(-\frac{1}{\rho}\frac{\partial\rho u'w'}{\partial z}, \frac{1}{\rho}\frac{\partial \rho u'}{\partial x}\) and \(\frac{1}{\rho}\frac{\partial \rho}{\partial x}\) are also averaged across the convective updrafts and downdrafts to determine the convective-scale be-
Momentum transport in idealised simulations

**Fig. 4.6:** Time-averaged terms, which are area-averaged across the updrafts and (a) pressure gradients $-\frac{1}{\rho} \frac{\partial p}{\partial x}$, (b) $-\frac{1}{\rho} \frac{\partial \rho u'}{\partial x}$, (c) $-\frac{1}{\rho} \frac{\partial \rho w'}{\partial z}$ and (d) the sum of $-\frac{1}{\rho} \frac{\partial \rho u'}{\partial x} - \frac{1}{\rho} \frac{\partial \rho w'}{\partial z}$ which represents the time-varying tendency, across updrafts for the 1 km horizontal resolution, medium domains with open boundary conditions. Dashed lines are the downdraft profiles and the solid lines are the updrafts.

**haviour and its contributions to the domain-mean momentum budgets.** Here updrafts are defined as regions within cloud that have vertical velocity greater than 1 m s$^{-1}$, and downdrafts are regions within cloud with vertical velocity less than -1 m s$^{-1}$. Figure 4.6 shows how these small-scale contributions to the tendency change as the convective system evolves. During the initial stages (i.e. the first two hours), the pressure gradient across updrafts is positive, changing to negative at low levels during the later stages as the system becomes organised. This suggests the pressure gradient acts to accelerate the low level flow in the downshear direction in the early stages and in the upshear direction in the mature stages. At low-levels the magnitude of this term increases as the system develops. The pressure gradients across the downdrafts generally show a weaker tendency of the opposite sign to the updraft tendencies.

The total cross updraft/downdraft momentum flux $\rho u'w'$ divergence (Fig.
4.3. Results

4.6(c)) increases in magnitude as the system matures. The tendency across the updrafts acts to increase the zonal flow below about 4 km and decrease the flow above 4 km, consistent with net negative momentum flux and downgradient transport. The momentum flux tendency associated with the updrafts is larger than that from the downdrafts, which is mostly confined below 4 km. This result is consistent with intense convective cores and evaporatively driven downdrafts, which act to partially offset the tendency from updrafts.

Weisman and Rotunno (2004) explain how mesoscale convective systems evolve from being downshear tilted to upshear tilted as the baroclinically generated vorticity from the cold pool begins to dominate the environmental shear at low-levels. This explanation is consistent with the evolution of the convective system herein, with (Fig. 4.3 (a)-(c)) showing the transition from downshear to upright to upshear tilted over about 3 hours. However, the tendency from the updraft/downdraft momentum flux divergence (Fig. 4.6(c)) does not undergo this transition and maintains its vertical structure throughout the evolution. The only significant change from the early stages to the mature stages is the change in the cross-updraft pressure gradient. Later in the storm evolution the upshear-tilted mesoscale circulations work in concert with the convective-scale transports.

An interesting point to note is that the $\frac{1}{\rho} \frac{\partial (\rho u^2)}{\partial x}$ term (Fig. 4.6(b)) across up/- downdrafts, reinforces the momentum flux term (i.e. strengthens the tendency) at low altitudes. In the domain mean, however, the low altitude tendency from this term is small, which suggests the domain-mean mesoscale contribution balances the convective-scale contribution. Moreover, the total up/downdraft contributions from the three dominant terms (Fig. 4.6(d)) are consistent in sign with the domain-mean contributions at low altitudes ($< 4$ km). However above this level, the sign is opposite to the domain-mean terms in Figure 4.5(a). This difference in sign above 4 km suggests that the mesoscale tendencies oppose and dominate the convective-scale tendencies above the background shear layer.

4.3.2 Effect of domain size on momentum budget

Figure 4.7 compares the momentum budgets for the medium and large domains. The large domain convective system has a similar size (both horizontally and vertically) to that from the medium domain and also develops at a similar rate resulting in almost identical dynamical structures and orientation. For the larger
Momentum transport in idealised simulations

domain the mean tendencies are smaller, due to the larger averaging area. As expected, the horizontal derivatives becoming less important the farther the domain boundary is from the convective area and tending to zero with increasing domain size. For the large domain \( \frac{1}{\rho} \frac{\partial \rho u^2}{\partial x} \) is negligible. However, an important feature of Fig. 4.7(c) is that even in the large domain, the domain-mean pressure gradient is still an important component of the momentum budget. The pressure gradient generally reduces the magnitude of the tendency caused by the momentum flux. This mesoscale pressure gradient is an influential component of the momentum budget, even on scales much larger than the convective system, and acts to maintain the organised mesoscale circulation. The relevance of this result is that current CMT parameterisation schemes neglect this mesoscale pressure gradient term and only attempt to represent the impact of the convective-scale pressure gradient term on the in-cloud velocities (an example is the GKI scheme).

The tendencies for the cyclic simulations, however, are comprised solely of the momentum flux divergence \( \frac{1}{\rho} \frac{\partial \rho u^2 w'}{\partial z} \), as is assumed for all parameterisation schemes. This is because the horizontal derivatives - \( \frac{1}{\rho} \frac{\partial p}{\partial x} \) and \( \frac{1}{\rho} \frac{\partial \rho u^2}{\partial x} \) - are identically zero. Of interest is that for each domain size, the \( u \) tendency for the domains with open boundary conditions (Fig. 4.7 (left column)) are similar to those for cyclic boundary conditions (right column). This is perhaps not surprising for the large domain as the horizontal gradient terms are small (see discussion above). However, this result is not necessarily expected for the medium domain. In particular, for the open boundary conditions the tendency from the momentum flux is large in magnitude and offset by the horizontal gradient terms. For the cyclic boundary conditions the momentum flux tendency (which is equal to the total tendency) is much smaller, allowing the \( u \) tendency for each simulation to be very similar. It might be as the net effect of the cyclic boundary conditions (and stronger descent, see discussion later) is to reduce the strength of the convection and hence the momentum flux, with the agreement in \( u \) tendency being fortuitous.

4.3.3 Resolution sensitivity

The momentum budgets for the different horizontal resolutions are compared (Fig. 4.8) and it is evident that the resolution does not greatly affect the relative contributions from the terms that comprise the momentum budgets. Most of the terms from Eqn. 3.11 contribute to the total tendency for the medium domain. The strength of the domain mean momentum flux and pressure gradient tendencies in
Fig. 4.7: Momentum budgets for 1 km medium (top) and large (bottom) domains with open (left) and cyclic (right) boundary conditions. The total tendency $\frac{\partial \bar{u}}{\partial t}$ is shown in red.
the upper troposphere is largest for the 3 km model (Fig.4.8(c)) compared to the finer grid spacing simulations. At lower altitudes the differences are smaller. This upper-level sensitivity is likely because there is less entrainment and mixing in the 3 km simulation and therefore there is greater vertical transport of horizontal momentum beyond the middle troposphere.

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{fig4.8.png}
\caption{Momentum budget comparison for simulations with grid spacing of 500 m (left), 1 km (centre) and 3 km (right) models, for medium domain with open boundaries. Domain average values of terms shown in key.}
\end{figure}

Direct comparison of the domain mean tendencies (Fig. 4.9(a)) show the largest differences above the low-level shear layer in regions where convective updrafts and mesoscale circulations are strong. The 3 km grid spacing simulation shows a region of positive tendency at about 12 km, which is likely related to enhanced overshooting updrafts. There is no clear convergence of the results in terms of the mean tendency.
For the convective-scale contributions to the tendency, viz. $\rho u'w'$ averaged across the up/downdrafts, the simulations do show convergence between the 1 km and 500 m grid spacing simulation (Fig. 4.9(b)). The 3 km case shows enhanced convective-scale fluxes above about 7 km altitude, consistent with less entrainment and larger undiluted updrafts (see e.g., (Bryan et al., 2003)). This apparent convergence at the convective scales, but not the domain-mean tendency suggests that at least some of the resolution sensitivity arises from changes in the mesoscale circulations with resolution.

These results compare favourably to those of Weisman et al. (1997), who found a similar behaviour for 2 km and 4 km grid spacing, which produced slightly greater momentum flux when compared to the 1 km simulation. Overall, the fluxes compared well due to the coarser resolution model runs producing mature, upshear-tilted circulations. They imply that resolutions up to 12 km will produce mesoscale structures and therefore CMT need not be parameterised in such cases. However, those results are specifically focused on the mesoscale structure of squall
4. Momentum transport in idealised simulations

lines and not the convective-scale structures as shown in Fig. 4.9(b). As discussed by Bryan et al. (2003), care should be taken not to use the model output from 1 km resolution simulations to represent a benchmark or control, as much finer resolution is needed, i.e. $O(100 \text{ m})$, to produce turbulent flows including entrainment.

Another interesting point to note is the opposite sign of the momentum flux of the updrafts below 4 km and the downdrafts above 4 km. Figure 4.9(b) shows that the momentum flux from the downdrafts is positive above this height, and the updrafts are negative. Referring back to Fig. 4.3(c), the mesoscale momentum transport due to the front-to-rear flow and the rear-inflow jet (RIJ) have the same sign of $\rho u'w'$ (negative), which is consistent with the dominant updraft flux at middle and upper levels and the downdraft flux at low-levels. The low-level countergradient updraft flux and upper-level countergradient downdraft flux likely arise from circulations that are tilted in the opposite direction to the convective system; Fig. 4.3(c) suggests that these circulations are located ahead of the convective system in the region of forward-tilted mesoscale circulations.

4.3.4 Effect of domain size on simulated convective systems

The sensitivity of the momentum transport to the size of the domain was investigated earlier to determine whether the assumptions underlying typical momentum transport parameterisation scheme are justified. Therefore, is it appropriate to approximate the tendency by the momentum flux only (as represented by Eqn. 2.3)? Also, is it appropriate to use small domains to study CMT as they struggle to represent the mesoscale transports, which we have shown in the previous section to be important? This is particularly relevant as it has become common to use long-running radiative-convective equilibrium (RCE) simulations with periodic boundary conditions to examine the structure and impact of moist convection, including momentum transport.

To explore this sensitivity, Hovmöller diagrams of cloud mixing ratio during the six hours simulation for all domain sizes with periodic boundary conditions were used to demonstrate how the different domains allow for the development of convection. In order to provide a direct comparison between the domains, for analysis purposes the domains were trimmed to the same size. The large and medium domains were trimmed to 100 km width and the square domain (50 km x 50 km) was duplicated. The cyclic simulations were also beneficial for this analysis.
because for open boundary conditions with small domains the convective system moves out of the domain. The largest two domains (Fig. 4.10 (top)) show the convective systems are maintained throughout the 6-hour simulation and beyond, and show very similar horizontal and temporal structure. Figure 4.10 (bottom) show that the small and the square domains have lifespans of around 2-3 hours only, dissipating thereafter. It is instructive to explore why these systems decayed earlier than those in the larger domains.

![Hovmöller diagrams of cloud for various domains](image)

**Fig. 4.10:** Hovmöller diagrams of cloud for the 1 km cyclic, large and medium domains (top left and right respectively) and small and square domains (below left and right respectively). The square domain was doubled to have the same 100 km width as the other domains (the large and medium domains are trimmed to 100 km). The blue shading shows the average mixing ratio (cloud, rain, ice, snow and graupel) in g kg$^{-1}$.

To determine what is causing these short-lived systems and why the convection in these small domains is consequently suppressed, the mean vertical velocity for the same (trimmed) regions shown in Fig. 4.10 was calculated. Figure 4.11 shows that the smaller domains have much stronger downward velocities, which suppresses the convection within those domains by stabilising the environment. The reason for this is that for periodic boundary conditions the domain-mean
vertical velocity must be zero, and for a given convective system the strength of this subsidence must be larger for a smaller domain. As shown here for this environment, once the domain gets too small (i.e. $\leq 100 \text{ km x 50 km}$), the longevity of the convective system is reduced and it is unable to maintain a long-lived organised system.

Fig. 4.11: Mean vertical velocity out of cloud across trimmed domains (100 km width) for all cyclic domains at 1 km resolution.

One alternate explanation for the differences in organisation in the different sized domains is that the structure (and tilt) of the convective systems are different in their early stages, which cause them to fail to organise in the smaller domains. To determine whether the convective-scale behaviour of the convective systems in the smaller domains are unduly affected by the domain size at the early stages of their lifecycle, the momentum flux, the pressure gradient and the $\frac{1}{\rho} \frac{\partial \rho \mathbf{u}^2}{\partial x}$ terms averaged across the up/downdrafts for the small, medium and large domains were calculated during the first 2 hours of each simulation. As Fig. 4.12 indicates there is very little difference between the profiles from the various size domains. This implies that all the domains produce convective systems that are similar in structure during the initial stages of the simulation. Therefore in the first two hours all simulations have convective systems with similar convective-
4.3. Results

**Fig. 4.12:** Time-averaged (a) pressure gradients $-1/\rho \frac{\partial p}{\partial x}$, (b) $-1/\rho \frac{\partial \rho u'u'}{\partial x}$, (c) $-1/\rho \frac{\partial \rho w'w'}{\partial z}$ and (d) $-1/\rho \frac{\partial \rho u'u'}{\partial z} - 1/\rho \frac{\partial p}{\partial x} - 1/\rho \frac{\partial \rho u'u'}{\partial x}$. All terms are averaged across updrafts (solid) and downdrafts (dashed) for the 1 km horizontal resolution domains with cyclic boundary conditions during 0-2 hours (represented by $< >$).

Scale behaviour, and the small domains produce an adequate representation of the convective scales. It is the smaller domains, with their stronger compensating subsidence, that inhibits the transition of the convection to organised systems that can be maintained beyond a few hours.

Some recent studies (e.g. Romps (2012)) studied CMT using small domains approximately the size of our square domain and ran simulations to RCE. One conclusion of Romps’s study was that the horizontal pressure gradient is not particularly important for the parameterisation of convective momentum transport. The results here show, however, that the small domain can produce systems with a notable cross-updraft pressure gradient (Fig. 4.12(a)), but these systems cannot be maintained. It unlikely such a system would form spontaneously in a small-domain RCE simulation, which would instead be dominated by unorganised convection with weaker cross-updraft pressure gradients.
4.3.5 Evaluation of the Gregory parameterisation scheme

The GKI parameterisation scheme (Gregory et al. (1997)) is popularly used to parameterise CMT (with shallow and deep convection often parameterised separately (Stratton et al. (2009))). The GKI scheme was developed using results from Kershaw and Gregory (1997), which was used to estimate the momentum transports from various regimes of deep convection. Parameterised transports are determined using the mass-flux from the model’s cumulus parameterisation scheme as well as a parameterisation of the effects of cross-updraft pressure gradients. A key aspect of this parameterisation is the assumption that the convective-scale pressure gradients (i.e., across the updrafts and downdrafts) are proportional to the product of the mass flux and the mean vertical wind shear, such that (for updrafts)

$$-\frac{\partial}{\partial x} \left( \frac{p' \rho}{\rho} \right) \bigg|_{u} \sim C_u M_u \frac{\partial \bar{\pi}}{\partial z}$$

where $C_u$ is a constant, $M_u$ is the mass flux across the updraft given by $\rho \bar{w} u$, where $\bar{()}$ represents the averages over the area covered by convective updrafts, and $\bar{\pi}$ is the domain mean background wind. This relation was determined by linear theory and assumes that a high pressure anomaly exists on the upshear side on an updraft (see Rotunno and Klemp (1982)), akin to a plume in shear flow.

Figure 4.13 examines the terms in Eq. 4.2 across the updrafts, throughout the duration of the simulation using the medium domain with open boundary conditions. The mass flux $M_u$ (Fig. 4.13a) is strictly positive (by definition) and accordingly the $M_u \frac{\partial \bar{\pi}}{\partial z}$ term (Fig. 4.13b) is positive within the shear layer ($< 4$ km), though varies in magnitude with time. Above the shear layer this term is small. The convective-scale pressure gradient (Fig. 4.13c) changes with both time and height, evolving from being negative at early times to being positive within the shear layer later, while remaining negative further aloft. This pressure variation, is broadly consistent with Eq. 4.2 in the first two hours, i.e. with the pressure gradient force being directed downshear, but as the system develops the pressure gradient force in the shear layer is directed upshear. This reversal of the low-level across-updraft pressure gradient is related to the development of the mesolow behind the leading convective line, as described by LeMone (1983), and is inconsistent with Eq. 4.2. This mesolow is also evident in Fig. 4.3(d), as
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indicated by the red contours.

The parameter $C_u$ (Fig. 4.13(d)) (which is only shown here below 3.5 km as

![Graph showing terms in the relation used in the Gregory parameterisation scheme.](image)

Fig. 4.13: Terms in the relation used in the Gregory parameterisation scheme, $-\frac{\partial}{\partial z} \left( \frac{\rho p}{\rho w} \right) \propto C_u M_u \frac{\partial U}{\partial z}$ - shows (a) $M_u = \rho w$, (b) $M_u \frac{\partial U}{\partial z}$, (c) $-\frac{\partial}{\partial x} \left( \frac{\rho p}{\rho w} \right)$, and (d) $C_u$ where $C_u = -\frac{\partial}{\partial x} \left( \frac{\rho p}{\rho w} \right)$ but only up to 3.25 km - (the height of the shear level) for the 1 km resolution medium open domain. These terms are averaged across time periods and across updrafts ($> 1 \text{ m s}^{-1}$) for each height level.

it is poorly defined further aloft) varies both in magnitude and sign. The time variation in the pressure gradient is mainly responsible for the fact that $C_u$ is neither constant with height nor one-signed, contrary to an assigned fixed value. Moreover, $C_u$ is only positive in the early stages of the system evolution before the system is organised. Grubišić and Moncrieff (2000) also found a variation of the value of $C_u$ with height and though they found the GKI approximation to be accurate in the mid-levels, it lacked accuracy elsewhere. Grubišić and Moncrieff (2000) suggest that $C_u$ should be a function of height $z$, as it varies between levels. The results presented here support this result, but further suggest that the GKI
formulation is inadequate for organised systems.

Previously $C_u$, which is a tunable parameter, has been assigned various values depending on the model in which it is implemented. Values have ranged from 0.7 (Gregory et al. (1997) which is also used by UKMO (Stratton et al. (2009)) and an earlier version of NCAR’s Community Atmosphere Model (CAM) 3.0 (Richter and Rasch (2008)); 0.55 (Zhang and Wu (2003)); and recently, 0.4 in CAM 5.1 (Neale et al. (2010)). Over the 6-hour model run shown in Fig. 4.13(d) the average value of $C_u$ equals approximately -0.3 within the shear layer, which is smaller (in magnitude) and of the opposite sign to all of the above values. In the first two hours of the simulation, however, $C_u$ is approximately 0.5, which is consistent with the above values.

As noted by Neale et al. (2010), the magnitude of $C_u$ (and correspondingly $C_d$) has an important control on the strength of parameterised convective momentum transport as the pressure gradient is a ‘sink’ term; as $C$ increases the strength of the parameterised momentum transport decreases. However, as shown earlier (e.g., Fig. 4.6) the pressure gradient can actually act in concert with the up-/down draft fluxes, such as at low-levels in the early stages of the evolution of the convective system. Ultimately, this pressure gradient evolves with the convective system and as shown for this organised system is poorly represented by Eq. 4.2. This poor representation is because Eq. 4.2 does not take into account the up-shear directed pressure gradient associated with the elevated mesolow behind the leading edge of organised systems.

4.3.6 Evaluation of the Moncrieff analytic models of momentum transport

The Moncrieff analytic models of organised moist convection are based on the conservation of mass, entropy, total energy, vorticity and horizontal momentum generation for inviscid steady flow in Lagrangian space (Moncrieff (1981)). These nonlinear models are approximately exact because the sole assumption is that the buoyancy is a separable function of the vertical velocity, which is valid for moist adiabatic motion. These models have been comprehensively evaluated by cloud-system resolving model simulation data and field-campaign measurements (e.g. Houze (2004), Houze (2014), Moncrieff (2010)). In particular, the two-dimensional analytic models are solutions of the elliptic integro-differential
4.3. Results

Vorticity equation:

$$\nabla^2 \psi = G(\psi) + \int_{z_0}^{z} \left( \frac{\partial F}{\partial \psi} \right)' \, dz'$$  \hspace{1cm} (4.3)

where $\psi$ is the streamfunction, $G$ is the environmental shear and the integral is the vorticity generated by the horizontal gradient of buoyancy.

The far-field solution of Eq. 4.3 provides open lateral boundary conditions for the two-dimensional models. Analytic models of three-dimensional propagating squall lines (Moncrieff and Miller (1976)), and upshear/downshear propagating convective bands (Lane and Moncrieff (2015)/Moncrieff and Lane (2015)) are tractable only in the far-field. However, the Lagrangian formulation of the analytic models (see below for the two-dimensional models) enables the three-dimensional transports to be calculated from the far-field solutions.

The two-dimensional Moncrieff (1992) archetypal model (Fig. 4.14), the minimalist mesoscale system, is governed by the quotient of work done by the horizontal pressure gradient and the inflow kinetic energy, $E = \Delta p_s / \frac{1}{2} \rho U_s^2$, where $\Delta p_s$ and $U_s$ are the surface pressure change across the system and the surface inflow speed, respectively i.e., the ratio of the two terms of the Bernoulli pressure. With application to MCSs, squall lines, and density currents the archetypal models have a distinguished rearward-tilted circulation, and the mesoscale momentum transport has sign opposite to that of the propagation speed, e.g., momentum transport is negative when the propagation speed is positive, and vice versa.

The Lagrangian based two-dimensional analytic models enables the momentum transport to be derived from the far-field solutions of the vorticity equation. The two-dimensional archetypes is as follows. Referring to Fig. 4.14, the momentum transport is for the total relative flow $(u_m, w_m)$. Define the difference operator $\Delta = [\ ]^L_m$ and the horizontal averaging operator $< > = \frac{1}{L_m} \int_0^{L_m} ( \ ) dx$. Integration of the steady Eulerian momentum equation in the x-direction

$$\frac{\partial}{\partial x} \left( u_m^2 + \frac{p_m}{\rho} \right) + \frac{\partial}{\partial z} (u_m w_m) = 0$$  \hspace{1cm} (4.4)

gives

$$\frac{\partial}{\partial z} \langle u_m w_m \rangle = - \frac{1}{L_m} \Delta \left[ u_m^2 + \frac{p_m}{\rho} \right]_0$$  \hspace{1cm} (4.5)

In the above, $u_m$ and $w_m$ are normalised by $U_s$, and $\frac{p_m}{\rho}$ by $U_s^2$. The mesoscale momentum transport divergence is proportional to the cross-system change of
Bernoulli pressure and the aspect ratio ($\frac{1}{L_m}$) is the constant of proportionality. Integration of Eq. 4.5 with $w_m = 0$ at the horizontal lower boundary gives the mesoscale momentum transport at height $z$.

$$\langle u_m w_m \rangle = -\frac{1}{L_m} \int_0^z \Delta \left[ u_m^2 + \frac{p_m}{\rho} \right] dz'$$

(4.6)

i.e., the horizontal momentum is vertically redistributed but no net-generation of momentum can occur.

There are important physical and dynamical distinctions between the mesoscale momentum transport based on the Lagrangian analytic models and the traditional eddy-based momentum transport, where an eddy is defined as a deviation from the horizontal mean (see Eq. 3.4). In other words, the eddy flux is replaced by the mesoscale transport divergence where $u_m$ and $w_m$ are total (i.e., eddy + mean) flow components. Therefore, the mesoscale acceleration of the mean flow vertical profile is

$$\frac{\delta}{\delta t} \langle u_m \rangle = -\alpha_m \frac{\partial}{\partial z} \langle u_m w_m \rangle = \alpha_m \frac{1}{L_m} \left[ u_m^2 + \frac{p_m}{\rho} \right]_0$$

(4.7)

The $w_m = 0$ boundary condition at the horizontal upper boundary provides a powerful integral constraint on the mesoscale momentum transport. It follows that the Bernoulli pressure constrains the momentum tendency

$$\int_0^1 \Delta \left[ u_m^2 + \frac{p_m}{\rho} \right] L_m dz = 0$$

(4.8)

i.e., the horizontal momentum is redistributed but the net-momentum generation is zero.

There are important distinctions between the mesoscale momentum transport based on the analytic models and the traditional eddy-based momentum transport, where an eddy is defined as a deviation from the horizontal mean (see Eq. 3.4). In other words, the eddy flux is replaced by the total mesoscale transport divergence, i.e., eddy + mean components. Therefore, the mesoscale acceleration of the mean flow vertical profile is

$$\frac{\delta}{\delta t} \langle u_m \rangle = -\alpha_m \frac{\partial}{\partial z} \langle u_m w_m \rangle = \alpha_m \frac{1}{L_m} \left[ u_m^2 + \frac{p_m}{\rho} \right]_0$$

(4.9)
where $\alpha_m$ is an amplitude function or closure (Moncrieff (1992), Eq. 24). Fig. 13, showed that the momentum transport divergence for the archetypal model accelerates/decelerates the mean flow in the lower/upper troposphere, respectively.

The above formulae can be evaluated using the numerical simulations herein with open lateral boundary conditions. The terms in Eqn. 4.5 are represented in Fig. 4.15 for the medium open domain with 1 km grid spacing. It can be seen that the relationship between $\frac{\partial}{\partial z} \langle u_m w_m \rangle$ and the terms $-\frac{1}{L_m} \int_0^z \Delta \left[ \frac{u_m^2 + \rho_m}{\rho} \right] dz'$ holds true for the results from this WRF model simulation.

An important structural characteristic of the archetypal models is that the upshear tilt of the mesoscale circulation means that the sign of the mesoscale transport is opposite to the propagation vector, e.g., a system propagating in the positive x-direction has negative momentum transport. The archetypal model is controlled by $E = \frac{\Delta p}{\frac{1}{2} \rho U_s^2}$, the ratio of the two components of Bernoulli pressure at the lower boundary. Figure 2 of Moncrieff (1992) shows the morphology of the relative flow for systems in the range $-8 \leq E \leq \frac{8}{5}$. The lower and upper limits represent a system with no downdraft and no overturning updraft (strict propagation), respectively. Systems featuring deep downdrafts and deep overturning updrafts are associated with small absolute values of E. This structure pertains to the numerical simulations because $\Delta p$ is small and $U_s = -18$ ms$^{-1}$ so E is approximately zero. Vertical integration of Eq. 17 in Moncrieff (1992) gives the momentum transport for $E = 0$ as

$$\langle u_m w_m \rangle = \frac{4}{3} \left( \frac{z}{h} \right)^3 - 2 \left( \frac{z}{h} \right)^2$$

(4.10)

for $0 \leq z \leq h$, and provides values for $h \leq z \leq 1$ due to the symmetry of the profile. The archetypal model momentum transport profile shown in Fig. 4.16(a) resembles the transport calculated from the numerical simulation in Fig. 4.16(b).

In dimensional units the minimum value of $\langle u_m, w_m \rangle$ in the archetypal model is -13.6 m$^2$s$^{-2}$, using $L_m = 200/12 = 16.7$ and $U_s = -18$ ms$^{-1}$. This is in good agreement with -15.5 m$^2$s$^{-2}$ for the simulation results (Fig. 4.16(b)).
Fig. 4.14: a) Schematic description of the Moncrieff (1992) archetypal model based on the conservation of mass, total energy, vorticity in a Lagrangian framework, and the integral constraint on momentum transport. b) Numerical solution of the two-dimensional vorticity equation shows the characteristic rearward-tilt of the airflow. The dotted and broken lines denote free-boundaries whose shapes are part of the two-dimensional solution, subject to boundary conditions defined by the far-field solution. Fig. a) is from Moncrieff (1992).
Fig. 4.15: Terms in the relationship $\frac{\partial}{\partial z} \langle u_m w_m \rangle = -\frac{1}{L_m} \Delta \left[ u_m^2 + \frac{v_m^2}{\rho} \right]_{L_m}$ for the open lateral boundary conditions for medium domain. Averaged over 0400-0500 hours during the mature stage of evolution. Sum of the terms on the RHS are shown by the solid red line.
Fig. 4.16: Top - a) Profile of $\langle u_m w_m \rangle$ for archetypal model in Moncrieff (1992). Below - b) Profile of $u_m w_m$ from the numerical simulation using the medium open domain.
4.4 Synthesis and Conclusion

This study on the momentum budget of idealised convective systems focused on aspects relevant for the parameterisation of convective momentum transport. Specifically, there are important sensitivities of the simulated momentum budget to model domain size and resolution, as well as the contributions from convective-scale transports to the overall budgets.

Simulations with varied domain size revealed the key role of horizontal gradient terms in contributing to the momentum budget. In particular, even for relatively large domains (viz. the ‘medium’ domain that was 400 km long) the horizontal pressure gradient was a notable contributor to the mean flow tendency. This mesoscale pressure gradient arose entirely from the simulated convective system, acting to maintain the organised circulation associated with that system. This mesoscale pressure gradient mostly opposes the tendency from the convective momentum flux divergence, and for the ‘medium’ domain was about half the magnitude of the momentum flux tendency. Even for the ‘large’ domain (length = 800 km), the mesoscale pressure gradient was about one quarter the magnitude of the momentum flux term. This specific result has important implications for CMT parameterisation as these domain sizes are larger than most global model grid boxes; this mesoscale pressure gradient is not represented in any scheme, but it clearly plays an important role for mesoscale systems. Even when the work done by the static pressure term is identically zero; namely $E=0$, the system still has a distinguished vertical tilt (see Moncrieff (1992); Fig. 4) that provides mesoscale momentum transport.

The ‘small’ domain simulations considered here were not large enough for the convective systems to evolve and maintain themselves realistically. As shown in Fig. 4.12, the systems across all domains develop in a similar fashion for the first few hours. The pressure gradient and the structure of the convective system in these small domains are similar during the first two hours of the simulation (Fig. 4.6 and Fig. 4.10). However, when the domains are too small $\mathcal{O}(\leq 100$ kms), then the convection is suppressed by overly strong subsidence and is unable to become properly organised. This has important implications for the interpretation of momentum fluxes associated with convection in small-domain RCE simulations e.g., Romps (2012), as they are unable to properly represent organisation.

The effect of model resolution on the CMT showed the 3 km grid spacing model producing larger fluxes and more convective overshoots than higher reso-
Momentum transport in idealised simulations. This result was consistent with previous studies that attribute some of the sensitivities to insufficient entrainment at these convection-permitting resolutions. Though, at least for those simulations presented here, there seemed to be convergence of the convective-scale fluxes at grid spacings of 1 km.

Comparison of the convective-scale transports, i.e., those associated with individual updrafts and downdrafts, and those associated with the domain mean finds that in the early stages of evolution the convective-scales work against the mesoscales and result in a near-zero tendency. At mature stages, after the system tilts upshear, the convective-scale and mesoscale transports act in concert in a downgradient manner. As part of this evolution the low-level cross-draft pressure gradients change sign during the system evolution associated with the development of organised mesoscale circulations.

The most common parameterisations for convective momentum flux, which are based on the GKI scheme, use entraining plume models and incorporate a simple representation of the effects of the cross-updraft pressure gradient on the momentum tendency. As shown here, this representation poorly reproduces the evolution of the pressure gradient, as the assumed constant of proportionality between the pressure gradient and the product of the mean shear and the mass-flux actually changes sign. This difficulty arises due to the inability of such schemes to properly represent the processes associated with organised systems’ notably vertical tilt. This underlines the importance of an improved representation of the roles of mesoscale transports in CMT parameterisations, which should differ in their formulation from downgradient-based representations of the transport properties of unorganised plumes.

Finally, the findings from the medium domain idealised simulation are then used to evaluate Moncrieff’s mesoscale momentum transport models in which the mesoscale momentum transport divergence is determined to be proportional to the cross-system change of the Bernoulli pressure term (comprised of the surface pressure change and the surface inflow speed) and the aspect ratio. It is found that momentum transport profile from this archetypal model closely resembles that from the numerical simulation using the WRF model.
5 Momentum transport during TWP-ICE

5.1 Introduction

As discussed in the previous chapter, nearly all convective parameterisations and CMT parameterisations are based primarily on convective-scale dynamics. We have seen that mesoscale processes can act in a different manner to the convective-scales and consequently may play an important role in the overall momentum budget. As stated in Chapter 4, convection is multiscalar, being composed of dynamical processes that operate at many different spatial and temporal scales. This chapter continues this analysis into the different temporal and spatial scales involved in organised convection, with particular focus on the momentum transport.

The main aim of this chapter is to understand how the profiles of $\rho u'w'$, for both mesoscale and convective-scales, change as the systems develop and to examine the multiscalar nature of the transports using a real case simulation over five different horizontal resolutions. This is achieved by employing spectral-filtering to perform scale separation of the momentum transports, which identifies the circulations that are driving the momentum transport at given times and how they are represented on the different gridscales. The profiles from each of the 5 model runs are also compared to see how resolution affects the representation of the momentum transports. A four and a half day period is studied during the monsoon break from the Tropical Warm Pool - International Cloud Experiment (TWP-ICE).

To accomplish these aims, a series of model simulations of tropical convection are run, based on the TWP-ICE (Fig. 5.1). A description of the TWP-ICE campaign is found in Section 5.2 and the model configuration is described in Section 5.3. The results section details how the spectral-filtering is applied to the model data and is followed by an examination of the data to see how the transports change through time, as well as the effect of the horizontal resolution on
5.2 The TWP-ICE Campaign

The experiment took place over an area encompassing Darwin in the Northern Territories, Australia and occurred during January and February 2006, over the monsoon season (May et al. (2008)). During this period, the monsoon has active phases with widespread (but weak) convection from the larger-scale forcing. There are also break periods when the storms have a prominent diurnal cycle and are locally forced by convergence due to storm outflow and/or sea-breezes,
which can initiate squall lines (Keenan and Carbone (1992)). An example of this is shown in Fig. 5.2, where convective lines are initiated along the coastlines and are oriented in the same direction as the land/sea interface.

![Fig. 5.2: Radar reflectivity at 2.5 km height showing mature-stage squall line at 1705 UTC 09 Feb 2006.](image)

The monsoon and break regimes have different shear environments. Whereas the active phase of the monsoon is characterised by a westerly low-level shear, the break period’s dominant shear is the 3 km easterly wind maximum (Fig. 5.3) (Keenan and Carbone (1992)). Studies of the GARP (Global Atmospheric Research Program) Atlantic Tropical Experiment (GATE) by (LeMone and Zipser (1980), Szoke and Zipser (1986)) have shown that the continental convective systems associated with the break season have much greater mean updrafts and downdrafts than the oceanic systems, and therefore greater electrical activity (Rutledge et al. (1992)).
The dates studied, are 09 -13 February 2006, which are part of the monsoon break period, and this was chosen as it is characterised by localised, intense convective activity with a number of organised systems developing and dissipating within the immediate area. This phase of the monsoon is representative of diurnally forced continental convection, Fig. 5.4 shows a number of systems identified through precipitation during the monsoon break, which therefore enables a range of convective systems to be observed. It also enables a further comparison with observations to be conducted - this is the focus of Chapter 6. The WRF model has also been used to successfully simulate TWP-ICE during this particular period. Caine et al. (2013) found that WRF produced higher convective cells than observations using a 1.25 km resolution. This study also showed that using a low threshold to define the convective areas (25 dBZ) underestimated the size of the largest convective cells, and that increasing the resolution to 416.7 m helped to improve the representation of the intense convective cores. Wapler et al. (2010) also using a horizontal resolution of 1.25 km, demonstrated that the model repro...
duced realistic rain rates and rainfall distribution, providing that the mesoscale organisation is adequate. However, the simulated break period poorly produced the diurnal cycle due to the use of the parameterisation scheme for the outer domain which produced overly strong large-scale descent.

By examining these examples of convective activity over this break period, we can observe squall-lines such as those identified in the previous chapter, as these break conditions are conducive to the initiation and development of these organised systems, as well as an MCS known as Hector. This chapter analyses the multiscalar nature of momentum transport associated with these two distinct types of MCS.

A Hector is an isolated, short-lived, vigorous, convectively-driven thunderstorm which occurs on most days of the simulation over the Tiwi Islands (Fig. 5.5) approximately 50 km north of Darwin (for further discussion on the geographical region see Keenan et al. (1990)). This storm commonly occurs during the transition (November - December, February - March (Carbone et al. (2000)) and break seasons (which occur intermittently from mid-December to mid February (Zhu et al. (2013)), as a consequence of the diurnal heating (Keenan et al. (1990)). It evolves over a fairly flat area of terrain and is classified into two types by Carbone et al. (2000) with the forcing mechanisms deciding which type - either
sea-breeze fronts from the north and south coasts producing a type A, or type B, resulting from the interaction of a sea-breeze front and a gust front from existing convection (Gentile et al. (2014)). The lifecycle of Hector is relatively short (5-7 hours (Keenan and Rutledge (2000))) compared to that of the squall lines (lasting up to a day). The squall lines that occur during the break period are examples of continental convection and tend to develop during the afternoon and evening during the break season. They often form as isolated cells which merge, evolve and organise themselves into squall lines (Keenan and Carbone (1992)). Additionally, the squall lines that occurred during the break period formed over land and are possibly initiated on the ranges approximately 100 - 200 km SE of Darwin.
5.3 The numerical model

5.3.1 Model configuration

This simulation of tropical convection uses the Advanced Research Weather Research and Forecasting Model (ARW-WRF) v.3.1.1 (Skamarock et al. (2008)). It is initiated on 8 February at 1200 UTC and runs until 13 February 1200 UTC 2006, with the first 12 hours discarded as spinup. The simulation has five domains, with an outer domain and four one-way nested domains. The locations and extent of these domains are shown in Fig. 5.6. The horizontal grid spacings are 33.75, 11.75, 3.75, 1.25 kms with the smallest domain having a grid spacing of 416.7 m. A summary of these domains is shown in Table 5.1. Each model domain has 64 vertical (sigma) levels with a model top of 10 hPa and the top 5 km features a sponge layer to absorb reflected gravity waves.

Fig. 5.6: The 5 domains used for the real case study of TWP-ICE.
The model runs include the Lin (Purdue) (Lin et al. (1983)) ice microphysics scheme (water, cloud, ice, rain, snow and hail). Other runs in Chapter 6 use the Thompson scheme (Thompson et al. (2008)). Longwave radiation is represented via the Rapid Radiative Transfer Model (RRTM) (Mlawer et al. (1994)) and the shortwave by the Goddard shortwave radiation scheme (Chou and Suarez (1994)). Land surface physics use the Noah scheme (Chen and Dudhia (2001)) and boundary layer physics employs the Mellor-Yamada-Janjić scheme (Mellor and Yamada (1982)). Domains 1 & 2 use the Betts-Miller-Janjić cumulus parameterization scheme (Betts and Miller (1986), Janjić (1994)), whereas the three higher resolutions domains treat convection explicitly as they are convection-permitting (Δx, Δy < 4 km). The largest domain is initialised using initial & boundary conditions from ECMWF ERA-Interim reanalysis data. For further information about the model configuration and background, see Caine et al. (2013), who also compared model data to radar observations and found errors in the depth and size of the convective cells, with simulated cells attaining a greater height than observations at horizontal resolutions > 1.25 km, possibly due to a lack of explicit entrainment (Bryan and Morrison (2012)). Wu et al. (2009) found that using the Lin scheme overestimates graupel, leading to larger rain rates than observations.
5.3. Scale separation

To gain an insight in to the dynamics and momentum transport of the systems at different spatial scales, a Fourier Transform is applied to the perturbations from the domain-averaged wind i.e., to the $u'$, $v'$ and $w'$ components. For the Hector and squall line, the averaging to define the perturbations occurs across the trimmed domains shown in Fig. 5.1. As with the previous chapter, variables were interpolated to height levels before analysis.

The $u'$, $v'$, and $w'$ variables were passed through the Fourier Transform function horizontally, and an ideal filter applied. Filtering separates certain frequencies within a signal (Bowman (2006)). The ideal filter used here is a step function (i.e. uses values of either 0 or 1) in wavenumber which therefore passes the selected range of wavenumbers unchanged and ignores the rest. An example of this is shown in Fig. 5.7, where the selected wavenumbers are contained within the RHS of the graph, where the filter has a value of 1. This is known as filtering in the spectral domain. This high-pass filter identifies the high wavenumbers produced by the FFT, with the filtered output passed through an inverse FFT. The resulting values are the convective signal of the variable (as the convective scales are the large wavenumbers). This process is then repeated for each height level.

![Figure 5.7](image)

**Fig. 5.7:** Example for a high-pass band filter, which uses a step function to pass selected wavenumbers through.
Here a threshold of 20 km is chosen to separate the convective scales from the mesoscales. This scale is appropriate based on standard definition of “mesoscales”, albeit admittedly somewhat arbitrary. Nonetheless, tests with a filter cut-off of 10 kms and 15 kms produced similar results (not shown).

The $u$ and $w$ velocity components can therefore be represented as

$$u' = u'_c + u'_m \quad w' = w'_c + w'_m$$

(5.1)

the momentum flux is therefore,

$$u'w' = u'_c w'_c + u'_c w'_m + u'_m w'_c + u'_m w'_m$$

(5.2)

where $c$ and $m$ are the convective scale and mesoscale components respectively, and the cross-correlated terms represent the interaction between the two scales. However, the mean of these cross-correlation terms is zero as the separation is completed in spectral space and the individual components are orthogonal. The resulting non-zero terms are:

$$u'_c w'_c = u'_c \ast w'_c \quad u'_m w'_m = u'_m \ast w'_m$$

(5.3)

$$u'w' = u'_c w'_c + u'_m w'_m$$

(5.4)

So for each vertical level, the total momentum transport $\rho u'w'$ is represented as a convective and mesoscale $\rho u'w'$ for each gridpoint, which is then averaged across the relevant domain at each height level. The same method of scale separation and calculation of momentum transport also applies to $\rho v'w'$.

5.4 Results

The inner-most domain was divided into two smaller sub-domains one that incorporated just the mainland and parts of the Timor Sea/Arafura Sea for the squall lines and a sub-domain that was centred over the Tiwi Islands for the Hector (Fig. 5.8). It is useful to consider these two distinct convective systems, as they have different morphologies and mesoscale organisation.

Although simulated Hector events occurred almost daily, with a number of squall lines also passing through this domain over this period, only one example of each was chosen for detailed analysis. The selected Hector occurred at 0200 - 1000 UTC on 10 February 2006 and the selected squall line at 0600 - 1600 on 09
**Fig. 5.8:** The inner domain d05 (dashed line) with the two sub-domains for the analysis of the Hectors and the squall-lines.

February 2006.

Figure 5.9 shows the outgoing long wave radiation (OLR) during Hector for domain d05 at 0320 UTC on 10 February 2006, with Fig. 5.10 indicating the wind profile across the whole inner (d05) domain. The mean wind profile is typical for this time of year as is shown by the vertical profile of the wind in the 315° direction (south-easterly wind) which shows a maximum around the 3 km level. Figure 5.11 shows two cross-sections with (a) showing the updrafts in the centre of the system and (b) shows the negligible tilt of the system which is nearly symmetric, with no strong preferred direction of orientation.

By contrast, the squall line is shown in Fig. 5.12 with its distinctive convective line approaching the coastline from the south-east. As it propagates and develops, it tilts backwards and this is shown by the cross-section in Fig. 5.13. This is therefore an upshear-tilted system similar to those discussed previously. Studies by Smull and Houze (1987), Lafore et al. (1988), LeMone and Moncrieff (1992) and LeMone and Moncrieff (1994) all found mesoscale transports to be the opposite sign to the shear, with the fully-developed system demonstrating an upshear tilt, however the magnitudes were different due to various observational and simulation datasets.
Fig. 5.9: Outgoing long wave radiation (white contours) over the Tiwi Islands (solid line) near Darwin during a Hector at 0320 UTC on the 10 Feb 2006 (domain d05). The rain mixing ratio shows (dotted) contours every 0.5 g kg$^{-1}$.

Fig. 5.10: Mean wind averaged across whole domain throughout the monsoon break period. a) U wind component, b) V wind component, c) wind in the 315° direction and d) wind direction using cardinal points.
Fig. 5.11: Cross-section of Hector (solid line) at 0330 UTC on the 10 Feb 2006 (domain d04) - (a) contours show the values of \( w \) and (b) \( u'w' \), with positive values shown by solid contours and negative by dashed.
Fig. 5.12: Outgoing long wave radiation (white contours) near Darwin during of squall line at 1030 UTC on the 09 Feb 2006 (domain d05). The rain mixing ratio shows (dotted) contours every 0.5 g kg$^{-1}$. Location of cross-section in Fig. 5.13 is shown by the red line.

Fig. 5.13: Cross-section of squall line during TWP-ICE monsoon break period at 1030 UTC 09 FEB 2006 from domain d05 Solid line shows the cloud 0.01 g kg$^{-1}$ contour. Black lines are $\rho u'w'$ (solid, positive values and dashed, negative).
5.4.1 Convective and Mesoscale Contributions

When calculating $\rho u'w'$, $u'$ is the component of wind in $315^\circ$-$135^\circ$ direction as the squall lines propagate in this direction which is aligned with the shear. Figure 5.10 shows the shear is acting in the south-easterly direction (towards the $315^\circ$ direction) which is typical of the monsoon break environment. Therefore, $315^\circ$ is in the negative $u'$ direction and $135^\circ$ is in the positive. The $v'$ wind component is calculated along the $225^\circ$-$45^\circ$ line. We use the components, $u'135^\circ$ and $v'45^\circ$, to represent the across-line and the along-line structures, respectively. Positive values of $\rho u'w'$ therefore indicate that the system has a backwards tilted system (upshear) and a negative sign refers to a downshear tilt with the corresponding transports being downgradient and countergradient, respectively.

Firstly, to examine the entire domain over the whole break period, $\rho u'w'$ was separated into convective and mesoscale components and averaged over the 4$\frac{1}{2}$ day period. Figure 5.14 shows the contributions from the mesoscale and convective-scale momentum transports are approximately equal. Although the convective-scale momentum transport occurs over a smaller horizontal area, the updrafts cause much stronger upward motion and fluxes than the mesoscale transports. It is important to realise that even though the transports have similar profiles, the nature of these scales are profoundly different. The convective scale is influenced more heavily by the vertical velocity wind component $w$ due to the updraft, whereas the mesoscale is influenced to a greater extent by the horizontal wind component and its degree of tilt. There is also the fact that convective “disorganised” events occur more often than organised systems (Caine et al. (2013) identify approximately 70% from observations); these would have the characteristic of a convective signal with a weak mesoscale component. The momentum fluxes along the convective line (notated $\rho v'w'$) are weaker (Fig. 5.14, right) with the mesoscale fluxes having about half the magnitude of $\rho u'w'$ over the same period and this is why $\rho u'w'$ is of interest.

The momentum transport from Hector is then analysed by averaging the values of $\rho u'w'$ over the smaller ‘Hector’ domain throughout the system’s lifecycle. Figure 5.15 shows that the convective scale transports are greater than the mesoscale around 3 km. This is due to the strong convective ‘burst’ that Hector produces in the mature phase and its powerful updrafts. Although the spatial extent of Hector is mesoscale, it has a strong convective signature; this is due to the sea breeze convergence over the Tiwi Islands causing strong convective updrafts.
Fig. 5.14: Momentum transport separated into convective scale and mesoscale components for $\rho u'w'$ (left) and $\rho v'w'$ (right), averaged over the whole d05 domain during the break period.
and that the system does not organise itself into a sustained, long-lived mesoscale circulation. The fairly symmetrical nature of this thunderstorm also negates a tilt in a specific direction and so the mesoscale transports partially cancel each other out (i.e., the more upright a system, the lower the net magnitude of $\rho u'w'$). Therefore a system with no tilt has $\rho u'w' = 0$ by definition.

Fig. 5.15: Momentum transport separated into convective scale and mesoscale components for $\rho u'w'$ (left) and $\rho v'w'$ (right), averaged over the Hector d05 domain from 0200 - 1000 UTC 10 FEB 2006.

In contrast, the mesoscale transports of the squall line (Fig. 5.16) dominate those at the smaller, convective scale. This is due to the organised flows from the mature mesoscale system, which develop in a manner similar to that described for a TS system in Section 2.1.3 with the associated inherent structures. This system (and all of the systems produced by the model) propagates towards the north-west and is tilted upshear when fully-developed. This orientation of the
mesoscale system accounts for the overall sign of $\rho u'w'$. It would be expected that the convective transports would be much smaller as they tend to be upright, however the fact that they have a positive sign occurs because they are embedded within the larger mesoscale circulation. They are also both positive as the mature phase of the squall line endures for a longer time and it should be noted that these transports could possibly change sign at different stages of the lifecycle. Referring back to Fig. 5.10, the low-level wind shear is negative in the $(315° - 135°)$ $u$ direction and therefore the transports are downgradient.

\begin{center}
\includegraphics[width=\textwidth]{Fig5.16.png}
\end{center}

\textbf{Fig. 5.16:} Momentum transport separated into convective scale and mesoscale components for $\rho u'w'$ (left) and $\rho v'w'$ (right), averaged over the squall line d05 domain from 0600 - 1600 UTC 09 FEB 2006.
5.4.2 Time Evolution

The scale-separation is then applied to different stages in each of the two convective system’s lifecycle to ascertain the evolution of the contributions from the various scales. Figure 5.17 shows these at two-hourly stages in Hector’s evolution. The momentum transport values are domain-averaged at each level, and the average values for each two hour period is shown. Figure 5.18 shows the structure of the cloud at each time period, with contours of $\rho u'w'$. During the early stage of development, the convective transports dominate the mesoscale, and grow stronger during the 2 - 4 hour time, which is due to the development of a cold pool (Fig. 5.19). Both then die away as the system starts to decay with the mesoscale enduring longer (til approximately 0700 UTC) with a negative sign at the top of the shear layer as the environmental wind exerts it influence over the system as it dissipates from 0800 - 1000 UTC.

Fig. 5.17: Scale-separated momentum transport during Hector 0200 - 1000 UTC 10 FEB 2006. The different coloured lines show the time-averaged values for $\rho u'w'$ for various stages of the system’s lifecycle.
Fig. 5.18: Zonal cross-section through the centre of Hector and its associated cold pool - defined by the $\leq -1^\circ$C temperature perturbation at 0300 (top left), 0500 (top right), 0700 (below left), and 0900 (below right) UTC on 10 FEB 2006 from domain d04. The solid black contour represents the 0.01 g kg$^{-1}$ cloud mixing ratio. The other contours are $\rho u'w'$ at 5.0 N m$^{-2}$ levels for 0500 UTC and 0.5 N m$^{-2}$ for all other times.

The squall line however, with its more organised circulations, has a stronger mesoscale signal (Fig. 5.20). In the early stages Fig. 5.21 shows initial convective-scale development entering the domain around 0800 UTC and there is no obvious organisation apart from convective scale updrafts as shown by the contours of rain. The squall line is partially developed as it enters the south-east corner of the domain, and this also explains the significant convective scale contributions in the first few hours (Fig. 5.22). The mesoscale overwhelms the convective scales during the mature phase (2 - 6 hours); as convection develops, the system organises and the mesoscale features (such as convective cells, the formation of the jump updraft, the forward anvil and a trailing stratiform region) evolve (Fig. 5.13). As the system becomes organised (after $\sim$ 2 hours), the mesoscale circulations are established and this is reflected in the dominant mesoscale transport profile. In the decaying stages, both transports are weakly negative in the low-level shear layer as the circulations decay, the wind perturbations diminish and are replaced by the environmental wind.
Fig. 5.19: Zonal cross-section through the centre of Hector and its associated cold pool - defined by the ≤−1°C temperature perturbation - at 0330 UTC on 10 FEB 2006 from domain d04. The solid black contour represents the 0.01 g kg\(^{-1}\) cloud mixing ratio. Contours are \(w\) every 0.5 ms\(^{-1}\), solid is positive and dashed are negative.
In summary, there are a number of differences in the characteristics of the two systems studied. The transports from Hector are smaller overall and shows predominantly convective-scale characteristics. The squall line conversely, has a mesoscale signature as the larger scale momentum transports dominate the overall momentum transport. Scale separation is therefore a useful technique in understanding the dynamics of two different forms of convection. Using this method to analyse the evolution of the storms at these two different scales reveals that the underlying circulations are well-represented and improves our understanding of the dynamical structure.

**Fig. 5.20:** Scale-separated momentum transport during a squall line 0600 - 1600 UTC 09 FEB 2006. The different coloured lines show the time-averaged values for $\rho u'w'$ for various stages of the system’s lifecycle.
Fig. 5.21: Outgoing Longwave Radiation during the lifecycle of the squall line every two hours from 0600 - 1600 UTC 09 FEB 2006. The rain mixing ratio shows (dotted) contours every 0.5 g kg\(^{-1}\). The red lines indicate the locations of the cross-sections in Fig. 5.22.
Fig. 5.22: Cross-section of squall line during developing (0800 UTC), mature (1000 UTC) and decay (1200 UTC) stages of life cycle on 09 FEB 2006 from domain d05. The solid black contour represents the 0.01 g kg$^{-1}$ cloud mixing ratio. The other contours are $\rho u'u'$ at 5.0 N m$^{-2}$ levels for 1000 UTC and 0.5 N m$^{-2}$ for all other times.
5.4.3 Resolution Sensitivity

The previous analyses were all carried out using a 416.7 m resolution domain (d05). The effects of resolution on the CMT are now investigated by comparing the resolved transports from each domain. Each of the larger domains were “trimmed” to the size of domain d05 and the two sub-domains containing Hector, which centres on the Tiwi Islands and secondly the mainland area south of Darwin containing the land-based squall-lines which decay rapidly near the coast. These datasets were also interpolated to the gridpoints from the highest resolution domain which affords a direct comparison to the previous results (i.e., the analysis code is identical for each domain). Moreover, as the multidomain structure is one-way nested, this facilitated this direct comparison. Of course, differences in convective systems are intertwined with differences in fluxes.

Across the entire monsoon break period studied, the magnitudes of the convective and mesoscale transports converge to a similar contribution of 0.04 Nm$^{-2}$ at the top of the shear layer (Fig. 5.23). The convective scale is not properly represented by grid spacing above $\sim 4$ km. It is also of interest that the 3.75 km domain produces total CMT which is around the same magnitude as that from the 1.25 km and 416.7 m domains, due to the overestimation of the mesoscale and underestimation of the convective-scale transport.

The momentum transport from all the models are compared for Hector, and Fig. 5.24 shows both scale transports for all 5 domains. The convective scale transports are not represented by the 2 coarsest model runs and barely by domain d03 (3.75 km resolution) i.e., the 20 km cut-off is only $6\Delta x$. This is because the convective-scale is too small for the models to resolve dynamic processes at this level. For the two smallest domains (d04 and d05), the convective-scale is well-represented with stronger transports at the highest resolution. The mesoscale transports are captured in some respects by the all simulations. This illustrates why Hector is not adequately resolved by domains d01 and d02, as these coarsest resolution models do not sufficiently resolve convection produced by these MCSs.

The squall line during its mature phase (at 1000 UTC, 4 hours into its lifecycle) is investigated Fig. 5.25. Each domain is shown from the coarsest resolution at the top down to the 417 m at the bottom (e). The white contouring represents OLR (i.e., the cloud), the solid line is the coastline (0.01m height) and the dotted lines show the areas of convective updrafts. Each domain is trimmed to the inner domain and once again the data is interpolated to this domain. As expected
the convective areas are not resolved by the two most coarse domains, and only weakly at 3.75 km (hence convection permitting is determined to be defined as grid spacing < 4 kms (Prein et al. (2015))). There is also a slower propagation at greater spacing; part of this could be due to the lack of development of the convective updrafts which would therefore inhibit the development of the convective system, the cold pool and the organisation of the flows, momentum transport and ultimately the upscale growth. Coherent flows and mesoscale organisation rely on the balance between environmental shear, evaporative cooling, which generates and sustains the cold pool. This is also maintained by adequate latent heating supplied by the microphysics scheme. However at lower resolutions, the downdrafts are weaker and this explains the lack of organisation with increasing grid spacing (Moncrieff and Liu (2006)).
Fig. 5.24: Scale-separated momentum transport $\rho u'w'$ during a Hector 0200 - 1000 UTC 10 FEB 2006, for all model domains.
Fig. 5.25: Comparison of the squall line (shown by outgoing longwave radiation) at all resolutions at 1000 UTC 9 FEB 2006. The rain mixing ratio shows (dotted) contours every 0.5 g kg$^{-1}$. 
Figure 5.26 shows that for the squall line, the three highest resolution domains represent the convective-scale with the 1.25 km and 417 m resolution models converging, although Bryan et al. (2003) have suggested that for convergence to occur, models would have to have resolutions $O(100 \text{ m})$. The mesoscale transports also appear to converge at 1.25 km grid spacing as domains d04 and d05 have almost identical profiles. The domain d03 (3.75 km grid spacing) has the strongest mesoscale transports, because at those scales the minimum resolvable scale is such that all explicit circulations are classed as “mesoscale”. It is also of interest that this resolution also produces total CMT is similar to the smallest domains due to the overestimation of mesoscale and underestimation of convective-scale.

![Scale-separated momentum transport during a squall line 0600 - 1600 UTC 09 FEB 2006, for all model domains](image)

**Fig. 5.26:** Scale-separated momentum transport during a squall line 0600 - 1600 UTC 09 FEB 2006, for all model domains

This figure also suggests that 1.25 km resolution is sufficient for representing the transports at both scales for a squall line and also the dynamics and organised flows of the typical trailing stratiform MCS. It is generally accepted that to
resolve mesoscale convection, a *maximum* grid spacing of around 1 km is needed (Weisman et al. (1997)). This might not completely resolve all the convective elements within organised systems, but does allow them to develop in a realistic manner (Moncrieff (1992), Caine et al. (2013)). Domain d01 produces the wrong sign for mesoscale transports at the mid (5 km height) and upper (8 - 10 km height) levels and this is analysed further by a cross-section.

Figure 5.27 shows the north-south averaged $\rho u'w'$ for the 33.75 km (top) and the 1.25 km grid spacing (below) for the squall line. The coarser resolution simulation produces weak, negative momentum transport in the mid-levels of the troposphere, which would indicate a downshear tilt with weak organisation. The 1.25 km model in contrast represents an upshear-tilted system with strong tilted circulations that indicate a high degree of organisation. This result is very similar to that from Moncrieff and Liu (2006) (their Fig. 10). The consequence is that coarse resolution models such as climate models etc, can produce the wrong sign of momentum transport and this has implications for the effects of the convection on the tendency of the mean wind and on the strength of convection and the longevity of the MCS itself as well as upscale growth, which can be strengthened or diminished by the transports. As convective - but no CMT - parameterisation was used in the largest domain, it shows that without the parameterisation of CMT specifically, the momentum transport that being represented is not correct compared to the domains where the momentum flux is being explicitly resolved.

### 5.5 Synthesis and conclusions

Using the Fourier Transform to apply scale separation to the wind perturbation components has shown itself to be a useful tool in identifying the multiscale characteristics of two convective systems which also have distinct structures due to their individual morphology. The two systems evolve quite differently, with the Hector having a mostly upright orientation due to the sea-breeze convergence from all sides of the Tiwi Islands and strong convective-scale transports (therefore with less of a mesoscale signature), revealed by the scale separation. A comparison of the main mesoscale circulations identified from the cross-sections reveals that Hector has mostly symmetrical structures, whereas the squall line has those similar to Moncrieff’s archetypal triple-branch model (Fig. 5.28).

Fig. 5.27: Cross-sections showing the sign of momentum transport in domains d01 (top) and d04 (below) - red contours are positive values and blue are negative. The averaged value across the domain for each level are shown on the right of each cross-section. Note: the scales are different for both domains, with d04 having 100 times the magnitude.
Fig. 5.28: Comparison of the main mesoscale circulations identified in Hector - 0330 UTC 10 Feb 2006 (top) and the squall line - 1030 UTC 09 Feb 2006 (below). Contours in the top diagram show $\rho u'w'$ whereas the bottom diagram shows contours of $w$ - (solid, positive values and dashed, negative).
The squall line conversely, has a definitive backwards tilt, is orientated in an upshear direction driven by strong convection in the initial phase, which then as the flow becomes organised, is sustained by the mesoscale circulation. The means of evolution, due to the traditional development of the cold pool and the flows which are driven by the balance between the evaporative cooling and the wind shear, are expressed through investigation of the dynamical processes at these two scales. Although both of these systems are classified as MCSs, they are quite different in how they develop and are sustained. The method used to investigate their dynamics using the momentum transport profiles at each stage, allows for easier identification and interpretation of the type of system we are looking at without the need to necessarily characterise the main flows and cross-sections. This technique also highlights how convection is simulated by the model at different grid spacings.

One of the main findings is that the coarsest model does not resolve either mesoscale or convective-scale transports which infers that MCSs are not produced in a meaningful way by the 33.75 km model. At this resolution, models produce the wrong sign of momentum transport, resulting in a short-lived system with a downshear tilt. This system does not produce the organised flow of a typical trailing stratiform convective system and can also impact on the environmental flow. Obviously at this resolution, precipitation and surface winds would not be accurately portrayed by this model as it does not allow for realistic systems to be explicitly resolved.

Another key result is that at around 10 km grid spacing, the mesoscale circulations are resolved consistent with previous studies (e.g., Stephens et al. (2008)), however they are over-emphasised. The implications for this finding is that models run at approximately this horizontal resolution may produce organised systems with a greater upshear tilt than is realistic, though much of this may be offset by under-resolved and weaker convective transports. This means that the transports, which in this case are downgradient, thereby reducing the shear (mixing) too strongly and would likely shorten the lifecycle of the system although a direct comparison is hard to make visually as the systems are offset in timings (with the d03 system travelling slower than d05 as it approaches the coast).

A consequence of convective-scale transports for the 11.25 km domain being too weak, is that they are obviously crucial in the development of the organisation of the mesoscale circulations. The strength of the convection is important for the height of the jump updraft and this is fundamental to the morphology of the
triple-branch model as well as the pressure field (and obviously, the momentum transport), as these all determine the convective regime. This result confirms that regional models, for example, which run at similar horizontal resolutions (i.e. around 10 kms), do not resolve the convection elements and do not therefore have sufficient grid spacing to properly resolve the cold-pool. Mesoscale circulations are also enhanced by the subgridscale heating which produces MCS-type circulations, unlike those from coarser resolution models (Zhang et al. (1989), Gao et al. (1990), Moncrieff (1992)) which do not.

Convergence of both scale transports occurs at 1.25 km resolution for the squall line, however Hector is comprised of smaller-scale processes. The squall line convergence indicates that this resolution of ~ 1 km is sufficient to accurately resolve transports associated with a convective system. Therefore this model is able to produce both the convective-scale processes and the mesoscale organisation associated with a squall line. The mesoscale systems produced by the 416.7 m and 1.25 km grid simulations both have similar structures as well as, almost identical lifecycles and propagation speeds.

Convective scales are not represented in coarse resolution domains (i.e >10 km) and is much weaker in the 3.75 km domain; this has implications for the initial stages of the formation of an organised system, as the mesoscale is only weakly represented for both and so for coarse resolution models (10 kms or more). Even with convective parameterisation, the coherent circulation is not represented by these models. The mesoscale momentum transports provide a significant contribution to the total momentum budget of the MCS circulations, which suggests that even if parameterisation of CMT is employed, it raises the question of whether the mesoscale transports would be appropriately represented by GCMs and global models.
6 Momentum Transport from Radar during TWP-ICE

Thus far in this study, the research has solely relied on the data obtained from simulations using the WRF model, with the highest resolution domains explicitly resolving the smaller-scale dynamical processes in a presumably realistic manner. Although, as noted in Section 4.4, caution must be taken, as turbulent processes which can affect entrainment and detrainment need horizontal grid spacing of $O(100\text{m})$, and therefore resolutions coarser than this, will not necessarily produce realistic results. The availability of observational data during the TWP-ICE sampling period is therefore extremely useful as it allows us to compare the systems produced by the model, with those identified by observations. Traditionally, it has been difficult to obtain explicit measurements of momentum flux at sufficient horizontal and temporal resolution, along with difficulties in measuring horizontal and vertical velocities simultaneously. In this chapter, a new dataset is examined that uses dual-Doppler retrievals to determine the three-dimensional velocities, and hence characterise the kinematics, including CMT characteristics, of observed systems.

6.1 Introduction

Work by LeMone (1983) used observational data during the GARP Atlantic Tropical Experiment (GATE), which took place near the west African coast in 1974. Repeated aircraft passes across the system measured the three wind components, $u$, $v$, and $w$ and these were used to calculate the momentum flux both across and along the convective line. In another study, LeMone et al. (1984) also used the same dataset to analyse the mesoscale circulations within the system. They found that the vertical flux of horizontal momentum normal to the line may transport in either an upgradient or downgradient manner whilst parallel to the convective line.
the transport is downgradient in all observed cases. One limitation of such data is that it is spatially limited as there are only measurements where the aircraft makes its passes and as such there is not a complete dataset encompassing the entire convective system at a given time period.

This thesis has only used data from the WRF model thus far and, it would be a useful exercise to determine whether the types of convective system that have been studied in the previous chapter, (i.e., trailing stratiform, upshear-tilted MCSs), are similar to those observed systems during TWP-ICE. Unfortunately, it is not possible to obtain a full 3D dataset of wind data as measurements are either sparsely recorded across a given area or concentrated over a short time period. Typically, observations are obtained from various sources, such as automatic weather station (AWS) sites and wind profilers (which rely on the convective system passing directly overhead), a single Doppler radar, and individual radiosondes. This research phase uses a radar dataset encompassing large areas of 3D wind data during the TWP-ICE campaign (described in the previous chapter), which incorporates the area around Darwin. This dataset uses both directly recorded horizontal wind from a dual-Doppler analysis using the radars in the Darwin area located in the Northern Territory, Australia (Collis et al. (2013)) as well as derived vertical wind components.

It is the aim of this chapter to analyse this new radar dataset to determine the observed systems that cross the Australian mainland during this sampling period, enabling a comparison between the similarities in the dynamics of the systems from both the model and the observations. This would also confirm the veracity of our assumption that the high resolution model runs are realistic in terms of their organisational characteristics and transports. Conversely, any systems that are observed, but not represented in the model will be investigated to determine why they are not being reproduced in WRF, providing a comparison between observed and simulated events.

This is a new exploration of the ability to estimate the CMT from 3D-radar derived winds using this dataset. To achieve these aims, the events that occurred in the same four day period simulated in the previous chapter, are identified and their momentum transport profiles calculated from the 3D wind dataset. Following this, a couple of these events are then chosen for further analysis depending on their type. A system is identified from the radar data that is different to the typical upshear-tilted (leading convective) trailing stratiform convective systems produced by the real case in WRF (Chapter 5). This observed system is then
analysed to determine its kinematics during different stages of its evolution.

This chapter begins with a description of the radar used for data collection and a brief explanation of the method used to calculate the vertical wind motions. It is then followed by a description of the observed systems during the sampling period and a brief investigation into the vertical winds from both the observations and the model. Finally an analysis of the wind profiles and the thermodynamics - relative humidity and temperature - is carried out, using both observations and WRF to clarify the reasons for the differences in observed and simulated characteristics.

6.1.1 The Doppler radar dataset

The data was compiled using two scanning Doppler weather radars which record the motion of hydrometeors over a large area, with each radar measuring the radial wind component from the precipitation within this volume (Collis et al. (2013)). This method of retrieval provides a less direct measurement than either in situ measurements from aircraft passing through a storm or those from vertically pointing profiling radars but they can capture the entire mesoscale system passing over the local area and allow for a direct comparison with model output. The data is collected from an operational Bureau of Meteorology Doppler radar located at Berrimah, Darwin and a C-band polarimetric (C-POL) research radar (Keenan et al. (1998)) at Gunn Point (Fig. 6.1). These two radars conducted multipass variational wind retrieval and performed a volumetric scan every 10 minutes. Each scan required 8 minutes to perform all plan position indicator (PPI) scans. PPI scans horizontally and varies the azimuth angle (compass direction), while holding the elevation angle constant. The radar reflectivity measurements cover the domain in Fig. 6.1 completely.

Unlike convectional radar which locates hydrometeor targets in its beam and measures their reflectivity, Doppler radar also measures the radial velocity of those targets as well. Obviously, with a single radar, the transverse component is unknown, meaning that the total velocity and therefore the direction are both unknown. However, when two Doppler radar are used together, both the horizontal speed and horizontal direction can be determined. The unambiguous wind data is found within areas known as dual Doppler lobes and these data lie in an area bounded by the angle of intersection between the two beams of 30° and 150°.

The location of these lobes used to compile the TWP-ICE dataset are shown
shaded in red in Fig. 6.1. The dataset which contains both reflectivity as well as 3D wind data that was used for the analysis is provided by Alain Protat from the Bureau of Meteorology and used his method for deriving the vertical velocities, described fully in Protat and Zawadzki (1999). One possible error source is the movement of the MCS during the time taken to complete a full weather volume (in this case 8 minutes). Therefore to reduce errors, two successive scans are then used and a linear interpolation of the data is undertaken. For example if there are two scans, one on the hour and one at ten minutes past the hour, the resulting interpolated file will be representative of 5 minutes past the hour.

The vertical wind is also calculated using the anelastic continuity equation
\[
\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{1}{\rho} \frac{\partial \rho w}{\partial x} = 0,
\]
which assumes that density does not vary in horizontal space or in time, only in the vertical. The density is calculated using the relationship between density and height,
\[
\rho(z) = \rho(z=0) \times e^{-z/H},
\]
where \(H\) is the air density scale height. The value for \(H\) is set at 8000 m, which is the mean value obtained from the Darwin soundings (sensitivity tests showed that this value could be used throughout the sampling period).

**Fig. 6.1:** Location of the two Doppler radar - the CPOL radar at Gunn Point and that at Berrimah - near Darwin during TWP-ICE. The dual Doppler lobes are indicated in red. Note that the reflectivity dataset fully covers this domain. Adapted from Collis et al. (2013).
6.1. Introduction

The vertical wind component is calculated by integration of the airmass continuity equation, in both directions (integrating upwards and also downwards) from the ground assuming $w = 0$ at the ground and at the storm top (the upper-level boundary). Finally the weighted average of both resulting calculations of $w$ are calculated in order to minimise the errors.

In order to calculate the precipitation rate from the radar data, an algorithm derived by Bringi et al. (2001) was used. This algorithm was chosen as it has been calibrated using data from the Gunn Point radar. The 2.5 km height reflectivity data is converted to the rain rate in mm hr$^{-1}$ using the relationship $Z = 305R^{1.36}$, where $Z$ is the absolute reflectivity and is calculated by converting the reflectivity datasets (a dimensionless quantity with units of dBZ) using the formula $dBZ = 10 \log_{10}Z$.

The main dataset has a domain covering 130.103°E to 131.871°E and 11.4392°S to 13.1717°S with a horizontal resolution of 2.5 km and a vertical resolution of 500 m. Note the focus is on the squall-line events as this domain does not cover the Tiwi Islands and therefore, comparison to the “Hector” events in the previous chapter, cannot be done. Therefore the focus is on the squall-line events over the mainland. This dataset encompasses the whole four day sampling period used for the WRF simulations in the previous chapter. A second dataset is used to analyse more closely the dynamics of one of the systems; this data has a finer horizontal resolution of 1.5 km.

To calculate the perturbation values of the winds, the domain average value for each height level are obtained from the Atmospheric Radiation Measurement (ARM) datasets during the TWP-ICE campaign. This work uses the Constrained Variational Objective Analysis Data (Zhang and Lin (1997), Zhang et al. (2001)) to ascertain the background wind ($u$, $v$, and $w$). This is achieved by analysis of soundings of wind, temperature and water vapour measurements with minimal adjustment of these variables to conserve mass, moisture, energy and momentum. This dataset has a time resolution of every 3 hours, a vertical resolution of 10 mb averaged over the TWP-ICE Campaign area (Fig. 5.1), which results in a single wind profile for these wind components as well as relative humidity, which is used for further analysis of the cold pool (Section 6.2.5).
6.2  Results

This sections outlines how the individual events that occurred during the monsoon break period are identified and why they were chosen. These events are identified using space-time data that covers the entire sampling period from 09 - 13 February 2006. These convective systems are then studied in more detail by looking at the reflectivity data at a height of 2.5 km across the whole domain to identify more clearly when they form, how they evolve and decay, as well as their dynamical characteristics.

6.2.1  Identification of events

To identify the various convective events that pass through the domain during the break, a Hovmöller diagram of reflectivity is produced across the four day period from 0000 UTC on 09 Feb 2006 until 0000 UTC 13 Feb 2006 (Fig. 6.2). The column maximum reflectivity across the domain is shown, and the events are chosen by recognising a continuous strip of reflectivity passing from the right boundary (i.e., the east) to the left (west), each representing an organised long-lived system passing through the domain. Six coherent events are observed during this period and Table 6.1 summarises the approximate start and finish time (also duration) of each of these identified occurrences of organised convection.

The background wind profiles for each event are compared in Fig. 6.3. These profiles are obtained from the ARM dataset and are averaged over the time periods in Table 6.1. They show that for all the events, the values of $u$ are negative and mostly positive for $v$. The momentum flux (per unit mass) profiles ($u'w'$ and $v'w'$) are calculated from the 3D wind data within the Doppler lobes and are shown in Fig. 6.4. These profiles show that the first two events have negative values of $u'w'$ throughout the lower to mid-levels (below 8 km), which, when compared to Fig. 6.3, suggest that there is a downshear tilt to each of these two systems. These are of particular interest as the wind shear in the lower levels is negative and this suggests that these systems could be oriented downshear. This mode of MCS is in contrast to the upshear-tilted trailing stratiform systems, which are the only type of squall line identified using the WRF model dataset (described in Chapter 5).
Fig. 6.2: Hovmöller diagram showing maximum 2.5 km reflectivity during the monsoon break period during TWP-ICE, 09 FEB - 13 FEB 2006.
Fig. 6.3: Mean wind profiles of convective events averaged over the four and a half day period from the ARM (top) dataset. The black lines represent the average zonal wind \( u \) and the blue lines, the meridional wind \( v \). Each line style represents an event shown in Fig. 6.2 and is averaged over the duration of each system, as defined in Table 6.1.
Fig. 6.4: Profiles of $u'w'$ (left) and $v'w'$ (right) for all events from 2.5 km Doppler radar dataset during TWP-ICE 09 - 13 FEB 2006.
A second WRF model simulation was run (following the Lin dataset used in the previous chapter) using the Thompson microphysics scheme. This allows a comparison with the Lin scheme as well as the observations. Using the WRF data, five squall line events were identified, two from the Lin scheme and three from the Thompson scheme. A summary of 3-hour mature period of each event in both simulations is shown in Table 6.2; these periods are used for subsequent analyses. An example of each system is shown using reflectivity data derived from the WRF model both for the Lin simulation (Fig. 6.5) and for Thompson (Fig. 6.6).

<table>
<thead>
<tr>
<th>Event #</th>
<th>Start Time</th>
<th>Finish Time</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0910 UTC 09 February 2006</td>
<td>1210 UTC 09 February 2006</td>
</tr>
<tr>
<td>2</td>
<td>1620 UTC 10 February 2006</td>
<td>1920 UTC 10 February 2006</td>
</tr>
<tr>
<td>1</td>
<td>0830 UTC 09 February 2006</td>
<td>1130 UTC 09 February 2006</td>
</tr>
<tr>
<td>2</td>
<td>2330 UTC 09 February 2006</td>
<td>0330 UTC 10 February 2006</td>
</tr>
<tr>
<td>3</td>
<td>0230 UTC 11 February 2006</td>
<td>0530 UTC 11 February 2006</td>
</tr>
</tbody>
</table>
Fig. 6.5: Reflectivity at 2.5 km height, during mature phase of the two events in WRF using the Lin microphysics scheme. In the lower diagram, a Hector can be seen developing over the Tiwi Islands.
Fig. 6.6: Reflectivity at 2.5 km height, during mature phase of the three events in WRF using the Thompson microphysics scheme.
Profiles of $u'w'$ from the radar data are then compared to the 1.25 km WRF simulations detailed above. The WRF model produces similar values for $u'w'$ and much weaker values for $v'w'$ at this 1.25 km resolution, compared to those fluxes identified from the radar (Fig. 6.7). Another interesting feature is the limited range of values throughout the profiles produced from WRF. This model produces positive values (indicating an upshear tilt) for all events, throughout all levels, whereas for Fig. 6.7, even for the upshear-tilted systems, negative values are found in the shear layer (consistent with the background wind shear). However it must be remembered that the winds are averaged over the Doppler lobes, which is a smaller area than those from the values from the model which are averaged over the cloud area, and as a result, the values would be higher in the radar CMT profiles than the model.

![Fig. 6.7: Profiles of $u'w'$ (left) and $v'w'$ (right) for all events from WRF real dataset during TWP-ICE 09 - 13 FEB 2006 for the d04 - 1.25 km domain.](image)
When deciding which specific events to investigate in detail, each event was then investigated further using the 2.5 km height reflectivity data to see their overall structure, how they evolved, and how they became organised. An example of each system during the mature phase of their lifecycle is shown in Fig. 6.8.

Events 1 and 3 are chosen for further examination as these systems pass directly through the lower Doppler lobe. As a result, there should be more wind data available to analyse their development over as long a time period as possible, which provides a thorough analysis of the structure of each system during their mature phase. Moreover, Event 1 is chosen as it has a negative value for $u'w'$ and Event 3 as it has a positive value and both of these have coherent mesoscale structures as seen from the reflectivity, whereas many of the other systems are comprised of numerous cells developing and merging which complicates any analysis.

### 6.2.2 Vertical wind

Comparing the maximum and minimum vertical wind velocities from the two events chosen from the radar data and those from the WRF model (Fig. 6.9), it can be seen that the derived vertical winds from the radar dataset are weaker than those from the WRF model. The domain maximum values of $w$ from the Doppler radar are mainly between $5 - 10 \text{ m s}^{-1}$ throughout, which is approximately one order of magnitude lower than those from WRF. The radar-derived minimum values are also 2 - 3 times smaller than WRF.

Although it is acknowledged that the vertical winds derived from the radar dataset are weaker than observations from other sources such as profilers and soundings (Protat and Zawadzki (1999)), it should also be noted however that the WRF model may also be producing stronger than realistic values. Varble et al. (2014) suggests the reason that the model produces these strong intense updrafts could be due to the parameterised microphysics, as fine resolution models produce more realistic vertical velocities. Therefore the magnitudes of the momentum transport must not be taken as accurate, although the sign of the transport would still be correct.
Fig. 6.8: All events during mature phase of lifecycle. Coloured contours show reflectivity from radar dataset at a height of 2.5 km. Wind vectors are shown, with reference vector on LHS of each diagram.
Fig. 6.9: Time series of maximum (top) and minimum (below) vertical wind velocities for the 2.5 km resolution dual-Doppler radar dataset and the Lin and Thompson WRF model runs over a three-hour period during the mature stage of each system.
6.2.3 Event 3: Upshear-tilted trailing stratiform convective system

Event 3 is analysed using vertical cross-sections across the convective line (i.e. normal to the direction of propagation) to identify the mesoscale circulations in this system. Figure 6.10 shows two such cross-sections, both of which show system-relative vectors; the top diagram shows the cross-section through the centre of the system (along a latitude of 12.65°S, shown in Fig. 6.11), while the bottom shows an averaged cross-section of approximately 22 km width. This system propagates in a westward direction with the convective line oriented in a north-south line. The values of the vertical wind component $w$ have been exaggerated (by a factor of 15) in order to determine the direction of the circulations more clearly. It can be seen that there is a front-to-rear flow (from left-to-right) similar to a jump updraft and a downdraft at the rear of the system, with overturning at the front of the system (seen more clearly in the bottom diagram); these are summarised in Fig. 6.12. The mesoscale circulations seen in this system are similar to the standard model described by Moncrieff (1992), in which the typical system has an upshear tilt and front-to-rear flow.

Figure 6.11 also shows a horizontal plot of reflectivity at the same time at a height of 2.5 km. The main system enters the domain from the east and as it approaches the coastline, it eventually merges with another convective system to the west. Unfortunately, this larger system is outside the range of the Doppler lobes and the wind data does not occur in or near the convective line/core. The tilt and the observable dynamics of this MCS concur with the momentum transport profile which suggests an upshear-tilted system with a positive sign of momentum transport.

The Steiner classification (Steiner et al. (1995)) is used to separate convective and stratiform parts of the MCS (Fig. 6.13). The dark blue regions indicate convective areas, mid-blue is stratiform, and the palest blue is out-of-cloud (in other words, unclassified areas of low reflectivity). This classification applies an algorithm to radar data to separate the radar echoes into convective and stratiform regions. This is based on the method by Rosenfeld et al. (1995) which searches for stratiform precipitation using the bright band signature to determine the type of precipitation. These areas are defined by the intensity of the echo and the peaks. This separation of convective and stratiform areas was originally developed by Steiner and Houze (1993) and is a modification of an earlier paper by Churchill and
Houze (1984), who formulated a method for separating the types of precipitation.

The convective areas are of a similar size to those indicated by the higher reflectivity in (Fig. 6.11) - i.e. > 40 dBZ - this is because part of the definition of convective precipitation is for areas with a reflectivity of > 40 dBZ to be counted as a convective centre. The stratiform areas defined by the Steiner classification are much smaller however, than those shown by reflectivity. This is due to the different definition of stratiform cloud by the Steiner classification, which is based on air motion and precipitation as well as the bright band, whereas Fig. 6.11 only uses a threshold of 10 dBZ. As a consequence, light rain from warm stratiform clouds may not be counted within the Steiner stratiform classification and may therefore account for any such discrepancies between the figures. The Steiner classification is more comprehensive and takes into account the behaviour of the
Fig. 6.11: Event 3 during mature phase at 1215 UTC 010 FEB 2006. Vectors shown are u and v in the Doppler lobe regions. The asterisks indicate the location of the two radar and the red line indicates the cross-section shown in Fig. 6.10.

microphysics as well as the reflectivity. Finally, the pale blues areas, which are areas of very low reflectivity, negligible vertical motion and low precipitation are therefore classified as out-of-cloud, compared to Fig. 6.11 which uses a definition of < 10 dBZ.
Fig. 6.12: The main mesoscale flows identified from cross-section of wind data for Event 3 at 1215 UTC 10 Feb 2006.

Fig. 6.13: Event 3 during mature phase at 1215 UTC on 10 Feb 2006; convective areas (dark blue) and stratiform areas (mid blue) are shown using Steiner classification. The pale blue is out-of-cloud and white areas are missing data.
6.2.4 Event 1: Downshear-tilted convective system

Event 1 is shown by the reflectivity data at a height of 2.5 km during its mature phase (Fig. 6.14) at 0715 UTC on 09 February 2006. This system travels in approximately a westward direction, where it merges with other smaller regions of convection to grow and then eventually dissipate near the coastline. From this figure, it seems that this system could possible be a parallel stratiform system as described in Section 2.1.3, the stratiform area of the system is positioned to the right of the convective region.

![Figure 6.14](image1)

**Fig. 6.14:** Event 1 during mature phase at 0715 UTC on 09 February 2006. Vectors shown are $u$ and $v$ in the Doppler lobe regions. The asterisks indicate the location of the two radar and the red line indicates the cross-section shown in Fig. 6.16

Again, the system is represented by areas of convection and stratiform, as defined by the Steiner classification (Fig. 6.15). The convective areas tally well with the reflectivity (Fig. 6.14) however, due to a different method of classifying
stratiform areas, the stratiform areas in mid blue are smaller. It appears from both these diagrams that this system could still be a parallel stratiform convective system or possibly a hybrid of a parallel and a leading stratiform convective system. In order to understand the mesoscale circulations, cross-sections through this MCS are investigated, to determine the type of system, and the dynamics of the MCS (i.e., the mesoscale circulations that define and maintain it).

**Fig. 6.15**: Event 1 during mature phase at 0715 UTC on 09 February 2006; convective areas (dark blue) and stratiform areas (mid blue) are shown using Steiner classification. The pale blue is out-of-cloud and white areas are missing data.

Two cross-sections are shown in Fig. 6.16; the top diagram is a transect of reflectivity along the 12.9°S line of latitude at 0715 UTC on 09 February 2016, the diagram below is a 0.2° latitude (approximately 22 km), meridionally-averaged section centred on 12.9°S. The reason that this line was chosen for the cross-section is because the system is propagating in an almost westward direction with the convective line oriented in a north-south direction in the early stages changing to a WNW-ESE direction in the latter stages. Therefore this transect captured the cross-convective line dynamics most clearly as it is virtually perpendicular to the convective line. Only a part of this cross-section is shown in order to see the circulations in more detail.

The wind vectors show an updraft occurring; air is travelling towards the west.
(from right to left) in the mid-levels, around 5 km height with the jump occurring at approximately 131.58°E. This is similar to the jump updrafts investigated in the previous sections, however this jump is a mirror image of those other flows and had a rear-to-front flow rather than the typical front-to-rear seen earlier i.e. the jump is orientated in the same direction as the trailing convective systems in the previous section (6.2.3). There is also an overturning branch located at the rear (RHS) of the system and a downdraft at the front at low-levels. Figure 6.17 shows a representation of the main mesoscale circulations identified from Fig. 6.16. This configuration of mesoscale circulations has not been observed in the TWP-ICE model simulations and points to a different species of MCS to the trailing stratiform systems typically present with the WRF squall lines. In the mid-latitudes, such a system would be considered uncommon (Parker and Johnson (2000)), however there is no consensus on how frequent such systems are in tropical regions.

Another cross-section of the same convective system along the same degree of latitude at 0715 UTC on 09 February 2006 is shown in Fig. 6.18, using wind data, but with a higher horizontal resolution of 1.5 km. The top left diagram shows the reflectivity at a height of 2.5 km with the blue horizontal line showing the position of the cross-section. The top right diagram shows the reflectivity along the 12.9°S line of latitude, and the wind vectors are system-relative u and w. The \( w' \) value multiplied by a factor of 10 to show the updrafts more clearly, this is less than the 2.5 km resolution, as the 1.5 km dataset has improved values of \( w \). Unfortunately, the system is in a slightly different position to that shown in the lower resolution dataset. The lower diagram in Fig. 6.18 is north-south averaged reflectivity and wind across a cross-section with a width of 0.2° latitude.

The dynamics of this system, although at first glance seem to be different from those shown in Fig. 6.16, however they are in fact quite similar. The updraft is still there, however it is irregular and deviates from its path slightly; i.e., it has a slight ‘s’-shape as it moves from rear-to-front and deviates slightly backwards before continuing towards the front of the system. The reason for this is that the higher resolution would resolve more small-scale dynamics and would likely produce this flow. The easterly mid-level flow ascends in the convective updraft at a longitude of approximately 131.63°E. So the overall path of the circulation is the same as that seen in Fig. 6.16. The overturning branch at the rear of the MCS however is missing. The lower diagram shows the overturning branch located at the front of the system as evident from the 2.5 km data.
Fig. 6.16: Event 1 during mature phase at 0715 UTC 09 FEB 2006. Top cross section is along the 12.9°S line of latitude using the 2.5 km resolution radar dataset. System-relative wind vectors of $u$ and $w$ (x 15) are shown.

Fig. 6.17: The main mesoscale flows identified from cross-section of wind data for Event 1 at 0715 UTC 09 Feb 2006. Note: this is almost a mirror-image of Fig. 6.12.
Fig. 6.18: Event 1 during mature phase at 0715 UTC 09 FEB 2006 using higher resolution dual-Doppler data with a horizontal resolution of 1.5 km. All wind vectors are system-relative. (Top left) - map of the area and reflectivity at a height of 2.5 km. The red line indicates the cross-section in the top right diagram (along 12.9°S line of latitude). (Top right) - the wind vectors are $u \times 10$. (Bottom) Averaged reflectivity (meridionally) through the cross section of 0.2° latitude.
6.2.5 Comparison to the WRF model results

The downshear-tilted system that has been identified using the radar dataset is different to those that have been observed in both the idealised and real case studies using the WRF model. All the systems produced by WRF for the TWP-ICE period (see Chapter 5) are leading convective, trailing stratiform (LCTS), whereas the observations show at least one downshear system, over the somewhat limited four day period, so it is hard to draw any conclusions as to how frequent these are. The reason why WRF is not producing this type of system is now investigated.

As discussed in Section 2.1.3, the direction that an organised system tilts is influenced by the balance between the environmental wind shear and the vorticity generated by the cold pool. Therefore, the reason the model is not simulating a downshear-tilted system could be because either the wind shear is weaker in the model or the strength of the cold pool is too strong, (which ‘pulls’ the system into an upshear-tilted orientation) or a combination of the two.

The wind profiles are compared first to see whether there are any differences between the wind shear in the model and the observational ARM data during TWP-ICE. As we do not have direct observational measurements of the cold pool (i.e. mesoscale temperature variations), the potential differences in the cold pools (between WRF and radar) are compared using other variables. The relative humidity is evaluated as a lower relative humidity implies a greater potential evaporation which effects cold pool generation. The precipitation intensity is also investigated to determine whether the model could produce a stronger cold pool due to more potential evaporation.

Work by Caine et al. (2013) has indicated that one of the main differences between the two microphysics schemes is that Thompson scheme produces smaller rain rates than the Lin simulation and this is probably due to the differences in the amounts of graupel and snow at higher levels. The Thompson simulation produces a much smaller flux of graupel, and greater amounts of snow than Lin; which due to graupel’s higher terminal velocity, means that Thompson would precipitate less water as it would retain a greater mass of hydrometeors aloft. The WRF model produces much stronger updrafts than seen in the observations, possible due to a lack of entrainment, thereby transporting an excess of water aloft. This could be due to stronger cold pools over a smaller area, which would generate a greater updraft and more intense convective cores; this will be addressed in (c).
### (a) Wind Shear

Figure 6.3 shows the observed wind profiles during three-hour periods during each of the events for the ARM datasets. Although they change throughout the sampling period, the profiles are very similar with the low-level jet located at the expected 3 km height for $u$, and is the typical wind profile during the break period. Figure 6.19 has a stronger low-level wind shear for both microphysics schemes, which would be more conducive to the development of a downshear-tilted system, suggesting the wind shear is likely not the reason for the lack of downshear-tilted systems and suggests that the strength of the cold pool is the determining factor why the WRF model runs have preferentially produced the upshear systems. Consequently, the thermodynamic processes of the model could be responsible, rather than the background wind, and are now investigated.

Fig. 6.19: Time-averaged wind profiles of convective events from the Lin (left) and Thompson (right) datasets at 416.7 m resolution, averaged across the TWP-ICE domain. The black lines represent the average zonal wind $u$ and the blue lines, the meridional wind $v$. Each line represents a separate event.
(b) **Relative Humidity**

The relative humidity during each MCS from the model and from the ARM data are compared using a time series of low-level relative humidity (Fig. 6.20). “Low-level” is defined as 3 km height as this is the height of the sub-cloud layer (shown for each WRF system in Figs. 6.21 and 6.22), and therefore the lowest 3 km has been used to calculate the mean relative humidity. The red boxes indicate the periods of convective activity. The lower the relative humidity, the less saturated the environmental air and therefore the more likely that evaporation of precipitation will occur, thereby cooling the area and producing a stronger cold pool. The ARM data shows that the relative humidity during the events ranges from 80% to 90%, whereas the models shows values ranging from 75% - 85% for the Lin scheme and 60% - 85% for the Thompson scheme. These results indicate that on average during events, the model boundary layer is drier, which could lead to greater evaporation and stronger cold pools than those observed.

![Fig. 6.20](image_url)

**Fig. 6.20:** Time series of low-level relative humidity obtained by (a) the Lin and (b) the Thompson microphysics schemes in WRF and (c) observational data from ARM, during the monsoon break period - 09 FEB - 13 FEB 2006. Red boxes indicate the convective events that occurred during the period studied.
Fig. 6.21: Cross-section of cold pools from the Lin simulation. The blue shading represents the -1.0 °C temperature perturbation and below. The cloud outline is the 0.1 g kg⁻¹ cloud mixing ratio contour.
Fig. 6.22: Cross-section of cold pools from the Thompson simulation. The blue shading represents the -1.0 °C temperature perturbation and below. The cloud outline is the 0.1 g kg$^{-1}$ cloud mixing ratio contour.
(c) Precipitation and Reflectivity

The precipitation and reflectivity of both are compared to determine whether the cold pool is the likely reason for the WRF model preferentially producing upshear-systems. A time-series (Fig. 6.23) shows the reflectivity at 2.5 km height during the three-hour mature phase of each of the organised systems that passed through the domain during the four and a half day sampling period. The top diagram in Fig. 6.23 shows the domain-averaged reflectivity at a height of 2.5 km. The lower diagram shows the maximum reflectivity across the domain at the 2.5 km level. Although the radar has stronger reflectivity overall (as the averaged reflectivity is higher compared to the WRF model), the model produces much higher maximum reflectivity. These maximum values would be located in the region of the convective updrafts, suggesting the model produces more intense systems than the observations. As the average value is less for WRF, this indicates that the WRF systems encompass a smaller stratiform area.

![Fig. 6.23](image)

_Fig. 6.23:_ (top) Domain-averaged reflectivity during 3-hour mature phase of all the events in WRF and from Doppler at 2.5 km height with maximum reflectivity for each event (below). The time axis represents the three hours of the mature phase of each system.
The rain rate averaged across the domain is shown in Fig. 6.24; this is for non-zero dBZ reflectivity gridpoints. The radar observations show a much lower rain rate than the model data for all events. When compared to Fig. 6.23, it is clear that although the radar has a greater domain-average reflectivity, the non-zero rain rate is much lower. This lower rain rate suggests that the convective area of the systems identified from the radar are much weaker than WRF, especially as the WRF model has greater maximum values for the reflectivity. Figure 6.24 also suggests that the radar encompasses a larger area, as the domain-average reflectivity is large but the non-zero rain is much less so therefore the radar events are larger, weaker systems that those identified in the model.

Fig. 6.24: Domain-mean non-zero rain rate in mm hr$^{-1}$ from Lin, Thompson and Radar data. The time axis represents the three hours of the mature phase of each system.

These findings indicate the intensity of each of the systems is different in observed and simulated systems. It is likely that the WRF events would have stronger cold pools, as the large rain rates should produce more evaporation and stronger cooling. It is therefore important to compare the size and intensity of the systems.
6.2. Results

Assessing the systems qualitatively, Fig. 6.5 and Fig. 6.6 shows all of the systems at maturity. At first glance, the WRF systems are the same size or smaller than Event 1 from the radar (Fig. 6.14), whilst the radar event 3 is much larger than any of those from WRF. The other feature that is evident is the larger, more intense convective cores from both the Thompson and Lin microphysics than those identified from the radar reflectivity.

To see quantitatively, whether the systems have different sizes, Fig. 6.25 shows the areas covered by stratiform (defined as being 25 - 45 dBZ) and convective (defined as > 45 dBZ) precipitation and total (> 0 dBZ). These diagrams show that although the radar encompasses smaller areas for stratiform and for convective, the total area is much greater which indicates and very large area of weak reflectivity (> 0 to < 25 dBZ). This proves that the rainfall is spread over a much larger area than WRF with large, sprawling regions of very weak/light rain and that the two radar systems are similar in magnitude to each other.

Fig. 6.25: Comparison of reflectivity during a three hour period of the mature stage for various events. The reflectivity is separated into stratiform (25 - 45 dBZ)(top), convective (> 45 dBZ)(middle) and total area (> 0 dBZ)(bottom).
When the reflectivity is averaged over the size of the system itself rather than the entire domain, as with the rainrate, Figure 6.26 shows the reflectivity averaged for gridpoints with reflectivity greater than 0 dBZ and it is clear that once the reflectivity over the convective system is calculated, then the WRF model has stronger systems than those observed. Figure 6.21 shows a cross section through the two Lin systems with the cold pool (shown by the blue contours showing the temperature perturbation at each level) below the cloud. The top figure shows a clearly defined cold pool below the cloud with the head of the cold pool in the area of the hydraulic jump of the front-to-rear flow (as shown in Fig. 5.13).

\[Fig. 6.26: \text{Reflectivity averaged across convective area (i.e. } > 0 \text{ dBZ)}\]

The cold pools from the Thompson scheme are shown in Fig. 6.22. The middle figure has the strongest cold pool which seems to precede the main system and is in the proximity of where the convective cells are generated. The strength of this cold pool could be the reason for the definite upshear tilt to the system. The other figures have slightly weaker cold pools but again have the typical tilt and outline of a trailing stratiform system.

The respective strengths of the cold pool and the wind shear are compared, to gain an understanding of which way the convective system will tilt, with the assumption that the minimum buoyancy varies linearly within the cold pool. This
6.3 Synthesis and conclusions

The use of this unique dataset of 3D winds and reflectivity has revealed a system that has not been identified thus far in the WRF model simulations of TWP-ICE. This method of producing wind data has been extremely useful as a tool for identifying the dynamics of a real convective system. Also, using momentum transport profile as an initial analysis has been shown to be a very useful way to identify the type of convective system. To date we have been using high definition model simulations as observations. This downshear-tilted system has shown that the model is not revealing the true diversity of systems and also allows for evaluation of the model as well.
The momentum flux calculated from the radar data is of similar magnitude to that realised by the WRF model simulations, however the the radar shows a wider variety of values, as the early events (1 & 2) produce negative fluxes in the lower to mid-levels whereas WRF produces only positive values throughout the vertical levels. Even for the upshear-tilted systems identified by the radar data, the bottom most levels in the shear layer produce negative flux, consistent with the wind shear at that time. Although the magnitude of the values is similar between the model and the observations, with the radar dataset producing fluxes that are twice as strong as those represented by WRF at a height of 4 - 6 km. Overall, the WRF model seems to be producing realistic fluxes.

The WRF model’s microphysics schemes are over-producing rain and the WRF environment is too dry, both of which likely creates cold pools that are too strong. The systems that are resolved by the two microphysics schemes are also much smaller with stronger reflectivity in the location of the convective updrafts. These are all ingredients that appear to cause the model to preferentially generate upshear-tilted systems. These MCSs therefore have more intense convective cores producing stronger cold pools, and it is this that hinders the model from representing downshear-tilted systems. Although these downshear tilted systems may not be common in reality, they still represent part of the diverse range of species of systems (Bluestein and Jain (1985), Blanchard (1990), Houze et al. (1990), Parker and Johnson (2000)).

Another difference between the radar and WRF data is that the land-based squall lines produced by the model tend to propagate in a north-west direction (in approximately the same direction of as background wind) and typically dissipate near the coast, whereas the radar reflectivity shows that many of the systems propagated in a more westward direction and continued to exist well over the Arafura Sea. There also appears to be smaller regions of convection which are initiated near the coastline during events 1 and 3, which could be due to convergent wind along the coastal area; these are not apparent in either the Lin or the Thompson WRF model simulations. This work shows that these radar observations, although limited in (horizontal) area, can give useful insights to the types of systems, and can help improve models in turn, as by studying the range of organised convection, we can ensure that these are reproduced by the models.
7 Conclusions and recommendations for future work

7.1 Summary of results

The aim of this project was to better understand the characteristics of momentum transport during all stages of the lifecycle of organised deep convective systems. A number of idealised three-dimensional squall-lines simulations were conducted using a numerical model to investigate the momentum transport associated with deep or tropical convection. These simulations revealed the importance of the mesoscale processes on the momentum budget. A real case simulation was then analysed using a scale separation technique to identify the relative contributions from convective-scale and mesoscale momentum transport and to identify their sensitivity to horizontal resolution. The largest horizontal resolution domain model produced the wrong sign of momentum transport. This thesis concluded with a comparison of the real case with a dataset derived from Doppler radar measurements over Darwin, which identified a different type of MCS to those produced by the model in the first two parts of this study. These investigations sought to answer the following questions:

A. Which terms are most influential to the momentum transport profile and do these findings affect the way that momentum transport is represented in a parameterisation scheme?

B. What are the relative contributions of mesoscale and convective transports in organised convective systems?

C. Does observational data from radar reveal differences in convective momentum transport characteristics in real systems to those identified in models?
A summary of the main findings from this thesis is presented below.

**Main findings from idealised simulations**

1. Different domain sizes have revealed the role of horizontal gradient terms in the overall contribution to the momentum budget. Even for large domains ($\sim 400$ km) the horizontal mesoscale pressure gradient significantly contributed to the budget.

2. Small domains are not large enough to allow convective systems to evolve and sustain themselves in a realistic way. The pressure gradients and convective structures are similar in the early stages for all cyclic model runs, thereby suggesting that the domain size does not on its own, restrict the development of the systems. The mean vertical velocity also suggests that the convection is suppressed by strong subsidence.

3. Comparison of the different horizontal resolutions found larger fluxes from the 3 km model. This was due to a lack of turbulent processes being resolved and a lack of entrainment in the model. Convergence occurred around 1 km for convective-scale fluxes.

4. Convective-scale transports oppose mesoscale transport in the early stages; when the mesoscale transport acts in a countergradient manner, the convective transports are downgradient. At later times, they both act in a downgradient manner.

5. The Gregory, Kershaw and Inness (GKI) parameterisation is based on an entraining plume model and represents cross-updraft pressure gradient on the momentum tendency. The relationship between the pressure gradient, the shear, and mass flux changes sign and this is not taken into consideration using the tunable parameter $C_u$ which only allows for a negative pressure gradient, seen only during the initial stages of the system’s development. However the pressure gradient reverses over time, possibly due to the appearance of a mesolow which develops behind the convective line and this change of sign is not possible with the existing relationship in the GKI scheme.

6. The findings from the idealised simulations are then used to evaluate Moncrieff’s mesoscale momentum transport models in which the mesoscale mo-
mentum transport divergence is proportional to the cross-system change of the Bernoulli pressure term and the aspect ratio. It is found that the mesoscale momentum transport (and therefore the tilt of the convective system) are accurately represented by this archetypal model and closely match the findings from the WRF model.

Main findings from real simulations

1. One of the main findings from the real case is that the coarsest model does not resolve either mesoscale or convective-scale transports which infers that MCSs are not produced in a meaningful way by the 33.75 km grid spacing. At this resolution, the model produces the wrong sign of momentum transport, resulting in a short-lived system with a downshear tilt. This system does not produce the organised flow of a typical trailing stratiform convective system and can also impact on the environmental flow. This result is similar to that from Moncrieff and Liu (2006). Obviously at this resolution, precipitation and surface winds would not be accurately portrayed by this model as it does not allow for realistic systems to be explicitly resolved.

2. At around 10 km grid spacing, the mesoscale circulations are over-emphasised, producing organised systems with a greater upshear tilt than at higher resolutions. Therefore the downgradient transports reduce the shear and there is overly strong mixing occurring which would likely shorten the lifecycle of the system. The lack of convective-scale transports, which are obviously crucial in the development of the organisation of the mesoscale circulations, affect the height of the jump updraft and the morphology of the triple-branch model, as well as the pressure field. This indicates that regional models, for example, which have similar size meshes do not resolve the convection elements and do not therefore have sufficient grid spacing to resolve the cold-pool properly.

3. The squall line shows convergence at 1.25 km resolution which means that this model is able to produce both the convective-scale processes and the mesoscale organisation of a squall line. The systems produced by the 417 m and 1.25 km grid simulations both have similar structures as well as almost identical lifecycles and propagation speeds.

4. Convective scales are not well resolved in the 3.75 km resolution model and not at all, for grid spacing > 10 kms. The mesoscale is only weakly
represented in the coarsest resolution models (grid spacing 10 kms or more),
and this shows that even with convective parameterisation, the coherent
circulation is not represented by these models.

Main findings from radar

1. This unique dataset of 3D winds and reflectivity has revealed a system that
has not been identified thus far in the WRF model simulations. This method
of producing wind data has been extremely useful as a tool for identifying
the dynamics of a real convective system. Also, using momentum transport
profile as an initial analysis has been shown to be a very useful way to
identify the type of convective system.

2. The momentum flux calculated from the radar data is of similar magnitude
to that realised by the WRF model simulations, however the the radar shows
a wider variety of values, as some events produce negative fluxes in the
lower to mid-levels whereas WRF produces only positive values throughout
the vertical levels. Even for the upshear-tilted systems identified by the
radar data, the bottom most levels in the shear layer produce negative flux,
consistent with the wind shear at that time. Although the magnitude of the
values is similar between the model and the observations, with the radar
dataset producing fluxes that are twice as strong as those represented by
WRF at a height of 4 - 6 km. Overall, the WRF model seems to be producing
realistic fluxes.

3. The WRF model’s microphysics schemes are over-producing rain and the
WRF boundary layer environment is too dry which likely creates cold pools
which are too strong. The systems that are resolved by the two microphysics
schemes are also much smaller with stronger reflectivity in the location of the
convective updrafts. These are all ingredients that appear to cause the model
to preferentially generate upshear-tilted systems. These MCSs are shown to
have larger convective cores with higher reflectivity than those observed by
Doppler, which also provide higher rain rates and stronger cold pools; it
is this that hinders the model from representing downshear-tilted systems.
Although these downshear-tilted systems are not common in reality, they
still represent part of the diverse range of species of systems.

4. Another difference between the radar and WRF data is that the land-based
squall lines produced by the model tend to propagate in a north-west di-
retraction (in approximately the same direction of as background wind) and typically die out near the coast, whereas the radar reflectivity shows that many of the systems propagated in a more westward direction and continued to exist well over the Arafura Sea.

5. It is clear the performance of the WRF model is sensitive to the choice of setup - such as domain configuration, physics parameterisation and forcing data - this should be understood when attempting to reproduce a realistic range of convective systems.

7.2 Future work

The findings from this study have generated suggestions for a number of future investigations.

1. Mesoscale pressure gradients need to be represented in the parametrisation schemes; in the GKI scheme, the parameter $C_u$ should be a function of height $z$ as we have shown that it only has realistic values in the shear layer. But also this work has shown that as a convective system develops, the small-scale pressure gradients change sign and therefore the parameter needs to also vary with height and time, to represent this change due to organised circulations.

2. Continuing to study mesoscale transports by using the method used to separate the mesoscale and convective scale contributions, as the mesoscale momentum transports provide a significant contribution to the total momentum budget of the MCS circulations. Therefore, even if parameterisation of CMT is employed, it raises the question of whether the mesoscale transports would be appropriately represented by GCMs and global models. This needs to be studied further.

3. The increasing availability of Doppler data and the improvement in the calculation of the vertical winds will allow a wider sampling of systems and also better 3D wind datasets, to allow real systems to be analysed and then comparison made to the model results. The increasing availability of radar would also allow for larger areas to be analysed and allow to track the dynamical development of systems that pass through the scanning area, thereby allowing a wider range of real systems to be analysed.
7.3 Concluding remarks

This thesis contributes to the growing research of convective momentum transport due to deep convection by analysing the behaviour of CMT within organised mesoscale convective systems. It has also demonstrated the importance of the multiscale nature of momentum transport by highlighting some of the deficiencies in the way that CMT is treated through parameterisation. Although high resolution model simulations can resolve momentum adequately, there is, and will remain a need for parameterisation. This work has shown the importance of the mesoscale organisation of convective systems and brought attention to the need to include this knowledge within the parameterisation schemes. It has also highlighted some the WRF model’s weaknesses and suggests using both model and observational data when attempting to reproduce real cases.
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Momentum transport by organised deep convection

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