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Predictability and Representation of Convection in a Mesoscale Model

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I confirm that this is my own work and the use of all material from other sources has been properly and fully acknowledged.
Abstract

The validity of convective parameterisation breaks down at the resolution of mesoscale models and the circumstances for which a parameterisation should be used are not fully understood. Two mesoscale convective systems are chosen for their apparent difference in predictability using forecasts with different initialisation times. Convection in the non-predictable case is then shown to be in equilibrium with the large-scale forcing whereas convection in the predictable case is shown not to be in equilibrium. For both cases, three experiments are performed with different partitionings between parameterised and explicit convection. The two extremes lead to fully explicit and fully parameterised convective solutions.

For both cases, the intensity, location, horizontal scale and evolution of convection are all highly sensitive to the partitioning of convection. Despite only small differences in the total latent heating between the experiments for the equilibrium case, the large-scale impact of convection is very different for both cases. Explicit convection responds in a similar manner for both cases and localised intense regions of precipitation develop that track with the mid-level flow. In contrast, the equilibrium-based parameterisation scheme behaves differently in the two cases. For the equilibrium case, the scheme produces the smooth field of convection expected from an equilibrium response but for the non-equilibrium case, the scheme fails dramatically and develops bows of parameterised precipitation that were not observed.

An ensemble is designed to investigate the predictability of the triggering locations of explicit convection within the bounds of uncertainty due to unobserved small-scale variability. The timing and magnitude of the perturbations are found to be crucial and suggest a given ensemble technique may need to incorporate some case dependency. Consistent with the above hypothesis, the ensemble captures a range of possible triggering locations for the equilibrium case but shows high predictability of triggering locations for the non-equilibrium case.
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Thanks to the other PhD students for providing a fun and lively place to work. I shall miss the daily weather sessions looking out of the window in 1U07, the pantomime rehearsals, the random ensemble (a musical group, not to be confused with the work of this thesis) and general banter. I am also grateful to my many housemates, past and present, for making my time in Reading so enjoyable.

Finally, thanks to my family for providing places to escape to and their support.
Quantitative precipitation forecasts lose skill with time more rapidly than forecasts of any other surface parameter (e.g. Olson et al., 1995). Atmospheric models are frequently used to drive hydrological models for flood forecasting where the accurate prediction of timing and amount of precipitation becomes crucial. Often, flash flooding results from deep convective events, particularly organised convective systems, and it is under precisely these convective conditions that quantitative precipitation forecasting (QPF) is particularly difficult (e.g. Stensrud, 2001; Stensrud et al., 2000).

The representation of deep convection in numerical weather prediction models is a very difficult task. Perhaps the greatest challenge is the numerical representation of the continuous spectrum of temporal and spatial scales that exists within a convective circulation from micro-scale turbulence up to the more slowly-varying primary circulation (Emanuel, 1994). Furthermore, numerical weather prediction models are poor at developing convection at the correct time and location since the near-grid scales that act to trigger convection can have large uncertainty. The complex and highly nonlinear interaction between convection and the synoptic scales introduces further difficulty for numerical modelling.

The aim of this chapter is to identify the critical issues for the numerical representation of deep convection, specifically in mesoscale models. In the first half of the chapter, the convective circulation and the organisation of convection are described. The current approaches to the representation of convection in mesoscale models are then discussed and recent work is reviewed. The importance of uncertainty in the initial conditions in mesoscale models, particularly at the near-grid scales, is then identified for simulations of convective events. Recent case study based sensitivity
studies are then divided according to the predictability of the net properties of convection such as cumulative precipitation, heating and moistening. The distinguishing features of the convective environments for predictable and non-predictable cases are then identified. The chapter closes with a summary that motivates the thesis aims.

1.1 Convection

1.1.1 The Convective Circulation

Upright convection is a response to buoyant instability involving complex and highly non-linear processes. The net effect of convection is to warm and dry the atmosphere. An overview of the mature convective circulation is given below. The numbers refer to the schematic shown in Fig. 1.1.

1. Vertical transport of heat, moisture and momentum from lower-levels within the updraught forms the ascending branch of the convective circulation.

2. Latent heating associated with complex microphysical processes within the updraught excites gravity waves that spread this heating across the convective environment resulting in environmental subsidence.

3. Environmental air is entrained into the updraught and cloudy air is detained into the environment. This mixing modifies both updraught and environment temperature, moisture and momentum.

4. Precipitation rates largely depend upon updraught velocity and complex microphysical processes. Precipitation reaching the ground implies a net heating and drying of the environment.

5. Precipitation drag, water loading and cooling due to phase-changes of water act to reduce updraught buoyancy and often initiate downdraughts during the mature phase of a convective cell. These cold pools propagate away from their source at the surface and can be important in initiating new cells (e.g. Thorpe et al., 1982).
1.1.2 Organised Convection

In this section, the mechanisms for organisation, the structure, the impact on the large-scale environment and the frequency of organised convective systems are presented. Particular emphasis is given to a subset known as Mesoscale Convective Systems (MCSs). Simulations of two MCSs are used as the basis for numerical investigations throughout this thesis.

Deep conditional instability, vertical shear of the horizontal wind and low-level moisture convergence are necessary for deep, long-lived, organised convection (e.g. Rotunno et al., 1988; LeMone et al., 1998; Emanuel, 1994). A multitude of organised convective phenomena occurs within conditionally unstable regions with unlimited, often subtle, variations in wind-shear and thermodynamic profile. The process by which single convective cells organise and grow upscale is known as mesoscale organisation (as described by Houze, 1989; Cotton and Anthes, 1989). The defining characteristics of propagation and upscale growth are discussed below.

(a) Propagation and Upscale Growth

Propagation describes motion relative to the mean environmental flow and results from the internal dynamics of convection. Propagating convection can process more air than would otherwise be available had the convection moved purely with the mean environmental flow. In modelling studies of a mid-latitude two-dimensional squall line by Thorpe et al. (1982), the mechanism found to be essential for its propagation was the lifting of boundary-layer air through an inversion due to convergence at a cold pool boundary. Yang and Houze (1995) suggested the convective cells that compose a squall line arise as gravity-waves forced by continuous low-level convergence at the
cold pool boundary. Tompkins (2001) suggested the rôle of cold pools in low wind-shear, tropical ocean conditions was thermodynamical. Downdraughts injected cold and dry air into a cold and moist sub-cloud environment resulting in a spreading cold pool that was more moist just inside the cold pool boundary. The radius at which new convection initiated was determined by the time taken for surface fluxes to remove the negative buoyancy of the moist air at the cold pool boundary. Mapes (1993) explained the gregarious nature of tropical convection through gravity-wave lifting at low-levels due to convective heating that generates conditions favourable for additional convection in its mesoscale vicinity.

Mesoscale pressure systems generated by organised convective systems can act to increase low-level convergence and supply additional mass and moisture to the convective cells by a positive feedback (Fritsch and Chappell, 1980). Surface meso-lows are often observed close to a meso-high beneath the mature convective system (e.g. Cotton and Anthes, 1989). Hoxit et al. (1976) suggested a mechanism to explain observed surface pressure falls of 2 - 4 mbhr⁻¹ in the vicinity of organised convection. Compensating subsidence of order 10cms⁻¹ in the 500 - 100mb layer in an atmosphere of lapse rate less than dry-adiabatic locally warms the upper atmosphere. The resulting circulation removes mass from a vertical column in the region of subsidence and reduces surface pressure. The increase in surface pressure beneath the mature convective system results from mesoscale circulations that replace warm and moist inflow air with cold and dense downdraft air from mid levels.

Small-scale intense convective cells in the early stages of organisation grow upscale and develop mesoscale regions of precipitation (e.g. Maddox, 1980; Rutledge et al., 1988). The development of a common stratiform region of mesoscale dimension is seeded by ice particles detrained from the convective cells. High resolution radar analyses of organised convective systems (e.g. Fankhauser et al., 1992; Jorgensen et al., 1995; Hildebrand et al., 1996) indicated the presence of a continuous spectrum of convective structures. Organised convective systems are characterised in satellite imagery by a cirrus shield at the tropopause-level.

(b) Mesoscale Convective Systems

Houze (1993) defined an MCS, an example of organised convection, as a continuous cloud-system of thunderstorms associated with precipitation greater than 100km in at least one horizontal direction. MCSs are destructive convective phenomena that can produce severe localised winds,
damaging hail and flash flooding. Indeed, there is evidence from surface synoptic station data to suggest the devastating Lynmouth floods of 15th August 1952 (as documented by Marshall, 1952) resulted from an MCS. MCSs form a substantial part of the regional climate of the tropics (Dudhia et al., 1987) and are prevalent over mid-latitude continents (Laing and Fritsch, 1997). Britain, being a relatively small maritime landmass, experiences an average of just under two MCSs a year (Gray and Marshall, 1998), although in recent years as many as five have been observed. This may not necessarily represent an increase but may reflect the difference between analysing an historical data archive and real-time data. Furthermore, Gray and Marshall (1998) based their identification on surface synoptic stations and archived satellite imagery where available. It is only with remote sensing data of high temporal resolution (< 6 hours) that coherent convective systems can be identified with confidence.

(c) Impact on the Large-Scale Environment

The impact of convection on the large-scale environment and its subsequent downstream evolution are sensitive to the vertical heating and moistening profiles (e.g. Kuo and Reed, 1988). The vertical profile of heating differs between individual convective cells and organised convection (Houze, 1982; Johnson, 1984). Although diabatic processes such as melting and radiation are important, the net heating is dominated by condensation associated with vertical motion. Characteristic profiles of latent heating and divergence for regions of individual convective cells and organised convection are shown in Fig. 1.2. Net upward mass transport within deep convective cells produces net latent heating at all levels and results, from mass continuity and supported by observations (e.g. Mapes and Houze, 1995), in horizontal convergence at low-levels and horizontal divergence at upper-levels. Within the stratiform region of organised convection, however, net cooling resulting from melting and evaporation at mid and low-levels can initiate mesoscale downdraughts. Mass continuity results in convergence at mid-levels and divergence at upper and low-levels as observed (e.g. Mapes and Houze, 1995). It is therefore of great importance for a numerical weather prediction model to provide accurate vertical profiles in simulations of convection. Hartman et al. (1984) showed that an organised system heating profile reproduced tropical divergent circulations in a linear steady-state model more accurately than heating profiles of single cells.

Organised convection can be described in terms of Potential Vorticity (PV) structures. PV is used
Figure 1.2: Characteristic profiles of latent heating and horizontal mass divergence in regions of single cell and organised tropical convection (adapted from Houze, 1997).

in chapter 4 to identify the large-scale impact of convection. The Rossby-Ertel PV (Ertel, 1942) is defined as:

\[ PV = \frac{1}{\rho} (\zeta \cdot \nabla \theta), \tag{1.1} \]

where \( \rho \) is the density, \( \zeta \) is the absolute vorticity and \( \theta \) is the potential temperature. PV is conserved for adiabatic and frictionless flow and can be created or destroyed by diabatic processes such as latent heating associated with deep convection. The pattern of PV generated by a steady state heating source, that is, the heating continues over time periods of the same order as the vertical advection, as for organised convection, is shown schematically in Fig. 1.3. An area of negative PV at upper levels is generated together with a mid-level positive PV partner co-located with the region of maximum heating. PV anomalies are associated with anomalies in circulations and static stability. For organised convection, negative PV at the tropopause is associated with cold and warm anomalies above and below the tropopause and anticyclonic rotation centred at the tropopause.

Perturbations to upper-level jet structure through PV interaction can lead to significant downstream
modification to static stability and large-scale circulation. Raymond and Jiang (1990) proposed a mechanism by which pre-existing PV anomalies can interact with environmental shear and generate regions of mesoscale ascent favourable for MCS development. The mechanism was found to play a key role in forecast evolution in case studies by Gray (2001) where the impacts of convectively generated PV were found to be greatest in baroclinic zones where pre-existing large-scale ascent was either strengthened or weakened. Moreover, MCSs have often been observed to preferentially trigger downstream of dissipating convection in convectively modified flow (Mike Gray, personal communication).

1.2 Numerical Representation of Convection

In this section, the key issues for the representation of convection in numerical weather prediction models are identified. The section begins with the ideology of parameterisation and identifies criteria for its validity. The problems with including a parameterisation in mesoscale models are discussed and the results of previously published literature, using a variety of representations of convection, are reviewed.

The objective of numerical weather prediction is to predict the future state of the atmosphere from knowledge of its present state by use of numerical approximations to the dynamical equations. Differential terms are approximated by finite differencing on a grid. The finite difference equations can only describe scales significantly greater than two grid lengths. Subgrid-scale fluxes of
heat, mass and moisture can have considerable impact on the grid-scale flow. Although subgrid-scale processes are not included in a model, their statistical properties can be. A parameterisation describes the statistical properties in terms of resolved variables.

Arakawa and Schubert (1974) recognised that convection acts on a much faster timescale than that of the large-scale atmosphere and that a grid-scale is required to be large enough to contain a large ensemble of convective clouds but small enough to resolve the large-scale environment that forces convection. This spatial and temporal scale separation led to the quasi-equilibrium hypothesis that has formed the basis for many parameterisations. If the cloud-scale is small compared to the scale of the forcing, the errors incurred by the neglect of fluctuations about the ensemble mean are small. The large-scale forcing and the ensemble vary in a coupled way such that the statistical properties of the ensemble vary on the spatial scale of the forcing. Under a steady-state large-scale forcing, the average effect of the cloud ensemble is to produce an equilibrium state on a timescale that is long compared to the timescale of convection. If the large-scale is changing with time, no equilibrium is reached and convection depends on the history of the large-scale forcing. When the large-scale timescale is sufficiently longer than the convective timescale, the past history can be represented by the current large-scale forcing. The ensemble therefore follows a series of quasi-equilibrium states. The large-scale forcing and the ensemble therefore vary in a coupled way such that the statistical properties of the ensemble vary on the temporal and spatial scales of the forcing.

The choice of convective representation in numerical weather prediction models is determined by the relative scale of atmospheric convection to the grid-scale. For current climate models, the scale of individual convective clouds, 1-30km, is small compared to a grid-scale of approximately 200km. Deep convection is unresolved and needs to be parameterised. Such models provide no information on the details of convection such as precise location, propagation, size and maximum intensity because they occur on sub-grid scales. At the other extreme, for cloud-resolving models with a grid-length between 500m and 2km, most convective clouds are large compared to the grid-scale and the dynamics of the individual clouds themselves require explicit formulation. These models are at sufficiently high resolution to remove the need for a parameterisation of convection. The unresolved turbulent motions within convection, however, may still need to be parameterised. Cloud-resolving model simulations of deep convection have achieved more realistic mesoscale structure and higher precipitation maxima (e.g. Mass et al., 2002). This approach, however, is limited by the computational expense and the difficulty in initiating such a high resolution model with real data. As such, they are not suitable for operational forecasting but remain a powerful
1.2.1 Convection in Mesoscale Models

Mesoscale models have a resolution such that they are computationally viable for operational forecasting and have the capability to provide useful information on the smaller scale details of convection including location, intensity and propagation (examples are given later in this section). Convection in nature, however, is often not much smaller than the grid scales of current mesoscale models (10-30km). In the absence of a clear separation of scale, the validity of parameterisation breaks down and makes it difficult to know whether a parameterisation scheme should be included in mesoscale models. In a summary of cumulus parameterisation in mesoscale models, Kuo et al. (1997) recognised that even though the theory of parameterisation breaks down, the need for a parameterisation does not since a convection scheme permits buoyant instability to be removed on sub-grid scales as observed. Most mesoscale models include a parameterisation of convection and an explicit scheme that represents condensation and precipitation associated with the large-scale resolved flow. The representation of propagation and upscale growth associated with mesoscale organisation in a mesoscale model presents perhaps one of the greatest challenges (e.g. Dudhia, 1989; Kain and Fritsch, 1998). It is possible that the upscale growth will be represented explicitly to some extent. Despite significant effort, the question still remains as to how the physical processes should be partitioned between parameterised and explicit components in a mesoscale model.

(a) An Explicit Representation

Errors involved in the application of a parameterisation of convection in a mesoscale model may mean the explicit method (without a parameterisation of convection) is more useful. The problems associated with the explicit method are well documented, however, and are discussed below.

Explicit convection requires grid-scale saturation and aliases convection onto the smallest resolvable scale leading to spurious delays in the onset and subsequent over-prediction of convection (e.g. Zhang et al., 1988; Kato and Saito, 1995). Furthermore, air ascends at the grid-scale vertical
velocity (Molinari and Dudek, 1986) which is much smaller than observed velocities within deep convective updraughts. Wang and Seaman (1997) disregarded the explicit method as a viable option for mesoscale models on the basis of this slower and stronger convective response. A review of the representation of convection in mesoscale models (Molinari and Dudek, 1992) found the majority of successful explicit simulations of mesoscale organisation were those associated with strong large-scale forcing for convection. In such cases, resolved vertical motions were sufficient to minimise delays in onset.

Rosenthal (1978) recognised that explicit convection allows a broad spectrum of interactions between the convective (to the extent it is resolved) and the larger scales. However, a lack of timescale separation between convection and the large-scale forcing can result in the simultaneous occurrence of buoyant instability and explicit convection leading to artificial convective modes (Zhang et al., 1988). The explicit method is further limited by the predictability timescale of hours for atmospheric convection (Emanuel, 1994). A direct application of Lorenz’s analysis of non-linear systems (see Lorenz, 1969) to the observed atmospheric spectrum indicated the ultimate predictability of 20km wavelengths to be 2 hours (Lilly, 1990).

(b) A Parameterised and Explicit Representation

It has been argued that the small-scale convective cores embedded within mesoscale updraughts is a direct analogy to parameterised and explicit convection at mesoscale resolution (Kuo et al., 1997). Kain and Fritsch (1998) recognised that a convection scheme is important to prevent rapid unrealistic growth on the smallest resolvable scales of the model. Simulations at 25km resolution of a mid-latitude squall line showed that as the partitioning shifts in favour of the parameterised convection, the ability of the model to capture the linear structure and the system translation speed deteriorated. They concluded that the optimal solution for current numerical models is one in which convection is partitioned between parameterised and explicit processes. Based on explicit and parameterised simulations of a mid-latitude squall line and an MCS at 12.5 and 25km resolution, Zhang et al. (1988) concluded that even at grid spacings of order 10km a parameterisation of subgrid-scale convection is very important to retard excessive circulations. They also noted that convective rainfall would not occur in the presence of saturation on scales ~ 10km in nature.

To remove the spurious delay in explicit convection and the subsequent over-prediction of vertical mass transport, Molinari and Dudek (1986) suggested that a subgrid-scale source of detrained
precipitation particles was needed to develop independently of the parameterisation over subsequent timesteps. Kuo et al. (1997) suggested this was the models’ manifestation of the continuous transition from convective to mesoscale overturning that occurs in nature. However, Zhang et al. (1994) and Liu et al. (2001) showed that simulations of mesoscale organisation were sensitive to the magnitude of moisture detrainment rather than the phase. There is great uncertainty for the formulation of a parameterisation scheme for mesoscale models. Moreover, the location and amount of convective precipitation is highly sensitive to the convection scheme (e.g. Kuo et al., 1996; Spencer and Stensrud, 1998; Wang and Seaman, 1997; Gallus, 1999; Bélair et al., 2000). Kuo et al. (1997) hypothesised the differences were related to the ability of the scheme to interact with the grid. Furthermore, a single scheme that consistently outperforms others in a wide range of meteorological situations does not currently exist.

(c) Partitioning

The partitioning of convection between explicit and parameterised components appears to be a key issue for mesoscale simulations. Sensitivity studies using different convection schemes by Wang and Seaman (1997) and Gallus (1999) showed that the partitioning of precipitation was very sensitive to both the convection scheme and the convective environment. Three of the four parameterisation schemes investigated by Wang and Seaman (1997) were found to be only weakly sensitive to horizontal resolution in the range 12-36km. The Betts and Miller (1986) adjustment scheme (discussed briefly on page 14), however, was found to be sensitive to resolution. This scheme employs a constant adjustment timescale to specified reference profiles. It was hypothesised that the sensitivity of partitioning to resolution was a result of using the same timescale at both resolutions. Betts and Miller (1993) suggested the timescale should be reduced accordingly with grid size to maintain appropriate grid-scale subsaturation but did not specify any strict guidelines.

The impact of horizontal resolution on the relative rôles of parameterised and explicit convection was examined by Bélair and Mailhot (2001) in simulations of a mid-latitude squall line. Results from a 2km horizontal resolution simulation were used to evaluate simulations at 6, 18 and 50km. At 6km, the leading convective line was both parameterised, using the Kain and Fritsch (1990) scheme, and explicit, as shown in Fig. 1.4, with greater explicit rainrates. At 18km, parameterised convection formed the front edge of the squall line and the explicit convection formed a broader
CHAPTER 1: Introduction

Parameterised and explicit rainrates valid at 0300 UTC 8 May 1995 at horizontal resolutions of 6, 18 and 50km (Bélair and Mailhot, 2001). The shadings represent rainrates greater than 1, 5, 10 and 25mmhr$^{-1}$.

Figure 1.4: Parameterised and explicit rainrates valid at 0300 UTC 8 May 1995 at horizontal resolutions of 6, 18 and 50km (Bélair and Mailhot, 2001). The shadings represent rainrates greater than 1, 5, 10 and 25mmhr$^{-1}$.

The hydrostatic approximation greatly simplifies the equations of motion and considerably reduces computational expense. This has made it an attractive option for mesoscale models but it has limitations. The validity of the hydrostatic approximation in mesoscale models is discussed in appendix A.

(d) Summary

Mesoscale models divide convection into explicit and parameterised components. The problems associated with an explicit representation and a simultaneous parameterised and explicit representation are well documented. However, the general consensus as to the most useful partitioning of convection leans in favour of a simultaneous explicit and parameterised treatment. The details of

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convection are very sensitive to the formulation of the convection scheme and the partitioning of convection between the explicit and parameterised components appears to be a key issue.

1.3 Parameterisation Schemes

The requirements and formulation of parameterisation schemes are discussed in this section. Current parameterisation schemes used in mesoscale models can generally be classified as either adjustment or mass-flux schemes. An overview of the two formulations is presented together with an outline of some recent schemes that have attempted to relax the assumptions that do not hold in mesoscale models.

1.3.1 Requirements of a Parameterisation Scheme

Convective parameterisation formulates the cumulative effects of convection in terms of large-scale prognostic variables. A convection scheme is intended to:

- Determine the region where convection occurs.
- Predict the overall magnitude of energy release. This 'closure' assumes a relationship between convection and its environment.
- Vertically distribute this energy as some combination of heating, moistening and momentum transport. Exactly how the energy is distributed in the vertical is poorly understood (Houze, 1989).

1.3.2 Approaches to Parameterisation

(a) Adjustment Schemes

Smagorinsky (1956) and Manabe et al. (1965) developed the first convective adjustment schemes for use in global general circulation models. Dry and moist adiabatic adjustment is applied to
unstable dry and saturated layers respectively. Such schemes are based on the physically plausible notion that upright convection is a response to buoyant instability. Unknowns include the rate of relaxation and the equilibrium profile of water vapour. This adjustment implicitly determines vertical transport of heat and moisture by convection. All condensed water is assumed to fall out as precipitation in most schemes. A soft adjustment is often applied in which adjustment is assumed over a fractional area, $\sigma$, of a grid-column. Final values of an adjusted variable, $\phi$, are weighted as:

$$\bar{\phi} = \phi^{\text{environment}} (1 - \sigma) + \sigma \phi^{\text{cloud}}. \quad (1.2)$$

The Betts and Miller (1986) scheme adjusts a model’s temperature and moisture profiles towards the moist virtual adiabat observed in tropical deep convective environments. The precipitation, $P$, resulting from this adjustment is defined as:

$$P = \int_{P_B}^{P_T} \frac{q_f - q \, dp}{\tau \, g}, \quad (1.3)$$

where $q$ is the model’s specific humidity, $q_f$ is the reference profile of specific humidity, $\tau$ is the timescale over which the adjustment occurs and $P_T$ and $P_B$ are the pressures at cloud top and bottom, respectively. A more recent version (Betts and Miller, 1993) included a parameterisation of downdraughts.

(b) Mass-Flux Schemes

The convective transport can be written in a vertical eddy flux form, $\overline{w \phi'}$, where $w$ is the vertical velocity, $\phi$ represents a scalar and the overline represents a grid-box average. The heat source, $Q_1$, and moisture source, $Q_2$, due to convection may be written as:
where $Q$ is condensation minus evaporation, $\theta$ is the potential temperature, $q$ is the specific humidity, $p$ is the pressure, $L$ is the latent heat of condensation, $C_p$ is the specific heat of dry air and $\Pi$ is the exner function. The first term in each equation is due to condensation and the second term is due to eddy transport. For use in a parameterisation scheme, the eddy flux term needs to be written in terms of resolved quantities. Mass-flux theory, first formulated by Ooyama (1971), assumes that the vertical transport of any scalar, $\phi$, is well represented by the product of the mass-flux, $M_c$, and the bulk updraught scalar excess:

$$\overline{\rho w' \phi'} \sim M_c (\overline{\phi} - \overline{\phi})$$

where $\overline{\phi}_c$ is the in-cloud value. The idea of using convective mass-flux to parameterise vertical transports by convective draughts is now widely used in mesoscale models. Equation 1.6 assumes the area of cloud is small compared to the grid-box area so the environmental value of $\overline{\phi}$ can be approximated by the grid-box average. Equations 1.4 and 1.5 now become:

$$Q_1 = \frac{L \overline{Q}}{C_p \Pi} - \frac{\partial \overline{w' \theta'}}{\partial p};$$

$$Q_2 = -\overline{Q} - \frac{\partial \overline{w'q'}}{\partial p},$$
A cloud model, formulated around mass continuity and the thermodynamic equation is used to calculate the in-cloud quantities: $M_c, \overline{\vartheta^c}$ and $\overrightarrow{q}$. Owing to uncertainty in its formulation, the details of the cloud model differ between schemes. Vertical heating and moistening profiles are severely constrained by prescribed entrainment and detrainment profiles and therefore the impact of parameterised convection on the resolved flow is constrained. Inverted cloud models are often used to represent downdraughts. Kain and Fritsch (1998) recognised that these cloud models represent deep convection as individual elements, rather than strongly interacting multi-scale convection as observed and do not allow for vertical development.

The mass-flux at cloud-base needs to be determined in terms of the resolved variables. Many closures are based on quasi-equilibrium and set the cloud-base mass-flux proportional to the large-scale forcing. Tiedtke (1989), for example, set the cloud-base mass-flux equal to the water vapour tendency integrated over the sub-cloud layer. An adjustment closure, however, relaxes quasi-equilibrium by setting the cloud-base mass-flux proportional to the difference between the large-scale structure and some reference structure. Large-scale ascent, for example, cools the atmosphere and increases $\overline{\vartheta^c} - \overline{\vartheta}$, thus increasing convection. Convection then warms the atmosphere and decreases $\overline{\vartheta^c} - \overline{\vartheta}$, leading to less convection. The physical processes that determine the timescale over which this adjustment occurs are not well known. Recent work, however, has shown it scales with the cloud-spacing (Cohen, 2001). Such equilibrium-based schemes are expected to produce a smooth field of convection co-located with regions of CAPE.

Recent cloud-resolving modelling studies by Swann (2001) showed the net heating and moistening due to convection can be well represented by the mass-flux approach. The Gregory and Rowntree (1990) mass-flux scheme, used in this study, is described in detail in chapter 2.

(c) Prognostic and Stochastic Schemes

Kuo et al. (1997) suggested that the approach to parameterise convection in mesoscale models needs to change from diagnostic to prognostic to incorporate the history of convection. As an alternative to the quasi-equilibrium closure of Arakawa and Schubert (1974) that relied heavily on a temporal scale separation, Pan and Randall (1998) explicitly predicted the cloud-base mass-flux through a convective kinetic energy closure. This provided convection with a memory and removed the need to define a large-scale forcing.
Recent work has shown the variance in total convective mass-flux is inversely proportional to the mean number of clouds in the ensemble (Cohen, 2001). In order to relax the spatial scale separation assumption of quasi-equilibrium, the variance of the convective ensemble needs to be included in a parameterisation. Lin and Neelin (2002) hypothesised that there exists variability in convection that may influence the large scales, but which is not well described by the ensemble mean. They recognised two approaches to include higher order moments of parameterised convection. The first method, investigated by Lin and Neelin (2000), stochastically parameterises the physics of the convective elements themselves. The second method, investigated by Lin and Neelin (2002), specified a distribution of convective heating from which the value put into the model was randomly selected. However, results showed the atmosphere heavily modified the input variance so this direct control ignored important feedbacks. They showed that a substantial portion of the large-scale variance in precipitation distribution and winds in a general circulation model may be a response to smaller scale variability.

1.4 Explicit Schemes

The majority of mesoscale simulations employ both a parameterisation of convection and an explicit scheme that represents condensation and precipitation processes associated with the large-scale resolved flow such as frontal cloud. In this section, the main components of explicit schemes are outlined.

Cloud and precipitation processes are parameterised using the grid-scale specific humidity, temperature and vertical velocity. These variables determine the condensation rate and therefore the supply of liquid water content through processes such as large-scale ascent and diabatic cooling. Most schemes are based on the work of Sundqvist (1978) who modelled precipitation by instantaneously removing a portion of the condensate. Without a cloud scheme, condensation would only occur when grid-scale supersaturation was reached. Excess water vapour would then be converted directly to precipitation. A cloud scheme allows condensation of water before the artificial constraint of grid-scale supersaturation. A cloud scheme may also allow the effects of cloud amount, thickness and height to be included primarily on temperature via radiative effects but also on precipitation.
Owing to computational constraints, a detailed simulation of cloud microphysics is not practical. Instead, cloud processes are often parameterised using a bulk water technique where liquid water is subdivided into cloud water and precipitation water (Tiedtke, 1993). The assumption that moisture and condensate are spread evenly across a grid-box is poor in mesoscale models. It is desirable yet extremely difficult to quantify the subgrid-scale nature of the microphysical processes. The different schemes are characterised by their number of microphysical variables such as cloud water, ice, graupel and snow and their conversion mechanisms. Most schemes remove rain water from the system in a single timestep and allow physical processes to act on the falling rain. The Wilson and Ballard (1999) explicit cloud and precipitation scheme, used in the numerical simulations for this thesis, is described in chapter 2.

1.5 Accounting for Uncertainty

The issue of the numerical representation of convection is not only a matter of partitioning between parameterised and explicit components. A deterministic mesoscale forecast is only one of a range of possible solutions within the bounds of uncertainty due to error in model physics and uncertainty due to the unobserved scales. Indeed, deterministic simulations using a variety of mesoscale models and representations of convection have repeatedly failed to capture the precise timing and location of deep convection (e.g. Bernardet et al., 2000; Stensrud, 2001; Ferretti et al., 2000; Romero et al., 1998). In this section, the impact of uncertainty in simulations of organised convection is explored and methods to account for this uncertainty are reviewed.

Uncertainty and the inherent limits to atmospheric predictability (Lorenz, 1969) have motivated the use of ensembles as a forecasting tool. Ensemble forecasting samples uncertainty and can predict the probability of a future state of the atmosphere (Leith, 1974). Mesoscale model simulations of 20 MCSs using different convection schemes and initialisations by Gallus and Segal (2001) showed a lack of consistent improvement in QPF skill scores for any one combination. They suggested ensemble forecasting may be of more value than a deterministic forecast. Success with global ensembles in the 1990s (e.g. Toth et al., 1997) led to experimentation with short-range mesoscale ensembles. A 12km resolution ensemble generated from five slightly different analyses (a multianalysis ensemble) by Grimit and Mass (2002) was demonstrated to be as skillful in terms of wind direction forecast errors as a 4km deterministic forecast. Indeed, for a case of explosive
cyclogenesis, Du et al. (1997) found the improvements in root mean square error and equitable threat score statistics due to ensemble forecasting exceeded any due to doubling either horizontal or vertical resolution.

Short-range mesoscale ensemble techniques have been designed to sample analysis error (Grimit and Mass, 2002; Du et al., 1997), errors in model physics (Stensrud et al., 2000) and both analysis and model error simultaneously (Hou et al., 2001; Stensrud et al., 1999b, 2000). The most useful approach may depend on the application. If a model is capable of representing the convection, a data sparse region may benefit more from a multianalysis ensemble (Grimit and Mass, 2002) where analysis errors are likely to dominate model error. A well-observed region, however, may benefit more from a multimodel ensemble where poor model assumptions and parameterisations largely dominate the errors in the solution. The use of ensembles may not always lead to improved forecasts. For example, Stensrud et al. (2000) showed no members of a multianalysis ensemble were able trigger an observed MCS and suggested this was the result of the absence of a convectively-generated meso-high in the initial conditions.

Numerical weather prediction models are poor at developing convection at the correct location and time partly because of the small scales that can act to trigger convection in nature (Kain and Fritsch, 1992; Stensrud and Fritsch, 1994). At mesoscale resolution these scales are near-gridscale and as such are poorly resolved, if included at all. In addition, these scales cannot be verified by the current coarse resolution observation networks. Indeed, Stensrud and Fritsch (1994); Zhang and Fritsch (1986); Gallus and Segal (2001) showed the importance of mesoscale details in the initial condition for the location and timing of deep convection. Stensrud et al. (1999a) found the inclusion of cold pools in the mesoscale initial condition produced significant improvements not just in location but in rainfall totals. Techniques for perturbing these small scales have previously included singular vectors (Molteni et al., 1996), bred modes (Toth and Kalnay, 1993), and the Monte-Carlo method (Leith, 1974). The most useful method, however, is not clear (Anderson, 1996) and may depend on the application. In addition, the magnitude and structure of small-scale error can only be estimated because of a lack of validating observations. Moreover, Mullen and Baumhefner (1989) proposed case-to-case variability of analysis uncertainty. Ensembles exploiting analysis uncertainty have generally involved some perturbation at analysis time (e.g. Hou et al., 2001; Du et al., 1997; Stensrud et al., 2000).
1.6 Importance of the Large-Scale Environment

Zhang et al. (1988) hypothesised that success of parameterised versus explicit convection depends on the particular precipitating mode; convective versus stratiform, and on the nature of the large-scale forcing for upward motion; quasi-stationary versus propagating and weak versus strong. The results of a number of studies that have explored the sensitivity of the convective solution to either resolution, the convection scheme or small-scale details in the initial conditions are reviewed. A result common to all these studies is that the sensitivity is highly case-dependent. The cases are divided according to the predictability of the net properties of the convective solution such as precipitation, heating and moistening and key features of the large-scale environments are identified.

For two cases of organised convection that triggered at a cold pool boundary, Spencer and Stensrud (1998) showed the maximum 24 hour rainfall totals and locations were very sensitive to the representation of the downdraft in the Kain and Fritsch (1990) parameterisation scheme. Cold pool boundaries propagated through a region of weak upward motion characterised by Convective Available Potential Energy\(^3\) (CAPE) of 3000\(\text{Jkg}^{-1}\) and Convective Inhibition (CIN) of 56\(\text{Jkg}^{-1}\) (Schwartz et al., 1990). A sounding ahead of the boundary had a lifting condensation level of 862mb and a level of free convection of 750mb. Dynamical lifting at the boundary was sufficient for a parcel to reach its level of free convection. For three other cases of organised convection associated with strong synoptic or frontal forcing, the 24 hour maximum rainfall totals were only weakly sensitive to the details of the convection scheme. For one of these events, convection developed as low-level air ascended over a quasi-stationary east-west oriented surface warm front (Elsner et al., 1989). Synoptic-scale ascent was provided by a 500mb trough. The other two events were associated with warm moist southerly advection beneath cool advection aloft south of a deep low pressure system.

Wang and Seaman (1997) simulated a squall line using four different parameterisation schemes and one explicit representation. The synoptic situation (described by Brandes, 1990) was characterised by a weak surface front and weak thermal advection below 700mb. Profiles ahead of the propagating squall line had large CAPE and some CIN. An hourly timeseries of total precipitation rates implies the total rain amount was very sensitive to the representation of convection. Liu et al. (2001) performed 7 day simulations of tropical convection forced by objectively analysed

---

\(^3\)CAPE is defined in equation 2.41 on page 44
time-varying large-scale advection of temperature and moisture. Simulations were performed with and without the Kain and Fritsch (1990) convection scheme at two resolutions within the range of current mesoscale models. During periods of strong large-scale forcing the sensitivity of the 6 hour domain average accumulated precipitation to both the representation of convection and resolution was much less than for weakly forced periods. For periods of weak large-scale forcing, the subgrid-scale fluxes became a significant source of CAPE.

For a case of organised convection that developed at a convergence line in a warm sector and north of the surface warm front associated with flow over the boundary, Gallus (1999) showed that the domain integrated precipitation was remarkably constant with horizontal resolution in the range 12 - 78km and was similar between simulations using different convection schemes. A sounding showed 1500Jkg\(^{-1}\) and almost zero CIN. For a second case that developed north of a slow-moving east-west oriented surface warm front, the domain integrated precipitation was found to be sensitive to both resolution and the convection scheme. An environmental sounding showed 2000Jkg\(^{-1}\) of CAPE and 25Jkg\(^{-1}\) of CIN. For a seven day period, Stensrud et al. (1999a) examined sensitivity to cold pools in the initial conditions. Under sustained strong large-scale forcing, the 24 hour rain amount was only weakly sensitive. Under weak large-scale forcing, however, the suggestion from the field of 24 hour rain amount is that the domain average 24 hour rain amount was very sensitive. Stensrud and Fritsch (1993) described a period of 2 days in which 5 MCSs developed in a region characterised by weak thermal gradients, weak thermal advection, weak upward motion, CAPE and CIN. Deep convection initiated where dynamical lifting associated with low-level boundaries eliminated the CIN. Simulations of this event by Stensrud and Fritsch (1994) showed the domain averaged precipitation were highly sensitive to both the initial conditions and the formulation of the trigger in the convection scheme.

**Summary**

The case studies for which the net properties of the MCS were predictable were characterised by strong large-scale forcing for upward motion. For these cases, the region of CAPE was co-located with approximately zero CIN. Under strong large-scale forcing, CAPE is largely determined by the large scales and, in the absence of CIN, convection is free to act. The case studies for which the net properties of the MCS were not predictable were characterised by weak synoptic forcing. For these cases, the region of CAPE was co-located with CIN. Under weak large-scale forcing,
CAPE and CIN are strongly modulated by mesoscale details such as boundary layer convergence lines and cold pools. Moreover, this mesoscale structure is often generated by the convection itself (e.g. cold pools).

### 1.7 Chapter Summary

As discussed in section 1.2, mesoscale models divide convection into explicit and parameterised components. The problems associated with an explicit representation and a simultaneous parameterised and explicit representation are well documented. However, the general consensus as to the most useful partitioning of convection leans in favour of a simultaneous explicit and parameterised treatment. There is, however, poor understanding of the circumstances for which a parameterisation should be used. The small-scale details of convection such as precise location, size, intensity and propagation are very sensitive to the formulation of the convection scheme and the partitioning of convection between the explicit and parameterised components appears to be a key issue.

The issue of convective representation in mesoscale models is not only a matter of partitioning. A successful forecast of convection ultimately depends on the ability of the model to capture the location and timing of convection. This, however, can be strongly affected by uncertainty in the initial conditions due to unobserved scales and errors in the model physics. The initial location of convection and subsequent rain amounts have been found to improve on using finer scale initialisation.

The feedback of convection on the large-scale flow is sensitive to the vertical heating and moistening profiles of convection. These profiles are significantly different between single convective cells and organised convective systems. Accurate representation of the vertical profiles of organised convection may be important for the large-scale impact and downstream flow.

Cases of organised convection for which the net properties are predictable all appear to be characterised by strong synoptic forcing. For these cases, the region of CAPE is co-located with approximately zero CIN and convection is free to act. It is likely that for these cases, the rate of stabilisation of the atmosphere by convection balances the rate of destabilisation by the large-scale forcing and convective equilibrium is satisfied. Cases of organised convection for which the net properties are not predictable all appear to be characterised by weak synoptic forcing. For these
cases, the region of CAPE is co-located with CIN and convection is not always free to act. It is likely that for these cases, convection does not come into balance with the large-scale forcing and convective equilibrium is not satisfied. It is hypothesised that the behaviour of explicit and parameterised convection including the location, horizontal scale, intensity, propagation and the impact on the large-scale environment depends on whether or not the system is in convective equilibrium.

1.8 Thesis Aims

To improve the poor understanding of the circumstances for which a parameterisation scheme should be used in mesoscale models, this thesis identifies the importance of convective equilibrium for the behaviour of parameterised and explicit convection. More specifically, the following questions, motivated by the summary in the previous section, are answered:

**Q1.** How does the parameterisation scheme behave under equilibrium and non-equilibrium conditions?

**Q2.** Can the explicit representation of convection provide information beyond that acquired using the parameterisation scheme?

**Q3.** Is the large-scale response to convection sensitive to the partitioning of convection for the duration of a mesoscale forecast?

**Q4.** Is the precise triggering location of convection predictable?

Chapter 2 describes the Met Office mesoscale model used for the numerical simulations throughout this thesis and the initialisation procedure. Particular emphasis is given to the explicit cloud and precipitation scheme and the equilibrium-based parameterisation scheme. Some idealized modelling studies are discussed to aid the interpretation of results presented in chapters 4 and 5. Details of experiments used to answer the first three questions are outlined. The three experiments provide three different partitionings between parameterised and explicit convection; a fully explicit solution, a fully parameterised solution and a solution with significant amounts of both.

Simulations of two MCSs are used as the basis for numerical investigations throughout this thesis.
In chapter 3, an overview of the MCSs is provided using observational and model data. The fields of CAPE and CIN indicate that for one case, equilibrium is satisfied and for the other case, equilibrium is not satisfied.

Chapters 4 and 5 present the main results of this thesis. Firstly, the state of equilibrium, or otherwise, is verified for the two case studies. The main aim of chapter 4 is to identify the behaviour of parameterised and explicit convection for both cases. The sensitivity of the large-scale flow to the partitioning between explicit and parameterised convection is also explored. Specifically, the ability of parameterised and explicit convection to capture the small-scale details of convection including location, timing, size, intensity and propagation characteristics is determined. The magnitude, vertical structure and horizontal extent of the large-scale modification is then determined for parameterised and explicit convection and compared with observations.

In the first half of chapter 5, the mechanisms for triggering deep convection in the model are explored. In the second half of the chapter, the ability of the model to reproduce the precise location and timing of the two convective events is determined using an ensemble approach. The ensemble explores the range of possibilities consistent with the observed large-scale environment by sampling small-scale uncertainty. Finally, the main conclusions of this thesis are presented in chapter 6 together with some ideas for future work.
Mesoscale model simulations of two MCSs form the basis for numerical investigations in this thesis. In this chapter, the main components of the Met Office unified mesoscale model version 4.5, including the options selected for this investigation, are presented. Particular emphasis is given to the calculation of buoyancy and the representation of precipitation processes for both explicit motions and the parameterisation scheme. Some hypothetical idealized experiments are discussed to examine how the model partitions convective mass-flux between the parameterisation scheme and explicit motions. The method used to answer the first three questions posed in chapter 1 is then described and the model initialisation procedures are outlined.

2.1 The Met Office Unified Model

The Met Office unified mesoscale model version 4.5 is formulated around the quasi-hydrostatic primitive equation set. The equations are very similar to the primitive equations but include vertical Coriolis force terms in the horizontal momentum equations and exclude the shallow atmosphere approximation. The large-scale equation set presented here is reproduced from Unified Model Documentation Paper No.10 (Cullen et al., 1993). The notation is that defined in Unified Model Documentation Paper No.5 (Wilson, 1993). The vertical coordinate \( \eta = \eta(p, p^*) \), where \( p \) is the pressure and \( p^* \) is the pressure at the surface, is defined such that \( \eta(0, p^*) = 0 \) and \( \eta(p^*, p^*) = 1 \). The equations are written in spherical coordinates \((\lambda, \phi)\) and reproduced here for a general position of the coordinate pole.
CHAPTER 2: Modelling Strategy

The horizontal momentum equations are defined as:

\[
\frac{\partial u}{\partial t} + \frac{u}{r_s \cos \phi} \frac{\partial u}{\partial \lambda} + \frac{v}{r_s} \frac{\partial u}{\partial \phi} + \eta \frac{\partial u}{\partial \eta} + \frac{1}{r_s \cos \phi} \left( \frac{\partial \Phi}{\partial \lambda} + \left( \frac{R(T_v + \mu T_v)}{p} \right) \frac{\partial p}{\partial \lambda} \right) - f_3 v + f_2 \bar{w} - \frac{u}{r_s} (\nu \tan \phi - \bar{w}) = F_u, \tag{2.1}
\]

\[
\frac{\partial v}{\partial t} + \frac{u}{r_s \cos \phi} \frac{\partial v}{\partial \lambda} + \frac{v}{r_s} \frac{\partial v}{\partial \phi} + \eta \frac{\partial v}{\partial \eta} + \frac{1}{r_s} \left( \frac{\partial \Phi}{\partial \phi} + \left( \frac{R(T_v + \mu T_v)}{p} \right) \frac{\partial p}{\partial \phi} \right) - f_3 u + f_1 \bar{w} - \frac{u^2}{r_s} (\tan \phi + \frac{\nu \bar{w}}{r_s}) = F_v, \tag{2.2}
\]

where \(u\) and \(v\) are the horizontal wind components, \(r_s\) and \(T_v\) are the pseudo radius and basic state temperature taken from a stable reference atmosphere (as defined by Cullen et al., 1993) to ensure conservation properties, \(\Phi\) is the geopotential height, \(\bar{w}\) is an approximate vertical velocity, \(R\) is the gas constant for dry air and the virtual temperature \((T_v)\) is defined as \(T(1 + (\epsilon^{-1} - 1)q)\) where \(\epsilon\) is the ratio of molecular weights of water and dry air. \(f_1, f_2\) and \(f_3\) are components of the Coriolis parameter and \(\mu\) is defined as \((f_2 u - f_1 u + (u^2 + v^2)/r_s)\).

The hydrostatic equation is defined as:

\[
\frac{\partial \Phi}{\partial \eta} = - \left( \frac{R(T_v + \mu T_v)}{p} \right) \frac{\partial p}{\partial \eta} - C_p \theta_s \frac{\partial \Pi}{\partial \eta}, \tag{2.3}
\]

where \(\theta_s\) is the basic state potential temperature, \(C_p\) is the specific heat of dry air and \(\Pi\) is the exner function.

The thermodynamic equations are defined as:

\[
\frac{\partial \theta_1}{\partial t} + \frac{u}{r_s \cos \phi} \frac{\partial \theta_1}{\partial \lambda} + \frac{v}{r_s} \frac{\partial \theta_1}{\partial \phi} + \eta \frac{\partial \theta_1}{\partial \eta} - \frac{1}{\pi} \left( \frac{L_c \theta_1}{C_p T} \right) \frac{RT \omega}{C_p p} = F_{\theta_1}, \tag{2.4}
\]
\[
\frac{\partial q_{l}}{\partial t} + \frac{u}{r_{c} \cos \phi} \frac{\partial q_{l}}{\partial \lambda} + \frac{v}{r_{c}} \frac{\partial q_{l}}{\partial \phi} + \frac{\partial q_{l}}{\partial \eta} = F_{q_{l}}, \tag{2.5}
\]

where \( L_{c} \) is the latent heat of condensation, \( \omega \) is the pressure vertical velocity and the thermodynamic variables are defined as:

\[
\theta_{l} = \theta - \frac{L_{w}}{C_{p} \prod} q_{cl}, \tag{2.6}
\]

\[
q_{r} = q + q_{cl}, \tag{2.7}
\]

where \( q_{cl} \) is the cloud liquid mixing ratio as diagnosed by the large-scale cloud scheme (described later). The mixed phase precipitation scheme requires a separate prognostic equation for ice \((q_{cf})\) defined as:

\[
\frac{\partial q_{cf}}{\partial t} + \frac{u}{r_{c} \cos \phi} \frac{\partial q_{cf}}{\partial \lambda} + \frac{v}{r_{c}} \frac{\partial q_{cf}}{\partial \phi} + \frac{\partial q_{cf}}{\partial \eta} = F_{q_{cf}}. \tag{2.8}
\]

Finally, the continuity equation is defined as:

\[
\frac{\partial}{\partial \eta} \left( r_{c} \frac{\partial p}{\partial t} \right) + \frac{1}{\cos \phi} \left( \frac{\partial}{\partial \lambda} \left( u r_{c} \cos \phi \frac{\partial p}{\partial \eta} \right) + \frac{\partial}{\partial \phi} \left( v r_{c} \cos \phi \frac{\partial p}{\partial \eta} \right) \right) + \frac{\partial}{\partial \eta} \left( \eta^{2} \frac{\partial p}{\partial \eta} \right) = 0. \tag{2.9}
\]

The quantities \( F_{u}, F_{v}, F_{\theta_{l}}, F_{q_{l}}, \) and \( F_{q_{cf}} \) are source terms from parameterisations and include any diffusion required to ensure computational stability. Parameterised convection, described in section 2.3, couples with the large-scale equations through the source terms in the thermodynamic equations. The parameterisation scheme provides increments of potential temperature and water vapour mixing ratio only and does not provide an increment to \( q_{cf} \). Horizontal momentum sources due to parameterised convection are neglected but may have significant effects on the large-scale atmosphere through upscale transport of energy and momentum (Kuo et al., 1997).
(a) Buoyancy

In applying the equations of motion to convective situations the effect of cloud condensate (as diagnosed by the large-scale cloud scheme described later in section 2.2) on buoyancy should be considered. In nature, condensate loading is one process that contributes to negative buoyancy and can be sufficient to reverse the updraught in upright convection. Since liquid and ice particles in the air quickly achieve their terminal fall speeds, the frictional drag of the air on the particles comes into balance with the gravitational force acting on the particles. The drag of the particles on the air can be expressed as \(-g (q_{cl} + q_{cf})\) and should be included as an additional acceleration term in the equation set but is not included in the Unified Model. An equivalent way to incorporate condensate loading is to define the density term in the hydrostatic equation (the first term on the right-hand side in equation 2.3) to include the mass of condensate as well as the mass of the air itself but this is not done in the Unified Model. On the other hand, the contribution to positive buoyancy due to water vapour is included through the use of the virtual temperature. As water vapour condenses the buoyancy decreases through the removal of water vapour but not through the increase in condensate. This is likely to result in an excessively buoyant model.

The vertical momentum equation is approximated by the hydrostatic equation which assumes balance between the vertical pressure gradient force and the gravitational force and the prognostic equation for vertical velocity is lost. The process by which buoyancy forces ascent is therefore not as straightforward as in the complete momentum equations since there is no direct relation between buoyancy and vertical acceleration. The following exercise provides insight to the problem and a diagnostic expression for vertical velocity is derived. For simplicity and without significantly changing the problem a simplified hydrostatic primitive equation set (with the shallow atmosphere approximation) given below is used (adapted from White (2000)). The horizontal momentum equations:

\[
\frac{Du}{Dt} = f v + \frac{uv tan \phi}{a} - \frac{1}{\rho a cos \phi} \frac{\partial p}{\partial \lambda} + F_e, \tag{2.10}
\]

\[
\frac{Dv}{Dt} = -f u - \frac{u^2 tan \phi}{a} - \frac{1}{\rho a} \frac{\partial p}{\partial \phi} + F_{\phi}, \tag{2.11}
\]

where \(a\) is a mean value for the radius of the Earth.
The hydrostatic equation:

\[ g + \frac{1}{\rho} \frac{\partial \rho}{\partial z} = 0 \]  

(2.12)

The continuity equation:

\[ \frac{D\rho}{Dt} = -\rho \nabla \cdot \mathbf{u} \]  

(2.13)

The thermodynamic equation:

\[ \frac{D\theta}{Dt} = \left( \frac{\theta}{C_v T} \right) Q, \]  

(2.14)

where \( Q \) is the heating rate. By writing the continuity equation (2.13) in terms of \( \partial \rho / \partial t \) and using the hydrostatic equation (2.12) the flow \( \mathbf{u} \) can be separated into a horizontal component \( \mathbf{v} \) and a vertical component \( w \):

\[ \nabla_H (\rho \mathbf{v}) + \frac{\partial}{\partial z} \left( \rho w - \frac{1}{g} \frac{\partial \rho}{\partial t} \right) = 0, \]  

(2.15)

where \( \nabla_H \) indicates the horizontal divergence. Integrating (2.15) over the interval \([z, \infty]\) gives:

\[ \frac{\partial \rho}{\partial t} = \rho gw - g \int_z^\infty \nabla_H (\rho \mathbf{v}) \, dz \]  

(2.16)

Equation 2.16 states that the time rate of change of pressure at height \( z \) is equal to \((g \times)\) the rate of convergence of mass into the column above \( z \). Another expression for \( \partial \rho / \partial t \) can be obtained by using (2.13) and the perfect gas law to write the thermodynamic equation (2.14) as:

\[ \frac{D\rho}{Dt} = -\gamma p \nabla \cdot \mathbf{u} + \frac{\rho R Q}{C_v} \]  

(2.17)

Hence (using (2.13):

\[ \frac{\partial \rho}{\partial t} = -\mathbf{v} \cdot \nabla_H p + \rho \mathbf{w} \cdot \mathbf{v} - \gamma p \nabla \cdot \mathbf{u} + \frac{\rho R Q}{C_v}. \]  

(2.18)

The right sides of 2.16 and 2.18 must balance giving the diagnostic expression:

\[ \gamma p \frac{\partial w}{\partial z} = \gamma p \left( \frac{Q}{TC_p} - \nabla_H \cdot \mathbf{v} \right) - \mathbf{v} \cdot \nabla_H p + g \int_z^\infty \nabla_H (\rho \mathbf{v}) \, dz, \]  

(2.19)

known as Richardson’s equation from its use in the first numerical weather prediction experiment (Richardson, 1922).
The following hypothetical situation provides insight into how buoyancy results in vertical motion for the hydrostatic primitive equation set. Consider a finite bubble of air less dense than the surrounding air in an otherwise horizontally uniform and motionless atmosphere. It can be anticipated intuitively that buoyancy cannot exist without a simultaneous distribution of the pressure field. At the height of the base of the bubble the pressure is lower in the bubble than in the environment and we consider here a Gaussian pressure perturbation as shown by the red line in Fig. 2.1. The pressure perturbation is zero above the top of the bubble and increases throughout the depth of the bubble to a maximum at the base of the bubble. It follows from the hydrostatic relation that the maximum pressure perturbation extends throughout the depth of the atmosphere beneath the bubble.

The perturbation to the wind field can be expressed as $\Delta u \sim (-1/\rho) \nabla P \Delta t$ and so follows the form of the negative of the derivative of the pressure perturbation as shown in yellow in Fig. 2.1. Environmental air will accelerate horizontally towards the maximum in pressure perturbation throughout the depth of the atmosphere beneath the bubble. The horizontal divergence term in equation 2.19 ($\nabla_H \cdot \mathbf{v}$) follows the form of the negative of the second derivative of the pressure perturbation (shown in blue in Fig. 2.1) and, assuming all other terms in equation 2.19 are small.

![Figure 2.1: A Gaussian perturbation in the horizontal x-direction (red), the negative of the first derivative (yellow) and the negative of the second derivative (blue).](image-url)
(not shown) and the heating rate is zero, the vertical gradient of vertical velocity is positive beneath the centre of the bubble and negative around the edge.

The vertical gradient of vertical velocity is zero above the top of the bubble where the pressure perturbation is zero. Therefore the vertical velocity field remains constant with height at the maximum magnitude reached at the top of the bubble. The centre of the column within which the bubble exists will ascend and the surrounding air will descend. It follows that the higher the bubble is in the atmosphere the faster it will ascend.

(b) Model Integration

Computations are on a rotated latitude-longitude grid allowing for a grid-spacing of approximately 12.5km over a mesoscale domain of 146 × 182 grid lengths covering the UK (see Fig. 3.9 on page 66 for an example). Vertical resolution is on 38 η levels. Moisture variables are carried on the lowest 35 levels. Resolution is highest in the boundary layer where vertical gradients are expected to be large with 14 levels below approximately 800mb. Computations are on an Arakawa B grid using a conservative split-explicit finite difference scheme (Cullen and Davies, 1991). This splitting technique means the horizontal advection terms in the governing equations are integrated with a timestep limited by the windspeed, whilst the terms which describe gravity-inertia oscillations are integrated with a succession of shorter timesteps. Model physics are calculated with a timestep of 5 minutes that contains 4 advection timesteps. A more comprehensive description of the integration scheme than given here is provided by Cullen (1993).

The domain is forced by lateral boundary conditions generated by the Met Office global model updated every hour using a rim width of 8 grid points. The global model was run at a resolution of 0.83° longitude and 0.55° latitude giving 432×325 grid points. At the latitude of the UK, the dimensions of a global grid-box are approximately 60km by 60km. The global model has 30 vertical levels and model physics are calculated on a timestep of 10 minutes that contains 3 advection timesteps. The same parameterisation scheme using the same CAPE-closure timescale of two hours is used in the global and mesoscale models. The philosophy of this configuration is discussed later in section 2.5 and in chapter 4 section 4.5.2a.

The longwave and shortwave radiation scheme, based on the two-stream Edwards and Slingo (1996) code, is called every hour. The scheme includes the single scattering properties of liq-
uid and ice clouds for both longwave and shortwave radiation. The boundary-layer scheme uses 1st order eddy diffusion with a top-of-mixed-layer entrainment scheme and is non-local in unstable conditions. The surface exchange uses the Penman-Monteith formulation including four sub-surface levels. The model uses ancillary files for vegetation parameters, soil parameters, sea surface parameters, ozone and orography including an orographic roughness length to parameterise the effect of pressure anomalies due to flow over orography.

2.2 Explicit Cloud and Precipitation Processes

Large-scale cloud and precipitation processes are represented in the model by the Wilson and Ballard (1999) microphysically based mixed phase explicit precipitation and cloud scheme.

(a) The Cloud Scheme

The cloud scheme diagnoses \( q_{cl} \) and liquid cloud fractional cover for use with the precipitation scheme from input values of the cloud conserved variables \( \theta_l \) and \( q_l \) (defined in equations 2.4 and 2.5). The calculation of liquid cloud fraction and \( q_{cl} \) assumes a distribution of the cloud conserved variables about their grid-box mean as shown in Fig. 2.2. Fluctuations about the gridbox mean are not necessarily due to turbulence alone but rather to any unresolved motion. The scheme, based on the method of Smith (1990), assumes a triangular distribution of \( q_t \) with a width determined by the critical relative humidity (\( RH_c \)) and temperature. The critical relative humidity, typically 85%, is that required for the generation of \( q_{cl} \). For a given \( q_t \), lower values of \( RH_c \) broaden the distribution of \( q_t \) and increase \( q_{cl} \).

The cloud fraction (\( C \)) can be written in terms of the grid-box mean relative humidity (\( \overline{RH} \)) defined as \( q/q_{e}(T, p) \). For \( \overline{RH} \leq RH_c \), \( C = 0 \) and for \( RH_c < \overline{RH} < (5 + RH_c)/6 \), \( C = R^2/2 \) where \( R \) satisfies:

\[
R^3 - 6R + 6 \left( \frac{\overline{RH} - RH_c}{1 - RH_c} \right) = 0. \tag{2.20}
\]

The physically realistic root of this cubic gives:

\[
C = 4 \cos^2(\pi/3 + \phi/3), \tag{2.21}
\]
Figure 2.2: An example probability density function of $q_t$ within a gridbox (blue line) where $\overline{q}$ is the gridbox mean value and $q_s$ is the saturation specific humidity. The cloud scheme diagnoses $q_{cf}$ as determined by the shaded region.

where:

$$
\phi = \cos^{-1}\left[\frac{3}{2^{3/2}} \left(\frac{\overline{RH} - RH_c}{1 - RH_c}\right)\right]. \tag{2.22}
$$

For $(5 + RH_c)/6 \leq \overline{RH} \leq 1$:

$$
C = 1 - \left[\frac{3}{2^{3/2}} \left(\frac{1 - \overline{RH}}{1 - RH_c}\right)\right]^{2/3}. \tag{2.23}
$$

At $\overline{RH} = (5 + RH_c)/6$ both equations 2.21 and 2.23 give $C = 0.5$ and so provide a continuous function. The impact of the critical relative humidity on the stability of an example thermodynamic profile is shown schematically in Fig. 2.3. The example shows a situation for which the inclusion of $RH_c$ changes a stable profile to an unstable profile. If air from 1000mb with a temperature of 298K and a specific humidity of $8\text{gkg}^{-1}$ is forced to ascend it will reach 100% relative humidity at 800mb (determined using Normand’s construction) and the air will saturate. Further forced ascent will follow a moist adiabat with a wet-bulb potential temperature ($\theta_w$) of 289.0K (shown by the red line in Fig. 2.3) but the parcel will not achieve positive buoyancy at any level. If $RH_c$ is set to 85% then for the same parcel some water vapour will convert to liquid water above 850mb. Subsequent forced ascent will not follow a moist adiabat at $\theta_w = 290.5$K (shown by the green line in Fig. 2.3) since the parcel is not saturated. Subsequent parcel ascent will follow a profile with
a lapse rate between that of moist and dry adiabat profiles and, for the example thermodynamic
profile given in Fig. 2.3, the parcel will become positively buoyant above 800mb.

In addition to $q_{cl}$ and a liquid cloud fraction the cloud scheme also calculates an ice cloud fraction
using a similar approach ($q_{cf}$ is predicted by equation 2.8). When mixed phase cloud exists,
both liquid and frozen ice cloud fractions will be non-zero. The total cloud fraction is calculated
assuming minimum overlap between the liquid and ice clouds. The total cloud fraction, the liquid
water content and the ice water content are used in the radiation scheme for clear-sky calculations.

(b) The Precipitation Scheme

Explicit precipitation is generated by a physically-based, mixed-phase transfer scheme represent-
ing four water species; vapour, liquid droplets, raindrops and ice. Owing to computational con-
straints, the ice variable describes all frozen water that can occur in nature including snow, ice
crystals and rimed particles. The transfer terms linking these species, detailed in Wilson and
Ballard (1999), are shown schematically in Fig. 2.4. Some schemes (e.g. Tiedtke, 1993) use tem-
perature to partition condensate into liquid and ice whereas a more physical method is used here.
It is recognised that liquid water responds to changes in supersaturation on a timescale small com-
pared to the model timestep and as such its diagnostic treatment is justified and is diagnosed by the
cloud scheme. Ice, however, adjusts to equilibrium on a timescale of tens of minutes and as such
is treated prognostically by equation 2.8. Ice is allowed to fall between layers and be advected by
the model’s tracer advection scheme.

The dominant microphysical processes involved in determining surface precipitation in regions of
deep convection are likely to include the autoconversion of rain from liquid water, the accretion of
liquid water by raindrops and the evaporation of rain. Cloud droplets combine to form drops large
enough to fall as rain. This process is extremely difficult to quantify and depends on many factors,
such as the age of the cloud. The autoconversion rate uses the formulation of Tripoli and Cotton
(1980):

$$
\frac{dq_{cl}}{dt} = \frac{4\pi g E_c \rho^{1/3} 3^{7/3} q_{cl}^{7/3}}{18(4\pi/3)^{1/3} \mu (N_c \rho_{liq})^{1/3}},
$$

(2.24)

where $E_c$ is the collision/collection efficiency (assumed to be 0.55), $g$ is the acceleration due
to gravity, $\mu$ is the dynamic viscosity of air, $N_c$ is the number of droplets per cubic metre and
$\rho_{liq}$ is the density of water. An approximate value of $N_c = 3 \times 10^8 \text{m}^{-3}$ is adjusted slightly to take
Figure 2.3: A schematic tephigram showing profiles of environment temperature and dew-point temperature (black solid and dashed lines). The red lines show adiabatic parcel ascent from the surface and the blue lines show a parcel ascent from the surface that includes some conversion of water vapour to liquid water above a critical relative humidity of 85%. The blue and red lines are coincident below 850mb.
Figure 2.4: A schematic diagram showing the four water species and the modelled transfers between them (Wilson and Ballard, 1999).

into account land and sea variations in the amount of cloud condensation nuclei. The minimum water content before autoconversion can occur is given by considering a critical drop size. When the radius of a typical drop exceeds $7 \times 10^{-6}$ m, autoconversion can occur. Autoconversion is not allowed to decrease liquid water below this minimum liquid water content.

Raindrops falling through liquid water cloud can collect droplets when collisions occur. Assuming all collisions result in collection and the collision cross-section is that of the falling raindrop, the expression for the rate of change of mass contained in rain ($m$) is:

$$\frac{dm}{dt} = \frac{\pi}{4}D^2 v_{\text{rain}}(D) \rho_{\text{fl}},$$  \hspace{1cm} (2.25)

where $D$ is the diameter of the raindrop and $v_{\text{rain}}(D)$ is the fall-speed of the raindrop (from Sachidananda and Zrinč, 1986). This is integrated across the rain drop size distribution (from Marshall and Palmer, 1948) to obtain a rate of change of rain mass. In addition, the rate of change of rain mass is affected by evaporation in subsaturated air. This is represented by a diffusion term and depends on the sub-grid supersaturation calculated using the liquid cloud fraction from the cloud scheme. Finally, the change in rain mass is converted to a change in rain rate using the
diagnostic expression:

\[ \Delta \text{Rainrate} = \frac{\rho \Delta z \Delta \text{Rainmass}}{\Delta t}. \] (2.26)

Starting from the top level, working downwards, the transfer terms calculate new values of vapour, liquid and ice and diagnoses precipitation. Ice is allowed to fall to lower levels. The Wilson and Ballard (1999) precipitation scheme is based on that of Rutledge and Hobbs (1983) who presented a parameterisation of growth processes involved in the enhancement of precipitation in 'seeder-feeder' type situations such as warm frontal rainbands. However, the important mechanisms for precipitation growth under widespread lifting on the order of 0.1 ms\(^{-1}\) over \(\sim 100\) km are likely to be quite different from those in deep convective systems. Indeed, Lin et al. (1983), using a complex microphysical scheme in a cloud-resolving model, found that a snow variable is crucial in predicting the correct rainfall characteristics for a deep convective system.

### 2.3 Parameterised Convection

The Gregory and Rowntree (1990) mass-flux scheme calculates the effects of an ensemble of either shallow, mid-level or deep convection on the grid-scale variables. The scheme can represent both moist and dry convection and was originally formulated for use in lower resolution models. The scheme currently runs operationally in the Met Office mesoscale model.

**(a) The Trigger Function**

To reduce computation a stability test is performed to select grid points to pass to the trigger function. Looping over model levels from the surface upwards, grid points are selected for which convection may be possible either because stability is low enough or because convection is already occurring as determined by the previous loop. The stability criterion is determined using an approximate parcel potential temperature at level \(k+1\) \((\theta_p^{k+1})\) calculated on the assumptions that no entrainment occurs during lifting from level \(k\) and that the mixing ratio of the parcel, if saturated in layer \(k+1\), is equal to the saturation mixing ratio at the environment potential temperature \((q_s(\theta_E^{k+1}))\):
CHAPTER 2: Modelling Strategy

\[ \theta_P^{k+1} = \theta_E^k + \text{MAX} \left( 0, \left( q^k - q\theta_E^{k+1} \right) \left( \frac{L_v}{C_P \Pi^{k+1}} \right) \right). \]  \hspace{1cm} (2.27)

If:

\[ \theta_P^{k+1} - \theta_E^{k+1} > -1.5 \]  \hspace{1cm} (2.28)

then it is deemed that convection may occur. The trigger function is then applied to the selected points. For convection initiating within the boundary layer a test ascent is carried out first to ascertain, using the depth of the cloud, whether deep, mid-level or shallow convection is possible.

- Shallow Convection: a parcel reaching neutral buoyancy in the boundary layer.
- Mid-level Convection: a parcel initiating above the boundary layer.
- Deep Convection: a parcel maintaining positive buoyancy above the boundary layer.

The scheme includes an option to vary parameters in the trigger function and the convective mass-flux calculation according to the type of convection. For this study parameters are set independent of convection-type as in the standard scheme. Convection is triggered at grid points where a parcel taken from the environment and given a 0.2K buoyancy excess remains buoyant by 0.2K or greater in the next layer after ascent. Entrainment of environmental air and latent heating due to any phase changes of water are included in the parcel ascent. Buoyancy \( B \) is defined as the difference in virtual potential temperature between the parcel and the environment:

\[ B = \theta_P^{k+1}(1 + (\epsilon^{-1} - 1)q_P^{k+1}) - \theta_E^{k+1}(1 + (\epsilon^{-1} - 1)q_E^{k+1}) \]  \hspace{1cm} (2.29)

where \( \theta_P \) and \( \theta_E \) are the parcel and environment potential temperatures, \( q_P \) and \( q_E \) are the parcel and environment mixing ratios and \( \epsilon \) is the ratio of molecular weights of water and dry air.

The triggering of convection is determined by the change in buoyancy over one model layer. Consider a hypothetical profile with a step function in \( \theta \) that represents a layer of CIN. If we hypothesise a perfect model then increasing the model vertical resolution will better resolve the vertical profile as shown in Fig. 2.5. The potential temperature at model level \( k \) (\( \theta^k \)) is obtained by the simple averaging operator:

\[ \theta^k = 1/2(\theta^{k+1/2} + \theta^{k-1/2}). \]  \hspace{1cm} (2.30)
For an example resolution of \( d/5 \) the model well resolves the layer of CIN of depth \( d \) as shown in Fig. 2.5a. As the resolution is decreased the number of model levels within the layer of CIN decreases and the layer becomes less well resolved, as shown in Figs. 2.5b and 2.5c. As resolution decreases the likelihood of triggering the convection scheme increases. Owing to the decrease in maximum magnitude of \( \theta \), a parcel lifted from below the layer of CIN will be more buoyant at lower resolutions. This effect may be sufficient to trigger convection at low resolutions that was suppressed at high resolutions. To achieve resolution independence in a perfect model the trigger function must be based on vertically integrated CIN so as to remove dependence on the model grid scales. For an imperfect model, the situation becomes more complicated because triggering will also be affected by any significant differences in the structure of the CIN layer at different resolutions.

The vertical spacing between levels in the Unified Model at the top of the boundary layer is approximately 200m. The model will only resolve scales significantly greater than two vertical grid lengths or at least 1km at the top of the boundary layer. Shallow layers of CIN that can be sufficient to suppress convection in severe storm environments (e.g. Stensrud and Fritsch, 1993; Brandes,

![Figure 2.5](image-url)

Figure 2.5: Schematic showing how a step function in \( \theta \) (black line) of depth \( d \) is approximated on a vertical grid (red crosses and red dashed lines) using three different vertical grid spacings (\( \Delta Z \)). The blue circles and blue dashed line shows the effect of offsetting the vertical model levels with respect to the step function.
1990) will be poorly resolved. The simple example shown in Fig. 2.5 gives an indication that the model may only capture an indication of CIN, if at all, in such situations and the convection scheme may trigger for situations where convection was suppressed in the real atmosphere.

(b) The Cloud Model

Once triggered, the rest of the parcel ascent and the effect of convection on the large-scale atmosphere is calculated using a cloud model. A one-dimensional bulk cloud model based on a single one-dimensional entraining and detraining plume is claimed to represent the summation over an ensemble of clouds with different characteristics. The updraught mass-flux represents the sum over the ensemble and parcel characteristics represent the ensemble average. The cloud model equations are given here with $\sigma$ as the vertical co-ordinate (as in Gregory and Rowntree, 1990) because the scheme was originally coded for a previous version of the mesoscale model that used $\sigma$. The vertical co-ordinate is not important since the scheme uses layer thickness in terms of pressure. Equations for the mass-flux ($M_e$), potential temperature ($\theta$), specific humidity ($q$) and cloud liquid water ($l$) for the bulk model are defined as:

$$\frac{\partial M_e}{\partial \sigma} = (E - N - D),$$  \hfill (2.31)  

$$\frac{\partial \theta e M_e}{\partial \sigma} = (E \theta e - N \theta_n - D \theta_r) + \frac{LQ}{C_p},$$  \hfill (2.32)

$$\frac{\partial q e M_e}{\partial \sigma} = (E q e - N q_n - D q_r) - Q,$$  \hfill (2.33)

$$\frac{\partial l e M_e}{\partial \sigma} = -(N l_n + D l_r) + Q - PR,$$  \hfill (2.34)

where

$E$ = entrainment rate  
$N$ = mixing detrainment rate  
$D$ = forced detrainment rate  
$\phi_c = \phi$ in cloudy air
\( \phi_e = \phi \) in environmental air
\( \phi_n = \phi \) on mixing detrainment
\( \phi_r = \phi \) on forced detrainment

\( Q = \) conversion of water vapour to liquid water and ice
\( PR = \) liquid water and ice precipitated

The vertical integration of the bulk equations requires knowledge of the cloud-base mass-flux and of the entrainment and detrainment rates. The cloud-base mass flux is calculated using a closure assumption and is described later. The entrainment and detrainment rates are given by:

\[
E = \epsilon M_c, \quad N = \nu M_c, \quad D = \delta M_c, \quad (2.35)
\]

where \( \epsilon, \nu \) and \( \delta \) are the fractional mass entrainment and detrainment coefficients. The entrainment profile is set as \( \epsilon = 3A_E\sigma \) where \( A_E = 1 \) in the lowest layer and \( A_E = 1.5 \) in all other layers. At the surface, \( \sigma = 1 \) and \( \epsilon = 4.5 \) and near cloud top, \( \sigma = 0.2 \) and \( \epsilon = 1 \). Strong entrainment at low-levels reflects turbulent shallow convection and weak entrainment at upper levels represents less rigorous mixing of deep convection. The mixing detrainment formulation, \( \nu = \epsilon \sigma (1 - A_E) \), is such that \( \nu = 0 \) in the lowest layer. With the above formulations of entrainment and mixing detrainment into the cloud, mass-flux increases with height which is expected of a buoyant parcel. This increase in mass-flux with height is restricted for the ensemble mean by the addition of a second detrainment process. Terminal detrainment acts only at the highest level of the plume and represents the termination of weaker clouds in the ensemble. If, after ascent including entrainment, mixing detrainment and moist processes, the excess buoyancy of the parcel is less than a critical value, \( b \), terminal detrainment is initiated. The portion detrained into the environment is sufficient to allow the parcel to be buoyant by the amount, \( b \). In this sense, cloud varying characteristics are taken into account. Although the value of \( b = 0.2K \) is based on explicit modelling studies of tropical convection (Tao et al., 1987), single column model tests by Gregory and Rowntree (1990) showed the scheme was robust to the value of \( b \). The bulk parcel is totally detrained into the environment at the level of neutral buoyancy for an undilute parcel or when the mass-flux falls below a critical magnitude.

As mentioned in chapter 1, there are significant advantages to schemes that directly represent moist convective downdraughts. A simple one-dimensional inverted entraining and detraining plume model represents transport of heat and moisture within convectively driven downdraughts.
The entraining and detraining plume equations are similar to those used for the updraught and so are not reproduced here. A buoyancy analysis of an equal mixture of cloudy updraught and unsaturated environment air is carried out at each level. Unsaturated parcels are brought to saturation through evaporation and/or sublimation of precipitation. A parcel then possessing negative buoyancy triggers a downdraught. Negative buoyancy and saturation are maintained through evaporation and melting of falling precipitation. The effects of water-loading are only crudely simulated by allowing the downdraught to continue with small positive buoyancy. The initial downdraught mass-flux is proportional to the updraught mass-flux at a reference level:

\[ M_{\text{init}}^{DD} = \alpha M_{\text{ref}}^{UD}, \]

where \( \alpha \) is chosen to be 0.05 based on explicit cloud modelling results.

Downdraught entrainment rates are constant with model level but are enhanced at the freezing level, justified by the observation that the majority of downdraughts initiate at this level where a minimum in equivalent potential-temperature often exists. As with updraughts, downdraughts interact with their environment through entrainment and detrainment of heat and moisture and compensating vertical motion within clear air. Finally, if a downdraught penetrates to within 100mb of the surface, the entrainment is set to zero while the portion of downdraught air detrained into each layer is proportional to the layer thickness.

(c) Cloud Microphysics

The development of precipitation within deep convection in nature occurs mainly by the coalescence of smaller drops of water with larger ones. The rate at which this occurs depends upon the distribution of the aerosol size spectrum. Clouds that form in regions with greater numbers of large nuclei (\( r > 20 \mu m \)), as in maritime air-masses, precipitate earlier than clouds that form in continental air-masses that typically have higher concentrations of small nuclei (Ludlam, 1980). In addition, precipitation is formed by the growth of hail from the freezing of supercooled droplets in deep clouds whose tops fall below 263K. These observations are reflected in the convection scheme through suppressed precipitation of condensed water until the cloud exceeds a critical depth (\( D_{\text{crit}} \)) and cloud condensate exceeds a critical value (\( l_{\text{min}} \)). \( l_{\text{min}} \) is set to \( 1 gkg^{-1} \) or the saturation specific humidity where it is less and \( D_{\text{crit}} \) is defined by:
Until these criteria are met no precipitation occurs and cloud condensate is stored in the updraught. The effects of water loading on parcel buoyancy due to the stored condensate are neglected which may result in excessively buoyant parcels. The amount of precipitation produced in lifting the bulk parcel from level \textit{k} to level \textit{k+1} (\(P^k\)) is defined as:

\[
P^k = (l_p^{k+1} - l_{min})M^{k+1} \frac{p^*}{g} \quad (2.36)
\]

The parcel therefore retains the amount \(l_{min}\) of condensate. Total precipitation falling from cloud-base is diagnosed by a simple summation of precipitation generated by lifting each layer individually. Freezing and melting of falling precipitation occurs at the freezing level of the environment and associated temperature increments are added accordingly. A proportion of the precipitation created in the updraught is assumed to fall through the downdraught, undergoing evaporation as it does so. The proportion transferred at any level is based on maintaining continuity of precipitation mixing ratios between the updraught and downdraught. The remaining precipitation is assumed to fall through cloudy air and is allowed to evaporate below cloud-base only at a rate proportional to the subsaturation of the environment. Evaporation is constructed by integrating the fractional evaporation of precipitation (\(J\)) over a layer of depth \(dz\) as:

\[
\frac{dJ}{J} = \beta(q - q_s)\rho \frac{dz}{Q} \quad (2.37)
\]

where \(\rho\) is the density of air, \(\beta\) is a constant set to 10^{-3}s^{-1} (based on Kessler (1969)) and \(Q\) is the rainfall rate estimated by assuming precipitation covers 10% of the grid area. The value of 10% is chosen because it represents a typical value from an analysis of GATE phase III ship and radar rainfall data (Gregory and Rowntree, 1990). The rainfall rate \(Q\) is therefore ten times the grid-box rainfall at cloud base (\(J_B\)). In subsaturated air the precipitated water (\(J^k\)) reaching level \(k\) from level \(k+1\) is given by:

\[
J^k = J^{k+1} \exp \left( -\beta p^* \Delta \sigma q_s \frac{T^k_{E}}{10J_B g} \right). \quad (2.38)
\]
Finally, the mixing ratio and temperature increments due to evaporation over a timestep ($\Delta t$) are defined as:
\[
\Delta q^k = \frac{(j^{k+1} - j^k) g \Delta t}{p^* \Delta \sigma^k}, \quad (2.39)
\]
\[
\Delta T^k = -\frac{L}{C_p} \Delta q^k. \quad (2.40)
\]

Evaporative cooling may be sufficient to introduce a negatively buoyant anomaly. The large-scale dynamics will respond to this anomaly over a timescale of hours for explicit motions through the mechanism described in section 2.1a resulting in descent. This will stabilise the grid element profile to both parameterised and explicit convection. The impact of the assumptions made in the calculation of parcel buoyancy (including the treatment of precipitation) is discussed later in section 2.4.2c and is compared with the buoyancy assumptions made in the large-scale equations.

**(d) Closure**

An adjustment closure is applied such that the mass-flux is scaled to reduce CAPE to zero over a given CAPE-closure timescale, $\tau$. Once triggered, the cloud-base mass-flux is therefore calculated from relaxing the atmosphere back to a stable state. This closure is based on the work of Fritsch et al. (1976) who recognised that convection removes CAPE on a timescale small compared to its generation by large-scale processes. CAPE is defined as:
\[
CAPE = \int_{\text{cloud}} \left( \theta^\text{v,parcel} - \theta^\text{v,environment} \right) \frac{dp}{\rho}, \quad (2.41)
\]
where $\theta_v$ is the virtual potential temperature. Parcel ascent follows a moist virtual adiabat and the exclusion of condensate drag on parcel buoyancy may result in excessive CAPE. Assuming steady state clouds, the rate of change of CAPE with time due to convective activity is given by:
\[
\left( \frac{\partial \text{CAPE}}{\partial t} \right)_\text{conv} = -\int_{\text{cloud}} \frac{\partial \theta^\text{v,environment}}{\partial t} \frac{dp}{\rho}. \quad (2.42)
\]
Finally, the net changes to grid-scale potential temperature, $\theta$, and specific humidity, $q$, over a timestep are calculated from the effects of compensating subsidence and forced and mixing detrainment of heat and moisture. Changes of phase and evaporation of precipitation also contribute to changes in grid-scale $\theta$ and $q$. The changes to either $\theta$ or $q$ (written as $\phi$ here) due to convection can be expressed as:

$$\frac{\partial \phi}{\partial t} \bigg|_{\text{convection}} = \frac{\phi^E - \phi^\theta}{\tau},$$

(2.43)

where $\phi^\theta$ is the value of a grid variable before convection, $\phi^E$ is the grid value in an equilibrium state and $\tau$ is the timescale over which the adjustment occurs.

### 2.4 Idealized Studies

To aid the understanding and interpretation of model simulations of the two case studies presented in chapters 4 and 5 a set of hypothetical idealized experiments are discussed. The aim is to determine how the model divides the convective mass-flux between explicit motions and the convective parameterisation scheme. The first experiment determines how absolute instability associated with a saturated boundary layer anomaly is removed in the model. The second experiment determines the effect of increasing the magnitude of a warm and moist boundary layer anomaly on the thermodynamic profile and the model’s response.

#### 2.4.1 A Saturated Boundary Layer Anomaly

Consider a horizontally homogeneous motionless initial state. The state has a conditionally unstable profile, with CAPE, but also significant CIN. A boundary layer anomaly is introduced, rendering a group of grid cells saturated and thus absolutely unstable (i.e. no CIN). The situation is shown schematically in Fig. 2.6. Two hypothetical extreme cases are considered; one in which
the model removes all instability via the parameterisation scheme and the other in which the model removes all instability via explicit motions.

(a) **All instability removed via the parameterisation scheme.**

If the CAPE-closure timescale of the parameterisation scheme is set to one model timestep the scheme behaves more like a true equilibrium scheme since convective mass-flux comes into balance with the large-scale destabilisation, in a vertically integrated sense, on a timestep by timestep basis. Under this formulation it is intended that the scheme will remove all instability via the parameterisation scheme. The response of the model follows the steps below.

- Firstly, the absolutely unstable profile will satisfy the low stability criterion prior to the trigger function as described on page 37.
- Secondly, the profile has zero CIN and will therefore trigger the scheme from the lowest

![Figure 2.6: Schematic showing the hypothetical conditionally unstable environment and a group of grid cells with a saturated boundary layer anomaly (shaded blue). The schematic profiles of temperature are based on a tephigram and show the environment temperature in black and the temperature for an adiabatic parcel ascent from the surface in red.](image-url)
model layer since such a parcel will satisfy the criterion of having buoyancy greater than 0.2K (using equation 2.29).

- Once triggered, the scheme calculates a cloud-base mass-flux that is consistent with the removal of vertically integrated CAPE. Assuming a perfect cloud model, the resulting profile will be stable. The cloud model distributes the convective mass-flux in the vertical and predicts the convective heating and moistening profiles due to forced detrainment, mixing detrainment and subsidence induced in the cloud environment that compensates for the updraught mass-flux. In addition, changes of phase of precipitation below cloud base, as discussed in section 2.3c, contribute to the heating and moistening tendencies.

- The large-scale thermodynamic equations 2.4 and 2.5 are incremented by the convective tendencies to $\theta$ and $q$ in the forcing terms $F_{\theta_i}$ and $F_{q_i}$. The prognostic equation for ice, given in equation 2.8, does not respond to parameterised convection directly because the convection scheme outputs water vapour and not condensate.

- A warm and moist anomaly exists at mid-levels and a cool and dry anomaly exists at low levels due to parameterised convection. At the second timestep explicit motions respond to parameterised convection through the modified thermodynamic profile. The dynamics will react to the horizontal gradient in buoyancy associated with the anomalies and, through the mechanism described in section 2.1a, will cause the positive anomaly to ascend and the negative anomaly to descend. This dynamical adjustment in stable conditions is associated with small vertical velocities and should not be thought of as explicit convection which requires instability.

- If the positive anomaly exceeds $RH_c$ as it rises the large-scale cloud scheme will diagnose $q_{cl}$ and the transfer terms in the large-scale precipitation scheme will act on the input values of $q$, $q_{cl}$ and $q_{cl,f}$ and calculate new values and diagnose precipitation.

Parameterised convective mass-flux is limited to the mass of the layer for parcels originating from the lowest model layer. For large CAPE the maximum mass-flux may only be sufficient to remove a fraction of the CAPE. The parameterisation scheme will then be called at subsequent timesteps, assuming the scheme can be triggered, to remove the remaining instability. If the parameterised convective mass-flux is insufficient to remove the instability within the timescale for explicit motions then explicit convection may develop as described in the next subsection.
A perfect cloud model would result in a stable thermodynamic profile over the CAPE-closure timescale of one timestep. However, the Gregory-Rowntree cloud model removes CAPE in a vertically integrated sense and the mass-flux profile is constrained through a prescribed entrainment profile. Therefore the cloud model is not likely to produce a stable thermodynamic profile. If the cloud model over-stabilises the upper troposphere and leaves an absolutely unstable profile at lower levels then the scheme will trigger at the next timestep. On the other hand, the cloud model may over-stabilise the lower levels and leave an absolutely unstable profile at mid-levels. Indeed, in single column model tests Gregory and Rowntree (1990) showed the scheme produced an excessively stable boundary layer due to the evaporation of precipitation below cloud-base only. This layer of CIN may be sufficient to suppress parameterised convection at the next timestep. If this were the case then explicit convection may develop as described in the next subsection.

(b) All instability removed via explicit motions.

One way to force the model to remove all instability via explicit motions is to set the CAPE-closure timescale to a value large compared to the timescale of hours for explicit motions. The model’s response to the saturated boundary layer anomaly follows the steps below.

- As for the previous case the parameterisation scheme will trigger at the grid points with the saturated boundary layer anomaly but the heating and moistening tendencies will be spread over many more timesteps such that for a single timestep the absolutely unstable profile will barely change.

- Through the mechanism described in section 2.1a the dynamics will respond to the absolutely unstable region and cause the saturated boundary layer anomaly to rise.

- This hypothetical finite bubble will rise to the level where the horizontal gradient of buoyancy is zero and compensating descent in surrounding grid points will develop through mass continuity. However, for a plume rising from a saturated boundary layer of infinite horizontal extent (a situation more similar to nature in large-scale unstable regions) the vertical extent of the region of ascent will increase on the timescale for explicit motions of hours.

- From the first timestep the saturated boundary layer anomaly will exceed $RH_c$. As a result the large-scale cloud scheme will diagnose cloud liquid water amount and the large-scale precipitation scheme will diagnose precipitation.
Throughout the development of explicit convection the parameterisation scheme will trigger at every timestep but its effect on the thermodynamic profile will be negligible.

Another mechanism by which the model may preferentially remove instability via explicit motions results from the inconsistent set of assumptions concerning parcel buoyancy between the large-scale equations and the convective parameterisation scheme. This affects the stability of a given profile to large-scale and parameterised convection. Perhaps the most serious difference is the ability of the large-scale dynamics to appreciate sub-grid variability through the cloud scheme. For the Wilson and Ballard (1999) scheme used in this study, this is achieved through the use of the critical relative humidity. Convective parameterisation parcel ascents often have appreciable CIN since the parcel starts with grid mean properties (albeit with slightly enhanced buoyancy in the standard scheme). The impact of a critical relative humidity on the stability of a given profile has already been discussed in section 2.2a. A profile can be conceived such as that shown in Fig. 2.3 which is stable to parameterised convection parcel ascent but which is unstable to the large-scale dynamics. Whereas the parameterisation scheme is suppressed the large-scale cloud scheme diagnoses condensate, some of which may convert to precipitation through the large-scale precipitation scheme. To reconcile this inconsistency it may be sensible to incorporate saturated in-cloud properties from the large-scale cloud scheme into the trigger function of the parameterisation scheme so that it may also appreciate subgrid variability. However, parameterised convection may not be suppressed over the timescale for explicit convective development since the thermodynamic profile will be modified at each timestep.

### 2.4.2 Positive Boundary Layer Buoyancy Anomalies

The discussion now moves on to a more complicated case in which a group of grid cells are given an increase in boundary layer temperature and an associated increase in specific humidity such as to maintain cloud-base height, introducing a positive buoyancy and CAPE anomaly. The effect of increasing the magnitude of this boundary layer anomaly on the thermodynamic profile and parcel buoyancy is shown schematically in Fig. 2.7. The discussion below covers the rôle of the trigger function, the closure and microphysics as the magnitude of the anomaly is increased. The aim of this exercise is to understand the rôle of components of the convection scheme and not to tune parameters.
CHAPTER 2: Modelling Strategy

Figure 2.7: Schematic showing a hypothetical conditionally unstable profile (black) and a parcel ascent from the surface. The impact of increasing the magnitude of the boundary layer temperature and specific humidity anomaly is shown for two example magnitudes (blue and red). Temperature is shown by thin solid lines, dew-point temperature is shown by dashed lines and parcel profiles from the surface (thick solid lines) follow adiabatic ascent.
(a) The Trigger Function

As the magnitude of the anomaly is increased, the magnitude of vertically integrated CIN decreases as shown in Fig. 2.7. In addition, the maximum magnitude of CIN within a model layer will decrease. As already discussed on page 39 a parcel lifted from below the layer of CIN will become increasingly buoyant at the next model level after ascent as the magnitude of CIN in the layer decreases. The trigger function requires that a parcel be positively buoyant by an amount $b$ (set to 0.2K in the standard scheme) at the next level after ascent. Parameterised convection will therefore trigger when the magnitude of the anomaly is such that CIN is reduced sufficiently to provide the parcel with buoyancy greater or equal to $b$. If $b$ is reduced and the parcel still receives an initial buoyancy excess of 0.2K at the previous level before ascent then the parameterisation scheme will trigger for smaller anomalies. In single column model tests Gregory and Rowntree (1990) examined the sensitivity of the scheme to $b$. In increasing $b$ to 1K the resulting profile was cooler by 1K - 1.5K than that using the standard scheme throughout much of the cloud layer. This compensates for the increase in forced detrainment due to the requirement of a parcel to be buoyant by greater than $b$ at each level.

For small magnitude anomalies that do not decrease CIN sufficiently to trigger the convection scheme the dynamics will respond to the imposed horizontal buoyancy gradient and cause the anomaly to ascend. If buoyant parcels are characterised by ascent then it may be sensible to formulate the trigger function to include a dependence on the grid element vertical velocity. This may provide a useful representation of subgrid-scale forcing since regions of ascent are likely to be more turbulent, and therefore have greater subgrid-scale variability, than regions of negligible vertical velocity. In a manner similar to the Kain and Fritsch (1993) parameterisation scheme, the trigger function could be formulated such that a parcel is given an additional buoyancy increment due to the grid element vertical velocity in the next layer after ascent. For the hypothetical anomaly the modified trigger would permit parameterised convection for smaller magnitude anomalies. However, the triggering may be delayed due to the timescale of hours for explicit motions.

(b) The Closure

Increasing the magnitude of the anomaly impacts the closure through the increase in CAPE as shown in Fig. 2.7. Increasing the magnitude of the anomaly simultaneously decreases CIN and increases CAPE. Although parcel ascents are shown for parcels originating from the surface it
is more likely that the convection scheme will trigger near the top of the anomaly since CIN is much less for parcels originating near the top of the anomaly than for parcels originating from the surface. This is because the effect of the anomaly is to change the buoyancy of a parcel above the anomaly only. For a constant CAPE-closure timescale and assuming the scheme triggers for each size of anomaly the effect increasing the magnitude of the anomaly will be to increase the magnitude of the cloud-base mass-flux. Therefore the cloud model will calculate larger convective heating and moistening tendencies in order to remove larger CAPE over a constant timescale.

(c) Microphysics and Water Loading

The impact of water loading on parcel profiles can be significant in reducing both CAPE and cloud-top height in nature. The scheme suppresses precipitation until liquid condensate reaches a critical value of $1 \text{g kg}^{-1}$ and yields 100% efficient precipitation thereafter, as discussed in section 2.3c. The discontinuous formulation will have no effect on the parcel ascent due to water loading since the neutral profile is based on a moist virtual adiabat and neglects water loading. As a result, the parcel CAPE and therefore the cloud-base mass-flux may be overestimated in the Gregory and Rowntree (1990) scheme for each magnitude of anomaly. On the other hand, neglecting the heat capacity of condensate for mixing ratios greater than $1 \text{g kg}^{-1}$ may have a significant impact on the parcel profile. This increment to latent heating may be stored in deep convective clouds in nature whereas the parameterisation scheme does not store condensate greater than $1 \text{g kg}^{-1}$ which may result in a profile that is too stable and too little CAPE. The impact of neglecting water loading and the heat capacity of condensate greater than $1 \text{g kg}^{-1}$ in the model will increase for increased magnitude of the anomaly due to the increased capacity of stronger updraughts in nature to store condensate.

The large-scale equations and the parameterisation scheme have different representations of microphysical processes (as presented in sections 2.2b and 2.3c). The parameterisations are necessarily complex which makes it difficult to hypothesise the net effect of both microphysics schemes. Perhaps an important similarity is the suppression of precipitation until a critical amount of condensate is reached. This critical value is a constant for the parameterisation scheme but is a function of the concentration of cloud condensation nuclei for the large-scale precipitation scheme. However, this difference will only impact buoyancy through the heat capacity of condensate and not through water loading since this is neglected in both schemes.
2.5 Method

The method used to answer the first three questions posed in chapter 1 is now described. The Gregory and Rowntree (1990) scheme adjusts to equilibrium over the prescribed timescale, $\tau$. This timescale parameter implicitly determines the partitioning between explicit and parameterised convection for a given environment. As mentioned in chapter 1, the physical processes that determine $\tau$ are not well known. The choice of $\tau$ for mesoscale modelling is therefore not clear. Using different values of $\tau$ in simulations of the two MCSs, the behaviour of parameterised and explicit convection in both an equilibrium and a non-equilibrium situation is explored. For each case study, three experiments are performed:

1. The 10 minute experiment: $\tau = 10$ minutes.
2. The 2 hour experiment: $\tau = 2$ hours.
3. The 1 day experiment: $\tau = 1$ day.

The 1 day experiment has a CAPE-closure timescale large compared to that of the large-scale forcing and is expected to produce a fully explicit convective solution. At the other extreme, the 10 minute experiment has a CAPE-closure timescale small compared to that of the large-scale forcing and is expected to provide a fully parameterised convective solution. An intermediate solution with significant amounts of both parameterised and explicit convection is expected from the 2 hour experiment. Boundary conditions in all experiments are generated by the global model using the Gregory and Rowntree (1990) convection scheme and a 2-hour CAPE-closure timescale as described in section 2.1b. The boundary conditions are also influenced by errors due to the convection scheme and are therefore not entirely independent. It is expected that the solution will be more severely constrained by the boundaries for the 1 day and 10 minute experiments than for the 2 hour experiment due to the difference in CAPE-closure timescale. Experimentation with the position of the western boundary of the mesoscale domain for case 1 showed only weak sensitivity of the details of convection. For case 2, convection developed near the centre of the domain where boundary constraints are small. However, 18 hours after initiation convection in nature tracked close to the eastern boundary where solutions are expected to be more constrained.

As discussed in chapter 1, the small-scale details of convection including location, timing, intensity and evolution are compared between the experiments and observations. The magnitude and
vertical structure of the large-scale modification is then compared between the 10 minute and 1 day experiments and observations. Results are presented in chapter 4.

Two further experiments test the sensitivity of the convective solutions to the trigger function in the parameterisation scheme. The experiments are designed to identify any constraints imposed by the trigger function on the timing and location of convection in the 10 minute, 2 hour and 1 day experiments. The details of the experiments and results are presented in chapter 4.

### 2.6 Model Initialisation

A start time such that convection develops after the period of model spin-up is essential to this investigation. Appendix B explores the effect of model spin-up for case 1 by comparing timeseries of hourly rain amounts between simulations from two start times. To ensure convection initiated after the period of model spin-up, the model was initialised at 1800 GMT 28 May 1999 for case 1. No Met Office analyses were available at 1800 GMT 28 May 1999 so an ECMWF analysis at 1.0° horizontal resolution and on 50 vertical levels was used. Initial conditions for the mesoscale model were derived from an interpolation of ECMWF analysis data onto the mesoscale domain. No additional data were assimilated because of the great difficulty involved. The model was forced by boundary conditions generated from the global Unified Model which was itself initialised from the ECMWF analysis.

For case 2, the model was initialised at 0600 GMT 11 September 2000 from a Met Office mesoscale analysis, 12 hours prior to the development of convection thereby avoiding any period of model spin-up. Data were assimilated using the 3DVAR approach (described in Lorenc et al., 2000). The final analysis is a product of minimising the square of the differences between the state and the observations and the state and the background state over the entire domain. Gravity wave activity in the forecasts is controlled by using the incremental analysis update initialisation scheme, in which the increments produced by the 3DVAR analysis are added gradually to the evolving fields every timestep from T-1 to T+1.

Precipitation rates from the Network Radar were assimilated for case 2 using a latent heat nudging approach Jones and Macpherson (1997) in which model profiles of latent heating are scaled by the ratio of observed and model precipitation rates. This causes the model to adjust so that the
diagnosed precipitation rate agrees more closely with observations. A four-hour window of latent heat nudging is centred at the analysis time. A parabolic weighting falls off with time away from the analysis time from 1 to zero. Latent heat nudging will not influence the time and location of convection since this occurred 12 hours into the simulation well outside the nudging window.
CHAPTER THREE

Two Case Studies

Two MCSs that remained entirely contained within the model domain throughout their lifecycle have provided the basis for numerical simulations throughout this thesis. A brief summary of their synoptic environments and lifecycles is presented here using observations and large-scale dynamics from the mesoscale model. The chapter closes by identifying the key differences between the cases.

The cases were chosen because of the apparent difference in predictability. A small ensemble of forecasts using different initialisation times, at 6 hour intervals, suggested that for case 1, the initial location of convection was sensitive to the initialisation time but for case two, the initial locations were not sensitive.

3.1 The 29 May 1999 Mesoscale Convective System: Case 1

Synoptic Situation

The meteorological situation known as the Spanish plume, described by Morris (1986) and McCallum and Waters (1993), occurred during the 28 and 29 May 1999. The synoptic evolution, described below, provides favourable conditions for deep, organised convection over the UK.

- An upper-level trough moves eastward and forces ascent over Iberia. A decreasing surface
pressure over a pre-existing low-level temperature gradient results in thermal advection and a further reduction in surface pressure. The Met Office synoptic analysis at 0000 GMT 28 May 1999, shown in Fig. 3.1a, shows the low pressure system associated with the upper-level trough over the Atlantic Ocean.

- A low-level warm front, resulting from thermal advection, forces ascent ahead of it and descent behind as it tracks north with the large-scale flow over France towards the UK. The Met Office synoptic analysis at 0000 GMT 29 May 1999, shown in Fig. 3.1b, shows the low-level warm front as a trough extending over northern France from the deepening low centre over the Bay of Biscay.

- The warm moist plume, below 750mb, tracks north beneath a cooler drier mid and upper-level south-westerly flow over the UK resulting in deep conditional instability.

- Significant vertical shear of the horizontal wind and a tropospheric-deep conditionally unstable profile provide necessary conditions for deep, organised convection.

**Initial Development**

Convection developed at 0000 GMT southwest of the UK as seen in Meteosat infrared imagery (not shown). Deep convection clustered in this region but remained to the west of the UK. Further convection triggered over the English Channel between 0330 and 0400 GMT. Meteosat infrared imagery, shown in Fig. 3.2a at 0630 GMT 29 May 1999, shows these convective cells within a region of vertical shear as determined by the model. Some validation of the model dynamics is provided by observed wind profiles over southern England that agreed well with the corresponding model wind profiles (not shown).

The following description of the large-scale flow is based on an analysis of the large-scale model dynamics. A southerly flow tracked beneath a mid and upper-level south-westerly flow resulting in a veering of the horizontal wind, as shown in Fig. 3.2a. Convection developed on a plume of high wet-bulb potential-temperature, $\theta_w$, air that tracked north at low-levels towards southern England, as seen in Fig. 3.2b. The front edge of the low-level warm moist plume was elevated above a low-level easterly flow resulting in a potentially unstable band over northern France. A schematic vertical cross-section through the low-level plume (line AB in Fig. 3.2b) is shown in Fig. 3.3.
CHAPTER 3: Two Case Studies

Figure 3.1: Met Office analyses at (a) 0000 GMT 28 May 1999 and (b) 0000 GMT 29 May 1999.
Figure 3.2: (a) Meteosat infrared imagery after initial development at 0630 GMT 29 May 1999. The arrow points to new convective cells over the English Channel. Contoured are 850mb geopotential height ($Z$) (thick, $\Delta Z=15$ m) and 500mb $Z$ (thin, $\Delta Z=15$ m) at 0400 GMT from the model. (b) Geopotential height at 850mb (as in (a)) and model $\theta_w > 289$ K (shaded) at 850mb at 0400 GMT 29 May 1999. The observed convection (marked by a cross) developed within the region of warm moist air at low-levels. The 850mb warm-front as determined by the model is also marked.
The plume and the easterly jet tracked north with the large-scale dynamical evolution maintaining the forcing for convection. An analysis of system relative flow on isentropes and $\theta_w$ (not shown) indicated significant moisture convergence along the elevated northern edge of the warm moist plume. The model diagnosed ascent along the front edge of the plume at 0400 GMT of 4 - 6cms$^{-1}$. Browning and Hill (1984) observed an MCS develop in a similar synoptic environment in this region. They emphasised the importance of a surface warm front at the southern limit of a low-level easterly flow in providing the forcing for convection.

The necessary conditions of deep conditional instability, vertical shear of the horizontal wind and low-level moisture convergence necessary for deep, long-lived, organised convection as described in chapter 1 were satisfied over a large-scale region over the north coast of France and the English Channel. The observations are of too coarse resolution to determine the triggering mechanism in the real atmosphere. Small-scale variability within this potentially unstable band is hypothesised to be the triggering mechanism. The initial location of convection is marked by a cross in Fig. 3.4 within a band of CAPE and almost zero CIN. A large part of the region of CAPE was co-located

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1CAPE and CIN are calculated offline based on pseudo-adiabatic parcel ascent. Input temperature and specific humidity profiles are on pressure levels every 50mb. An outer loop cycles through parcel origin levels from 1000mb to 750mb and values of CAPE and CIN are taken from the parcel ascent with the largest value of CAPE-CIN. The assumptions concerning buoyancy for pseudo-adiabatic ascent are not the same as those made by the parameterisation scheme or the large-scale equations as discussed in chapter 2. In addition, the vertical resolution of the input data will be too coarse to capture small-scale variability that may provide a significant contribution to CAPE and CIN. For these reasons, Figs. 3.4 and 3.13 are shown only to provide information on the general stability of the large-scale atmosphere.
with low CIN. It is likely that the ascent at the front edge of the plume suppressed any increase in CIN that may have otherwise occurred. The observed vertical profile of temperature and dew-point temperature at Camborne in southwest England at 0500 GMT, shown in Fig. 3.5, provides some validation of the model’s large-scale destabilisation of the atmosphere. An easterly flow in the lowest 80mb, above the nocturnal inversion, lay beneath the plume of higher $\theta_w$ air between 920 and 800mb. Significantly colder air in the mid-troposphere provided 1100Jkg$^{-1}$ of CAPE and 30Jkg$^{-1}$ of CIN for initiation from the 920 to 800mb layer which are similar values to the model derived CAPE and CIN in this region.

The model captured a second elevation of the northward propagating plume behind the first. This second plume, shown in the schematic cross-section in Fig. 3.3, is evident as a northwest-southeast extension of the region of high $\theta_w$ air at 850mb in Fig. 3.2b. The increase of meridional wind with height sheared out the elevated plume. The model diagnosed 3 - 5cms$^{-1}$ of ascent along this second plume consistent with flow along sloping isentropic surfaces. Later in the morning, convection was also observed on this second plume.
CHAPTER 3: Two Case Studies

62

Figure 3.5: Observed profile of temperature (solid) and dew-point temperature (dashed) at Camborne (location marked on Fig. 3.4) at 0500 GMT 29 May 1999.

Structure and Evolution of the Convective System

The evolution of the MCS is shown in a sequence of Meteosat IR imagery shown every three hours in Fig. 3.6. During the early stages, shown in Fig. 3.6a and Fig. 3.6b, convective cells were observed at 0300 GMT 29 May 1999 to the southwest of the UK and at 0600 GMT over the English Channel. The cells clustered and developed a common cirrus shield by 0900 GMT, shown in Fig. 3.6c. The mid-level southerly flow diverged at a confluent asymptote that was located across southern England. The subsequent track of convection ENE over southern England is likely to have been a consequence of its initiation to the east of the axis of divergence. The mature MCS tracked ENE over southern England with the mid-level environmental flow during the day (Fig. 3.6d, Fig. 3.6e and Fig. 3.6f).

As the system matured, surface stations observed coherent mesoscale pressure anomalies. Troughs of low pressure ahead and behind the system developed either side of a ridge of high pressure beneath the region of heavy precipitation. This tri-pole pattern in the surface pressure field tracked with the system across the country and is shown in the 1400 GMT surface pressure analysis in Fig. 3.7a. Maximum anomalies of $\pm$ 2mb occurred on the length-scale of the convecting region. As
Figure 3.6: Meteosat IR imagery (K) every 3 hours at (a) 0300 (b) 0600 (c) 0900 (d) 1200 (e) 1500 and (f) 1800 GMT 29 May 1999.
described in chapter 1, mesoscale pressure systems can act to increase low-level convergence and supply additional mass and moisture to the convective cells by a positive feedback.

As discussed in chapter 1, mesoscale circulations can replace warm moist inflow air with cold and dense downdraught air from mid levels resulting in an increase in surface pressure beneath the descending, diverging cold pool. Assuming the observed profile at 0500 GMT from Camborne (see Fig. 3.5) is typical, a saturated parcel from 650mb brought adiabatically to the surface will have a temperature of 13°C. The 2m temperature analysis at 1400 GMT, shown in Fig. 3.7b, shows a temperature of 13°C beneath the rear of the MCS. An observed 1.2m temperature difference of 8°C between the environment and downdraught air masses marked the front edge of the MCS. Surface observed dew-point depression of 7°C in the environment contrasted with 0°C for the saturated downdraught air.

Coherent parallel convective lines of northwest-southeast orientation developed on the southeastern side of the MCS between 0600 and 1000 GMT. Nimrod radar data at 5km horizontal resolution at 1015 GMT, shown in Fig. 3.8, show lines of precipitation extended 150km over the western English Channel and the Brest Peninsula. The mechanism for the linear organisation cannot be determined from the coarse observation network. Hypothesised mechanisms include a pre-existing parallel line-structure to the band of forcing or organisation through convectively-generated gravity waves. A wavelength of 57km was estimated as the distance between the convective lines shown in the radar data and a phase speed of 25ms⁻¹ was estimated using a mean layer wind of 13ms⁻¹ and a scale height of 7.5km estimated from radiosonde data. These values are typical for atmospheric gravity waves (Holton, 1992). Using high resolution visible and infrared imagery Browning and Roberts (1995) observed parallel lines of convection of similar dimensions to those observed in case 1 associated with a cold front. Differential lifting in the lower troposphere by mesoscale convergence lines resulted in rope-like clouds and provided the trigger for convection.

Met Office lightning data, shown in Fig 3.9, indicate intense convection remained along the south-
Figure 3.7: Rainrate > 2mmhr⁻¹ (shaded) at 1400 GMT 29 May 1999 together with (a) the surface pressure analysis (contoured every 1mb from 1013 to 1016mb) showing the tri-pole pattern and (b) the 2m temperature analysis (contoured every 2°C from 14 to 24°C). The analyses were generated by hand using data from surface synoptic stations.
Figure 3.8: Nimrod radar data at 1015 GMT 29 May 1999 showing parallel convective lines on the south-eastern side of the MCS.

Figure 3.9: Met Office lightning data for the 24 hour period from 2230 GMT 28 May 1999 to 2230 GMT 29 May 1999.
ern edge of the MCS; a result of the warm moist southerly inflow. Nimrod radar data estimated rainrates greater than $64\text{mmhr}^{-1}$ in these cells. The most easterly convective line remained active as the more westerly lines dissipated. The line re-intensified at 1400 GMT as a propagating coherent feature, possibly a squall line, over southern England. The coherent line, 50-80km long and 15km wide, on the southern side of the convecting region moved in a direction distinct from the mid-level environmental flow. The line propagated eastward at a constant $11\text{ms}^{-1}$ across southern England. Nimrod radar data at 1515 GMT, shown in Fig. 3.10, shows the line at the southern edge of the MCS. The front edge of the line is well defined by a strong gradient in Meteosat IR imagery at 1500 GMT, shown earlier in Fig. 3.6e.

The mature MCS tracked parallel to the mid-level environmental flow east of the northward moving low-level warm moist plume and subsequently dissipated.

provide error bars on the location by correlating different sets of three. When compared with commercially available systems, the mean separation between the location of lightning strikes between the two systems was only 16km (Hamer, 1996).

Figure 3.10: Nimrod radar data at 1515 GMT 29 May 1999 showing linear convective structure at the southeastern edge of the MCS.
Case 1 Summary

- A Spanish plume event.

- Convection developed within a large-scale potentially unstable band caused by the elevation of the front edge of a southerly warm moist flow above a cooler easterly flow.

- A large part of the region of CAPE was also a region of almost zero CIN.

- Propagating lines of convection.
3.2 The 11 September 2000 Mesoscale Convective System:  
Case 2

Synoptic Situation

A synoptic-scale wave-cyclone below 600mb tracked over the UK on 11 September 2000 and is shown in the two Met Office synoptic analyses at 0000 GMT 11 September 2000 and 12 September 2000 in Fig. 3.11. The low-level wave-cyclone together with the warm and cold fronts were tied to a 650mb jet. Hourly 1.2m dry-bulb temperature analyses (not shown) showed three distinct air-masses associated with the cyclone; the warm sector was bordered to the west by cold air behind the cold front and to the north by a cool air-mass ahead of the warm front. Superimposed on this was the diurnal cycle and sea-breeze effects. The warm sector was characterised by northward advection of air at 26°C at 1.2m that contrasted with a temperature of 15°C north of the surface warm front. The low-level northward advection of warm moist air of the warm sector was overlain by mid-level northeast-ward advection of cooler drier air associated with the mid-level jet. This large-scale environment provided the necessary conditions for deep, organised convection, as described in chapter 1.

Initial Development

Convection developed over northwest England between 1715 and 1800 GMT 11 September 2000 (determined by lightning data, Meteosat infrared imagery and Nimrod radar data) and is shown in Meteosat infrared imagery at 1830 GMT in Fig. 3.12a. Convection developed within a region of veering horizontal wind (determined by the model), also shown in Fig. 3.12a. Fig. 3.12b shows convection developed within the region of high $\theta_w$ in the warm sector, just ahead of the eastward moving surface cold front. Surface observations (not shown) indicate the first cells developed in a trough of low pressure associated with the approaching cold front. The vertical shear of the horizontal wind together with a conditionally unstable profile provided favourable conditions for organised convection over northwest England.

The model showed the warm sector was characterised by a capping inversion. This prevented
Figure 3.11: Met Office analyses at (a) 0000 GMT 11 September 2000 and (b) 0000 GMT 12 September 2000.
Figure 3.12: (a) Meteosat infrared imagery after initial development at 1830 GMT 11 September 2000. The arrow points to the first convective cell. Contoured are 850mb geopotential height ($Z$) (thick, $\Delta Z=15m$) and 500mb $Z$ (thin, $\Delta Z=15m$) at 1800 GMT from the model. (b) Geopotential height at 850mb (as in (a)) and model $\theta_w >289K$ (shaded) at 850mb at 1800 GMT 11 September 2000. The observed convection (marked by a cross) developed within the region of warm moist air at low-levels. The 850mb warm and cold fronts as determined by the model are also marked.
Figure 3.13: Case 2: Model CAPE >500Jkg$^{-1}$ (enclosed by thick contour) and CIN >10Jkg$^{-1}$ (shaded) at 1800 GMT 11 September 2000. The cross marks the initial location of convection within a region of CAPE and low CIN.

The widespread outbreak of thunderstorms. The first convective cell, as observed by Meteosat, triggered within the region of high CAPE and outside the region of high CIN, as shown in Fig. 3.13. CIN decreased towards the eastward moving cold front close to northwest England at 1800 GMT 11 September 2000. In contrast to case 1, most of the region of CAPE was co-located with high CIN.

Structure and Evolution of the Convective System

The sequence of Meteosat IR imagery shown in Fig. 3.14 shows the clustering of convective cells over northern England. A common cirrus shield grew rapidly to 100km in diameter by 2130 GMT. The triggering mechanism of further observed convection and clustering cannot be determined due to the coarse resolution of observations. The system, still in its early stages, propagated ENE across northern England towards the North Sea. New development occurred at the inflow on the western side of the mesoscale convecting region as shown in Fig. 3.14b. Nimrod radar data at 2100 GMT, shown in Fig. 3.15, shows convective cells with rainfall rates greater
than 16mm\text{hr}^{-1} \text{ and in some locations, greater than } 64\text{mmhr}^{-1}. \text{ There is also an indication of a developing mesoscale region of precipitation of more homogeneous intensity on the eastern side of the region of convection.}

The mesoscale convecting region moved ENE relative to the low-level cyclone with the mid-level flow. \text{A jet maximum of } 20\text{ms}^{-1} \text{ at 650mb as determined by the model advected the convecting region through the cyclone and up the sloping low-level warm front. Advection of warm moist air over the cool easterly flow at the warm front provided a second elevated source of warm, moist air necessary for the additional convection observed north of the pre-existing convection at 2130 GMT shown in Fig. 3.14b. The lightning data shown in Fig. 3.16 shows the development of further convection over the northern Pennine hills and the additional convection north of the analysed surface warm front between 2015 and 2100 GMT. The track of the first convective cell from northwest England to the east coast is clearly shown by a path of high density lightning strikes. The elevated convection north of the surface warm front is weaker than the lower level convection to the south. Nimrod rainfall data shown later in Fig. 4.10 on page 93 supports this. Smull and Augustine (1993) and Trier and Parsons (1993) documented the influence of frontal boundaries on MCS evolution. For both case studies, convection to the north of the front was typically weaker, as observed for this case study, and higher-based than convection to the south. They proposed that the accurate placement and slope of the frontal boundary would be crucial for successful simulations of the MCSs.}

\text{No mesoscale surface pressure anomalies were observed since these are only likely to be detectable once the system generates mesoscale circulations. This occurred as the system tracked over the North Sea away from observing stations. The observation network was too coarse to detect any surface thermodynamical signature of the convective cells over land. A vertical profile of temperature and moisture at Boulmer in northeast England at 0000 GMT 12 September 2000, shown in Fig. 3.17, together with nearby surface stations, confirmed the presence of a northward sloping warm front with height. A layer of cold saturated air 50mb deep lay beneath warm moist air between 950mb and 750mb. This marked the warm frontal boundary at 950mb. The profile had zero CIN for a parcel ascending from within the 900 to 825mb layer and approximately 2000Jkg^{-1} of CAPE; an absolutely unstable profile. The cells intensified and formed the new centre of the MCS as it tracked over the North Sea, as shown in Fig. 3.14c. An oval cirrus shield increased to a diameter of 300km in the upper troposphere in association with the clustering convective cells. The track of the system is shown by lightning data in Fig. 3.18 from the clustering of intense}
Figure 3.14: Meteosat IR imagery (K) approximately every 3 hours at (a) 1830 and (b) 2130 GMT 11 September 2000 and (c) 0100 and (d) 0330 GMT 12 September 2000.
**Figure 3.15:** Nimrod radar data over northern England at 2100 GMT 11 September 2000. Approximately 10 centres of convection can be identified and a mesoscale region of precipitation has developed on the eastern side.

Convective cells over northern England to a more coherent convective system over the North Sea. The lightning data shows that intense convection remained along the southern edge of the MCS; a result of the warm moist southerly inflow. The system propagated towards Denmark overnight (see Fig. 3.14d) and subsequently dissipated as the cyclone matured and the supply of warm moist air at low-levels was gradually cut off after 1200 GMT 12 September 2000.

**Case 2 Summary**

- A low-level wave cyclone.
- Convection developed ahead of the cold front.
- A large part of the region of CAPE was also a region of high CIN.
- The MCS was advected through the low-level cyclone and a second source of potentially unstable air resulted from the elevation of warm moist air up the forward sloping low-level warm front.
CHAPTER 3: Two Case Studies

Figure 3.16: Met Office lightning data for the period 1630 GMT to 2230 GMT 11 September 2000 every 45 minutes with orography at 12km resolution contoured every 100m and surface analysed fronts (thick solid lines).

3.3 Conclusions

In this section, the key differences between the two case studies are identified and some hypotheses are formed.

Convective Equilibrium

For case 1, a large part of the region of CAPE was co-located with almost zero CIN. In the absence of CIN, convection is free to respond to changes in CAPE due to large-scale processes. Under such conditions, the rate of stabilisation of the atmosphere by convection balances the rate of destabilisation by large-scale processes. It is hypothesised that convective equilibrium was satisfied over the region of CAPE. For case 2, however, a large part of the region of CAPE was co-located with CIN. In the presence of CIN, convection is not free to respond to changes in CAPE. Under such conditions, the rate of stabilisation of the atmosphere by convection does not balance the rate of
Figure 3.17: Absolutely unstable profile observed at Boulmer, northeast England at 0000 GMT 12 September 2000. Absolute instability and vertical wind shear below 600mb provided favourable conditions for deep, organised convection.

Figure 3.18: Met Office lightning data for the period 1630 GMT 11 September to 1630 GMT 12 September 2000 every 3 hours.
destabilisation by large-scale processes. It is hypothesised that convective equilibrium was not satisfied over the region of CAPE.

The mean magnitude of model CAPE for case 1 was less than that for case 2. This is consistent with the above hypotheses on convective equilibrium since for case 1, in the absence of CIN, convection is free to act and will not allow CAPE to accumulate. For case 2, however, in the presence of CIN, CAPE can accumulate in the absence of convection. The behaviour of model convection in the three experiments described in chapter 2 for both cases is described in chapter 4. Model solutions are compared to the observational data sets presented in this chapter.

**Mechanisms for Initial Development**

The elevation of the front edge of a warm moist southerly flow over a cooler easterly flow provided a large-scale potentially unstable region in case 1. It is hypothesised that for this case, convection was triggered by small-scale variability within the unstable band. It is hypothesised that the smaller potentially unstable region in case 2 resulted from forced ascent of warm moist air by local orography and the eastward moving low-level cold front. It is hypothesised that convection was triggered by small-scale variability within the potentially unstable region forced by the orography. The mechanisms in the model for triggering convection in both cases is investigated in chapter 5 together with an investigation of the predictability of the precise triggering locations.
In this chapter, the results of the experiments described in chapter 2 are presented for the two case studies. Firstly, the assertion made in chapter 2 that the partitioning between explicit and parameterised convection is sensitive to the CAPE-closure timescale is confirmed. The state of convective equilibrium, or otherwise, is then verified for both case studies by an analysis of the sensitivity of the total rainfall to the CAPE-closure timescale. An overview of the small-scale details of convection for the three CAPE-closure timescale experiments including location, size, intensity and evolution is presented for both case studies and is followed by a discussion. Two further experiments identify the sensitivity of the timing and location of parameterised convection to the trigger function. Finally, the sensitivity of the large-scale flow to the partitioning of convection is presented and compared with some observations.

4.1 Setup Verification

(a) Partitioning

All precipitation diagnosed by the explicit precipitation scheme in unstable regions is not necessarily associated with convection. The discussion of the hypothetical idealized experiments in section 2.4 showed that some explicit precipitation may be diagnosed during stable dynamical adjustment. The discussion suggested that vertical velocity can be used to diagnose explicit convection. Indeed, LeMone et al. (1998) defined convective cores as having vertical velocity greater
then $1\text{ms}^{-1}$. In order to determine the partitioning between parameterised and explicit convection it may be necessary to remove explicit precipitation not associated with convection. Removing explicit precipitation in regions of vertical velocity less then $1\text{ms}^{-1}$, say, will filter out explicit precipitation not associated with convection. However, there is only a weak correlation between vertical velocity in the free troposphere and explicit rainrate, as shown in Fig. 4.1 at 0400 GMT 29/5/99 (case 1) for the 2 hour experiment using 500mb vertical velocity. The general shape of the scatter plot is similar using vertical velocity at different model levels within the free troposphere and for fields at different times. There is no useful correlation by which to define explicit convection using vertical velocity. The precipitation fields shown later in section 4.2.2 show that explicit rainrates associated with stable dynamical adjustment are small compared to the localised intense centres of explicit precipitation associated with convection. In determining the partitioning between explicit and parameterised convection it is therefore assumed that the total explicit precipitation is equivalent to the total explicit convective precipitation to a good approximation over the potentially unstable regions.

The ratio of parameterised to total hourly rain amount within the region of CAPE and low CIN for the 3 CAPE-closure timescale experiments for case 1, presented in Fig. 4.2a, shows the parti-

![Figure 4.1: Scatter plot of 500mb vertical velocity versus explicit rainrate at 0400 GMT 29/5/99 (case 1) for the 2 hour experiment.](image-url)
tioning of precipitation is highly sensitive to the CAPE-closure timescale. The three experiments therefore provide three different representations of convection. The partitioning of precipitation also changes with time and three stages of partitioning between parameterised and explicit precipitation in the 2 hour experiment (using the naming convention for the experiments described on page 53) can be identified. The details of the convective solutions are shown later in section 4.2.2. Prior to the triggering of deep convection, the ratio is determined by a few grid-points with precipitation. The convection scheme triggered at T+3, shown by the abrupt change in partitioning in favour of the convection scheme. Explicit precipitation developed in the 2 hour experiment beyond T+6 and provided 80% of the total precipitation beyond T+10 as explicit convection developed. The results show a transition from 70% parameterised precipitation in the early stages (T+6) to 20% parameterised precipitation in the later stages.

Despite the fast response of parameterised convection to changes in CAPE, the 10 minute experiment generated approximately 20-30% of the total precipitation explicitly, also shown in Fig. 4.2a. For this experiment, all of the explicit precipitation was associated with stable dynamical adjustment. At the other extreme, precipitation in the 1 day experiment was almost exclusively explicit throughout the simulation. In the 1 day experiment, the slow response of parameterised convection to CAPE allowed the lower layers to saturate and explicit convection to develop.

![Graphical representation of precipitation partitioning](image_url)

**Figure 4.2:** Ratio of parameterised to total hourly rain amount for the 10 minute, 2 hour and 1 day experiments for (a) case 1 and (b) case 2.
Broadly similar partitioning between parameterised and explicit precipitation between the three experiments occurred in the region of CAPE and low CIN for case 2, shown in Fig. 4.2b. Convection began in all three experiments between T+11 and T+12. For the 10 minute experiment, convection was totally parameterised and the small amount of explicit precipitation resulted from stable dynamical adjustment. Explicit convection developed in the 2 hour experiment and later dominated the simulation. Almost all convection was explicit in the 1 day experiment. There was, however, difficulty in determining the partitioning after convection became embedded within the frontal zone after T+20 (not shown).

(b) Convective Equilibrium

The cumulative rainfall within the region of CAPE for the 3 experiments for case 1 is shown in Fig. 5.18a. The cumulative rainfall includes all rainfall in the model. For the purposes of determining the state, or otherwise, of convective equilibrium the total explicit precipitation associated with stable dynamical adjustment is assumed to be small compared to explicit precipitation associated with convection. Cumulative rainfall was only weakly sensitive to the partitioning of convection despite the high sensitivity of the small-scale details of convection (presented later in section 4.2). Most of the region of CAPE was co-located with low CIN (see Fig. 3.4 on page 61) and convection, either explicit or parameterised, was free to act. This enabled convection in all three experiments to realise the entire region of CAPE. The result is that expected for an equilibrium situation.

Although the cumulative rainfall was only weakly sensitivity to the partitioning of convection there are small differences that are consistent with expectation. The magnitude of cumulative rainfall was inversely related to the CAPE-closure timescale during the first 10 hours of simulation. All convection in the 10 minute experiment was parameterised. This convection is quick to respond to changes in CAPE and therefore the curve of cumulative rain amount can be thought of as tracing the large-scale destabilisation in a series of quasi-equilibrium states. Owing to the longer response time of parameterised convection in the 2 hour experiment, convection responded with weaker initial rain amount. Convection was weakest in the 1 day experiment before the grid became absolutely unstable and explicit convection developed. Later in the simulations, cumulative rainfall in the 2 hour and 1 day experiments equalled that in the 10 minute experiment to compensate for the earlier rainfall deficit. Later still, cumulative rainfall in the 2 hour and 1 day experiments exceeded that of the 10 minute experiment and the initial trend of cumulative rainfall with CAPE-closure
timescale reversed. It is possible that the larger cumulative rainfall in the 2 hour and 1 day experiments than in the 10 minute experiment can be explained by different responses of the large-scale flow between the three experiments. The sensitivity of the large-scale response to the partitioning of convection is investigated in section 4.5.

The cumulative rainfall within the region of CAPE for the 3 experiments for case 2, shown in Fig. 5.18b, shows high sensitivity to the partitioning of convection. For this case, the cumulative rainfall was largely determined by the partitioning of convection and not by the large scales as for case 1. Most of the region of CAPE was co-located with high CIN (see Fig. 3.13 on page 72) and convection was not free to act. The result implies that different amounts of CAPE were realised in the three experiments and is the result expected for a non-equilibrium situation. The cumulative rainfall within the region of CAPE from Nimrod Radar data averaged\(^1\) onto the mesoscale grid is also shown in Fig. 5.18b for comparison. The timeseries is given for the period within which the entire convective system remained within the domain of the radar. However, the timeseries for any of the three experiments were not expected to be similar to that observed for this non-equilibrium case. A similar timeseries from Nimrod data for case 1 could not be produced because convection developed outside the radar domain.

\(^1\)Any radar data point having a non-zero data value, contained in a cell centred on each mesoscale grid-point, is arithmetically averaged with others contained within that cell to give a rainrate value corresponding to that mesoscale grid-point.
(c) Summary

The three experiments with different CAPE-closure timescales provide three different representations of convection through different partitionings of convection between the explicit and parameterised components. For case 1, the cumulative rainfall was only weakly sensitive to the partitioning of convection whereas for case 2, the cumulative rainfall was highly sensitive. This provides strong evidence that case 1 was an equilibrium case and case 2 was a non-equilibrium case.

4.2 Small-Scale Sensitivity to Partitioning

In this section, an overview of the small-scale details of convection in the 3 CAPE-closure timescale experiments for both case studies is presented.

4.2.1 The 10 Minute Experiment

Case 1

The convective solution to the large-scale destabilisation was a region of parameterised convection with large-amplitude grid-scale noise coincident with the region of CAPE and low CIN. Fig. 4.4a shows the field of instantaneous parameterised rainrate at 0000 GMT 29 May 1999. The short adjustment timescale allowed parameterised convection to respond quickly to changes in CAPE and suppressed explicit convection. The explicit precipitation scheme diagnosed some precipitation, as shown in Fig 4.2 on page 81, during stable adjustment to the convective increments from the parameterisation scheme. It is hypothesised that the discontinuous trigger function in the convection scheme (see section 2.3 for details), caused the convection scheme to switch on and off at the grid-scale as the low-level parcel buoyancy oscillated about the prescribed threshold, resulting in grid-scale noise. The sensitivity of the field of convective triggering to the trigger function is examined later in section 4.4.
Figure 4.4: Instantaneous parameterised rainrate (mm hr\(^{-1}\), shaded contours) for the 10 minute experiment for (a) case 1 at 0000 GMT 29 May 1999 and (b) case 2 at 1900 GMT 11 September 2000.

Figure 4.5: Cumulative parameterised rain amount (mm, shaded) between 1800 GMT 11 September and 1000 GMT 12 September 2000 for the 10 minute experiment. Geopotential height at 650mb at 0600 GMT 12 September 2000 is contoured with an interval of 20m. The arrows indicate the direction of propagation of three bows of parameterised convection.
Case 2

The convective solution to the large-scale destabilisation was initially similar to that for case 1; parameterised convection with large amplitude grid-scale noise co-located with the region of CAPE and low CIN. Fig. 4.4b shows parameterised rainrates at 1900 GMT 11 September 2000. Again, explicit convection was suppressed. Later in the simulation, however, the scheme behaved differently. As the region of CAPE moved over the North Sea, parameterised convection organised bowed structures that were not observed in nature. The bowed structures propagated at 13ms$^{-1}$ at approximately 90° to the mid-level environmental flow. The tracks of three bows are shown in the field of accumulated precipitation in Fig. 4.5. One of the bows of parameterised precipitation that propagated into the region of high CAPE and high CIN is shown at 0100 GMT 12 September 2000 in Fig. 4.6. A vertical cross-section through the line marked AB on Fig. 4.6, shown in Fig. 4.7, shows convection at the front edge of a low-level cold pool.

![Figure 4.6](image_url)

Figure 4.6: Instantaneous parameterised rainrate (mm/hr$^{-1}$, shaded) at 0000 GMT 12 September 2000 for the 10 minute experiment. CAPE = 500Jkg$^{-1}$ is thickly contoured and CIN >10Jkg$^{-1}$ is shaded. The arrow indicates the direction of propagation of the bow of parameterised convection into the region of high CAPE and high CIN. CAPE and CIN were calculated off-line as described on page 60.
4.2.2 The 2 Hour Experiment

Case 1

The 2 hour CAPE-closure timescale allowed significant amounts of both explicit and parameterised convection to develop. Four stages of partitioning between parameterised and explicit precipitation have been identified and are shown by parameterised and explicit rainrates in Fig. 4.8. The parameterisation scheme triggered three hours before observed convection at 2100 GMT 28 May 1999, shown an hour later in Fig. 4.8a, over a larger region than observed, co-located with the region of CAPE and low CIN (not shown). The subsequent development of explicit precipitation, shown in Fig. 4.8b, resulted from the large-scale dynamics adjusting to the heating and moistening increments from the parameterisation scheme. The local intensification of explicit precipitation implies the development of explicit convection. This implies that the convective ten-
CHAPTER 4: Partitioning of Convection

(a) Stage 1: Parameterised convection. 2200 GMT 28 May 1999
(b) Stage 2: Explicit development. 0000 GMT 29 May 1999
(c) Stage 3: Simultaneous strengthening. 0300 GMT 29 May 1999
(d) Stage 4: Explicit locally dominates. 0500 GMT 29 May 1999

Figure 4.8: Evolution of parameterised rainrate (mmhr$^{-1}$, shaded) and explicit rainrate (mmhr$^{-1}$, thick contours with a contour interval 1.6 mmhr$^{-1}$ starting at 0.01mmhr$^{-1}$) in the 2 hour experiment for case 1.
dencies from parameterised convection were too small to be in equilibrium with the larger-scale destabilisation.

Simultaneous strengthening of explicit and parameterised precipitation occurred in localised regions at 0300 GMT 29 May 1999, as shown in Fig. 4.8c. Maximum model precipitation rates of $6\,\text{mmhr}^{-1}$ at this stage were well below the $29\,\text{mmhr}^{-1}$ observed. The explicit cells consisted of a single convective circulation. The ascending branch, an updraft core extending horizontally to approximately five grid points or 60km, was compensated by surrounding weaker descent on a larger horizontal scale. The region of localised explicit convection embedded within homogeneous parameterised convection tracked north with the large-scale flow. Simultaneous strengthening of explicit convection and weakening of parameterised convection occurred locally after 0500 GMT, as shown in Fig. 4.8d. Maximum explicit precipitation rates increased to $20\,\text{mmhr}^{-1}$ by 0500 GMT but remained only a fifth of those observed.

Figures 4.8c and 4.8d show the triggering of parameterised convection along the second elevated plume of warm moist air, south of the Brest Peninsula. In the preceding few hours, this second plume had been destabilised through shearing up potential temperature surfaces, thereby reducing CIN. During the next few hours, Nimrod Radar data showed parallel convective lines over this second band, as shown earlier in Fig. 3.8 in chapter 3. The model solution, however, progressed through the established stages of homogeneous parameterised convection, development of weak explicit precipitation, localised strengthening of parameterised convection and the development of explicit convection and finally, local explicit cells embedded within weak parameterised convection. The model failed to organise any linear convective structure and at 12km horizontal resolution the model cannot well-resolve such small scales.

Most importantly, the model failed to initiate locally intense convection over the English Channel as observed at 0400 GMT. Satellite, lightning and radiosonde data, shown earlier in chapter 3, showed these cells tracked north across the English Channel parallel to the mid-level flow and organised over southern England. Model CAPE at 0600 GMT was lower over the English Channel than further west. Homogeneous weak parameterised convection within this region (Figs 4.8b; 4.8c and 4.8d) was sufficient to suppress explicit convection. The model maintained intense convection west of England that did not move across southern England as observed.
Case 2

The convective solution evolved through the established stages of partitioning between parameterised and explicit precipitation identified for case 1, as shown in Fig. 4.9. Weak parameterised convection triggered at 1700 GMT, within an hour of the observed time of triggering in nature. Again, a larger convecting region than observed was co-located with the region of CAPE and low CIN, over northwest England for this case and is marked by an arrow in Fig. 4.9a. The maximum model rainrate of approximately $1\text{mmhr}^{-1}$ was well below that observed of $68\text{mmhr}^{-1}$ (averaged onto the mesoscale grid) in the early stages. As for case 1, explicit precipitation soon developed. As the region of CAPE propagated over the North Sea simultaneous intensification of explicit and parameterised precipitation occurred locally at two regions associated with the development of explicit convection within the area of homogeneous parameterised convection. Some parameterised convection penetrated into the region of CAPE and high CIN, as occurred in the 10 minute experiment, and is marked by an arrow in Fig. 4.9c. This precipitation was much weaker and dissipated much earlier than the bow of parameterised precipitation in the 10 minute experiment. As with case 1, the final stage was characterised by localised simultaneous strengthening of explicit convection and weakening of parameterised convection.

4.2.3 The 1 Day Experiment

Case 1

Explicit precipitation developed by 2200 GMT, later than parameterised precipitation in both the 10 minute and 2 hour experiments. Localised convective cells developed rainrates greater than $17\text{mmhr}^{-1}$ by 0200 GMT. In contrast to the 10 minute and 2 hour experiments, intense convection developed further east over the English Channel from 0400 GMT as observed. The explicit cells were similar in horizontal scale to the explicit cells in the 2 hour experiment. Accumulated explicit precipitation between 0300 and 1300 GMT 29 May 1999 and 700mb geopotential height at 0800 GMT is shown in Fig. 4.10. The independent convective cell triggered over the region of the Cherbourg Peninsula, further east than the location of the observed cells (marked by a cross in Fig. 4.10 and shown in Nimrod Radar data also in Fig. 4.4), and intensified as it tracked north with
CHAPTER 4: Partitioning of Convection

(a) Stage 1: Parameterised convection. 1800 GMT 11 September 2000

(b) Stage 2: Explicit development. 2000 GMT 11 September 2000

(c) Stage 3: Simultaneous strengthening. 0000 GMT 12 September 2000

(d) Stage 4: Explicit locally dominates. 0300 GMT 12 September 2000

Figure 4.9: Evolution of parameterised rainrate (mmhr$^{-1}$, shaded) and explicit rainrate (mmhr$^{-1}$, contoured with contours at 0.01mmhr$^{-1}$, 5mmhr$^{-1}$ and from there on with an interval of 5mmhr$^{-1}$) in the 2 hour experiment for case 2. The arrow in (a) points to the region of parameterised convection within the region of CAPE and low CIN. The arrow in (c) points to the region of parameterised precipitation that moved through the region of CIN.
the mid-level environmental flow. This cell tracked over central southern England and intensified, generating rainrates greater than 28mm hr$^{-1}$ and extended to 7×7 grid lengths by 1100 GMT. Over England the mid-level flow in the region of the cell turned to the east and intense convection was advected ENE across southern England as observed. The solution failed to organise the observed linear convective band over southern England.

The convection further west (not shown) remained over the Irish Sea where the mid-level southerly flow did not turn to the east at the confluent asymptote. At 0600 GMT an explicit cell triggered on the second potentially unstable band to the south in the same region as convection in the 2 hour simulation and intensified rapidly. As in the 2 hour experiment, this solution did not organise into the parallel convective lines observed.

**Case 2**

The convective response showed similar characteristics to that in case 1. The accumulated explicit precipitation between 1800 GMT 11 September and 0100 GMT 12 September and 700mb geopotential height at 2000 GMT 11 September is shown in Fig. 4.10. Three cells tracked parallel to the mid-level environmental flow. The cell tracks observed by Nimrod Radar, also shown in Fig. 4.10, also ran parallel to the mid-level environmental flow. The model triggered one explicit cell in the location of the observed initiation and another two hours early, marked by the number 2 in Fig. 4.10, both within the region of CAPE and low CIN over northern England. The model did not capture the relative strength and timing of the convective cells but did represent the overall history. The other explicit cell that was not observed in nature initiated approximately five grid points (60km) inland from the coast of central eastern England, marked by the number 3 in Fig. 4.10, within the region of CAPE and high CIN (see Fig. 3.13 on page 72). For the first few hours, maximum model rainrates of 1 - 5mm hr$^{-1}$ were well below the 45 - 68mm hr$^{-1}$ observed. Maximum model rainrates increased to 36mm hr$^{-1}$ over the North Sea as those observed fell to typically 20mm hr$^{-1}$.
Case 1: 0300 - 1300 GMT
29 May 1999

Case 2: 1800 GMT 11 September
- 0100 GMT 12 September 2000

Figure 4.10: Cumulative rain amount (mm, shaded) for the 1 day experiment and derived from Nimrod data interpolated onto the model grid for both cases. Model geopotential height at 700mb in the middle of the accumulation period is contoured at interval 5m over the model plots. The cross in the top two plots marks the observed location of initiation.


4.3 Discussion

For both cases, the small-scale details of convection including intensity, location, horizontal scale and evolution were highly sensitive to the partitioning of convection. The convective solution in each experiment displayed some common features between the case studies but the convection scheme in particular behaved differently between the cases. In this section, the similarities and differences in the small-scale details of convection between the cases are identified and some interpretation is provided.

(a) Similarities

For the 10 minute experiment, the solution was initially parameterised convection with large-amplitude grid-scale noise co-located with the region of CAPE and low CIN. The different behaviour of parameterised convection later in the simulation of case 2 is discussed in the second half of this section. The time averaged rainrate over 2 hours was a smooth field of precipitation co-located with the region of CAPE and low CIN and resembled the parameterised rainrates in the 2 hour experiment in terms of area coverage, as shown for both cases in Fig. 4.11 together with CAPE and CIN. Owing to its shorter response timescale, the rainrate from the 10 minute experiment for case 1 at 0000 GMT 29 May 1999 was larger than that from the 2 hour experiment, as shown earlier in Fig. 5.18a. This explains the larger rainrates in the 10 minute experiment than in the 2 hour experiment in Fig. 4.11 at 0000 GMT 29 May 1999. The smooth field of time averaged parameterised rain amount in the 10 minute experiment is that expected of convective equilibrium behaviour. The scheme is formulated to produce an equilibrium response and by setting the CAPE-closure timescale to 10 minutes, the solution traces the forcing in a series of quasi-equilibrium states. In a time average sense, the convection scheme behaved as expected but the noisy field of instantaneous rainrates implies the scheme was not numerically well-behaved.

Four stages of partitioning between parameterised and explicit precipitation was identified in the 2 hour experiment. Similar results have been shown by Kain and Fritsch (1998) who assessed how a mesoscale model partitioned convection in the simulation of a squall line. Parameterised convection was permitted prior to grid-scale saturation. Eventually parameterised drying tendencies were overwhelmed by moistening and destabilisation by resolved ascent and the grid saturated while the lapse rate was still greater than moist-adiabatic. As shown here, explicit convection soon
CHAPTER 4: Partitioning of Convection

Figure 4.11: Average parameterised rainrate over 2 hours (mm hr$^{-1}$, shaded) for the 10 minute experiment and instantaneous parameterised rainrate (mm hr$^{-1}$, shaded) for the 2 hour experiment. For case 1, CAPE $>10$ J kg$^{-1}$ is enclosed by the solid contour and CIN $>10$ J kg$^{-1}$ is enclosed by the dashed contour and cross hatched. For case 2, CAPE $>500$ J kg$^{-1}$ is enclosed by the solid contour and CIN $>10$ J kg$^{-1}$ is enclosed by the dashed contour and cross hatched.
dominated the parameterised convection. Although the solution of smooth parameterised precipitation was that expected of equilibrium behaviour, the convective tendencies were too weak to suppress explicit convection. Parameterised convection was therefore not in equilibrium with the large-scale destabilisation. Convection in the 2 hour experiment did not provide useful information on the size or intensity of the MCS or its propagation over southern England. Parameterised convection, however, tracked the region of CAPE and low CIN and provided guidance as to the region within which intense convection was observed.

The very slow response of parameterised convection in the 1 day experiment resulted in all buoyant instability being removed by explicit convection. For both cases, the convective response in the 1 day experiment was localised intense persistent regions of explicit convection steered by the mid-level flow. The explicit convection in the 1 day experiment provided useful information on the track of the observed MCSs and produced rainrates closer to those observed than in the 10 minute and 2 hour experiments. The horizontal scale was constrained at the lower bound by the horizontal grid-spacing of the model and at the upper bound by horizontal diffusion. Within this modelling framework the explicit cells extended to 7 grid lengths (approximately 80km).

(b) Differences

The equilibrium-based convection scheme was not expected to perform well in the non-equilibrium case. The scheme was not designed to respond quickly to the strong mesoscale forcing in case 2 (as described later in chapter 5) since it is only indirectly influenced by the effects of vertical motion. For case 2, the unobserved bows of parameterised convection that developed later in the 10 minute experiment and, to a lesser extent, in the 2 hour experiment increased the cumulative rainfall significantly. The bow propagated into the reservoir of CAPE and realised a significant amount of energy unavailable to convection in the real atmosphere. Roberts (2001) suggested a positive feedback mechanism involving a coupling between the convection scheme and the grid-scale dynamics that is consistent with the mechanism in these experiments. A cold pool, possibly a result of prior convection, caused convergence and ascent at its forward edge. This ascent modified the thermodynamic profile sufficiently to trigger the convection scheme. The equilibrium-based scheme responded instantaneously, producing cooling below the convection. This effectively propagated the cold pool forward and triggered the scheme at the next timestep allowing the system to propagate forward and realise the previously unavailable reservoir of en-
energy. The scheme does not represent a timescale for the development of convection and the match in timescales between convection and the cold pool allowed the two to become locked in phase. It is likely that the timescale for the development of convection in nature did not allow this phase-locking. Similar features often occurred in the Met Office mesoscale model in operational mode in convective situations.

In their review of recent work Kuo et al. (1997) addressed whether it would be preferable to force a convection scheme to remove all instability before convective overturning is permitted on the grid. They concluded that a convection scheme was necessary but did not always provide a sufficient representation of convection. This investigation has shown that the convection scheme alone provided no useful information on the size, intensity or track of the MCSs. Furthermore, the bows of precipitation that did not occur in nature are an example of the scheme failing dramatically in a non-equilibrium situation. The distribution of precipitation should therefore not necessarily be believed.

Although the characteristics of the convective response in the 1 day experiments were similar between the case studies, the accuracy of the triggering locations of the explicit cells was often poor. For the non-equilibrium case, the model triggered explicit convection in the observed triggering location and another explicit cell triggered two hours early. The model triggered a third cell that was not observed. For case 1, the model missed the observed triggering location by approximately 100km. The evolution of the temperature and moisture fields prior to the onset of convection in both cases are examined in Chapter 5 and the effects of surface features are isolated to determine the model’s triggering mechanisms. The predictability in precise location of triggering is then explored.

### 4.4 Sensitivity to the Trigger Function

A true equilibrium scheme removes CAPE instantaneously in the absence of CIN. The trigger function, in addition to the CAPE-closure timescale, prevents the Gregory-Rowntree scheme from being a true equilibrium scheme. It is possible that the trigger function constrains the small-scale details of convection including the range in timing and location of convection as determined in the previous section using extreme values of the CAPE-closure timescale. The sensitivity of the small-
scale details including timing and location of convection to the trigger function is investigated in two further experiments:

(a) The positive buoyancy experiment: Critical parcel buoyancy set to zero.
(b) The vertical velocity experiment: Additional parcel buoyancy in regions of positive vertical velocity.

The details of the two experiments are described in the next two subsections after this overview of the general method. To simplify the experiments the CAPE-closure timescale is set to 10 minutes. This allows an investigation of the sensitivity of the convection scheme without the complication of explicit convection. The experiments are carried out for case 2 because for this case convection was forced primarily by low-level mesoscale variability rather than large-scale ascent. Indeed, the low-level model fields of vertical velocity were characterised by mesoscale variability of approximate magnitude $0.2\text{ms}^{-1}$ in the region of convective triggering as shown later in section 5.1. It is likely that the timing and location of convection forced by low-level mesoscale variability will be more sensitive to the formulation of the trigger function than convection forced by large-scale, more homogeneous forcing as for case 1.

(a) The Positive Buoyancy Experiment

The stability test described in section 2.3a is removed so all grid points are passed to the trigger function. Owing to the code structure of the Unified Model it was not possible to remove the convective trigger entirely. Instead, the constraints imposed by the trigger function are minimised. Convection triggers at grid points for which a parcel taken from the environment and given no increment to potential temperature remains positively buoyant at the next model level after ascent (rather than the requirement of being buoyant by greater than 0.2K in all previous experiments). It is expected that this modified trigger function will be more likely to trigger convection for a profile with CAPE and some CIN than the original trigger function.

A situation can be envisaged for which the original scheme triggers (based on buoyancy over one layer), but then the parcel becomes immediately negatively buoyant in the next layer due to an inversion for example. The cloud model is formulated such that terminal detrainment acts when the parcel buoyancy falls below 0.2K (as described in section 2.3b). Terminal detrainment reduces
parcel mass-flux in regions of CIN and the scheme switches off with zero fluxes if the cloud depth fails to pass the critical depth. For the positive buoyancy experiment, the likelihood of the scheme switching off is reduced by setting the critical parcel buoyancy to continue parcel ascent to 0.0K.

(b) The Vertical Velocity Experiment

The stability test described in section 2.3a is removed so all grid points are passed to the trigger function. Convection is triggered at grid points where a parcel taken from the environment and given an increment to potential temperature due to the grid element vertical velocity remains positively buoyant at the next model level after ascent. The trigger function is similar to that in the positive buoyancy experiment except for the increment to parcel potential temperature due to the grid element vertical velocity which follows that of the Kain and Fritsch (1993) convection scheme. Parcel Buoyancy, $B$, at the next level after ascent is assessed using:

$$B = (\theta_p + \Delta \theta)(1 + (\epsilon^{-1} - 1)q_P) - \theta_{Ev},$$

(4.1)

where $\theta_p$ is the parcel potential temperature, $\Delta \theta$ is the potential temperature increment due to vertical velocity, $q_P$ is the parcel mixing ratio, $\epsilon$ is the ratio of molecular weights of water and dry air and $\theta_{Ev}$ is the environment virtual potential temperature. The first term on the right hand side is the parcel virtual potential temperature. The potential temperature increment crudely simulates subgrid-scale forcing as a function of the grid-scale vertical velocity ($w_E$) in the layer being lifted (Kain and Fritsch, 1993). The motivation for this formulation was provided by Chen and Orville (1980) who showed that thermals are stronger and larger when low-level convergence is present. The relationship between the potential temperature increment and vertical velocity is defined as:

$$\Delta \theta(w_E) = 4.64(w_E)^{1/3}.$$  

(4.2)

For the original Kain-Fritsch scheme, a running mean temporal filter is applied to the vertical velocity field before it is passed to the trigger function to produce a smoother field of convective
triggering. For simplicity, a 1-2-1 spatial filter is applied to the vertical velocity field for this experiment to approximate the running mean. In addition, the formula has been used with most success at grid lengths near 25km (Kain and Fritsch, 1990). The smoothed vertical velocity field is therefore adjusted to the equivalent value for a 25km grid-length assuming a linear dependence of vertical velocity on grid-length as suggested by Kain and Fritsch (1990). For the grid length of \( \sim 12.5\)km used in this investigation the vertical velocity field is divided by 2. The relationship in equation 4.2 is shown in Fig. 4.12. Additional buoyancy is given to parcels in regions of ascent and the temperature increment is zero in regions of descent. As a result of the \( w_E^{1/3} \) relation the temperature increment is very sensitive to the vertical velocity in the range 0.0 to 0.2ms\(^{-1}\) expected for regions of large-scale and mesoscale forcing.

For a preliminary experiment in which the buoyancy increment was added to parcels originating from any model level the scheme triggered in regions of ascent at the lowest model level due to the high sensitivity of the buoyancy increment to vertical velocity as shown in Fig. 4.12. The vertical velocity field at lower model levels was dominated by gravity wave activity which reflected unrealistically in the fields of convective triggering and contaminated the solution. To overcome this problem the increment to parcel buoyancy due to vertical velocity is only given to parcels triggering above model level 8 where the influence of gravity waves is much reduced. This restriction on the buoyancy increment has a counterpart in the Kain-Fritsch trigger function. For the Kain-Fritsch scheme a parcel from the lowest 100mb-deep layer with layer mean virtual temperature

![Figure 4.12: Relationship between the temperature increment and vertical velocity in the trigger function of the vertical velocity experiment (solid line). The dashed line indicates zero vertical velocity for reference.](image-url)
and specific humidity is lifted to its Lifting Condensation Level (LCL). Additional buoyancy is then given to the parcel at the LCL and if the parcel possesses positive buoyancy the scheme is triggered. The trigger function for the vertical velocity experiment follows this formulation in that the temperature increments are added above model level 8 (approximately 875mb) which includes the LCL across northern England of approximately 830mb.

4.4.1 Results

(a) The Positive Buoyancy Experiment

A comparison between the experiment with a CAPE-closure timescale of 10 minutes already discussed in section 4.2.1 using the default trigger (hereafter referred to as the default trigger experiment for this section) and the positive buoyancy experiment shows low sensitivity of both the general location, timing and magnitude of precipitation to the critical buoyancy in the trigger function. For both experiments the parameterisation scheme developed precipitation with large-amplitude grid-scale noise co-located with the region of CAPE and almost zero CIN. The range in timing and location of convection in the positive buoyancy experiment did not extend much beyond that of the default trigger experiment as shown in Fig. 4.13.

Figure 4.13: Accumulated parameterised rain amount (mm) between 1700 and 1800 GMT 11 September 2000 for (a) the default trigger experiment and (b) the positive buoyancy experiment.
The convection scheme in the positive buoyancy experiment produced light precipitation, typically less than 0.1mmhr$^{-1}$, over central and southern England and the North Sea, as shown in Fig. 4.13b, within the region of CAPE and CIN. This may be a result of the suppressed terminal detrainment allowing parcels to penetrate the layer of CIN. There are still regions of CAPE and CIN where convection is not triggered, over northeast England for example. It is possible that for these regions the magnitude of CIN is too great for parcels to penetrate even with the suppressed terminal detrainment.

In both the default trigger and positive buoyancy experiments convection organised bowed structures later in the simulations over the North Sea that did not occur in nature, as discussed in section 4.2.1.

(b) The Vertical Velocity Experiment

The small-scale details of convection (precipitation with large amplitude grid-scale noise) were similar to that for the default trigger experiment but the range in timing and location of convection was much larger. The convection scheme in the vertical velocity experiment triggered in the potentially unstable region over northwest England and over large areas outside this region as shown in Fig. 4.14b.

The field of convective triggering was strongly constrained to occur within regions of low-level ascent and CAPE rather than regions of low CIN and CAPE. CAPE and CIN were calculated outside the Unified Model (see page 60 for details of the calculation) and were not based on parcel profiles calculated by the convection scheme. The increment to parcel buoyancy due to the grid-scale vertical velocity acts to reduce CIN and trigger convection outside the region of triggering in the default trigger experiment. The low-level vertical velocity over northern England varied across the range for which the temperature increment in equation 4.1 is highly sensitive and as a result the field of convective triggering was dominated by the vertical velocity dependent trigger function. Convection preferentially triggered in regions of ascent as shown in Fig. 4.14. The field of convective triggering became smoother over the North Sea associated with the smoother field of low-level vertical velocity. Later in the simulation parameterised convection organised bowed structures over the North Sea, as occurred in the default trigger experiment, that propagated into the region of CAPE and CIN but were not observed in nature.
Figure 4.14: (a) Vertical velocity at 850mb at 1700 GMT 11/9/00 for the vertical velocity experiment (ms$^{-1}$, positive contours are solid and negative contours are dashed). (b) Accumulated parameterised rain amount between 1700 and 1800 GMT (mm, shaded) for the vertical velocity experiment with a contour of 850mb vertical velocity $= 0.0$ms$^{-1}$ at 1700 GMT.

In addition to triggering over northern England, convection in the vertical velocity experiment triggered earlier in the simulation between 1100 GMT and 1500 GMT 11/9/00 over central England within the region of CAPE and CIN. Convection did not occur in nature within this region. This was a region of low-level ascent, as shown in Fig. 4.15b, that provided parcels with sufficient buoyancy via the vertical velocity term in equation 4.1 to overcome CIN and trigger convection. Without the buoyancy increment due to vertical velocity no convection occurred within the region of CAPE and CIN, as shown in Fig. 4.15a. The buoyancy increment due to vertical velocity is an order of magnitude larger than the buoyancy increment given to parcels in the unmodified trigger function. It was therefore expected that convection in the vertical velocity experiment would trigger outside the range determined by the positive buoyancy experiment.

(c) Summary

The sensitivity of the timing and location of convection to the formulation of the trigger has been explored based on simulations of case 2. A simulation for which the constraints imposed by the trigger function were minimised (the positive buoyancy experiment) showed very similar timing and location of convection to a simulation using the original trigger function. It is therefore likely
A modified trigger function that included a dependence on the grid-scale vertical velocity, following the trigger function in the Kain-Fritsch scheme, resulted in an increase in the range of timing and location of convection. In addition to precipitation over the potentially unstable region over northwest England, the convection scheme triggered over regions where no convection occurred in nature. The Kain-Fritsch scheme was developed to aid simulations of mesoscale convection over mid-latitude continental land-masses where it is likely that large-scale convective environments are substantially different to those that occur over the UK. It may be appropriate to modify the formulation of the vertical velocity dependent trigger function for simulations over the UK to prevent the range in timing and location of convection extending well beyond the observed range.

4.5 Large-Scale Sensitivity to Partitioning

In this section, the response of the large-scale flow to convection is explored. A comparison between the large-scale response to parameterised and explicit convection is presented using results...
from the 10 minute and 1 day experiments. As described in chapter 1, mesoscale convection can be described in terms of PV structures. In the first half of this section, the signature of convection in the field of PV is identified. The dynamical and thermodynamical modifications associated with this convectively generated PV are then investigated and compared with observational data.

For case 2, the cumulative rainfall and therefore the net latent heating due to convection was shown in section 4.1 to be highly sensitive to the partitioning of convection. For this non-equilibrium case, the response of the large-scale flow is therefore expected be sensitive to the partitioning of convection. For case 1, however, the cumulative rainfall and therefore the net latent heating due to convection was shown to be only weakly sensitive to the partitioning of convection. For this equilibrium case, the response of the large-scale flow is not expected to be sensitive to the partitioning of convection.

### 4.5.1 PV Structure

**Case 1**

The signature of the explicit convective cells in the 1 day experiment at upper levels was a well-defined lens of negative PV, as shown in Fig. 4.16 at 1400 GMT 29 May 1999, 16 hours after the onset of convection. As discussed in chapter 1, the lens of negative PV is associated with cold and warm anomalies above and below the tropopause and anticyclonic rotation centred at the tropopause. The largest lens of negative PV, located over Wales and the Irish Sea, was generated by the explicit cells that remained to the west of England. The smaller lens, located over central England, was generated by the explicit cell that triggered in the region of the Cherbourg Peninsula, as shown earlier in Fig. 4.10 on page 93. Convection in the 10 minute experiment, however, generated a weaker region of negative PV at the tropopause-level co-located with the region of CAPE and low CIN, also shown in Fig. 4.16. This weaker parameterised convection failed to generate the lens structure or the magnitude of the upper-level region of negative PV seen in the 1 day experiment. This is an inevitable consequence of parameterised convection because it is inherently averaged over the region of CAPE and low CIN.

Vertical cross-sections of PV through the convecting region, also shown in Fig. 4.16, show the
Figure 4.16: Dry PV (Km$^2$kg$^{-1}$s$^{-1}$) at 1400 GMT 29 May 1999, 16 hours after the initial triggering of convection, for the 10 minute and 1 day experiments. PV=$2\times10^{-6}$Km$^2$kg$^{-1}$s$^{-1}$ is contoured in black.
1 day experiment

10 minute experiment

Figure 4.17: As in Fig. 4.16 but for case 2 at 0600 GMT 12 September 2000, 12 hours after the initial triggering of convection.
lens of negative PV in the 1 day experiment penetrated above the tropopause-level of the local environment. Convection in the 10 minute experiment, however, had much less an effect on the height of the tropopause (defined here as the PV = 2Km²kg⁻¹s⁻¹ surface). For this equilibrium case, the magnitude and distribution of convectively generated PV was sensitive to the partitioning of convection, against expectation.

Case 2

The characteristics of the convectively generated PV in the 10 minute and 1 day experiments were similar between the two case studies. The convectively generated PV for case 2 is shown in Fig. 4.17 in similar format to Fig. 4.16. As for case 1, convection in the 1 day experiment generated a lens of negative PV that penetrated above the tropopause-level of the local environment whereas convection in the 10 minute experiment generated a region of weaker negative PV co-located with the region of CAPE and low CIN with little impact on the tropopause height. For this non-equilibrium case, the magnitude and distribution of convectively generated PV was sensitive to the partitioning of convection, as expected.

4.5.2 Large-Scale Response

The large-scale dynamical and thermodynamical impact of the convectively generated PV is now explored. Owing to some fortuitous radiosonde ascents for case 2, the investigation continues only for this case. The similar characteristics of the convectively generated PV between the case studies allows for a generalisation of the results from case 2 to both cases.

(a) Dynamical Response

The effect of convection on a 250mb jet in both the 10 minute and 1 day experiments has been investigated by subtracting a background flow. The background flow was determined by performing a simulation with the latent heating due to condensation and freezing set to zero. For this model with dry dynamics, there was no convection and therefore no convective generation of PV. The background flow, shown at 250mb at 0600 GMT 12 September 2000 in Fig. 4.18, shows the
Figure 4.18: Wind speed (ms$^{-1}$, shaded) and wind vectors at 250mb for the background flow at 0600 GMT 12 September 2000.

Figure 4.19: The change in wind speed (ms$^{-1}$, filled contours, positive contours are solid and negative contours are dashed) at 250mb due to convection in a) the 1 day experiment and b) the 10 minute experiment at 0600 GMT 12 September 2000. Overplotted in (a) is a white contour of 250mb PV = $2 \times 10^{-6}$Km$^2$kg$^{-1}$s$^{-1}$ showing the lens shaped region of convectively generated PV. The arrows in (a) indicate anticyclonic rotation.
upper level jet. The jet increases within the rim width of 8 grid points from the eastern boundary. It is likely the flow was constrained at the eastern boundary since the boundary conditions were generated from the global model with moist dynamics in contrast to the mesoscale model with dry dynamics. As described in section 2.1b the global model uses the Gregory and Rowntree (1990) convection scheme with a 2-hour CAPE-closure timescale. The difference in flow between the two models in the region of the eastern boundary is likely to result from the exclusion of parameterised convective tendencies in the mesoscale model only.

Fig. 4.19 shows the change in 250mb wind speed due to convection in the 10 minute and 1 day experiments, 12 hours after initiation at 0600 GMT 12 September 2000. For the 1 day experiment, convection increased the background 250mb jet speed by up to 20ms$^{-1}$ along the northern edge of the PV lens. For the 10 minute experiment, convection increased the 250mb jet speed by 10ms$^{-1}$, approximately half that in the 1 day experiment due to the weaker upper-level anticyclone. In conclusion, the large-scale dynamical response to convection was sensitive to the partitioning of convection.

The impact of the eastern boundary can be seen in Figs. 4.19a and 4.19b by the decrease in the difference fields to almost zero over the rim width. The difference fields were expected to tend to zero at the boundary because here data from the global model using a 2-hour CAPE-closure timescale was subtracted from itself. It should be noted that Figs. 4.19a and 4.19b only show the top right corner of the mesoscale domain. The constrained large-scales may in turn feedback on the small-scale details of convection. Example magnitudes for the difference fields were taken at grid points away from the boundary where the solution is less constrained.

(b) Thermodynamical Response

The large-scale thermodynamical response to convection is compared between that in the 10 minute and 1 day experiments and the real atmosphere using radiosonde ascents. The radiosonde ascents shown in Fig. 4.21 from the Ekofisk oil platform in the North Sea (56.53°N, 3.22°E, marked by a cross in Fig. 4.20) provided evidence of a transient tropopause-level modification as the MCS moved over the site. The three profiles are at 1200 GMT 11 September 2000 (before), 0000 GMT 12 September 2000 (during) and 0000 GMT 13 September 2000 (after the MCS). The middle radiosonde ascended through the northeastern edge of the convectively generated cirrus shield as shown in Fig. 4.20. A transient warming below the 250mb tropopause and cooling above
Figure 4.20: Meteosat IR imagery at 0100 GMT 12 September 2000. The location of Ekofisk oil platform is marked by a cross.

250mb is consistent with an injection of mass into isentropic layers about the tropopause via deep convection.

Some observed examples of thermal modifications due to organised convection shown by Fritsch and Maddox (1981) were similar to those shown here. Cooling at 150mb, above the anticyclone centre, and warming at 300mb, below the anticyclone centre, were observed to extend greater than 700km from the anticyclonic centre. Bosart and Nielsen (1993) gave an example of a radiosonde penetration of a mesoscale anvil and is shown in Fig 4.22 together with two soundings from the surrounding environment. A comparison with the nearby profiles showed cold and warm anomalies above and below the tropopause as observed in this case. They provided evidence for undilute ascent from the sub-cloud layer to a level 2.6km higher than the surrounding undisturbed tropopause. The lower tropospheric profile at Ekofisk in Fig. 4.21 was not representative of the sub-cloud layer since the 0000 GMT 12 September 2000 radiosonde ascended through the north-eastern edge of the cirrus shield, north of the low-level source of warm and moist air.

A comparison with the equivalent model profiles relative to the convecting region is now presented. Model profiles representative of the atmosphere before, during and after convection are shown in
Figure 4.21: Observed temperature (solid) and dew-point temperature (dashed) profiles at Ekofisk oil platform 12 hours before (light grey), during (dark grey) and 24 hours after (black) the MCS. The profiles provide evidence of a transient upper-level modification due to organised convection.
**Figure 4.22:** Soundings (skew $T$-log$p$ format) at 1200 UTC 26 April 1991 for Lake Charles, Louisiana (LCH, solid lines) and two nearby soundings outside the convecting region. The 24°C wet-bulb potential temperature adiabat is dotted. Reproduced from Bosart and Nielsen (1993).

Fig. 4.23 for the 10 minute experiment and in Fig. 4.24 for the 1 day experiment. For the 1 day experiment, a transient cooling above 225mb and a transient warming below 250mb is consistent with the convectively generated PV and the observed modification. There is some transient cooling above 225mb in the 10 minute experiment but there is no evidence of any warming below. For the 10 minute experiment, the resulting vertical profile due to convection was a tropospheric deep layer of saturated adiabatic lapse rate consistent with removal of CAPE on a fast timescale capped by a higher than observed tropopause at 230mb.
Figure 4.23: Model temperature (solid) and dew-point temperature (dashed) profiles for the 10 minute experiment at Ekofisk 12 hours before convection (light grey), at Ekofisk during convection (dark grey) and downstream from convection (black). The arrow points to a profile with a saturated adiabatic lapse rate (SALR).
Figure 4.24: As in Fig. 4.23 but for the 1 day experiment.
Chapter Summary

For both the equilibrium and non-equilibrium cases, the small-scale details of convection including intensity, location, horizontal scale and evolution were highly sensitive to the partitioning of convection.

For case 2, the net precipitation and therefore the net latent heating due to convection were very sensitive to the partitioning of convection resulting in a range between $0.78 \times 10^{-8}$ and $3.53 \times 10^{-8}$ mm at T+18. For this non-equilibrium case, the equilibrium-based convection scheme failed dramatically. A coupling between the convection scheme and the grid-scale dynamics allowed parameterised convection to penetrate into the region of CIN and realise a significant amount of energy unavailable to convection in nature. The response of the large-scale flow was also sensitive to the partitioning of convection, as expected. Explicit convection generated a well-defined lens of tropopause-level negative PV whereas the inherently averaged parameterised convection generated a region of weaker PV co-located with the region of CAPE and low CIN. The vertical profile of the thermodynamic modification due to convection was shown to be different for parameterised and explicit convection. Moreover, the magnitude of the thermodynamic modification due to explicit convection was shown to be closer to that observed than for parameterised convection.

For case 1, convection was free to act in the absence of significant CIN and the cumulative precipitation and therefore net latent heating due to convection was very similar between the three experiments with a range between $2.00 \times 10^{-8}$ and $2.29 \times 10^{-8}$ mm at T+16. For this equilibrium case, the convection scheme behaved as expected and produced a smooth field of parameterised rainfall within the region of sustained large-scale forcing. Parameterised convection provided guidance as to the region of large-scale forcing within which organised convection was observed. The response of the large-scale flow, however, was sensitive to the partitioning of convection, against expectation. The impact of convection on the vertical thermodynamic profile was different between parameterised and explicit convection. The modification of the large-scale flow was therefore sensitive to the partitioning of convection for both the equilibrium and the non-equilibrium cases.

For the 1 day experiment, buoyant instability was almost entirely removed via explicit mesoscale cells and the convective response was similar for both the equilibrium and the non-equilibrium
case. Useful information was provided on the track of the MCSs and the rainrates were of a more similar magnitude to those observed than in the other two experiments. In addition, the large-scale response in the 1 day experiment was closer to the observed response in terms of the magnitude of the upper-level temperature modification than in the 10 minute experiment. The notable difference between the cases was the accuracy of the triggering location of deep convection. The predictability of the precise triggering locations is investigated in the next chapter.
In this chapter, the ability of the model to reproduce the precise location and timing of convection in the two cases is investigated. As discussed in chapter 3, the two cases were chosen for their apparent difference in predictability. In this chapter, predictability is examined formally using an ensemble technique to sample the range of possibilities in location of explicit convective cells in the 1 day experiments. In this aim, the important issues for the generation of the ensemble are identified. In the first section, however, the mechanisms for the initiation of convection in the model are investigated by looking at the evolution of the temperature and moisture fields prior to the onset of convection and isolating the effects of orography and land/sea contrasts.

5.1 Mechanisms for Initiation

Case 1

An explicit convective cell developed over the English Channel from 0400 GMT 29 May 1999, as shown earlier in Fig. 4.10 on page 93. Ascent due to advection along forward sloping isentropes with height at the low-level warm-front, shown earlier in the schematic vertical cross section in Fig. 3.3 on page 60, resulted in an increase in 775mb equivalent potential temperature, \( \theta_e \). At 0000 GMT the potentially unstable band of high \( \theta_e \) between 850 and 750mb over northern France (shown in Fig. 5.1a at 775mb) contained small-scale (approximately 5 grid lengths) maxima with magnitudes of approximately 0.1K greater than the local environment (as discussed later in section
Figure 5.1: Equivalent potential temperature (K, contoured with an interval of 0.3K) and vertical velocity (m s\(^{-1}\), shaded) at 775mb, the level of convective initiation at (a) 0000 GMT (b) 0100 GMT (c) 0200 GMT and (d) 0300 GMT 29 May 1999. The amplified small-scale maximum is marked by an arrow in (d). The observed triggering location is marked by a cross in (d).
Three hours later, a small-scale maximum had amplified to approximately 1.0K greater than the local environment, as shown in Fig. 5.1d. The effect of the amplification of the local maximum in $\theta_e$ on the vertical stability is shown in Fig. 5.2. The 775-700mb layer saturated and the middle troposphere cooled. The large-scale dynamics responded to this absolutely unstable structure near the grid-scale within the large-scale conditionally unstable region and explicit convection subsequently developed.

In the 1 day experiment described in the previous chapter, explicit convection triggered in the region of the Cherbourg Peninsula at the time of the observed convection but not at the observed location. It is hypothesised that the relatively low orography of the Cherbourg Peninsula acted as a forcing for the local maximum in instability in the model. A simulation with the peninsula replaced by sea, however, led to the development of a maximum in $\theta_e$ of the same magnitude in the same location as the simulation using the default orography and produced a nearly identical field of precipitation, as shown in Fig. 5.3. On the other hand, a simulation with the orography over northern France reduced to 0.1% of its original elevation, also shown in Fig. 5.3, failed to amplify a local maximum in low-level $\theta_e$ until the potentially unstable band tracked over southern England.

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{figure52.png}
\caption{Vertical profiles of temperature (solid line) and dew-point temperature (dashed line) at 0000 GMT (black), through the centre of the potentially unstable band before the amplification of a local maximum in $\theta_e$, and at 0300 GMT (blue), through the local maximum in $\theta_e$.}
\end{figure}
Figure 5.3: Orography (m), 775mb $\theta_e$ (K, contoured as in Fig. 5.1) and vertical velocity (ms$^{-1}$, shaded) at 0300 GMT and accumulated model precipitation (mm) between 0300 and 1300 GMT 29 May 1999 (Case 1).
Vertical velocity at 775mb was typically $0.01\text{ms}^{-1}$ compared to that of $0.03\text{ms}^{-1}$ using the default orography. Explicit convection triggered over southern England 6 hours later than convection in the simulation using default orography. The model created a layer of neutral stability in place of the removed orography. The effect of this layer on the solution will be negligible at the time of convective triggering using default orography, 8 hours into the simulation. The simulations imply that the circulations set up by the orography of northern France were necessary for the triggering of deep convection.

The change in stability between 0000 GMT and 0300 GMT for the two simulations with modified orography (in the location of the local maximum in $\theta_e$ in the experiment with default orography) are shown in Fig. 5.4. For both the simulation with unmodified orography (shown earlier in Fig. 5.2) and the simulation without the Cherbourg Peninsula the 700-775mb layer saturated and the middle troposphere cooled resulting in absolutely unstable profiles. In addition, low-level $\theta_e$ increased but by an amount too small to be apparent in the vertical profiles shown in Figs. 5.2 and 5.4a but large enough to trigger explicit convection. For the simulation with reduced orography over northern France the profile became more unstable by 0300 GMT but remained unsaturated at all levels as shown in Fig. 5.4b.

The model and nature triggered convection in different locations. This makes it difficult to infer anything about the triggering mechanism in nature from the model. The model contained only statistical information about the small-scale orography and flow, the details of which may have been important for triggering convection in nature. At 12km horizontal resolution the mesoscale model cannot resolve such fine structure and, more importantly, no observational data are assimilated at such small scales. The triggering mechanism in nature cannot be determined due to a lack of observations but there is evidence to suggest that it involved small-scale circulations within the potentially unstable region forced by the orography.

**Case 2**

In the 1 day experiment described in the previous chapter, two explicit convective cells developed over northwest England and another cell developed over eastern England at 1800 GMT. Prior to convection, a region of high $\theta_e$ at low-levels was located over much of England within the warm sector of the cyclone. The large-scale northward advection of high $\theta_e$ over England prior
Figure 5.4: Vertical profiles of temperature (solid line) and dew-point temperature (dashed line) at 0000 GMT (black), and at 0300 GMT (blue) at the same location as for the profiles in Fig. 5.2 for (a) the simulation without the Cherbourg Peninsula and (b) the simulation with reduced orography over northern France.
to convection is shown in Fig. 5.5. The region of high $\theta_e$ air at 900mb over northwest England resulted from ascent of order 0.1 ms$^{-1}$ ahead of the eastward moving cold front. This ascent is consistent with forced ascent caused by flow along backward sloping isentropic surfaces with height associated with the cold front (not shown). Fig. 5.5 also shows a strengthening of the $\theta_e$ gradient over the Irish Sea associated with the cold front and an increase in ascent over northwest England between 1600 and 1800 GMT. CIN increased away from the cold front. A small part of the region of CAPE was therefore also a region of almost zero CIN over northwest England, as shown earlier in Fig. 3.13 on page 72.

Ascent at low-levels forced by the eastward moving cold front and cold advection by the mid-level jet resulted in the potentially unstable region over northwest England. At 12km horizontal resolution it is hoped the mesoscale model well resolves accurate large scales. The model and nature developed convection within the same region and allows the mechanism for the destabilisation in nature to be inferred from the model.

Fig. 5.6 shows the growth of local maxima in 850mb $\theta_e$ within this potentially unstable region between 1600 and 1800 GMT. Convection developed at the locations of the two maxima shown

Figure 5.5: Equivalent potential temperature (K, contoured with a contour interval of 0.6K) and vertical velocity (ms$^{-1}$, shaded) at 900mb, close to the level of convective initiation at (a) 1600 GMT and (b) 1800 GMT 11 September 2000 prior to the development of explicit convection. The region of ascent and high $\theta_e$ over northwest England is marked by an arrow in (b).
in Fig. 5.6b from 1800 GMT. This saturated buoyant air at 850mb together with the cooling of the middle troposphere provided the trigger for deep convection. The change in the local vertical stability at the more southerly local maximum over northwest England between 1600 and 1800 GMT is shown in Fig. 5.7. The 900-800mb layer warmed and saturated and the 800-600mb layer cooled due to cold advection by the mid-level jet by up to 2°C. The increase in $\theta_e$ is much larger than that that occurred for case 1 and is therefore apparent in the vertical profiles shown in Fig. 5.7. The large-scale dynamics responded to the resulting absolutely unstable profile within the potentially unstable region and explicit convection subsequently developed.

It is hypothesised that the orography of northwest England was a necessary component of the triggering mechanism in the model. To test this hypothesis, experiments were performed with modified orography with the effects of orographic roughness in the boundary layer scheme removed. A simulation with the local orography of northwest England reduced to 0.1% of its original elevation triggered two convective cells in the same locations as the solution using the default orography, as shown in Fig. 5.8. A simulation with the orography of the interior of the mesoscale domain reduced close to zero, using a two-dimensional hanning function, had some impact on the intensity

![Figure 5.6: Equivalent potential temperature at 850mb over northern England at (a) 1600 and (b) 1800 GMT 11 September 2000. Two local maxima that triggered convection at the next timestep are marked by arrows in (b).](image-url)
of convection but only a small impact on the locations, also shown in Fig. 5.8. This experiment shows that the triggering locations of the explicit cells were not tied to the underlying orography. The small changes in convective intensity and location in the experiment with a flat interior domain may be a result of modified large-scale dynamics, specifically the cold front, by the removal of a large area of orography. The precise mechanism for the amplification of the local maxima in 850mb $\theta_e$ that triggered convection in the model cannot be determined from the experiments performed here and nothing can be inferred about the triggering mechanism in nature. The experiments do show, however, that the locations of convection were not sensitive to the orography.

Explicit convection also developed over eastern England, as shown earlier in Fig. 4.10 on page 93. A timeseries of low-level environmental wind perpendicular to the coast (not shown) shows cooler air moved inland between 1600 and 1800 GMT. Convergence at the leading edge of this sea-breeze front resulted in sufficient ascent to locally remove CIN and explicit convection developed from 1800 GMT. Convergence and a temperature gradient was observed by surface synoptic stations but no convection was observed. At 12km resolution, the model cannot well resolve such small scales and may be the reason for the false trigger. There are many other possible sources of error.
Figure 5.8: Orography (m) and accumulated model precipitation (mm) between 1700 and 2300 GMT 11 September 2000 (Case 2) without the effects of orographic roughness in the boundary layer scheme.
such as the artificial enhancement of vertical velocity by the model’s hydrostatic approximation or inaccurate sea-surface temperatures.

5.2 Ensemble Procedure

The range of possible triggering locations consistent with the observed scales is investigated using an ensemble technique. The general details of the method used for both cases are described first, followed by the case-dependent details. The importance of the timing and the amplitude of the perturbation field used in the generation of the ensemble member initial states is emphasised. Some verification of the ensemble procedure is provided and the section ends with some results.

Ensemble members are initialised from a modified initial state. The intention of this approach is to generate an ensemble of modified states that are consistent with the observed large scales. The ensemble is designed specifically to sample small-scale uncertainty and does not account for errors in the model physics. It is recognised that only the range of possibilities within the bounds of small-scale uncertainty will be sampled and that the range may be constrained by not sampling model error. The temperature field between model levels 7 and 16 (approximately 900 and 700mb), the layer of convective initiation, is modified by the addition of a field of spatially-coherent Gaussian perturbations obtained in the following way.

The two-dimensional circular Gaussian function of the form:

$$f(x, y) = e^{-[(x-\mu_x)^2+(y-\mu_y)^2]/2\sigma^2}, \quad (5.1)$$

where the maximum in $f(x,y)$ is located at grid-point $(\mu_x, \mu_y)$ and the standard deviation, $\sigma = 1.5$ grid points in both horizontal dimensions is applied over an area of $16 \times 16$ grid points. This Gaussian mask is centred at each grid-point in the model domain. The amplitude of each Gaussian is generated randomly between -1 and 1. In this aim, a number is generated from a uniform distribution between 0 and 1. The range is then extended to between -1 and 1 by multiplying by 2 and subtracting 1. An example field of these superimposed Gaussians is shown in Fig. 5.9. The field is such that its domain average is zero. The random field of superimposed Gaussian bumps was chosen because of its spatial correlation. If the random field was uncorrelated between grid
points, it would be heavily damped in the first few timesteps, reducing the amplitude and making it difficult to precisely control the amplitude of the perturbations that actually trigger the convection.

Although a horizontal scale for a Gaussian distribution is rather arbitrary, the distance ($\lambda$) between

**Figure 5.9:** Example field of randomly generated Gaussian perturbations over the model domain.

**Figure 5.10:** The distance ($\lambda$) between points that are a fraction ($1/N$) of the maximum for the function given in equation 5.2.
points that are some fraction \((1/N)\) of the maximum value of the Gaussian function is often used and is defined as:

\[
\lambda = 2\sigma\sqrt{2\ln N},
\]

where the standard deviation \((\sigma)\) is 1.5 grid lengths. The relationship in equation 5.2 is shown in Fig. 5.10. For this investigation, the scale is chosen to be the distance between points that are a quarter of the maximum value \((N = 4)\) since the effect of the increase in \(\lambda\) for \(N > 4\) is negligible for the results presented later. The horizontal scale of the superimposed Gaussians is equal to the horizontal scale of the single Gaussian mask and for \(N = 4\), \(\lambda = 5.0\) grid lengths or 62.5km. This horizontal scale is one that the model can represent without the need for large artificial diffusion and is an estimate of the scales that are not corrected by data assimilation. Under the Met Office 3D-VAR data assimilation scheme used in case 2, observational data are assimilated over a typical horizontal scale of 70km. Smaller scales are purely spun up by the mesoscale model and it is these scales that will be modified using the perturbation field shown in Fig. 5.9. As discussed in chapter 2, the model was initialised from an ECMWF analysis at \(1^{\circ}\) horizontal resolution for case 1 and no additional data were assimilated. It is therefore valid to modify the low-level temperature in case 1 using the same procedure as for case 2.

The Probability Density Function (PDF) of the perturbation field is also Gaussian and its amplitude is increased linearly from zero over model levels 7 to 9 and decreased linearly to zero over model levels 14 to 16 and is constant over model levels 10 to 13. The importance of the timing and amplitude of the perturbation field is identified in the discussion of the case-dependent details in section 5.3. Specific humidity is consistently modified by ensuring constant relative humidity during the modification. An example vertical profile of temperature and dew-point temperature before and after the modification, shown in Fig. 5.11, shows the effect of the modification is to change the stability of a parcel originating from the modified layer.

For both case studies, the timing and location of convection was not affected by flow through the boundaries. A simulation with the western boundary moved further West by 50 grid lengths for case 1 (not shown) showed the distribution and amount of precipitation were only very weakly sensitive to the position of this boundary. For case 2, convection developed close to the centre of the domain. It is therefore valid to force the domain with unmodified boundary conditions for both cases. Du et al. (1997) adopted a similar ensemble procedure to study the impact of initial condition uncertainty on QPF but allowed for land-sea and latitudinal variations in analysis.
uncertainty. Allowing for variations in small-scale uncertainty across the mesoscale domain used here for short-range forecasts is considered unnecessary. Du et al. (1997) also removed energy associated with inertial-gravitational modes using a non-linear normal mode initialisation. The perturbations used here are unbalanced and are not initialised because such small scales in the atmosphere are not necessarily in balance (as in Stensrud et al., 2000).

Assuming the large scales are accurate in the model, the ensemble of modified states is an estimate of the PDF for the observed state. Ideally, as much of the PDF as possible is independently sampled by the finite ensemble. The ensemble size is required to be sufficient to realise a large proportion of the solution of an infinite ensemble. Theoretical results from Leith (1974) showed that an ensemble size as small as 8 appreciably increased accuracy over the deterministic forecast. For a short-range five-member mesoscale ensemble, Du et al. (1997) increased the rank probability score of probabilistic QPF by 63% of the improvement attainable through an infinite ensemble. For 10 members, the increase was 90%. Here, 6 random perturbation fields are generated and inverted to initialise a 12 member ensemble. This technique constrains the ensemble mean perturbation field to be zero.
5.3 The Timing and Amplitude of the Perturbation Field

As discussed in the introduction, ensembles exploiting analysis uncertainty have generally involved some perturbation at analysis time. For these experiments, the sensitivity of the triggering locations to small-scale variability is investigated by modifying convective stability part way through the simulation prior to convection. The importance of the timing and amplitude of the perturbation field is demonstrated together with some verification of the ensemble technique.

The ensemble member solutions and the unmodified solution are required to be solutions to the same large-scale forcing. The intention of the ensemble procedure is to introduce a significant perturbation into the model using a method that does not change the large scales. This is achieved by introducing the perturbation field shown in Fig. 5.9 at an earlier time (a few hours before convection). For both cases, the perturbation that triggered convection was shown in section 5.1 to amplify in time in the hours before initiation. Therefore significant perturbations can be created by introducing the perturbation field with small amplitude into the model at an earlier time. The amplitude of the perturbation field is chosen to be the amplitude of the small-scale variability in the unmodified field. Errico and Baumhefner (1987) found diffusion affected small-scale perturbations in a limited area model. However, they found diffusion did not dominate the small scales and they hypothesised an opposing forcing mechanism to be some diabatic or adiabatic process. An external forcing mechanism (e.g. surface heterogeneities or the large-scale forcing) acts on the modified states to amplify a perturbation in time.

The effect of adding a perturbation field with larger amplitude, and therefore larger variance, than that of the small-scale variability in the unmodified state is to produce a modified state with increased variance. This widening of the PDF results in more grid points being potentially ‘convec
tive’. One consequence of this would be enhanced precipitation in the modified solution compared to the unmodified solution. This represents too large a modification to the behaviour of convection to be consistent with the observed large scales. The variance of the perturbation field is chosen to be that of the existing small-scale variance at the time of the modification.

Case 1

An analysis of the model’s triggering mechanism for case 1 (see section 5.1) showed the amplification of a small-scale maximum in $\theta_e$ between 850mb and 750mb prior to the development of
explicit convection from 0400 GMT 29/5/99. The perturbation field is introduced at 0000 GMT 29/5/99 at a time during which the large scales are providing a forcing for convection. A typical magnitude of the existing small-scale variance is determined by filtering the $\theta_e$ field at the level of convective initiation of 775mb. A high-pass Butterworth frequency filter was used to remove wavelengths greater than a critical wavelength. The critical wavelength represents a horizontal scale for the perturbation field shown in Fig. 5.9 and is chosen to be 5.0 grid lengths as described in section 5.2. Larger scales are removed from the solution. The histogram of the filtered solution over the potentially unstable region, shown in Fig. 5.12, shows a skewed distribution that has a peak probability density for magnitudes of approximately 0.1K and a variance of $6.4 \times 10^{-3}$K. The variance of the perturbation field is not set exactly equal to that of the unmodified state since the perturbation field is applied to temperature and not $\theta_e$.

For a given perturbation to temperature (and an associated perturbation to specific humidity), $\theta_e$ is perturbed by a larger magnitude. The perturbation to $\theta_e$ for unsaturated air may be approximated by:

$$\Delta \theta_e = \Delta \theta \left(1 + \frac{L_c \Delta q}{C_p \Delta T}\right). \tag{5.3}$$

where $\Delta \theta$, $\Delta T$ and $\Delta q$ are the perturbations to potential temperature, temperature and specific humidity, $L_c$ is the latent heat of condensation and $C_p$ is the specific heat of dry air. $\theta_e$ will receive a larger perturbation than that given to $\theta$ by an amount given by the additional term

![Figure 5.12: A histogram of the filtered field of 775mb equivalent potential temperature (K) at 0000 GMT 29/5/99 for the unmodified solution for the potentially unstable region.](image)
Using typical values of 775mb temperature and specific humidity within the potentially unstable band at 0000 GMT, a temperature perturbation of 0.1K results in a $\theta_e$ perturbation of 0.23K. The perturbation to $\theta_e$ is greater than twice the magnitude of the perturbation to temperature. In addition, the perturbation to $q$ for a given perturbation to $\theta$, assuming constant relative humidity, will increase exponentially with temperature (following the Clausius-Clapeyron relation). It follows that the variance of the modified $\theta_e$ field will be greater than the variance of the modified temperature field. The perturbation field must therefore have less variance than that of the unmodified $\theta_e$ field to ensure that the variance of the modified $\theta_e$ field is similar to the variance of the unmodified $\theta_e$ field. The variance of the perturbation field is chosen to be $1.6 \times 10^{-3} \text{K}$ which corresponds to 99% of the data points within the range ±0.1K. This field will be referred to as the 0.1K perturbation field.

The histograms for the filtered field of 775mb $\theta_e$ at 0000 GMT for the unmodified solution and an example ensemble member modified at 0000 GMT are shown in Fig. 5.13. The histogram of the 0.1K perturbation field used in the generation of the ensemble member initial conditions is also shown. Indeed, the variance of the modified $\theta_e$ field (dotted line) is greater than that of the 0.1K perturbation field (dashed line) that was added to the temperature field. The ensemble member $\theta_e$ field has a variance of $8.28 \times 10^{-3} \text{K}$ which is similar to that of the unmodified solution $\theta_e$ field of $6.4 \times 10^{-3} \text{K}$. A quantitative analysis reveals that 91% of the ensemble member $\theta_e$ field data points lie within the 95% limits of the unmodified $\theta_e$ field.

The difference between adding smaller amplitude perturbations earlier and larger amplitude perturbations later is shown in the fields of 775mb $\theta_e$ at 0400 GMT 29/5/99 in Fig. 5.14. The unmodified solution shown in Fig. 5.14a has a local maximum. The two example ensemble member solutions using a 0.1K perturbation field introduced at 0000 GMT, shown in Figs. 5.14b and 5.14c, both show a local maximum similar in magnitude to the unmodified solution but in a different location. For each ensemble member, it is expected that convection will trigger in a different location to that in the unmodified solution. Each of the 12 ensemble members were found to amplify a small-scale maximum in $\theta_e$ (sometimes two, depending on the perturbation field) of similar magnitude to that of the unmodified solution in different locations within the band of high $\theta_e$ (not shown). An example ensemble member solution in which a 1.0K perturbation field was introduced at 0400 GMT (using a perturbation field of similar amplitude to that of the existing small-scale variability at the time of the modification), shown in Fig. 5.14d, shows multiple maxima. Although the 1.0K perturbation field has similar amplitude of variability to the unmodified solution, it has different
Figure 5.13: Histograms for the filtered field of 775mb equivalent potential temperature at 0000 GMT 29/5/99 for the unmodified solution (solid line) and an example ensemble member (dotted line) using a 0.1K perturbation field at 0000 GMT (dashed line).

The four fields shown in Fig. 5.14 have been filtered using the same filter as for Fig. 5.12 to remove wavelengths greater than 5 grid lengths. The histograms of the four filtered fields are shown in Fig. 5.15. The histograms for the unmodified and the example ensemble members using a 0.1K perturbation field at 0000 GMT have a similar mean and variance. The ensemble procedure has modified small-scale $\theta_e$ without significantly changing its variance. The histograms for the unmodified solution and the modified solution using a 1.0K perturbation field at 0400 GMT shows increased variance in the modified state.

The impact of the time of the modification is seen more clearly in the difference in $\theta_e$ between the ensemble member and the unmodified solution at 0400 GMT shown in Fig. 5.16. The difference fields using a 0.1K perturbation field at 0000 GMT show a dipole associated with the relocation of the local maximum whereas the difference field using a 1K perturbation field at 0400 GMT returns the perturbation field. The histograms of the filtered difference fields, shown in Fig. 5.17, show that the difference fields using the smaller amplitude perturbation field earlier have much smaller
CHAPTER 5: Predictability of Convection

Figure 5.14: Equivalent potential temperature at 775mb at 0400 GMT for (a) the unmodified solution, (b) and (c) the solution modified at 0000 GMT using 0.1K perturbation field using 2 different perturbation fields and (d) the solution modified at 0400 GMT using a 1.0K perturbation field.
Figure 5.15: Histograms of filtered equivalent potential temperature (K) at 775mb at 0400 GMT for the four solutions shown in Fig. 5.14. The unmodified solution is shown by a solid line, the solutions modified at 0000 GMT using a 0.1K perturbation field are shown by dotted lines and the solution modified at 0400 GMT using a 1.0K perturbation field is shown by a dashed line.

The ensemble procedure does not increase the variance and therefore does not result in enhanced precipitation as shown in Fig. 5.18. Much later in the simulations the ensemble members diverged from the unmodified solution but this was considered unimportant for the purpose of the ensemble which was to determine the range of possible triggering locations. The cumulative precipitation from the solution using a 1.0K perturbation field at 0330 GMT shows increased precipitation soon after the introduction of the perturbation field. Such large amplitude perturbations were sufficient to change the large scales by increasing the size of the region within which convection can occur.

Case 2

Explicit convection developed from 1800 GMT 11/9/00. The perturbation fields were chosen to be introduced at 1600 GMT 11/9/00 to be within the period of mesoscale forcing. The amplitude for the existing small scales was determined by filtering the field of \( \theta_e \) at the level of convective initiation (850mb) to remove scales larger than the horizontal scale of the Gaussian perturbations (5 grid lengths). The filtered fields of \( \theta_e \) at the level of convective initiation for case 1 and case 2, shown in Fig. 5.19, show that scales less than 5 grid lengths have a much greater amplitude for case 2 than case 1.
CHAPTER 5: Predictability of Convection

(a) Using a 0.1K perturbation field at 0000 GMT. Ensemble member 8.
(b) Using a 0.1K perturbation field at 0000 GMT. Ensemble member 9.
(c) Using a 1.0K perturbation field at 0400 GMT. Ensemble member 8.

Figure 5.16: Difference in equivalent potential temperature (K) at 775mb at 0400 GMT between the unmodified solution and (a) and (b) the solution modified at 0000 GMT using a 0.1K perturbation field using 2 different perturbation fields and (C) the solution modified at 0400 GMT using a 1.0K perturbation field.
Figure 5.17: Histograms of the filtered difference fields of equivalent potential temperature (K) at 775mb at 0400 GMT for the three fields shown in Fig. 5.16. The difference fields using a 0.1K perturbation field at 0000 GMT are shown by the dotted lines and the difference field using a 1.0K perturbation field at 0400 GMT is shown by the dashed line.

Figure 5.18: Cumulative model precipitation averaged over the potentially unstable region (mm) for case 1 for the unmodified solution (thick solid line), two example ensemble members using 0.1K perturbations at 0000 GMT (red solid lines) and an example solution using 1K perturbations at 0330 GMT (dashed line).
Figure 5.19: Filtered equivalent potential temperature (K) at (a) 775mb at 0000 GMT 29/5/99 and (b) 850mb at 1600 GMT 11/9/00. An example black contour of unfiltered $\theta_e$ gives an indication of the region of high unfiltered $\theta_e$. 
A histogram of the filtered field in Fig. 5.19b for case 2 would not be representative of the underlying PDF due to the small sample size within the potentially unstable region over northwest England (as defined in Fig. 5.6 by the N-S orientated region of high $\theta_e$ ahead of the cold front between north Wales and Cumbria). However, it is clear from Fig. 5.19b that the small-scale magnitudes over northwest England were typically in the range ±1K. For the 0.1K perturbation field used in case 1, 99% of the values lie within ±0.1K and the field had a variance of $1.6 \times 10^{-3}$. Following a similar approach, the perturbation field used for case 2 is such that 99% of the values lie within ±1K. This 1K perturbation field has a variance of 0.135K. The ensemble of initial conditions were therefore generated using a 1.0K perturbation field at 1600 GMT 11/9/00 to minimise any impact on the variance of low-level $\theta_e$. It is not practical to show the equivalent histograms for case 2 as shown for case 1 due to the small sample size within the potentially unstable region.

### 5.3.1 Ensemble Bias

In addition to the problem of increased variance, an ensemble created by introducing the perturbation field closer to the time of convection will contain a bias to the location of the small-scale maximum in the unmodified solution (not shown) as predicted by the simple model presented below.

A simple model demonstrates the impact of having a pre-existing anomaly within an otherwise homogeneous field on the ensemble mean field. Consider an infinite horizontal grid of uniform forcing for convection. Grid-point A has enhanced instability and provides the trigger for convection. This is a simplification of the potentially unstable region in both cases prior to convection and is shown in Fig. 5.20. The ensemble procedure changes each grid-point by an amount chosen randomly between $\pm \epsilon$. The top-hat function used here is a simplification to the Gaussian function used in the mesoscale model but is sufficient for the purposes of this simple demonstration. Instead of considering the magnitude of the perturbations, let us consider the value normalised by the magnitude of the trigger and call it $n$.

For $n$ chosen randomly between $\pm 1$, no grid-point will attain greater instability than grid-point A. Grid-Point A, however, is also perturbed. Only those ensemble members that modify grid-point A by $\leq 0$ will contain grid points of greater or equal magnitude to the trigger. It follows that 50% of a large ensemble using $n$ chosen randomly between $\pm 1$ will trigger convection away from grid-point
A. Following a similar argument, 75% of a large ensemble using \( n \) between \( \pm 2 \) will contain grid points of greater or equal magnitude to the trigger and will therefore trigger convection away from grid-point A. Decreasing \( n \) introduces a discontinuity for \( n \) chosen randomly between \( \pm 0.5 \). For this modification, no grid points will be of greater or equal magnitude to grid-point A, even if grid-point A is modified by \( n = -0.5 \). All ensemble members will trigger convection at grid-point A.

The relationship between \( n \) and the bias to grid-point A of a large ensemble is shown graphically in Fig. 5.21. If there exists a pre-existing trigger (grid-point A in the simple model) a situation in which the triggering location is entirely dependent on the perturbations does not exist. The model shows that a random field added to a non-homogeneous field does not give a random field. The ensemble mean of a large ensemble will always have a bias to the pre-existing trigger. To remove the bias, the pre-existing trigger would need to be removed before the addition of the random field.

An upper limit to the magnitude of the perturbations when the ensemble procedure is applied to a real case study is shown by the dashed line in Fig. 5.21. The perturbations must not change the observed large scales because the ensemble members must be solutions of the same large-scale environment. The ensemble technique used for both case studies introduces the perturbation field, chosen randomly for \( n \) between \( \pm 1 \), a few hours before the development of convection. The amplitude for the perturbation field was determined by setting the variance of the perturbation field to be similar to that of the filtered unmodified solution. An external forcing mechanism then acts over time on the 12 modified states and, for case 1, a local maximum amplified to an order of magnitude greater than surrounding small-scale perturbations. This method reduces the problem of bias by
introducing a significant perturbation into the model using a method that does not change the observed large scales by allowing an external forcing to amplify small amplitude perturbations. The forcing mechanism for small-scale amplification was present in both the unmodified and ensemble member solutions. Fig. 5.3 on page 121 showed the forcing mechanism for case 1 was provided by the orography of northern France. The external forcing mechanism for the amplification of small-scale perturbations in case 2 could not be determined by the sensitivity studies performed in section 5.1. It is, however, hypothesised that the forcing was present in both the unmodified and ensemble member solutions and may have been a fixed surface feature (e.g. a land/sea boundary) or the unperturbed large scales (as described by Anthes et al. (1985)).

5.4 Ensemble Verification

Case 1

The ensemble samples the range of possibilities in triggering location consistent with the large-scale dynamics. The modification is intended to be one to which the location of convection may be sensitive but which leaves the large-scale environment unperturbed. For the equilibrium case, the cumulative precipitation over the region of CAPE and low CIN was shown in section 4.1 to be
controlled by the large scales rather than the partitioning of convection. If the large scales remain similar between the unmodified and ensemble member simulations, the cumulative precipitation is expected to be similar. Fig. 5.22 shows the cumulative precipitation for case 1 averaged over the region of CAPE, as shown in Fig. 3.4 on page 61, for the unmodified and the 12 ensemble member simulations. The cumulative rainfall was very similar between the ensemble member solutions for the duration of the mesoscale forecast. In addition, the cumulative rainfall for the ensemble member solutions did not diverge significantly from the cumulative rainfall of the unmodified solution. This suggests the ensemble members, including the unmodified solution, were all model solutions to the same large-scale environment and the ensemble procedure did not modify the large scales, as required.

Case 2

For the non-equilibrium case, the cumulative precipitation was shown in section 4.1 to be very sensitive to the partitioning of convection. It is therefore expected that the cumulative rainfall will
be sensitive to small-scale variability in the initial conditions. Despite adding a perturbation field which is thought to have similar amplitude to that of the small-scale variability in the unmodified solution, the cumulative rainfall shown in Fig. 5.23, shows a large ensemble spread about the unmodified solution. The potentially unstable region ahead of the cold front was small such that the area average perturbation field can be significantly different to zero. Adding a non-zero mean perturbation field will change the number of potentially ’convective’ grid points leading to significantly different rain amounts for this non-equilibrium case.

![Graph showing area average cumulative model precipitation](image)

**Figure 5.23:** Area average cumulative model precipitation (mm) for case 2 for the unmodified solution (dashed line), the ensemble members (thin solid lines).

The greater spread in cumulative rainfall than for case 1 does not imply that the amplitude of the perturbation field was sufficiently large to modify the large scales. The amplitude of the perturbation field was not sufficient to locally remove the inversion that existed over most of England (see Fig. 3.13 on page 72). Applying a perturbation field with larger amplitude than that of the small-scale variability in the unmodified solution was found to result in the localised removal of CIN over England (not shown) and intense convection realised the large reservoir of energy unavailable to the unmodified solution. For this ensemble, the ensemble member simulations behaved very differently from the unmodified simulation and the ensemble member initial states were not consistent with the large scales.
5.5 Results

Case 1

As shown in section 5.1, the growth of a small-scale maximum in $\theta_e$ within the band of high $\theta_e$ at low-levels provided the trigger for convection. Accumulated model precipitation for the unmodified solution for 10 hours after convective initiation, shown in Fig. 5.24a, shows the northward track of a single persistent cell across the English Channel. The cell triggered in the region of the Cherbourg Peninsula at the time of the observed cell but not at the observed location which is marked by a cross.

![Accumulated model precipitation (mm) between 0300 and 1300 GMT 29 May 1999 (Case 1) for (a) the unmodified solution and (b) the ensemble mean. The observed triggering location is marked by a cross.](image)

Figure 5.24: Accumulated model precipitation (mm) between 0300 and 1300 GMT 29 May 1999 (Case 1) for (a) the unmodified solution and (b) the ensemble mean. The observed triggering location is marked by a cross.

The ensemble mean field of accumulated precipitation, shown in Fig. 5.24b for the same time period as in Fig. 5.24a, shows greater spread in the east-west direction and therefore a lack of predictability. The maximum in the mean accumulated precipitation over southern England was only a third of that achieved by many of the individual ensemble member solutions. Each ensemble member triggered a convective circulation, sometimes two depending on the modification, within the potentially unstable band and the precise triggering location of deep convection was determined by the modification.
The triggering of convection did not occur randomly within the potentially unstable region but favoured the region of the Cherbourg Peninsula because the region of high $\theta_e$ at low-levels was characterised by a large-scale maximum in this region. This caused the ensemble to cluster in the region of the maximum rather than providing an even spread throughout the potentially unstable region. An ensemble generated using a perturbation field with larger amplitude (not shown) did indeed show a wider range of possibilities within the potentially unstable region and included the observed triggering location (marked by a cross in Fig. 5.24b). The modified initial states, however, were not consistent with the large scales and the ensemble was rejected.

The clustering of the ensemble away from the observed triggering location implies that errors also existed elsewhere other than at small scales. Owing to the initialisation from the coarse 1° horizontal resolution ECMWF analysis, error may have existed at larger scales. The 0000 GMT Met Office mesoscale analysis (not shown) showed much greater magnitude mesoscale structure in low-level $\theta_e$ with higher values along the band including the observed triggering location. The time of this analysis, however, was after the development of deep convection and did not allow for a clean simulation. A perturbation field that included a wider range of scales may have resulted in an ensemble that included the observed location of triggering.

**Case 2**

As shown in section 5.1, the growth of small-scale maxima within the potentially unstable region over northwest England provided the trigger for convection and further convection triggered over eastern England, on a poorly-resolved sea-breeze front. Accumulated model precipitation for the unmodified solution for 7 hours after convective initiation, shown in Fig. 5.25a, shows the north-eastward track of three independent, persistent convective cells, two over northern England and one over eastern England.

The ensemble mean accumulated precipitation, shown in Fig. 5.25b for the same time period as for Fig. 5.25a, shows the same three cell-tracks as in the unmodified solution. In addition, each ensemble member triggered convective cells in the same three locations as the unmodified solution. The three triggering locations were therefore not sensitive to the modification and showed high predictability. The magnitude of the convective cells, however, were sensitive to the local perturbation as shown by the different intensities of the three cells between the unmodified solution and the ensemble mean field. This is consistent with the high sensitivity of the area average
cumulative precipitation to the small-scale variability for this non-equilibrium case shown earlier in Fig. 5.23. In summary, the mechanisms for triggering were stronger than small-scale uncertainty in the model. Moreover, one of the locations of triggering was that observed (marked by a cross in Fig. 5.25).

Figure 5.25: Accumulated model precipitation over northwest England between 1800 GMT 11 September 2000 and 0100 GMT 12 September 2000 (Case 2) for (a) the unmodified solution and (b) the ensemble mean. The observed triggering location is marked by a cross. The same three cell tracks are present in both fields.
Chapter Summary

The ability of the model to reproduce the precise triggering location and timing of the two convective events was explored. An ensemble technique was developed to explore the range of possible triggering locations consistent with the observed large scales by sampling uncertainty due to unobserved small-scale variability.

For case 1, the downstream orography of northern France was shown to be necessary for the triggering of deep convection within the potentially unstable band at the time of the observed convection. The precise mechanism for the amplification of two local maxima within the potentially unstable region for case 2 could not be determined from the experiments performed here but the orography was shown not to be important. For case 2, further explicit convection triggered at a sea-breeze front that locally removed CIN. A sea-breeze was observed in nature but it did not trigger convection.

For case 1, the equilibrium case, the ensemble mean field showed a lack of predictability in precise triggering location and the model was able to trigger convection in different locations within the potentially unstable region. The range in triggering locations, however, did not include that observed in nature. This suggests an ensemble that only samples uncertainty at small scales may be misleading in the range of possible locations and highlights the difficulty in estimating uncertainty. Evidence was provided that the use of a perturbation field containing a wider range of scales may have resulted in a range of possibilities that included the triggering location in nature. For case 2, the non-equilibrium case, the ensemble mean field showed high predictability in all three triggering locations. One triggering location was that observed, one was two hours early but another was not observed and showed that error existed in the model other than at small scales.

The sensitivity of the cumulative precipitation and therefore the net latent heating to changes in the initial conditions using the same ensemble technique was different in the equilibrium and the non-equilibrium case. For the non-equilibrium case, the cumulative precipitation had a range between $0.78 \times 10^{-8}$ and $1.41 \times 10^{-8}$ mm at T+19 whereas for the equilibrium case, values ranged between $2.25 \times 10^{-8}$ and $2.40 \times 10^{-8}$ mm at T+19. This result is consistent with the result from chapter 4 that the cumulative precipitation was very sensitive to the partitioning of convection in the non-equilibrium case but less sensitive in the equilibrium case.
Mesoscale models have a resolution such that they are computationally viable for operational forecasting and have the capability to provide useful information on the smaller scale details of convection including location, intensity and propagation. In chapter 1, the critical issues for the representation of convection in mesoscale models were identified. The partitioning between parameterised and explicit convection was recognised as a key issue and the poor understanding of the circumstances for which a parameterisation scheme should be used was highlighted. The importance of an accurate large-scale response to convection for the downstream evolution was also identified. In addition, the initial location of convection was recognised as a key component but that its sensitivity to model physics and unobserved small-scale variability in the initial conditions often results in a poor forecast.

Recent case-study based studies were divided according to the predictability of the net properties of convection such as cumulative precipitation, heating and moistening. The common features of the large-scale environments were identified for each group. The strength of the large-scale forcing and the presence, or otherwise, of significant CIN within the region of CAPE divided the cases. This led to the hypothesis that the behaviour of parameterised and explicit convection depends on whether or not the system is in equilibrium and motivated the questions posed at the end of chapter 1. In their review of the representation of convection in mesoscale models, Molinari and Dudek (1992) suggested more studies of why cumulus parameterisation schemes succeed and fail are necessary. This work has explored the importance of equilibrium for the behaviour of
parameterised and explicit convection and for the large-scale response to convection. The main results are summarised here and discussed in the wider context of atmospheric research.

Simulations of two MCSs, presented in chapter 3, were used as the basis for numerical investigations throughout this thesis. In chapter 4, the basic setup of the numerical experiments was verified. The partitioning between parameterised and explicit convection was shown to be highly sensitive to the adjustment timescale in the equilibrium-based parameterisation scheme, thereby providing different representations of convection between the three experiments. Through an examination of the sensitivity of cumulative precipitation to the partitioning of convection, evidence was then provided that for case 1, the system was in equilibrium but for case 2, the system was not in equilibrium.

The basic setup of the experiments having thus been established, the structure of the thesis then followed the questions posed in the introduction. The questions are repeated here together with the answers and implications for future research.

**Q1.** How does the parameterisation scheme behave under equilibrium and non-equilibrium conditions?

For the equilibrium case, the parameterisation scheme produced the response expected of equilibrium behaviour but for the non-equilibrium case, the scheme behaved very differently. As discussed in chapter 1, the parameterisation scheme assumes both a temporal and spatial scale separation exists between convection and the forcing. This assumption is valid over a region in which a large number of clouds exist in a relatively uniform region of forcing and the forcing evolves slowly compared to the convection. Under such conditions, as in case 1, the parameterisation scheme produced a smooth field of precipitation that tracked the large-scale region of CAPE and almost zero CIN, the response expected from equilibrium behaviour.

For case 2, within a large-scale region of CAPE and CIN, a mesoscale region of almost zero CIN was generated by forced ascent ahead of the cold front. It is not expected that the forcing evolved slowly compared to the convection. If a lower bound for the spacing between clouds is given by the depth of the buoyantly unstable layer then it is neither expected that a large ensemble of clouds
can exist in this region of CAPE and low CIN. The application of the parameterisation scheme for this non-equilibrium case is not appropriate and the scheme failed dramatically. The instantaneous response of convection allowed for a coupling between the parameterisation scheme and the grid-scale dynamics. This set up bows of parameterised convection that propagated through the region of CAPE and CIN, outside the region of almost zero CIN.

In practical terms, this work suggests that the use of current equilibrium-based parameterisation schemes in mesoscale models is valid in equilibrium situations. An alternative formulation may be required for non-equilibrium situations. In particular, the schemes that attempt to relax the spatial and temporal scale separation assumptions by Pan and Randall (1998) and Lin and Neelin (2002), discussed in chapter 1, may be more appropriate. Alternatively, a scheme that responds directly to strong mesoscale forcing for upward motion such as the Kain and Fritsch (1993) scheme, which triggers as a function of grid-scale vertical velocity, may be more appropriate for regions of mesoscale forcing. Indeed, simulations with modified trigger functions presented in chapter 4 showed high sensitivity of the timing and location of convection to a vertical velocity dependent trigger function for case 2.

**Q2.** Can the explicit representation of convection provide information beyond that acquired using the parameterisation scheme?

For both cases, explicit convection was shown to provide useful information on the intensity and persistence of localised mesoscale convection. This representation provided more information on the small-scale details of convection than was provided by parameterised convection. Question 2 concerns the amount of useful information for applications such as flood forecasting. Despite the validity of the equilibrium-based parameterisation scheme in equilibrium situations, the field of parameterised precipitation provided no information on the small-scale details of the MCSs such as precise location, timing, size, intensity and propagation characteristics. Parameterised precipitation provided useful information only on the location of CAPE and low CIN. However, even this information was lost for the non-equilibrium case.

A solution of explicit convection avoids the assumptions of scale separation and responded in a similar manner to the potentially unstable regions in both the equilibrium and non-equilibrium
Localised mesoscale convective circulations achieved a precipitation intensity closer to that observed and provided useful information on the track and scale of the MCSs. However, the scale was determined unphysically by the horizontal resolution and diffusion. The accuracy in the location of triggering of the convective cells was often poor and this raised the issue of the predictability of convection which was addressed by question 4.

As discussed in chapter 1, the general consensus in previously published literature leans in favour of a simultaneous parameterised and explicit representation of convection in mesoscale models. Despite the useful information provided by explicit convection, this work has shown that there is still a need for a parameterisation of convection. The solution of explicit convection did not capture the propagation of convection, an essential component of organised systems. The explicit solution failed to capture the observed clustering of convective cells and, for case 1, the development of the linear convective structure. The mechanisms for such organisation are not well-resolved, if included at all at 12km horizontal resolution. It is possible that the temporal and spatial scales of explicit convection did not match those of the cold pools and could not lock in phase and propagate. There is a need for the parameterisation of these subgrid-scale processes.

**Q3. Is the large-scale response to convection sensitive to the partitioning of convection for the duration of a mesoscale forecast?**

For both cases, the large-scale response was shown to be highly sensitive to the partitioning of convection. In chapter 1, the importance of an accurate large-scale response to convection for the downstream evolution was identified. Question 3 concerns the importance of convective equilibrium for the sensitivity of the large-scale response to the partitioning of convection.

Owing to the high sensitivity of the net precipitation to the partitioning of convection for the non-equilibrium case, the response of the large-scale flow was also expected to be sensitive. For the equilibrium case, the net precipitation was only weakly sensitive to the partitioning of convection and the large-scale response was therefore also expected to be insensitive. Surprisingly, the large-scale response was found to be highly sensitive to the partitioning of convection for both the equilibrium and non-equilibrium cases. Explicit convection generated a well defined lens of tropopause-level negative PV whereas the inherently averaged parameterised convection gener-
ated a region of weaker PV co-located with the region of CAPE and low CIN. The vertical profile of the thermodynamic modification due to convection was shown to be different for parameterised and explicit convection. Moreover, owing to some fortuitous radiosonde data, the magnitude of the thermodynamic modification due to explicit convection was shown to be closer to that observed than for parameterised convection. This work has shown that the vertical profile of latent heating due to convection is as important as, if not more important than, the net value for the large-scale response.

In practical terms, this work suggests that the large-scale response is sensitive to the formulation of the cloud model in the parameterisation scheme, specifically, the entrainment and detrainment processes. This highlights a mechanism by which small errors can grow rapidly upscale via the convection scheme and contaminate the large-scale solution. As computing power increases, the resolution of global models may increase to that of current mesoscale models. For longer simulations over larger domains, the large-scale response to convection will become more important. The finding here suggests a more physically based cloud model will be necessary and, owing to our poor understanding of the mixing between clear and cloudy air, perhaps some more observational work is necessary.

Q4. Is the precise triggering location of convection predictable?

The equilibrium case showed a lack of predictability whereas the non-equilibrium case showed high predictability in all three locations. In chapter 1, it was recognised that an accurate forecast of organised convection ultimately depends on the ability of the model to capture the timing and location of its initiation. The importance of uncertainty in the initial conditions for the ability of the model to capture the precise location of convection was also identified.

The numerical simulations using an explicit representation of convection presented in chapter 4 showed generally poor accuracy in the precise location of convection. For the equilibrium case, the model triggered convection 100km from the observed location and both locations were within the potentially unstable region. For the non equilibrium case, the model triggered two explicit cells within the mesoscale potentially unstable region, one in precisely the location observed and
the other two hours early. However, the model also triggered a third cell within the region of CIN. For both cases, the triggers for convection in the model were found to be the amplification of a small-scale maximum in instability within the potentially unstable region. The spurious explicit cell in the non-equilibrium case triggered on a sea-breeze front, within the region of CIN.

An ensemble technique was designed to explore the range of possible locations consistent with the large-scale environments. In contrast to previous work using short range ensembles, an ensemble technique was developed to determine specifically the sensitivity of precise triggering location to unobserved small-scale variability. The timing and amplitude of the perturbation field were found to be crucial to ensure the modified states were consistent with the large-scale environments and to minimise bias to the location in the unmodified solution. For the equilibrium case, the model showed a lack of predictability and the range of possible locations did not include that observed. This suggests the ensemble that only sampled uncertainty at small scales was misleading in the range of possible locations and highlights the difficulty in knowing how best to design an ensemble. As discussed in chapter 5, the use of a perturbation field containing a wider range of scales may have generated a range of possibilities that included the observed location.

For the non-equilibrium case, the model showed high predictability for all three locations, one of which was that observed. One location, however, was well outside the region of almost zero CIN. Owing to the fact that the modified initial states were consistent with the large-scale environment, a trigger that is sufficient to locally remove significant CIN will, almost by definition, be shown to be predictable by this ensemble. This result suggests errors existed in the model other than in the initial conditions.

This lack of predictability in the equilibrium case and the spurious triggers in the non-equilibrium case may partly explain the Met Office mesoscale model’s failure to capture the location and timing of some deep convective events over the UK. Brooks et al. (1995) concluded that the major advantage of short-range ensemble forecasting over deterministic forecasting is that information on probability density functions and therefore forecast uncertainty can be obtained for any model parameter. Indeed, this work has shown an ensemble can provide additional useful information on the range of possibilities of the precise triggering location that may be useful for QPF and flood forecasting.
6.2 Discussion

For some general mesoscale model at 12km resolution this work has shown that the decision to include a parameterisation scheme will depend on the condition of convective equilibrium. The decision will also depend upon the users requirements. Knowledge of the region within which intense convection may occur can be acquired through an equilibrium-based parameterisation scheme for equilibrium situations only. On the other hand, information on the occurrence of intense persistent mesoscale convection and the impact of convection on the large-scales can be acquired by explicit convection but the location cannot be believed for both equilibrium and non-equilibrium situations. This work has shown that prior knowledge of convective equilibrium and therefore of the predictability of cumulative precipitation may be inferred from diagnosing the fields of CAPE and CIN. The utility of CAPE was recognised by Stensrud (2001) who suggested CAPE was able to provide useful guidance on the regions where convection was possible. This work, however, has shown that both CAPE and CIN together provide more accurate guidance, not only on the location but also on the predictability of cumulative precipitation. However, nothing can be inferred about the predictability of the small-scale details of convection, such as precise location, from knowledge of convective equilibrium.

Ensembles can highlight regions where the number of observations need to be enhanced to increase predictability. For equilibrium situations however, where there exists a large-scale region of uniform forcing, the results here suggest that subgrid-scale details may become important for the precise triggering location and increased observations may have limited impact. Mesoscale models may therefore only have predictability on the scale of the forcing and not on the distribution of convection within the potentially unstable region in equilibrium situations. The mean of an ensemble of convective clouds returned by the parameterisation scheme well describes the region within which deep convection may occur and represents the limit of predictability. The implication for flood forecasting is that the model has low predictability on the scale of the river catchment if it is smaller than the scale of the forcing in equilibrium situations. For the non-equilibrium case, the parameterisation scheme failed dramatically and explicit convection triggered within a region of CIN. The results here suggest that for non-equilibrium situations, predictability is lost on the scale of any potentially unstable region and the precise location of explicit convection should not necessarily be believed.
Despite the poor ability of explicit convection to capture the precise initial location, this representation did provide useful information on the possibility of localised intense persistent convection. It is desirable, then, to persuade the model to develop convection in the observed location. This could be achieved through some assimilation of radar or lightning data. The current approach of latent heat nudging (Jones and Macpherson, 1997) scales the model profile of latent heating by the ratio of observed to model rainrate over a period of 3 hours. The aim is to improve location rather than the intensity so it is proposed that the use of 5 minute lightning data may be of more use than radar data which often produces spurious echoes. This approach may work well in regions of uniform forcing for convection but less well for the heterogeneous forcing of non-equilibrium situations where the region of observed convection may not be a favourable region for convection in the model.

The increase in computational capability will allow for higher resolution numerical weather prediction models to be used operationally in the future. Indeed, the latest version of the Met Office mesoscale model is currently being developed at horizontal resolutions as low as 1km. Its non-hydrostatic formulation does not constrain the limit of horizontal resolution as a hydrostatic\(^1\) model would. The increase in horizontal resolution may improve simulations of organised convection largely via the removal of damage caused by inaccurate parameterisations. High resolution has, however, been shown to capture more detailed mesoscale structure and the propagation of organised convection (e.g. Mass et al., 2002; Moncrieff and Liu, 1999). Theoretically, for a large-scale region of uniform forcing, increasing horizontal resolution will not increase predictability of the precise location. For real case studies, however, there is scope for increased predictability in both equilibrium and non-equilibrium situations using increased horizontal resolution as the model resolves smaller scales. Unfortunately, the resolution of current observations is far coarser than the corresponding model resolution and the effect of increasing resolution may be to resolve smaller scale noise. More sophisticated techniques to assimilate remote sensing data offers the only real hope of increased predictability, particularly in equilibrium situations.

As discussed in chapter 1, the repeated failure to capture the timing and location of convection has led to much experimentation to include uncertainty in simulations. Techniques include mesoscale ensemble forecasting, statistical post-processing and the use of stochastic model physics. In developing the ensemble for this investigation, the timing and amplitude of the perturbation field were found to be crucial to avoid bias and generate an ensemble of states that were consistent with

\(^1\)The effect of the hydrostatic approximation on grid-scale vertical velocity is identified in appendix A
the large-scale environment. For mesoscale ensembles, this work suggests that not only may the
details of a given ensemble technique be case-dependent but traditional QPF skill scores may be
severely constrained by the condition of convective equilibrium.

6.3 Future Work

An obvious continuation of this work is to simulate the two case studies using the latest version
of the Met Office mesoscale model discussed in the previous section. The well documented case
studies could be used to explore whether or not a parameterisation should be included in the
non-hydrostatic model at 1, 2 and 4km horizontal resolution. At these resolutions, the model
will develop smaller scale convective structure and may be able to capture the propagation of
convection including the linear convective structure in case 1.

The failure of the Gregory and Rowntree (1990) scheme in the non-equilibrium case and the failure
of explicit convection to capture the propagation of the MCSs provides motivation to include an
alternative parameterisation in the version of the Met Office mesoscale model used here. The
Kain and Fritsch (1993) mass-flux scheme, based on the Fritsch and Chappell (1980) scheme,
was developed to facilitate simulations of MCSs at mesoscale resolution. The scheme is used
at mesoscale resolution for research and is currently considered for operational implementation
in many national centers, including those in Canada, the United States and France (Bélair and
Mailhot, 2001). The Gregory and Rowntree (1990) and Kain and Fritsch (1993) schemes are very
similar but do have some key differences. In chapter 1, the high sensitivity of the convective
solution to the details of the parameterisation scheme was identified. It is therefore expected that
the small-scale details of convection would be different between the two schemes.

In addition to its vertical velocity dependent trigger function, as discussed in chapter 4, the scheme
detrains cloud condensate to the grid which may allow for an evolution of convection through
delayed evaporative cooling. This may provide a more realistic coupling with the grid and prevent
the bows of parameterised convection in case 2. As discussed in chapter 1, however, recent work
has shown that simulations of mesoscale organisation are sensitive to the magnitude of moisture
detrainment rather than the phase. Further benefits may be derived from the entrainment and
detrainment profiles of the cloud model (Kain and Fritsch, 1990) that respond to the environment,
thereby relaxing the constraint of prescribed profiles (Kain and Fritsch, 1990). Furthermore, the scheme employs a similar adjustment closure to Gregory and Rowntree (1990), but allows the time taken to remove CAPE to vary as a function of the large-scale environment.

Considering the wider field of convection research, the imbalance between the modelling effort and observational studies needs to be addressed in order to improve the validation of high resolution model output. In addition to validation, further observations of processes such as the mixing between clear and cloudy air and the mechanisms for propagation will facilitate the formulation of such processes in high resolution models. Current atmospheric research is progressing in the directions of ensemble forecasting and high resolution deterministic prediction. The results from this thesis suggest the value of each may depend on the large-scale environment, particularly on the condition of equilibrium. Ultimately, the choice between ensemble or deterministic prediction depends on whether the aim is a useful or an accurate forecast. Owing to the increase in resolution of global models and the experimentation with mesoscale ensembles, the representation and predictability of convection in models of gridlength 10-30km will continue to be an important and challenging area of research.
Appendix A: The Hydrostatic Approximation

Consider the linearised two-dimensional Boussinesq equations of motion (adapted from Weisman et al., 1997):

\[
\frac{\partial}{\partial t} \left( \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x} \right) = -\frac{\partial B}{\partial x}; \quad (1)
\]

\[
\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0; \quad (2)
\]

\[
\frac{\partial B}{\partial t} = -N^2 w, \quad (3)
\]

where \( B \) is buoyancy, \( N \) is the Brunt-Väisälä frequency and we assume for simplicity there is no moisture, coriolis force, friction or mean wind. Assuming normal mode solutions:

\[
W_0 \approx \frac{k}{(k^2 + l^2)^{1/2}} N B_0, \quad (4)
\]

where \( W_0 \) is the vertical velocity and \( k = 2\pi/\lambda_x \) and \( l = 2\pi/\lambda_z \) are the horizontal and vertical wavenumbers respectively. For hydrostatic motions, the horizontal wavelength is assumed to be much larger than the vertical wavelength. This is equivalent to assuming the timescale of motion, \( \tau^2 \gg 1/N^2 \). Clearly, the hydrostatic approximation is not valid for \( N \sim 0 \) as for neutral and convectively unstable conditions. Equation 4 now becomes:

\[
W_0 \approx \frac{\lambda_z}{\lambda_x N} B_0. \quad (5)
\]

Assuming a typical scale height leads to, \( W_0 \propto 1/\lambda_x \) so as the grid-scale is reduced the amplitude of the vertical velocity increases proportionally. For non-hydrostatic motions the increase in vertical velocity is restricted by the horizontal wavenumber in the denominator of Equation 4. Fig.
1 shows the maximum vertical velocities for a range of horizontal grid spacings from full three-dimensional hydrostatic and non-hydrostatic models in the simulations of idealised squall lines by Weisman et al. (1997). The maximum vertical velocities for the non-hydrostatic simulations are comparable to the hydrostatic values for grid lengths of greater than 8km, but are significantly weaker at smaller grid lengths. Their results suggest 8km represents the grid-length below which non-hydrostatic effects become important.

![Figure 1: Maximum vertical velocities obtained for various resolution three-dimensional simulations for the full non-hydrostatic (solid) and an equivalent hydrostatic (dashed) model (Weisman et al., 1997). The dotted line represents the estimated vertical velocity for a purely hydrostatic model, assuming a $1/\Delta x$ scaling and a vertical velocity of $3ms^{-1}$ at 20km resolution.](image)

The hydrostatic approximation also affects vertical mass transport. We know that vertical mass-flux, $m \propto W_0/k$. The hydrostatic mass-flux can be expressed as:

$$m_H \approx \frac{B_0}{lN},$$

(6)

whereas the non-hydrostatic mass-flux,

$$m_{NH} \approx \frac{B_0}{(k^2 + l^2)^{1/2}N},$$

(7)

Therefore, for a given vertical scale, the hydrostatic mass-flux is independent of grid-length whereas the non-hydrostatic mass-flux increases with increasing grid-length. This is consistent
with the overprediction of convective mass-flux at coarser resolutions. The suitability of the hydrostatic approximation in models of horizontal gridlength less than 20km was investigated by Kato and Saito (1995) by examining explicit hydrostatic and non-hydrostatic simulations of idealised deep convection. The hydrostatic simulations overdeveloped system-scale circulations and overestimated precipitation area and amount.
Appendix B: Model Spin-Up

A start time such that convection initiates after the period of artificial model spin-up is essential for the numerical simulations in this thesis. For case 1, convection associated with the MCS had already triggered before the time of the 0000 GMT 29 May 1999 Met Office analysis. Further convection initiated two hours into the simulation, possibly within the period of model spin-up. The timing of convection and its subsequent evolution may have been adversely affected. Timeseries of area averaged total (explicit and parameterised) hourly rain amount for the 10 minute, 2 hour and 1 day experiments are shown in Fig. 2a. The averaging area contained the unstable region, defined as the region of CAPE, throughout the forecast. The figure shows the 0000 GMT 29 May 1999 Met Office mesoscale analysis had large rainrates in the initial conditions and did not allow a clean simulation.

Figure 2: Area averaged total (explicit and parameterised) rain amount (mm) within the region of CAPE for simulations initialised from (a) the 0000 GMT 29 May 1999 Met Office analysis and (b) the 1800 GMT 28 May 1999 ECMWF analysis (six hours earlier).
Simulations with an initialisation time of 1800 GMT 28 May 1999, six hours earlier than the 0000 GMT simulation ensured convection associated with the MCS was absent from the initial conditions. Convection then initiated well after any period of model spin-up. No Met Office analyses were available at 1800 GMT 28 May 1999 so an ECMWF analysis at 1.0° horizontal resolution and on 50 vertical levels was used. A timeseries of area averaged total hourly rain amount from this earlier simulation, shown in Fig. 2b, shows that convection was absent from the initial conditions and that the convection scheme triggered in the fourth hour of the simulation. Additional convection (not evident from Fig. 2b) initiated at 0200 GMT 29 May 1999 (T+8) after the period of model spin-up.

The 1 day experiments produced almost 100% of convective precipitation explicitly as shown in chapter 4. The simulations from both start times in Fig. 2 showed similar rates of increase of explicit rain amounts with time between 0000 and 0400 GMT. This implies the rain amounts from the 0000 GMT simulation were not affected by artificial model spin-up. The small differences in rain amount may be a result of the small differences in the large-scale dynamics between the forecasts.

For case 2, the model was initialised 12 hours prior to convective triggering, well after any period of model spin-up.
References


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