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Frame and Markowski (2010) have demonstrated the feasibility of using the tilted independent pixel approximation (TIPA) to account for the three-dimensionality of cloud shading effects in idealized numerical simulations of supercells in which radiative transfer effects were included (in addition to surface fluxes, which couple the surface temperature forcing to an atmospheric response). This thesis examines the results of simulations of a long-lived squall line in which cloud shading effects are included using the TIPA. The simulations are compared to control simulations that also include surface fluxes but do not experience cloud shading effects. In the simulations with cloud shading, particularly under the leading anvil, the cooling and stabilization of the lower boundary layer reduces vertical mixing, and gradients in buoyancy promote the generation of negative (northerly) environmental horizontal vorticity, both of which alter the low-level vertical wind shear in the inflow relative to the control simulations. It will be shown that the modifications of the hodograph and squall-line structure depend on the initial ground-relative wind profile.
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Chapter 1

Introduction

Recent research on convective storms has shown that cloud shading—that is, the attenuation of solar radiation by the storm, especially its anvil, which can extend for hundreds of kilometers downstream—can have a significant dynamical influence on the storms, particularly when the storms are long lived (Markowski and Harrington 2005; Frame and Markowski 2008; Frame 2009; Frame and Markowski 2010). This previous work has focused on supercell thunderstorms and the influence of cloud shading on the development of low-level rotation in the storm.

In this thesis, the effects of cloud shading on long-lived squall lines are investigated. Chapter 1 reviews squall-line structure and maintenance, and also describes recent attempts to include radiative transfer in numerical simulations of convective storms. Chapter 2 describes the methodology used in the simulations. Chapter 3 presents the results of the simulations. In Chapter 4, the results in Chapter 3 are related to previous theories of squall-line evolution and vorticity budgets are calculated to form a concise view of the effects of radiative transfer on squall-line evolution. Chapter 5 presents a summary and conclusion.

1.1 The characteristics and maintenance of squall lines

The squall line is one of the most common and extensively studied modes of deep moist convection. The AMS Glossary of Meteorology defines a squall line as “A
line of active thunderstorms, either continuous or with breaks, including contiguous precipitation areas resulting from the existence of thunderstorms.” Squall lines are distinguished from other types of MCSs by a larger length-to-width ratio. Squall lines can arise from frontal forcing (as observed in Newton 1950; Newton and Newton 1959; Sanders and Paine 1975; Park and Sikdar 1982; Carbone 1982), but can also evolve without the aid of a frontal surface (as observed in Ogura and Chen 1977; Ogura and Liou 1980; Smull and Houze 1985). However, the initial forcing mechanism does not necessarily dictate the subsequent evolution of the squall line. In such cases, the squall lines are driven by their own cold pools. Systems which are driven by their own pools (as opposed to by synoptic or frontal forcing) are known as type II mesoscale convective systems. The forthcoming simulations will examine type II squall-line evolution.

Attempts to explain squall-line dynamics must take into account the various forms that the squall line can have. According to a taxonomy by Parker and Johnson (2000), squall lines have been observed on different occasions to have leading stratiform, trailing stratiform, or parallel stratiform precipitation areas. The simulations and conclusions in this thesis will apply to the trailing stratiform squall line. A detailed cross section of trailing stratiform squall-line structure is given by Fig. 1.1. In general, the trailing stratiform squall line is characterized by front-to-rear flow in the mid to upper stratiform area trailing the leading convective line. The main updraft and cell regeneration occurs near the gust front and any old cells are advected to the rear of the system. The old cells deposit rainfall to the rear of the system and strengthen the system’s cold pool, and thus the boundary for lifting and cell regeneration. In systems with a strong surface cold pool, the updraft will also tilt rearward significantly. The buoyancy gradients associated with a rearward tilting updraft can induce a locally enhanced area of rear-to-front flow behind the system known as the rear-inflow jet. The rear-inflow jet can remain elevated or descend to the surface. Weisman (1992) found that the mode of the jet (elevated vs. descending) is dependent on horizontal buoyancy gradients near the back edge of the system. An elevated rear-inflow jet will lead to enhanced lifting through a deep layer and produce a more erect updraft, whereas a descending jet will lead to a more tilted updraft and sometimes damaging surface winds. The strength of the rear inflow increases with increasing CAPE and vertical wind shear.
Rotunno, Klemp, and Weisman (1988, hereafter referred to as RKW) used numerical simulations to propose an “optimal state” for squall-line maintenance, assuming that cell regeneration and realization of convective available potential energy (CAPE) would be maximized when a low-level updraft is vertically erect. In other words, the low-level updraft is not hindered by the shear or cold-pool circulations. RKW theory will be discussed here in great detail, as it will be referred to frequently in the upcoming sections. RKW used the Klemp-Wilhelmson 3D cloud model (Klemp and Wilhelmson 1978) in order to investigate the effects of low-level shear on squall-line structure and evolution. They found that the most erect updraft (the “optimal state”) in a squall line can be formed when the horizontal vorticity generated by the leading portion of the cold pool ($C$) equally balances the environmental vertical wind shear perpendicular to the line ($\Delta u$, westerly shear in this case) over a depth that tends to be a bit deeper than the mean cold-pool depth, which is typically 0–2.5 km (Fig. 1.2).

RKW found that if environmental shear is greater than the cold-pool parameter (i.e. $C/\Delta u < 1$), then the system will tilt downshear and the inflow will likely have to pass through a rain shaft, although this does not necessarily preclude a long-lived state (Rotunno et al. 1988, Parker and Johnson 2004b). However, if the environmental shear is less than the cold pool parameter ($C/\Delta u > 1$), then the system will tilt increasingly upshear, and updrafts will consequently lose vigor as they ascend along an increasingly front-to-rear slantwise trajectory. As long as
Figure 1.2. Schematic diagram outlining the basics of RKW theory [from Weisman 1992]. For systems in which $C/\Delta u < 1$, vorticity arguments necessitate a downshear tilting updraft. The ‘optimal’ state occurs when $C \sim \Delta u$, and produces a vertically erect updraft. In some suboptimal cases ($C/\Delta u > 1$, bottom panel), an elevated rear-inflow jet can be induced which also promotes deeper lifting at the gust front.

Convection can persist, an initially downshear tilted system will usually acquire increasing upshear tilt as the evaporational cooling and water loading eventually produces an increasingly stronger cold pool. The RKW model proposes a reasonable explanation for the observation of both upshear and downshear tilting systems. In Weisman (1992), the RKW model was modified to include the presence of the aforementioned rear-inflow jet, which increases rear-to-front flow in the later stages of a squall line’s life cycle. The jet forms after the cold pool strength has overcome the ambient environmental shear and caused the system to tilt upshear, thus creating the circulations needed to draw air to the gust front from the backside of the storm (see Fig. 1.2, bottom panel).

Significant opposition to RKW theory has focused on the idea that severe surface winds climatologically occur in environments with wind shears much less than predicted by RKW theory (Evans and Doswell 2001; Coniglio et al. 2004). This also calls into question the use of RKW theory in operational forecasting (Sten-
srud et al. 2005). Furthermore, RKW theory neglects the effect of deep wind shear, which has an impact on squall line longevity (Coniglio and Stensrud 2001; Coniglio et al. 2004). According to Markowski and Richardson (2010): “RKW theory probably places too much emphasis on the importance of the tilt of the gust-front updraft in attempting to explain squall-line longevity and severity”. Some of these concerns can be mitigated. For example, it should not be inferred that an optimal state is necessary to produce severe straight-line winds. In fact, a descending rear-inflow jet is favored in more sub-optimal scenarios, so the observation of derechos in less-than-optimal conditions is reasonable (Weisman et al. 2005). Despite the concerns presented above, RKW theory remains a well-known and proven conceptual model for explaining squall-line evolution. The effects of radiation on the terms used in RKW theory will be examined later in this thesis.

1.2 Potential cloud shading effects on squall lines

The impact of radiation on convective storms has been largely ignored and neglected in many simulations (e.g. Klemp and Wilhelmson 1978; Seitter and Kuo 1983; Weisman et al. 1988; Rotunno et al. 1988; Weisman 1993; Yang and Houze 1995; Fovell and Tan 1998). Studies of radiative effects on convective storms have produced mixed results in the past. In a simulation by Tao et al. (1993), the authors found that the inclusion of longwave radiation and accompanying cooling in the model did not alter the propagation speed, structure, pressure distribution, or characteristics of the squall-line systems. However, the authors did acknowledge that the addition of solar radiation and interaction with cloud microphysics could play a substantial role in storm evolution. Miller and Frank (1993) found that the addition of both shortwave and longwave radiation in the tropics influenced the destabilization of the atmosphere and subsequent location of the convective systems. Chin (1994) found that the inclusion of shortwave and longwave radiation had little impact on the multicell characteristics of squall lines, but the intensity of the rear inflow is altered by the inclusion of both radiation and ice microphysics.

More recent studies have explored the role of radiative transfer in modifying supercell behavior. Markowski et al. (1998) observed temperature gradients and associated baroclinicity between surface areas which have been shaded by optically
thick anvils and those that experienced no anvil cooling. Figures 1.3 and 1.4 illustrate the surface temperature deficits that were found underneath anvil-shaded areas in Markowski et al. (1998). The study proposed that parcels flowing through this baroclinic zone could acquire added vorticity that would modify storm rotation when tilted vertically in an updraft. In subsequent simulations, it has been shown that the inclusion of radiation and an optically thick anvil has significant effects on
surface fluxes, wind speeds, and surface temperatures (Markowski and Harrington 2005; Frame and Markowski 2008; Frame 2009; Frame and Markowski 2010).

If vertical wind shear in the pre-gust-front environment and temperatures in the cold pool are modified by the inclusion of cloud shading effects, then this could potentially impact squall-line evolution. Even in trailing stratiform squall lines, the inflow can be shaded for an appreciable distance by a downshear anvil. To the author’s knowledge, the radiative effects of anvil shading on a trailing stratiform squall line have never been taken into account. In the present study, a three-dimensional model will be used to examine the effects of anvil shading on squall-line evolution in a trailing stratiform squall line. This thesis will examine the following questions:

- What are the bulk differences in squall-line structure, strength, and evolution between simulations with cloud shading included and those without cloud shading?

- What are the horizontal vorticity modifications to anvil-shaded parcels which approach the gust front from the far field, and to what degree does cloud shading influence the cold pool strength? In other words, what is the impact of cloud shading on $C$, $\Delta u$, and $C/\Delta u$?
Chapter 2

Methods

In order to address the questions proposed in the preceding chapter, a numerical modeling study was devised to isolate the effects of radiation on convective storm simulations. Numerical modeling has been paramount in advancing the knowledge of squall lines and convective systems in recent decades. The model and experimental set-up described herein follows several of the methods of previous radiative sensitivity studies while isolating the variables that are most relevant to the present study.

2.1 Model design and initial conditions

Each model simulation was run using the Advanced Regional Prediction System (ARPS), version 5.1.5 (Xue et al. 2000, 2001). The simulation domain has a uniform grid spacing of 1 km in each horizontal direction, and a vertically stretched grid with a minimum grid spacing of 10 m at the lowest scalar grid point at 5 m above the surface, up to a spacing of 497 m at the top boundary, which is ∼18 km above the surface. The model contains 73 grid points in the vertical direction (z), 63 grid points in the north-south direction (y), and anywhere from 650–1400 grid points in the east-west direction (x) depending on the wind profile used. The total number of grid points in the x-direction is chosen so that the leading anvil does not move out of the domain. Open radiative boundary conditions are specified along the east and west sides of the domain, with periodic boundary conditions on the north and south sides of the domain. A rigid wall is employed along the top and
Table 2.1. List of parameters used in the present study

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value/Description</th>
</tr>
</thead>
<tbody>
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<td>$nx$</td>
<td>varies from 650 km to 1400 km depending on the wind profile</td>
</tr>
<tr>
<td>$ny$</td>
<td>63</td>
</tr>
<tr>
<td>$nz$</td>
<td>75</td>
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</tr>
<tr>
<td>$dy$</td>
<td>1000 m</td>
</tr>
<tr>
<td>$dz$</td>
<td>250 m (average, on a vertically stretched grid)</td>
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<tr>
<td>$dz_{min}$</td>
<td>10 m (at lowest grid point)</td>
</tr>
<tr>
<td>initialization date</td>
<td>May 20</td>
</tr>
<tr>
<td>initialization time</td>
<td>18:00 UTC (1200 LST)</td>
</tr>
<tr>
<td>lat/lon of domain</td>
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</tr>
<tr>
<td>large time step</td>
<td>3 s</td>
</tr>
<tr>
<td>small time step</td>
<td>0.5 s</td>
</tr>
<tr>
<td>radiation time step</td>
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</tr>
<tr>
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</tr>
<tr>
<td>deep soil temperature</td>
<td>285 K</td>
</tr>
<tr>
<td>skin moisture</td>
<td>0.3 m³/m³</td>
</tr>
<tr>
<td>deep soil moisture</td>
<td>0.3 m³/m³</td>
</tr>
<tr>
<td>E/W boundary cond.</td>
<td>open</td>
</tr>
<tr>
<td>N/S boundary cond.</td>
<td>periodic</td>
</tr>
<tr>
<td>top/bottom boundary cond.</td>
<td>rigid wall w/ 5-km deep sponge layer under the top boundary</td>
</tr>
<tr>
<td>momentum advection</td>
<td>4th order horizontal and 2nd order</td>
</tr>
<tr>
<td>turbulent mixing</td>
<td>vertical advection</td>
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<td>vertical mixing length scale</td>
<td>1.5-order TKE based scheme</td>
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<td>Sun and Chang (1986)</td>
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<td>horizontal computational mixing coeff.</td>
<td>4th order scaled by horizontal and vertical grid spacing</td>
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<td>vertical computational mixing coeff.</td>
<td>0.001 s⁻¹</td>
</tr>
<tr>
<td>microphysics</td>
<td>0.001 s⁻¹</td>
</tr>
<tr>
<td>surface physics</td>
<td>Schultz (1995) NEM ice microphysics</td>
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<td>soil type</td>
<td>fluxes calculated from stability dependent surface</td>
</tr>
<tr>
<td>vegetation type</td>
<td>drag coefficients and predicted temperature</td>
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<tr>
<td>vegetation fraction</td>
<td>and surface volumetric water content</td>
</tr>
<tr>
<td>roughness length</td>
<td>clay</td>
</tr>
<tr>
<td></td>
<td>grassland</td>
</tr>
<tr>
<td></td>
<td>0.3</td>
</tr>
<tr>
<td></td>
<td>15 cm</td>
</tr>
</tbody>
</table>
bottom boundaries with a Rayleigh sponge layer of 5 km below the top boundary. A large time step of 3 s is utilized, along with a small time step of 0.5 s to damp acoustic waves.

The model uses the microphysics parameterization of Schultz (1995) and includes five categories of condensate: cloud liquid, cloud ice, rain, snow, and precipitating ice (graupel, sleet, and hail). Eddy viscosities are determined by the prognosed turbulent kinetic energy and a mixing length scale, with the latter depending on the boundary layer depth according to the method of Sun and Chang (1986). The current boundary layer depth is diagnosed by the model, using the methods of Nieuwstadt and Tennekes (1981) for a stable boundary layer, and Gyrning and Batchvarova (1990) for an unstable boundary layer. Surface fluxes are computed using bulk aerodynamic formulae, with stability dependent surface drag coefficients and predicted surface volumetric water content and temperature. The soil model is a two-layer force restore model adapted from Noilhan and Planton (1989). The surface albedo is dependent on the solar zenith angle and soil type characteristics. Despite a lengthy run time in the simulations, the Coriolis force is not included in these simulations in order to keep the study as simple as possible. The squall lines simulated in this study are highly two-dimensional compared to the more three-dimensional squall lines which display large Coriolis accelerations. A comprehensive list of parameters and coefficients used is given in Table 2.1.

\section*{2.2 Radiative transfer model}

The forthcoming simulations utilize the NASA Goddard Radiative Transfer Parameterization (Chou 1990, 1992; Tao et al. 1996; Chou et al. 1998). This parameterization includes absorption and scattering of radiation due to ozone, carbon dioxide, water vapor, oxygen, clouds, and aerosols. The electromagnetic spectrum is divided into the UV and visible region and a longwave region. The absorption and scattering properties of the aforementioned atmospheric constituents are taken into account within several spectral bands in these shortwave and longwave regions. To the author’s knowledge, most numerical models allow for only vertical propagation of shortwave and longwave radiation. In order to capture the effects of shortwave radiation along a slantwise path based on a varying zenith angle,
a tilted independent pixel approximation was employed. The tilted independent pixel approximation (TIPA), described by Varnai and Davies (1999), uses a simple coordinate transformation shown in Fig. 2.1, in which relevant model data is interpolated to a new coordinate system based on the solar zenith angle. The radiation parameterization can then be run in such a way that surface locations not directly beneath a cloud will still experience accurate radiative attenuation from a slanted solar beam. Frame et al. (2009) showed that this approximation increased accuracy of shortwave radiation estimates in areas around the periphery of the cloud without significant computational cost. Original grid data are stored prior to the coordinate transformation, so only a horizontal interpolation of solar fluxes is required back to the original grid in order to calculate longwave fluxes. Longwave radiation is then calculated within vertical columns as has been done in previous models.

2.3 Experimental design

The base state in the model domain is horizontally homogeneous, and convection is initiated using a cold temperature perturbation (following the methods of Parker
and Johnson 2004a and Weisman et al. 1997). The perturbation is introduced everywhere west of the line at $x = 30$ km in the domain, and it has a surface potential temperature perturbation of $-12$ K, which increases linearly to 0 K at 2.5 km. Random 0.1 K temperature perturbations are added to the easternmost 10 km of the cold temperature perturbation in order to force three-dimensionality in the simulations. It was found that linear warm thermals were not able to initiate thunderstorms given a strong temperature inversion above the boundary layer. Weakening this ‘capping inversion’ would lead to spurious inflow convection and significant mixing of moisture in the boundary layer after only a few hours of heating, both of which would undesirably influence the inflow region of the squall line environment. Thus, the cold temperature perturbation was chosen to emulate a weakening cold front which is able to initiate thunderstorms in a linear manner despite a strong capping inversion. All simulations are initialized at 1800 UTC (1200 LST) on May 20 and are run to 7 hours. A start time of 1200 LST provides a developing squall line the opportunity to move into a boundary layer which has been heated for several hours by the mid-afternoon.

In order to isolate the effects of radiative attenuation in squall-line environments, two different radiative configurations are run in a variety of ground-relative wind regimes. First, a control simulation is specified in which incoming radiation is not attenuated by any cloud or hydrometeors (i.e. cloud ice, cloud water, rain water, hail, and graupel are each set to zero in the radiation code). These will be referred to as the ‘radiatively transparent’ (TRANS) simulations. Such a radiatively transparent simulation is necessary as a control so that the simulations with radiative attenuation by hydrometeors can be compared to a simulation with a similar clear-sky boundary layer evolution. The simulations with radiative attenuation by hydrometeors will be called the ‘radiative shading’ (SHADE) simulations, and is run as the test simulation, utilizing the TIPA described above. A schematic of the two radiation schemes is given by Fig. 2.2.

Given the potential impacts of anvil-induced cooling on the overall boundary layer shear, a number of different ground-relative wind profiles are tested in order to gauge the full spectrum of radiative impacts in different storm environments. Generally, magnitudes of lower and upper level shear are the same in each simulation, with the main differences being the surface wind speeds. In other words, the
Figure 2.2. Diagram of the two radiative attenuation schemes utilized in the model simulations. The top panel shows the attenuation scheme for the SHADE simulations, and the bottom panel shows the TRANS simulations.

origin of the hodograph is varied between the simulations. Three simulations are performed with $-15 \text{ m s}^{-1}$ (referred to as the $-1510$ wind profile), $0 \text{ m s}^{-1}$ (020 wind profile), and $15 \text{ m s}^{-1}$ (1535 wind profile) initial surface wind speed. These simulations were given $25 \text{ m s}^{-1}$, $20 \text{ m s}^{-1}$, and $20 \text{ m s}^{-1}$ of 0–2.5 km westerly wind shear respectively. This shear is modified throughout the simulation by surface fluxes. The $-15 \text{ m s}^{-1}$ initial wind profile is given more shear so that shear values could be relatively consistent over time with the other two simulations after surface drag modifies the initial profiles. Early in the simulations with added upper level shear, some of the squall lines display leading stratiform characteristics, but as surface drag begins to modify the wind profile near the surface, these systems acquire system-relative front-to-rear momentum and take on a trailing stratiform structure. A time series of the wind profile evolution for these simulations at the far eastern boundary is given by Fig. 2.3. In the simulations with $-15 \text{ m s}^{-1}$ and $15 \text{ m s}^{-1}$ surface winds, the surface drag modifies the initial profiles by about $10 \text{ m s}^{-1}$ towards the origin throughout the simulation. A second set of environmental profiles is also utilized in which the 0–2.5-km shear remains the same, but extra shear is added in the 7.5–10-km layer in order to increase the length of the anvil.
Figure 2.3. The evolution of the meridionally–averaged wind profiles from 0–5 km at the far eastern boundary of the domain for the three different initial surface wind speeds used. The 1535 panel (top right) represents wind profiles with 15 m s$^{-1}$ surface winds increasing to 35 m s$^{-1}$ at 2.5 km. The 020 panel (bottom left) represents wind profiles with 0 m s$^{-1}$ surface winds increasing to 20 m s$^{-1}$ at 2.5 km. The –1510 panel (bottom right) represents wind profiles with $-15$ m s$^{-1}$ surface winds increasing to 10 m s$^{-1}$ at 2.5 km.

In these ‘long anvil’ simulations, the boundary layer evolution of the wind profiles is nearly the same as those given by Fig. 2.3. Figure 2.4 gives an illustration of the six total wind profiles examined, and Table 2.2 gives a list of the simulations as they will be referenced from this point forward.

The initial sounding used, and a time series of sounding evolution at the far eastern boundary of the domain is given by Fig. 2.5. Care was taken to adjust the initial CAPE and CIN of the sounding so that convection could be initiated and sustained along the gust front for at least 5 hours, and random spurious convection could not develop ahead of the main line of storms. The sounding at the eastern boundary (as shown in Fig. 2.5) initially has 2428 J kg$^{-1}$ of surface-based CAPE and 32 J kg$^{-1}$ of surface-based CIN. Over time, CIN is eroded quickly and
Figure 2.4. Cross section of zonal wind profiles with height used in the squall line simulations. Solid lines represent wind profiles with increased upper level shear and dashed lines represent profiles without upper level shear, but with identical shear values in the boundary layer.

Figure 2.5. A time-series of the evolution of the sounding at the far eastern boundary of the domain (area not shaded by anvil) for the $-1510$ wind profile. Values of surface-based parcel CAPE and CIN are given above each sounding.
Table 2.2. Naming convention to be used for all simulations

<table>
<thead>
<tr>
<th>Name</th>
<th>Shear profile</th>
<th>Parameterization used</th>
</tr>
</thead>
<tbody>
<tr>
<td>TRANS−1510</td>
<td>−15 m s$^{-1}$ at 0 km to 10 m s$^{-1}$ at 2.5 km</td>
<td>Radiatively transparent</td>
</tr>
<tr>
<td>TRANS−1510_1030</td>
<td>same surface shear with 10 m s$^{-1}$ at 7.5 km increasing to 30 m s$^{-1}$ at 10 km</td>
<td>Radiatively transparent</td>
</tr>
<tr>
<td>TRANS020</td>
<td>0 m s$^{-1}$ at 0 km to 20 m s$^{-1}$ at 2.5 km</td>
<td>Radiatively transparent</td>
</tr>
<tr>
<td>TRANS020_2040</td>
<td>same surface shear w/ 20 m s$^{-1}$ at 7.5 km increasing to 40 m s$^{-1}$ at 10 km</td>
<td>Radiatively transparent</td>
</tr>
<tr>
<td>TRANS1535</td>
<td>15 m s$^{-1}$ at 0 km to 35 m s$^{-1}$ at 2.5 km</td>
<td>Radiatively transparent</td>
</tr>
<tr>
<td>TRANS1535_3555</td>
<td>same surface shear with 35 m s$^{-1}$ at 7.5 km increasing to 55 m s$^{-1}$ at 10 km</td>
<td>Radiatively transparent</td>
</tr>
<tr>
<td>SHADE−1510</td>
<td>−15 m s$^{-1}$ at 0 km to 10 m s$^{-1}$ at 2.5 km</td>
<td>Radiatively shaded</td>
</tr>
<tr>
<td>SHADE−1510_1030</td>
<td>same surface shear with 10 m s$^{-1}$ at 7.5 km increasing to 30 m s$^{-1}$ at 10 km</td>
<td>Radiatively shaded</td>
</tr>
<tr>
<td>SHADE020</td>
<td>0 m s$^{-1}$ at 0 km to 20 m s$^{-1}$ at 2.5 km</td>
<td>Radiatively shaded</td>
</tr>
<tr>
<td>SHADE020_2040</td>
<td>same surface shear w/ 20 m s$^{-1}$ at 7.5 km increasing to 40 m s$^{-1}$ at 10 km</td>
<td>Radiatively shaded</td>
</tr>
<tr>
<td>SHADE1535</td>
<td>15 m s$^{-1}$ at 0 km to 35 m s$^{-1}$ at 2.5 km</td>
<td>Radiatively shaded</td>
</tr>
<tr>
<td>SHADE1535_3555</td>
<td>same surface shear with 35 m s$^{-1}$ at 7.5 km increasing to 55 m s$^{-1}$ at 10 km</td>
<td>Radiatively shaded</td>
</tr>
</tbody>
</table>

CAPE increases above 3300 J kg$^{-1}$. Nearer to the gust front, subsidence warms the mid-levels and increases CIN while decreasing CAPE in the near-gust-front environment. The example shown by Fig. 2.5 occurs in the −1510 wind profile, although the other wind profiles show a similar sounding progression at the eastern boundary.

2.4 Matching the model to observations

In a study by Frame and Markowski (2010), the simulated temperature deficits produced under an anvil-shaded area were a bit less than the 5-K deficit that was observationally detected under an anvil shadow (Markowski et al. 1998). The authors did not explore all of the possible reasons for the underestimation of surface cooling. The ability to derive an accurate sounding underneath an anvil-shaded area depends on a number of parameterizations. For example, an accurate
profile depends not only on accurate microphysics and radiative parameterizations, but also on an accurate soil model. In the present simulations, output soundings showed sensitivity not only to the type of soil and vegetation specified, but also to deep soil moisture and temperature, which influences heat fluxes. In order to obtain the most realistic temperature profile in the surface layer, the deep layer of the soil model is given a dimensionless moisture value of 0.3 \((\text{m}^3/\text{m}^3)\) and a temperature of 285 K. This increased the latent heat flux and decreased the sensible heat flux toward more realistic values (based on ARM observations of surface fluxes in southwest Kansas on a severe outbreak day). According to Ren and Xue (2004), the temperature of the deep soil layer has been a topic of wide debate, so the author felt justified in adjusting these values to satisfy actual observations of fluxes. It was also found that the desired sounding evolution was produced with a clay soil type, covered sparsely by grassland with a roughness length of 15 cm.

In the Markowski et al. (1998) study, shortwave radiation, as well as temperature and dew point temperature, was observed at a few select locations under the influence of anvil shading. Figure 2.6 shows how actual shortwave radiation is attenuated compared to a location in the model runs which comes under the influence of the anvil at roughly the same time. The shortwave radiation in the model shows only minimal changes after anvil shading begins. In fact, at the onset of anvil shading, the shading simulation in the model (NO MODIFICATION, the red line in Fig. 2.6) shows no differences in shortwave radiation from the non-shaded simulations. The effect of radiation attenuation in the original shading scheme does not show up until \(\sim 2300\) UTC, nearly an hour after the onset of anvil shading. In Markowski et al. (1998), anvil shading induces a much greater and more rapid decrease in shortwave radiation. The underestimate of shortwave radiation attenuation in the model simulations could be the result of shortcomings in either the microphysics or radiative parameterizations. In order to nudge shortwave radiation (and surface potential temperature) values closer to those observed under anvils, the ice values in the anvil are artificially increased for the SHADE simulations. Every ice mixing ratio value between \(1 \times 10^{-7}\) kg kg\(^{-1}\) and \(2 \times 10^{-4}\) kg kg\(^{-1}\) is increased to \(2 \times 10^{-4}\) kg kg\(^{-1}\) in order to increase the optical thickness of the anvil. Studies of thunderstorm anvil ice water content by Lawson et al. (1998)
Figure 2.6. (Top) Model estimated shortwave radiation at a certain point in the −1510_1030 domain. FAR FIELD represents the progression of shortwave radiation far from the anvil under the clear sky. NO MODIFICATION represents simulations of anvil shading without artificially increased ice values. ICE MODIFICATION represents the simulations with artificially increased ice values. (Bottom) Actual shortwave radiation estimates [taken from Markowski et al. 1998]. The dash-dot line represents shortwave radiation. Solid and dotted lines represent temperature and dew point respectively.

and Heymsfield and Palmer (1985) indicate typical thunderstorm anvil-ice water content ranges between 0–2 g m\(^{-3}\), with a majority of anvil-ice concentrations near the bottom of that range, and ice concentrations decreasing significantly further away from the main updraft. Taking a typical observed ice concentration of 0.1 g m\(^{-3}\) and using the density of moist air for the sounding given, the resultant mixing ratio is 2.2 \(\times 10^{-4}\) kg kg\(^{-1}\). Therefore, our modifications, though possibly seen as drastic by some modelers, are actually conservative estimates of anvil-ice content which more accurately match observations. The changes produce a more
Figure 2.7. Cross section of potential temperature and height in the $-1510 \text{ to } 1030$ wind profile environment. Profiles are plotted in the clear sky area with no anvil shading (light line), near the gust front under the anvil with no ice concentration modifications (darker line), and near the gust front under the anvil with modified ice concentrations (darkest line).

Figure 2.8. Cross section of potential temperature and height observed both in clear sky and after 90 minutes of anvil shading [data based on soundings from Bryan and Parker 2010].

accurate shortwave radiation decrease (shown as the black line in Fig. 2.6). These changes also produce a much more realistic potential temperature profile (Fig. 2.7), as compared with an actual profile observed under an anvil by Bryan and Parker (2010), shown in Fig. 2.8. Each simulation referred to as a SHADE simulation in the upcoming results includes the aforementioned anvil-ice modifications.
Chapter 3

Results

Utilizing the ‘TRANS’ and ‘SHADE’ simulations, differences in squall-line strength are examined between the different radiative attenuation schemes. The forthcoming figures highlight these differences by showing meridionally-averaged cross sections of some common metrics for squall line strength: system-relative wind speed, updraft strength, perturbation potential temperature (cold pool strength), and a total cross-sectional view of cloud water and cloud ice. These figures show the structure of the line at $t = 4$ h, a time at which radiation has already impacted, and is still impacting, a well-developed squall line. In additional figures, a horizontal cross section is given to show the structure of the line, along with storm-relative wind vectors at 1 km and average 0–2.5-km updraft strength. Time series of the volumes of isosurfaces of zonal wind speed $> 20$ m s$^{-1}$ (or 30 m s$^{-1}$ depending on the wind profile used) and updraft strength $> 10$ m s$^{-1}$ are also given to assess overall system strength.

3.1 Characteristics of squall lines simulated with and without cloud shading effects

The simulations examine differences between radiative schemes in the six wind profiles that were described in the previous section. Wind profiles with initially identical surface winds will be grouped together, as the discussion of the systems with respect to the initial wind profile will be important in the upcoming chapter.
### 3.1.1 The \(-1510\) and \(-1510\_1030\) wind profiles

The differences between the TRANS\(-1510\) and SHADE\(-1510\) simulations are best illustrated by Fig. 3.1 and Fig. 3.3. It is important to note that each cross section is plotted over a different depth, a depth which is the most relevant for the given cross section. The cross sections are centered at the zonal gust front position. Examining the cross sections, one will notice a few differences between the TRANS\(-1510\) and SHADE\(-1510\) fields. The SHADE\(-1510\) simulation has a larger area of strong westerly boundary layer winds behind the gust front (Fig. 3.1a,b). These winds are co-located with a deeper and stronger cold pool in the SHADE\(-1510\) simulation (Fig. 3.1c,d). Figure 3.1e,f shows the updraft structures and strength to be relatively similar in each case, although there is a stronger downdraft present in the SHADE\(-1510\) simulation. Figures 3.1 g,h are included to show the length and position of the leading anvil, with the box indicating the zoom area for the upper-panel plots. The horizontal cross-sectional view of the line is given in the top panels of Figure 3.3. This cross section shows 0–2.5-km average updraft strength, 1-km system-relative wind vectors, and the position of the gust front (based on the surface 1-K temperature perturbation). The height of 1 km was chosen for the system-relative wind vectors in order to examine a height at which rear-to-front flow in the boundary layer is well represented. One kilometer is also a good height for representing the average inflow speed in the boundary layer. In Fig. 3.3, each simulation shows some variation in the \(y\)-direction, with some 1-km rear-to-front flow. The amount of lifting at the gust front is similar between the TRANS\(-1510\) and SHADE\(-1510\) simulations. Certain thresholds of zonal wind and updraft speed are also chosen as measures for system strength. The total volume enclosed by these threshold isosurface values are given in a time series by the bottom panels of Fig. 3.3. One relevant measure is the total volume of zonal wind exceeding 20 m s\(^{-1}\) in the bottom 0–2.5 km of the atmosphere from 0–50 km behind the gust front, which is given to show differences in rear-to-front flow behind the gust front. The total domain volume of vertical velocity exceeding 10 m s\(^{-1}\) is also given to assess uplift. In these simulations, there are few differences in vertical velocity. However, the zonal wind speed plots confirm the observation that the SHADE\(-1510\) simulation has a larger area of strong rear-to-front 0–2.5 km winds throughout the simulation. This could be an indication of a developing rear-inflow...
Figure 3.1. Meridionally-averaged vertical cross sections of (a,b) system-relative zonal wind speed in m s$^{-1}$, (c,d) perturbation potential temperature in K, (e,f) vertical velocity in m s$^{-1}$, and (g,h) cloud water/ice mixing ratio centered at the gust front in the $-1510$ wind profiles. The box in (g,h) indicates the horizontal zoom areas for the other plots. All plots occur at $t = 4$ h into the simulation.
Figure 3.2. Same comparison fields as given in Fig. 3.1 but for the −1510_1030 wind profile.
Figure 3.3. (top right) Horizontal cross section of averaged 0–2.5 km updraft strength in m s$^{-1}$ (contoured), 1 km system-relative wind vectors, and surface gust front position (solid black line) for the TRANS−1510 simulation (top left) and SHADE−1510 simulation (top right) at t = 4 h. (bottom left) Time series of total volume of zonal wind exceeding 20 m s$^{-1}$ in the bottom 0–2.5 km of the atmosphere from 0–50 km west of the gust front. (bottom right) Time series of total volume of vertical velocity exceeding 10 m s$^{-1}$ in the entire domain.

jet, which would be favored in a less optimal system.

The −1510.1030 profiles show similar results as the previous simulations. These results are illustrated in Fig. 3.2 and Fig. 3.4. The SHADE−1510.1030 simulation similarly shows a larger area of strong surface westerlies than in the TRANS−1510.1030 simulation. The SHADE−1510.1030 simulation also contains a stronger cold pool. In the SHADE−1510.1030 simulation, the main updraft has been severed from the lifting at the edge of the gust front. In the TRANS−1510.1030 simulation,
the main updraft area is more erect, and the lifting is connected to gust front lifting, although both updrafts are of comparable strength. Horizontal cross sections (Fig. 3.4) show more overall lifting along the gust front in the TRANS–1510_1030 simulation compared to the SHADE–1510_1030 simulation. The isosurface analyses in Fig. 3.4 show that the TRANS–1510_1030 simulation does have more total lifting and the SHADE–1510_1030 case has a greater overall area of strong boundary layer westerlies behind the gust front throughout the simulations. In both the −1510 and −1510_1030 simulations, the SHADE simulations have stronger westerly boundary layer winds behind the gust front and a stronger cold pool. In addition, the TRANS–1510_1030 simulation shows less total lifting along the gust.
front compared to the SHADE−1510_1030 simulation, which would indicate a less-optimal, but not necessarily weaker system. The updraft tilt appears similar, yet a stronger updraft is present near low levels in the TRANS−1510_1030 simulation and in upper levels in the SHADE−1510_1030 simulation.

### 3.1.2 The 020 and 020_2040 wind profiles

The wind profiles with initially calm surface winds show few differences between the SHADE and TRANS simulations. Figure 3.5 shows the cross sections of several model fields for the 020 wind profiles. There appears to be very little difference in the strength of zonal boundary layer westerly winds behind the gust front between the two simulations. Figures 3.5c,d show relatively similar cold-pool depths, although the SHADE020 simulation shows a bit stronger cold pool. The updraft strength is slightly larger near the gust front in the TRANS020 simulation, but the updraft is not necessarily less titled than in the SHADE020 simulation. Horizontal cross sections in Fig. 3.7 show 1-km rear-to-front for a brief distance behind the gust front, with convergence and widespread lifting along the gust front in both simulations. Both simulations exhibit signs of bowing in certain areas along the gust front position. The isosurface analyses of Fig. 3.7 show a slightly larger area of total lifting throughout the simulation in the TRANS020 case and also a slightly larger area of strong zonal boundary layer westerly winds behind the gust front. For the profiles with initially calm winds (020 and 020_2040), and initially westerly winds (1535 and 1535_3555), the isosurface threshold in evaluating the rear-to-front flow is increased from 20 m s$^{-1}$ to 30 m s$^{-1}$ to be consistent with an increase in overall system speed.

Figures 3.6 and 3.8 give a similar view of the squall lines with a longer leading anvil. In these simulations, both the zonal wind and updraft field once again appear very similar between simulations. The SHADE020_2040 simulation produces a stronger cold pool. The indication of a bit stronger updraft in SHADE020_2040 simulation in the lower levels near this frontal surface is given by Fig. 3.6f compared to Fig. 3.6e. Examining Fig. 3.8, one will notice a tendency for larger volumes of both total vertical velocity and zonal wind speeds in the TRANS020_2040 simulations. It is difficult to judge which simulation is more sub-optimal from these sets
Figure 3.5. Same comparison fields as given in Fig. 3.1 but for the 020 wind profile.
Figure 3.6. Same comparison fields as given in Fig. 3.1 but for the 020_2040 wind profile
of simulations. Zonal wind fields and updraft strengths are similar. The TRANS simulations show greater rear-to-front flow behind the gust front, but they also show more total domain lifting. Lifting near the gust front is slightly greater in the TRANS020 simulation (compared to the SHADE simulation), and in the SHADE020,2040 simulation (compared to the TRANS simulation). All in all, the simulations are very similar, and significant conclusions can not be made about differences between the simulations.

Figure 3.7. Same comparison fields as in Fig. 3.3 but for the 020 wind profile.
Figure 3.8. Same comparison fields as given in Fig. 3.3 but for the −020_2040 wind profile.

3.1.3 The 1535 and 1535_3555 wind profiles

Figure 3.9 provides a look at the cross-sectional differences between the SHADE and TRANS simulations for the 1535 wind profiles. In this case, the TRANS1535 simulation shows a slightly larger area of strong westerlies near the surface. The cold pool appears stronger in the SHADE1535 simulation, although the depth of the cold pool is similar between the two simulations. The total updraft is more tilted in the SHADE1535 simulation, and there is slightly stronger lifting near the gust front in the TRANS1535 simulation. Isosurface volume analyses (Fig. 3.11) show that the TRANS1535 simulation has a larger area of strong boundary layer westerlies behind the gust front, especially later in the simulation. The volume
Figure 3.9. Same comparison fields as given in Fig. 3.1 but for the 1535 wind profile.
Figure 3.10. Same comparison fields as given in Fig. 3.1 but for the 1535_3555 wind profile.
analysis also shows slightly more lifting in the TRANS1535 simulation.

The cross section and volume analyses for simulations with the longer anvil are given by Fig. 3.10 and Fig. 3.12. The TRANS1535_3555 simulations once again show a larger area of strong westerlies behind the gust front. The cold pool in each simulation is very similar for this wind profile. Also, the SHADE1535_3555 simulation indicates a stronger and more erect total updraft than the TRANS1535_3555 simulation. Figure 3.12 shows that the TRANS1535_3555 simulations have stronger boundary layer westerlies behind the gust front and also a bit larger volume of stronger vertical velocities. In the 1535 and 1535_3555 simulations, the TRANS simulations consistently show larger areas of zonal boundary layer westerly winds behind the gust front and more total uplift. The amount of uplift near the gust
front and the overall tilt of the updraft varies between simulations. The cold pool strength appears slightly larger in the SHADE simulations, but is of comparable strength in the TRANS simulations.

3.1.4 Summary of the differences

In order to assess the results of the previous simulations, one must define a criterion for system strength. The preceding vertical cross sections did show some differences in system strength between the SHADE and TRANS simulations at $t = 4$ h. Table 3.1 and Table 3.2 give the total rainfall and zonal surface wind maximum values of the various simulations.
Table 3.1. Total rainfall observed in each simulation

<table>
<thead>
<tr>
<th>Wind profile</th>
<th>Radiative scheme</th>
<th>TRANS</th>
<th>SHADE</th>
</tr>
</thead>
<tbody>
<tr>
<td>−1510</td>
<td>607,724 mm</td>
<td>625,843 mm</td>
<td></td>
</tr>
<tr>
<td>−1510_1030</td>
<td>577,634 mm</td>
<td>574,161 mm</td>
<td></td>
</tr>
<tr>
<td>020</td>
<td>605,228 mm</td>
<td>589,303 mm</td>
<td></td>
</tr>
<tr>
<td>020_2040</td>
<td>643,232 mm</td>
<td>597,907 mm</td>
<td></td>
</tr>
<tr>
<td>1535</td>
<td>744,410 mm</td>
<td>652,326 mm</td>
<td></td>
</tr>
<tr>
<td>1535_3555</td>
<td>835,863 mm</td>
<td>775,835 mm</td>
<td></td>
</tr>
</tbody>
</table>

Table 3.2. Maximum surface zonal wind speed observed in each simulation

<table>
<thead>
<tr>
<th>Wind profile</th>
<th>Radiative scheme</th>
<th>TRANS</th>
<th>SHADE</th>
</tr>
</thead>
<tbody>
<tr>
<td>−1510</td>
<td>20.3 m s(^{-1})</td>
<td>21.5 m s(^{-1})</td>
<td></td>
</tr>
<tr>
<td>−1510_1030</td>
<td>23.1 m s(^{-1})</td>
<td>19.1 m s(^{-1})</td>
<td></td>
</tr>
<tr>
<td>020</td>
<td>20.5 m s(^{-1})</td>
<td>18.9 m s(^{-1})</td>
<td></td>
</tr>
<tr>
<td>020_2040</td>
<td>22.7 m s(^{-1})</td>
<td>19.8 m s(^{-1})</td>
<td></td>
</tr>
<tr>
<td>1535</td>
<td>29.9 m s(^{-1})</td>
<td>24.9 m s(^{-1})</td>
<td></td>
</tr>
<tr>
<td>1535_3555</td>
<td>28.7 m s(^{-1})</td>
<td>30.7 m s(^{-1})</td>
<td></td>
</tr>
</tbody>
</table>

It was shown in volume analyses that the the TRANS simulations nearly always produce a larger area of vertical velocities, even in simulations with a more tilted and weaker updraft at t = 4 h. For this reason, the TRANS simulations produce more rainfall in every simulation except for those with the −1510 wind profile. The SHADE simulations nearly always produce a stronger cold pool at t = 4 h. The SHADE simulations show a larger total area of boundary layer rear-to-front flow behind the gust front in the −1510 and −1510_1030 simulations, while the TRANS simulation show larger areas of rear-to-front flow in the 020, 020_2040, 1535, and 1535_3555 simulations. However, the preceding figures indicate that maximum surface winds are not highly sensitive to the cloud shading. Furthermore, there are very few differences between the horizontal structure of the line and updraft strength/tilt between the simulations with and without cloud shading at t = 4 h. However, it should be stressed that the simulation fields were only examined at one
snapshot in time. In the upcoming chapter, a time-evolution of the post-gust-front cold-pool strength and pre-gust-front wind shear of each system will be assessed.
Chapter 4

Discussion

In Chapter 3, total system strength was evaluated in terms of total rainfall, updraft strength and tilt, cold pool strength, and boundary layer wind speed. These figures largely examined the properties of the line at one time ($t = 4$ h). In this chapter, we explore the predictions of RKW system optimality between the simulations for each wind profile. It is the aim of the current section to examine a time-series of $C$, $\Delta u$, and $C/\Delta u$ individually in order to further analyze the possible effects of cloud shading on squall lines.

4.1 Analysis of the dynamical differences between the squall lines simulated with and without cloud shading effects

The current section will provide 8-panel plots for each of the six simulations analyzed in the previous section. The plots in each panel are area-averaged values. The top left panel (a) will show the time evolution of cold-pool strength ($C$), averaged 10–20 km behind the gust front. This distance represents the typical location at which the air behind the gust front is stagnant relative to the leading edge of the gust front. The strength of the cold pool is proportional to the horizontal vorticity generation rate within the cold pool and is given by
where \( H \) represents a depth a bit larger than the cold pool (fixed at 2.5 km for these calculations) and \( B \) is the buoyancy

\[
B = g \left( \frac{\theta'_{\rho}}{\theta_{\rho, \infty}} \right).
\]

In the buoyancy calculation, \( \theta'_{\rho} \) is taken to be the difference between the density potential temperature \( (\theta_{\rho}) \) at a given point compared to the average density potential temperature at the far eastern boundary of the domain \( (\bar{\theta}_{\rho, \infty}) \). Similarly, the top right panel (b) shows the time evolution of \( \Delta u \), which is meridionally and zonally averaged from 10–20 km in front of the gust front. The two individual terms are then combined in (c) to show \( C/\Delta u \). The bottom panels of the figures (e-h) are used to plot both meridionally averaged \( u \) and \( \theta \) for each simulation at various distances from the gust front position at \( t = 4 \) h.

Figures 4.1 and Fig. 4.2 illustrate the differences between the SHADE−1510 and TRANS−1510 simulations from the standpoint of RKW theory. One will notice that the TRANS−1510 simulation is a more optimal simulation (i.e. \( C/\Delta u \) is closer to 1). The SHADE simulation has a slightly stronger cold pool (Fig. 4.1a), and Figs. 4.1b,c show that the pre-line vertical westerly shear (\( \Delta u \)) is greater in the transparent case, leading to a smaller \( C/\Delta u \) term. The SHADE−1510 simulation has experienced a 1-K temperature drop from the far field (Fig. 4.1g).

Figure 4.2 shows similar and more exaggerated results. In this case, the cold pool is much stronger in the SHADE−1510_1030 case. Once again, the pre-line westerly shear is greater in the TRANS−1510_1030 case. A combination of these factors leads to a more optimal system in the TRANS−1510_1030 simulations, with larger differences between the SHADE and TRANS simulations than in the short-anvil simulations. Positions near the gust front experience nearly 2.5 K of cooling from the far field. This cooling impacts the wind profile near the gust front in the SHADE−1510_1030 simulation compared to the TRANS−1510_1030 simulation. The SHADE−1510_1030 wind profile near the gust front has much greater surface layer easterly shear than does the TRANS−1510_1030 simulation.
Figure 4.1. Outline of each of the terms that influence RKW’s model of squall line evolution in the −1510 wind profile: a) The cold-pool term, b) the wind shear term and, c) $C/\Delta u$ evolution in time throughout the simulation, d) legend, e) and f) the meridionally averaged zonal wind and potential temperature profiles at various distances from the gust front in the TRANS−1510 simulations at $t = 4$ h, and in the g) and h) SHADE−1510 simulations at $t = 4$ h.
Figure 4.2. The same comparison field as given in Fig. 4.1 but for the $-1510_{1030}$ wind profile.
Figure 4.3. The same comparison field as given in Fig. 4.1 but for the 020 wind profile.
Figure 4.4. The same comparison field as given in Fig. 4.1 but for the 020_2040 wind profile.
(comparing Fig. 4.2f and 4.2h). This is due to weaker surface winds and stronger easterly winds near \( \sim 750 \) m in the SHADE–1510.1030 case. The impact of anvil cooling on the wind profile will be discussed later in this thesis.

Figures 4.3 and 4.4 show the RKW terms for the wind profiles with initially calm surface winds. The differences between the SHADE020 and TRANS020 wind profiles are much less noticeable than in the previous profiles examined. The cold pool strengths are relatively comparable, but with a slightly stronger cold pool in the SHADE020 simulation. The TRANS020 simulation has slightly more westerly shear throughout much of the simulation. Overall, the TRANS020 simulation is slightly more optimal throughout a large part of the simulation. The SHADE020 simulation shows near 1 K of cooling from the far field.

The cold pool is stronger in the SHADE020.2040 simulation than in the TRANS020 simulation. However, neither the \( \Delta u \) or \( C/\Delta u \) term is consistently greater for the SHADE020.2040 or TRANS020.2040 model runs throughout the 7 hour simulation. The SHADE020.2040 case has experienced 2 K of cooling from the far field, but this cooling does not seem to have much of an impact on the near gust front wind profile (compared to the TRANS020.2040 wind profile near the gust front).

Finally, Fig. 4.5 and Fig. 4.6 show the differences for wind profiles with initial westerly surface winds. Large differences are present between the SHADE1535 and TRANS1535 simulations, most noticeably in the wind shear term. The magnitude of the cold pool strength between the two simulations is variable. The TRANS1535 simulation is initially weaker, but then becomes much stronger in the final two hours of the simulation. Much less shear is present in the TRANS1535 simulation as the simulation progresses in time after \( t = 3 \) h. This influences the \( C/\Delta u \) term such that the TRANS1535 simulation becomes more sub-optimal than the SHADE1535 simulations. Inflowing air parcels experience near 1 K of cooling under the anvil shaded area compared to the TRANS1535 simulations.

Figure 4.6 shows similar and more enhanced differences as compared to the short-anvil simulation. In these simulations, the TRANS1535.3555 cold pool is stronger after 4 h, and there is less overall westerly shear as compared to the SHADE1535.3555 simulation. Similar to the short anvil simulations, the TRANS simulation is less optimal. The SHADE1535.3555 simulation shows about 2 K of cooling from the far field (compared to the TRANS1535.3555 simulation), and
Figure 4.5. The same comparison field as given in Fig. 4.1 for the 1535 wind profile.
Figure 4.6. The same comparison field as given in Fig. 4.1 for the 1535_3555 wind profile.
this cooling has a significant impact on the wind profile near the gust front. In both the SHADE1535 and SHADE1535_3555 simulations, the increased shear is the result of weaker westerly surface winds after 4 hours compared to the TRANS simulations (seen by comparing Fig. 4.5f and h, and Fig. 4.6f and h).

The investigation of the simulations’ progressions in terms of RKW theory shows a distinct trend based on the initial wind profile used. For easterly initial surface wind profiles, the SHADE simulations are more sub-optimal. For profiles with initially calm surface winds, there are few differences between the SHADE and TRANS simulations in either the long-anvil or short-anvil case. For westerly initial surface wind profiles, the TRANS simulations are more sub-optimal. The upcoming sections will discuss the reasons for these differences based on radiative modifications to the cold-pool and pre-line shear terms.

4.2 The cold pool term and shaded inflow simulations

A third suite of simulations was included to isolate the effects of anvil shading on the cold pool. In previous simulations, not only is the inflow shaded, but the cold pool is also shaded by trailing stratiform clouds. In order to shade the inflow without affecting the cold pool strength, a third set of runs called the ‘INFLOW’ simulations will now be presented. The ‘INFLOW’ simulations use the same shading technique for inflowing parcels as in the ‘SHADE’ simulations; however, any area behind the gust front does not experience radiative attenuation (i.e. all hydrometeors set to 0 kg kg\(^{-1}\) behind the gust front position in the radiation code). A schematic of this concept is given by Fig. 4.7.

Figure 4.8 gives an analysis of the cold pool strength and pre-line shear of the INFLOW simulations compared to the SHADE and TRANS simulations. The most noticeable aspect of Fig. 4.8 is that the INFLOW simulations are similar to the SHADE simulations in terms of cold-pool strength and westerly vertical wind shear. This implies that radiative effects have little or no effect on cold-pool strength, which promotes the idea that cold-pool air is generated and descends from aloft. Thus, the cold pool should not be significantly affected by any modified...
Figure 4.7. Schematic outlining the basic set-up for the INFLOW simulations. Areas ahead of the gust front are shaded, while those behind the gust front do not experience incoming radiation attenuation.

Figure 4.8. Total cold pool strength for the INFLOW, SHADE, and TRANS simulations for the (a) \(-1510\_1030\) wind profile, (b) \(020\_2040\) wind profile, (c) \(1535\_3555\) wind profile. Total wind shear for the (d) \(-1510\_1030\) wind profile, (e) \(020\_2040\) wind profile, and (f) \(1535\_3555\) wind profile.

surface fluxes due to radiative shading differences between simulations. Based on this observation, it seems as if cold-pool strength would be derived primarily from the amount of rainfall that falls to the rear of the storm (i.e. more trailing rainfall in more sub-optimal simulations) and/or increased evaporation from a rear-inflow jet (which is generally more prevalent in the more sub-optimal cases). In the
−1510 and −1510_1030 wind profiles, the more sub-optimal SHADE simulations show a stronger cold pool (Fig. 4.1a and 4.2a). In the 1535 and 1535_3555 wind profiles, the more sub-optimal TRANS simulations eventually have a stronger cold pool, although the cold pools are of comparable magnitude at $t = 4$ h (Fig. 4.5a and 4.6a). These observations are augmented by the apparent response of the cold pool to pre-gust front shear. This can be seen most easily in Fig. 4.5 and Fig. 4.6, comparing panels a and b. Significant decreases in shear in these simulations are followed by increases in cold pool strength at $\sim 1$ h later. If environmental shear is decreased, a system becomes more sub-optimal, and thus more trailing precipitation and increased rear-inflow evaporation would necessitate an increase in cold-pool strength. In conclusion, shortwave radiation has very little impact on cold-pool strength. However, the cold-pool strength is still sensitive to the radiative effects on the pre-gust-front shear, which will help determine the tilt of the updraft and the amount of precipitation advected to the rear of the system.

4.3 Evolution of the wind shear term

The amount of shortwave radiation received at the surface affects the shear profile ahead of the gust front through surface fluxes. Air parcels arriving at the gust front from the far field are subject to vorticity changes due to the cooling underneath the anvil. Previous analyses showed that $\Delta u$ is typically less in the SHADE case for initial easterly surface wind profiles, and less in the TRANS case for the initial westerly surface wind profiles. For the profiles with calm surface winds, the $\Delta u$ magnitudes are relatively similar between the SHADE and TRANS simulations. In each simulation, the SHADE case has a wind profile with seemingly less vertical mixing of momentum throughout the bottom 0–1 km than in its counterpart TRANS simulation (lower panels of Figs. 4.1 through 4.6). It is these differences in the wind profile evolution of the bottom 0–1 km of the atmosphere that appear to be exerting the most influence on the entire 0–2.5-km shear, and thus the optimality of the system as a whole in terms of $C/\Delta u$.

In order to investigate the processes responsible for the evolution of the environmental wind profile and 0–2.5-km shear, it is beneficial to consider the equation for meridional vorticity:
\[ \eta = \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}. \]  

(4.3)

In the inflow region, \( \frac{\partial w}{\partial x} \sim 0 \), so \( \eta = \frac{\partial u}{\partial z} (\Delta u) \). The modifications to parcel vorticity as air parcels approach the gust front through an anvil–shaded region lead to the differences between the profiles ahead of the gust front in the TRANS and SHADE simulations. The vorticity tendency of an individual parcel can be expressed as

\[ \frac{D\eta}{Dt} = -\frac{\partial B}{\partial x} + \hat{j} \cdot \nabla \times \vec{F} + \omega \cdot \nabla v. \]  

(4.4)

In this formulation, \( \eta \) is the meridional vorticity, \( \frac{D}{Dt} = \frac{\partial}{\partial t} + \vec{v} \cdot \nabla \), \( B \) is buoyancy, \( \vec{F} \) represents the viscous forces (turbulent viscosity + computational diffusion), and \( \omega \) is the vorticity vector. The horizontal variations of water vapor are negligible ahead of the gust front; therefore, the pre–gust front baroclinic term of the vorticity generation equation is approximated as

\[ -\frac{\partial B}{\partial x} = -\frac{g}{\theta_\infty} \frac{\partial \theta}{\partial x}. \]  

(4.5)

The \( \frac{D\eta}{Dt} \) and \( -\frac{\partial B}{\partial x} \) terms were computed from model output, and \( \hat{j} \cdot \nabla \times \vec{F} \) was obtained from \( \frac{D\eta}{Dt} + \frac{\partial B}{\partial x} - \omega \cdot \nabla v. \) To a good approximation, \( \hat{j} \cdot \nabla \times \vec{F} \) is equal to \( \frac{\partial^2}{\partial z^2}(K_{mv} \frac{\partial u}{\partial z}) \). The term \( K_{mv} \) indicates a vertical turbulent mixing coefficient for momentum which is related to the model-diagnosed turbulent kinetic energy (TKE \( \equiv (u'^2 + v'^2 + w'^2)/2 \)).

Figures 4.9 through 4.14 give the meridionally and vertically averaged horizontal vorticity tendencies over 0–2.5 km, as well as the meridionally and vertically averaged contributions from the baroclinic and turbulent terms in both the 0–1-km and 1–2.5-km layer at \( t = 4 \) h. It was found that the tilting/stretching term \( (\omega \cdot \nabla v) \) was orders of magnitude less than the other terms, as one would expect given the largely two-dimensional nature of the simulation. These plots begin 10 km to the east of the gust-front position and extend 400 km east, into the inflow region. Near the gust front, gradients in heating near the edges of convective towers produce large baroclinic and turbulent values. These gradients show up as large spikes in the vorticity tendency values typically within about 20 km of the gust front. These regions are too small (and thus parcel residence time is too short) to have
a significant impact of parcel vorticity. Instead, the significant impacts on parcel vorticity tendency due to anvil shading are seen in the 400-km inflow region. The variations in this region represent the horizontal vorticity changes as inflow parcels approach the gust front. A plot of meridionally averaged surface turbulent kinetic energy (TKE) is also given. Positions a) and b) on the figures mark important areas of the storm. Position a) indicates the location where $\frac{\partial \theta}{\partial z}$ changes from negative to positive in the lower 500 m of the atmosphere in the SHADE simulations (indicating static stability). Position b) indicates the edge of the anvil. It will be shown that these positions correspond with areas of substantial turbulent and baroclinic modification of vorticity. From this point forward, ‘horizontal vorticity’ will refer to meridional horizontal vorticity. By studying the meridional horizontal vorticity equation (Eq. 4.3), one can see that decreases in horizontal vorticity are reductions in westerly shear, while gains in horizontal vorticity increase westerly shear.

Figure 4.9 shows a 6-paneled image of these plots for the $-1510$ wind profile. Neither the baroclinic or turbulent vorticity tendency is consistently larger in the 1–2.5-km plots (Fig. 4.9b,c), as will be the case throughout the remaining simulations as well. Although the 1–2.5-km terms are influential in determining the overall magnitude of the 0–2.5-km vorticity tendency term, the differences between the SHADE and TRANS simulations are small in this region compared to the 0–1 km region. Thus, it is the 0–1-km terms which are largely responsible for the differences for parcel vorticity differences between the SHADE and TRANS simulations. In the 0–1-km terms, the baroclinic contribution to vorticity change in the SHADE$-1510$ simulation decreases rapidly after anvil shading begins (Fig. 4.9e). The 0–1-km turbulent terms in the SHADE$-1510$ simulation are relatively similar between the two simulations (Fig 4.9f). Low-level static stability in the SHADE$-1510$ simulation does not occur until a parcel reaches $\sim 15$ km ahead of the gust front. It is important to note that in this simulation (as well as in others), the turbulent term is on the order of $10^{-6}$, whereas the baroclinic generation term is an order of magnitude smaller ($10^{-7}$). Thus, the baroclinic term only accounts for a small part of the observed vorticity tendency; most of the vorticity tendency is the result of processes not related to baroclinic generation, advection, or tilting/stretching. These processes are represented by the residual
**Figure 4.9.** Vorticity tendency contributions for the $-1510$ wind profile at $t = 4$ h. Plots begin (at the left side of the figure) 10 km in front of the gust front position and extend zonally into the inflow region: a) meridionally and vertically averaged horizontal vorticity tendency from 0–2.5 km, b) meridionally and vertically averaged contributions to the vorticity tendency from the 1–2.5-km baroclinic term and c) from the 1–2.5-km turbulent term, d) surface TKE (throughout the domain, the box indicates the area examined in the other plots), e) meridionally and vertically averaged contributions to the vorticity tendency from the 0–1-km baroclinic term and f) 0–1-km turbulent term. The darker line on the TKE plot represents the shading simulation, and the lighter line represents transparent simulation. Position a) indicates the location where the bottom 500 m of the atmosphere becomes statically stable ($\frac{\partial \theta}{\partial z} > 0$) in the SHADE simulations. Position b) indicates the edge of the anvil.

Fig. 4.10 shows the results of the $-1510$,$1030$ longer anvil wind profile. In Fig. 4.10, the 0–2.5-km vorticity tendency shows a significant decrease in the SHADE$-1510$,$1030$ simulation approximately 120 km in front of the gust front. This decrease manifests itself in the turbulent generation of horizontal vorticity from 0–1 km, co-located with the onset of static stability. At this point, TKE also decreases substantially (Fig. 4.10d). There is some area between the point of initial static stability and the gust front in which the turbulent term in the SHADE simulation is not less than that of the TRANS simulation in the lower 0–1
km. This may indicate that vorticity tendency is affected rapidly at the onset of static stability, but after the parcel has been in a statically stable area for a long time, other turbulent processes (i.e., surface drag) may become more important. Once again, there is a strong indication of a rapid decrease in horizontal vorticity tendency in the lower 0–1 km in the baroclinic term underneath the anvil in the SHADE simulation. Overall, the combination of turbulent and baroclinic decreases in horizontal vorticity tendency under the anvil would lead to more negative 0–2.5-km horizontal vorticity tendencies in the SHADE−1510,1030 simulations than in the TRANS−1510,1030 simulations. The total 0–2.5-km horizontal vorticity tendency is similar between the simulations except near the area where low-level static stability begins, at which point the SHADE−1510,1030 simulation shows a more negative tendency. It is also important to note that the leading anvil, the change in static stability, and the change in baroclinic generation of vorticity begin farther ahead of the gust front in the −1510,1030 set of simulations than in the −1510 simulations.

Figures 4.11 and 4.12 illustrate the vorticity and TKE modifications for the
Figure 4.11. The same comparison field as given in Fig. 4.9 but for the 020 wind profile.

Figure 4.12. The same comparison field as given in Fig. 4.9 but for the 020,2040 wind profile.
020 and 020_2040 wind profiles respectively. The discussion of these figures will be grouped together, because relatively few differences are noticeable between the simulations. Nearly all terms for the 020 wind profile are identical between the SHADE and TRANS simulations. The exception is the 0–1-km baroclinic term, which still shows a strong decrease in horizontal vorticity tendency underneath the anvil for the SHADE020 simulations. For the 020_2040 wind profiles, there is a similar decrease in the 0–1-km baroclinic term for the SHADE runs, although in this case, the baroclinic modification begins farther in front of the gust front because of the longer leading anvil. The 0–1-km turbulent term in the SHADE020_2040 simulation shows a slight increase in turbulent generation of horizontal vorticity in the area of static stability. Turbulent kinetic energy also decreases in the shading case in the area of static stability.

Figure 4.13 shows the differences in vorticity generation for the 1535 wind profile simulations. The total horizontal vorticity tendency from 0–2.5 km is now greater in the SHADE1535 simulation, particularly within 30 km of the gust front. Looking at the contribution to this term, one will notice that the SHADE1535
Figure 4.14. The same comparison field as given in Fig. 4.9 but for the 1535.3555 wind profile.

The longer anvil simulation (Fig. 4.14) shows similar results as the 1535 wind profiles, but with differences extending farther away from the gust front owing to a longer leading anvil. The SHADE1535.3555 simulation shows an increase in the 0–1-km turbulent horizontal vorticity tendency term in the area of static stability, and this area of increased vorticity generation extends nearly 120 km in front of the gust front. The baroclinic term from 0–1 km shows a decrease in vorticity tendency for the SHADE1535.3555 simulation. There is overall more positive horizontal vorticity generation from 0–2.5 km for the SHADE1535.3555 simulation.
The results thus far indicate that the 0–1-km vorticity modifications play an important role in determining the total 0–2.5-km vorticity in front of the gust front, and these vorticity changes are seemingly co-located with changes in anvil shading and static stability. For initial easterly surface wind profiles, parcels in the SHADE simulations experience more of a decrease in positive horizontal vorticity as they approach the gust front in the long-anvil simulations due to a large area of static-stability-induced horizontal vorticity differences. Differences in parcel vorticity are small in the short-anvil case, due to the fact that static stability is only realized ~15 km ahead of the gust front, so the smaller magnitude baroclinic term is the only term that shows significant differences in vorticity generation between the two simulations. For calm wind profiles, there are only minor differences in vorticity change between SHADE and TRANS simulations. For initial westerly surface wind profiles, parcels in the SHADE simulations experience more of an increase in positive horizontal vorticity as they approach the gust front (most noticeably in the area where static stability begins). These results are expected given the $\Delta u$ analyses in the previous section. The following summarizes the effects of the baroclinic and turbulent vorticity generation terms with respect to cloud shading effects and differing initial wind profiles:

**Baroclinic vorticity generation**

The baroclinic term ($-\frac{\partial B}{\partial x}$) shows predictable changes for air parcels approaching the gust front in an anvil shaded region. In the TRANS simulations, $\theta$ should be the same in the far field as it is near the gust front. Therefore, the zonal gradient of $\theta$ (and therefore $B$) is zero. However, once shading is added to a simulation, $B$ becomes larger to the east (in the clear sky) than under the anvil (due to horizontal gradient in $\theta$). The term $\frac{\partial B}{\partial x}$ is now positive, so $-\frac{\partial B}{\partial x}$ is negative, and any parcel arriving at the gust front from the far field will experience a decrease in vorticity in the shaded region. An illustration of this concept is given by Fig. 4.15. This effect will be greatest for longer anvils, in which inflow parcels have a longer amount of time to cool, and there is a larger temperature deficit, and thus a larger buoyancy gradient between the clear sky and the anvil-shaded regions near the gust front.
For initial westerly surface winds
–Mixing effect
friction reduces initial surface winds;
vertical mixing weakens the environmental horizontal vorticity regardless of sign;
reduction of mixing under anvil leads to
more overall northerly environmental vorticity
+
– +h B
Baroclinic effect
weakens the southerly (positive) environmental horizontal vorticity
+
+++

Figure 4.15. Schematic illustrating the baroclinic generation of horizontal vorticity.

Turbulent vorticity generation
The turbulent term in the vorticity tendency equation involves the curl of the viscous force, which includes the effects of mixing, turbulent eddies, and generation of vorticity along a rough surface. The turbulent vorticity tendency term is more positive in the TRANS case for initial easterly surface wind profiles and less positive in the TRANS case for initial westerly surface wind profiles. In examining the averaged wind profiles given in Chapter 3, one can see that there is not as much mixing in the 0–1 km layer for the SHADE simulations. This reduction of mixing typically occurs around the location that static stability begins under the anvil. This can be seen in plots of the turbulence terms as a significant and sometime abrupt decrease in horizontal vorticity tendency for initial easterly surface wind profiles, and an increase in horizontal vorticity tendency for initial westerly surface wind profiles. The location of static stability is also associated with a significant decrease in TKE. So, the effect of the anvil shading on the turbulence term is to stabilize the boundary layer, thus suppressing the near-surface vertical mixing. For this reason, surface winds, which have been influenced by frictional drag, remain closer to the origin in all SHADE simulations because they do not mix with higher momentum air from aloft. Similarly, higher momentum air from above the surface layer does not mix down as readily, and is therefore not influenced as much by frictional drag in the SHADE simulations. These results are consistent with the findings of Frame (2009). For initial easterly surface wind profiles, the less mixing in the 0–1-km layer leads to stronger easterly shear near the surface and less 0–
2.5-km $\Delta u$ (westerly shear) in the shading case. For initial westerly surface wind profiles, less mixing leads to stronger westerly shear near the surface and greater 0–2.5-km $\Delta u$ in the shading case. For calm initial wind profiles, surface winds are zero initially, so the role of frictional drag is diminished, and the resulting evolution of the wind profiles between the SHADE and TRANS simulations show few differences. An illustration of mixing effects on the vorticity budget for each initial wind profile is given by Fig. 4.16 and Fig. 4.17.

The turbulent term is typically an order of magnitude larger than the baroclinic term in the lower 0–1 km of the atmosphere, therefore it has the largest effect on pre-gust front shear. The change in turbulent vorticity generation with the onset of static stability can be seen most noticeably in the long anvil cases, owing to in-
increased cooling and more suppressed vertical mixing. These results agree with the conclusions of Frame (2009), and Frame and Markowski (2010) that anvil cooling primarily impacts the wind profile ahead of the gust front by decreasing vertical mixing, and the magnitude of this decrease is dependent on the ground-relative wind profile. Given that the differences between the baroclinic and turbulent generation terms between SHADE and TRANS profiles are maximized with increased cooling (longer anvils), it is not surprising that the differences in the $C/\Delta u$ terms, and the vorticity terms are more noticeable in the longer anvil cases.
Chapter 5

Summary and conclusions

These idealized numerical simulations have yielded several important conclusions about the effects of cloud shading on squall lines. The most important conclusions can be summarized sequentially as follows:

- For the radiation parameterization used, only by artificially enhancing the ice crystal concentration of the anvils could realistic shortwave radiation reductions and low-level air temperature deficits be obtained within the regions shaded by the anvils.

- Radiative effects have little direct impact on the cold pool (radiation indirectly affects the cold pool by the aforementioned alteration of the convective region).

- Cloud shading most noticeably influences squall lines by modifying the pre-gust-front wind profile, specifically, its vertical shear and associated horizontal vorticity. Modifications of the pre-gust-front wind profile most directly affect the updraft tilt near the gust front, which is known to be sensitive to the relative strengths of the cold pool circulation versus the circulation associated with the environmental vertical wind shear.

- The horizontal vorticity in the inflow is modulated by both baroclinic and turbulent effects, with the latter being larger in magnitude than the former. Each of these effects acts most noticeably on the 0–1-km wind profile.
For the eastward moving/propagating trailing stratiform squall lines simulated herein, northerly (negative) horizontal vorticity (easterly vertical wind shear) is generated baroclinically in the inflow shaded by the leading anvil. This negative vorticity generation is due to an eastward-pointing horizontal buoyancy gradient in the lower 0–1 km of the atmosphere (i.e., a gradient that points from the shaded area toward the sunshine).

The sign of the turbulent mixing term in the vorticity equation (which tends to dominate) is sensitive to the initial wind profile. For an initial environment having easterly surface winds and westerly vertical wind shear, the reduction of vertical mixing near the surface owing to low-level stabilization within the shaded region reduces the 0–2.5-km westerly vertical wind shear (i.e., contributes to a negative meridional vorticity tendency in that layer). However, for an initial environment having westerly surface winds and westerly vertical wind shear, the shading-induced reduction of vertical mixing increases the 0–2.5-km westerly vertical wind shear (i.e., contributes to a positive meridional vorticity tendency in that layer).

In storms with longer leading anvils, differences between simulation with and without cloud shading are larger. With longer leading anvils, inflow parcels experience longer residence times under the anvil and therefore more cooling under the anvil, which decreases vertical mixing further ahead of the gust front and also leads to a larger gradient in baroclinicity between the anvil shaded area and the clear sky.

The greatest differences between clear sky and anvil–shaded horizontal vorticity occurs in an initial easterly surface wind profile with westerly shear aloft, where the baroclinic and turbulent terms work together to influence the vorticity tendency in an anvil–shaded area. In this case, stronger easterly surface winds (or stronger westerly shear above the surface) will ultimately experience greater differences between clear sky and anvil-shaded areas as strong winds will maximize vertical mixing differences in the surface layer. For initial westerly surface wind profiles, the baroclinic and turbulent terms oppose each other in anvil–shaded areas, although the turbulence term still dominates the overall tendency.
The cold pool strength and pre-gust-front wind profile of a squall line responds to baroclinic and turbulent forcings of an initial wind profile exposed to cloud shading. RKW theory remains a controversial way to explain system strength, and in this case, predictions of a more optimal system did not necessarily lead to a stronger system when examining some common metrics for system strength. However, the calculation of $C/\Delta u$ presented herein did not take into account the effect of the rear-inflow jet, which can alter the RKW balance terms (Weisman 1992). It is also important to note that lifted parcels in the SHADE and TRANS simulations did not have equal initial surface temperatures, which could impact the amount of CAPE that a parcel will experience given the same amount of gust-front lifting.

Although a variety of processes undoubtedly influence system strength, more enhanced differences in the cold pool and pre-line shear between simulations with and without cloud shading would likely lead to more noticeable differences in system strength. As stated earlier, the numerical simulations did not capture some of the larger temperature deficits under an anvil that have been observed in the field. It is reasonable to assume that enhanced pre-line cooling due to optically thicker anvils or longer parcel residence times would increase the baroclinic and turbulent differences mentioned above and lead to larger differences between the systems with and without cloud shading.

The attenuation of shortwave radiation does play a role in modifying some of the parameters which have been proven to control squall-line evolution. A further investigation of the factors that govern squall-line evolution and the improvement of microphysical parameterizations in the future will likely clarify the full magnitude of the effects which are presented in this thesis.
References


