Atmospheric Controls on Soil Moisture-Boundary Layer Interactions

by

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Submitted to the Department of Civil and Environmental Engineering in partial fulfillment of the requirements for the degree of

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Abstract

This thesis addresses the question of how the early morning atmospheric thermodynamic structure affects the interaction between the soil moisture state and the growth and development of the boundary layer (BL), leading to the triggering of convection. It is concluded that in mid-latitudes, for matters of convective triggering and response to land surface conditions, the critical portion of the atmosphere—approximately 1 to 3 km above the ground surface—is independent of geographic location and local synoptic setting. As long as the low levels of the troposphere are relatively humid but not extremely close to saturation, a negative feedback between soil moisture and rainfall is likely when the early morning temperature lapse rate in this region is dry adiabatic; a positive feedback is likely when it is moist adiabatic; and when there is a temperature inversion in this region, deep convection cannot occur, independent of the soil moisture. Additionally, when the low levels of the troposphere are extremely dry or very close to saturation, the occurrence of convection is determined solely by the atmospheric conditions.

Essential characteristics of the temperature structure of the early-morning atmosphere are captured by a new thermodynamic measure, the Convective Triggering Potential (CTP), developed to distinguish between soundings favoring rainfall over dry soils from those favoring rainfall over wet soils. Many measures of atmospheric humidity are effective at separating atmospherically-controlled cases from cases where the land surface conditions can influence the likelihood for convection, but $HI_{low}$, a variation of a humidity index, proved most effective.

A one-dimensional model of the planetary boundary layer (BL) and surface energy budget has been modified to allow the growing BL to entrain air from an observed atmospheric sounding. The model is used to analyze the impact of soil saturation on BL development and the triggering of convection in different atmospheric settings. Results from this 1D model and from the three-dimensional Fifth-Generation Penn State/NCAR Mesoscale Model (MM5) show a small but significant positive soil moisture-rainfall feedback in Illinois. This is consistent with an analysis of the distribution of early morning sounding values of $CTP$ and $H I_{low}$ from Illinois, though wind effects important in the MM5 simulations are not captured.
by the $CTP-HI_{low}$ framework. From the MM5 simulations, it is concluded that the land surface condition can impact the potential for convection only when the atmosphere is not already predisposed to convect or not to convect. This atmospheric predisposition can be determined by analyzing the $CTP$, the $HI_{low}$, and the vertical profile of the winds.

Analyses of $CTP-HI_{low}$ scatter plots from radiosonde stations across the contiguous 48 United States reveal that positive feedbacks are likely in much of the eastern half of the country. The only area showing a potential negative feedback is in the Dryline and Monsoon Region of the arid southwest. Land surface conditions are unlikely to impact convective triggering in the rest of the western half of the country. Use of the 1D BL model at four additional stations confirms that the $CTP-HI_{low}$ framework used in this nationwide analysis is valid for regions far removed from Illinois, where it was originally developed.

Thesis Supervisor: Elfatih A.B. Eltahir
Title: Associate Professor of Civil and Environmental Engineering
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Chapter 1

Introduction

The earth's land surface is a system with remarkable temporal and spatial variability. When solar radiation reaches the land surface, a portion is reflected away and the rest is consumed by heat flux into the ground, by evapotranspiration, and by sensible heat flux. These three fluxes act to increase the ground temperature, the specific humidity of the air, and the temperature of the air, respectively. The same amount of radiation reaching a desert or a rainforest will be utilized in dramatically different ways: over areas with ample water, evapotranspiration will dominate, while over areas without much water, the sensible heat flux will dominate. Consequently, soil moisture acts as a primary determinant in the partitioning of available radiation at the land surface. This partitioning, in turn, impacts the planetary boundary layer (BL) and the state of the atmosphere.

The planetary boundary layer, according to Stull (1988), is the part of the troposphere that is both affected by the characteristics of the land surface, and also responds to surface forcings on time scales of an hour or less. It is through the boundary layer that land surface effects are transmitted to the upper layers of the atmosphere, in the BL that most weather phenomena develop, and through the BL that these developments pass to reach the surface of the planet. The objective of this work is to improve the understanding of land surface-boundary layer interactions, by focusing on the role that soil moisture and local atmospheric conditions play in these interactions. Specifically, this thesis addresses the question of how the early morning atmospheric thermodynamic structure affects the
interactions between fluxes from the land surface (and thus the soil moisture state) and the growth and development of the boundary layer (BL) and the triggering of convection.

There are three main characteristics of the early morning atmospheric structure that significantly influence the characteristics of the boundary layer that will develop during the course of the coming day (Figure 1.1):

- The properties of the residual layer, since this air will quite likely be incorporated into the BL;

- The depth of the nocturnal stable layer, since this will determine the ability of surface fluxes to reach beyond the air of this near-surface stable layer and the time at which they do so; and,

- The height and strength of the inversion separating the mixed layer from the overlying free atmosphere, since this affects both the rate of entrainment of overlying air into the developing BL, and the buildup of moisture and moist static energy in the mixed layer.
A few studies have investigated the influence of varying one or more of these properties, notably Ek and Mahrt (1994), Chen and Avissar (1994), and Segal et al. (1995), which will be discussed below. There is need, however, for a measure that assesses the combined effects of these components of the early morning atmospheric structure on the potential for the land surface soil moisture to influence the development of convection.

Modeling results from global and regional climate models (GCMs and RCMs) have produced inconsistent reports on the degree and even the direction of the feedback between the soil moisture condition and subsequent rainfall. For example, the regional model of Giorgi et al. (1996) showed that dry soils enhance convection through the increase of turbulent mixing that accompanies increased sensible heat flux. Pan et al. (1996), in contrast, showed a positive feedback between soil moisture and rainfall in the United States during the drought of 1988 and the flood of 1993. Significantly, the work of Pal (1997) showed that the response of rainfall to soil moisture was dependent on the convection scheme used in the model.

Observational studies have also shown varied responses between soil moisture and rainfall, and many of these have noted the importance of the early morning atmosphere in these interactions. Wetzel et al. (1996) found evidence for atmospheric controls on soil moisture-boundary layer interactions in their analysis of one day from the FIFE experiment in Kansas, and one day from an Oklahoma summer. They determined that, "The primary reason for the difference in the response of the atmosphere to soil moisture between these two cases is the difference in the thermodynamic structure of the atmosphere over the two sites" (pp. 7361-2). Over the FIFE site, they found that clouds first formed over wet areas. In contrast, the Oklahoma case showed that clouds quickly formed over dry, sparsely vegetated areas. The primary difference in these two cases were the conditions of the stable and residual layers. In the Oklahoma case, there was a very shallow nocturnal inversion, which was easily eroded. In the FIFE case, the pre-existing stable layer was quite deep, leading to suppression of rising thermals. This suppression allowed for the buildup of moisture within the stable layer. Clouds then first formed over areas with the largest latent heat flux.

This example demonstrates the importance of the stable nocturnal layer in allowing for the buildup of moisture and moist static energy (MSE) within this near-surface zone. Segal et al. (1995) also note the importance of the layer nearest the surface, claiming that
under most conditions sensible heat flux plays only a secondary role on the development of precipitation. When there is a strong nocturnal boundary layer, however, this heating is crucial in the breakdown of the stable layer. In these cases, strong sensible heating can lead to spontaneous convection. After this surface inversion is eroded, however, the residual layer becomes important.

The studies of Rabin et al. (1990), Cutrim et al. (1995), and Rabin and Martin (1996) all focused on the development of shallow, fair weather cumulus clouds during relatively dry atmospheric conditions. During the dry seasons in Oklahoma, the Amazon, and the central US, respectively, areas of high sensible heat flux were seen to lead to earlier and more frequent shallow cumulus development than areas of high latent heat flux. These were, in general, non-precipitating clouds: as discussed by Mahrt (1997) and Mahrt and Pierce (1980), shallow convection tends to dry out the boundary layer before rainfall can develop.

In their investigation of the role of the capping inversion on the development of hail storms in northeastern Colorado, Mahrt (1997) and Mahrt and Pierce (1980) found that a weak capping inversion allowed widespread moist but shallow convection to develop. In these circumstances, many clouds were competing for limited moisture, preventing the development of a large severe storm. A somewhat enhanced inversion inhibited moist convection long enough for moisture and moist static energy to build up in the low levels of the troposphere. Once the larger-than-normal initiation energy was surpassed, an extreme storm event began. However, if the inversion was too strong, the required initiation energy was too great to be met and exceeded, and convection was fully suppressed.

Segal et al. (1995) also concluded that there is an intermediate range of inversion strengths most conducive to the development of precipitating convection. In addition, they explored the significance of the height of the capping inversion. When the cap height was high, entrainment was reduced because “the depth of the initial mixed layer [was] close to that of the afternoon mixed layer” (pg. 399). This lead to less dilution of moisture and MSE within the mixed layer and enhanced potential for deep convection. A shallower depth to cap, on the other hand, meant that the surface fluxes had greater relative impact, particularly in the early stages of the day. In this case, entrainment effects may be quite large, and the properties of the free atmosphere, as well as those of the residual layer, become quite
important.

The strength and height of the capping inversion were also shown to be important by Betts et al. (1996). They stress that the surface flux of MSE into the growing boundary layer is proportional to the sum of the sensible and latent heat fluxes, such that partitioning of available energy between these terms does not alter the total flux of MSE contributed from the surface. However, the diurnal fluctuations of MSE in the BL are closely tied to the surface sensible heat flux, since greater sensible heat flux leads to a deeper BL with more entrainment, both effects reducing the diurnal rise of MSE in the BL. The strength and height of the capping inversion will partially dictate the severity of this effect, as will the velocity of rising thermals.

Another important aspect of the early morning atmosphere is the humidity in the residual layer, as stressed by Chen and Avissar (1994). With their modeling analysis of humidity variations in an initial thermodynamic profile from the FIFE observations of 28 July 1989, they concluded that, “Depending on the atmospheric conditions, a significant variation in the land-surface moisture can produce either an increase, a decrease, or almost no change in the simulated cloud amount” (pg. 1397).

Ek and Mahrt (1994) used data from the HAPEX-MOBILHY experiment in addition to a one-dimensional model of the soil and boundary layer to look at the dependence of the relative humidity at the top of the BL on soil moisture, large-scale vertical motion, and the moisture and temperature stratification above the BL. They found that conditions favoring a negative feedback between soil moisture and cloud development occur when stratification above the BL is weak, while a positive feedback is favored when the air above the BL is strongly statified. They stress that results gained from individual experiments or case studies may be indicative of only one of these circumstances, and therefore may not be extendable to broad climate feedback arguments.

The role that soil moisture or vegetation play in the development of clouds and rainfall are important for an understanding of both the current climate, and the implications of future climate scenarios. Many modeling studies of the effects of increased atmospheric CO₂ (e.g., Manabe and Wetherald, 1987; Wetherald and Manabe, 1995; Rind et al., 1990; Mitchell and Warrilow, 1987) show general trends of higher summertime temperatures, higher potential
evaporation, and increased evapotranspiration outweighing increased precipitation. These effects lead to a general drying of soils, but there are regional variations which differ from these general trends. In order to fully understand the implications of these results, a better understanding of how the interactions between soil moisture and rainfall are controlled is needed.

These issues were addressed in earlier work that showed a small but significant positive feedback between soil moisture and subsequent rainfall in Illinois (Findell and Eltahir, 1997). Using soil moisture observations and near-surface air temperature, humidity, and pressure data from Illinois, Findell and Eltahir (1999) found that the feedback was not transmitted via a positive correlation between soil moisture and the moist static energy (MSE) of the air; nor was there evidence of a positive correlation between the MSE of the near-surface air and rainfall, as observed in the Amazon by Eltahir and Pal (1996), and discussed in theoretical terms by Eltahir (1998). There was, however, evidence of a significant negative correlation between soil moisture and the wet-bulb depression, and also between the wet-bulb depression and rainfall. These results led to the conclusion that a more complete analysis of the structure and development of the entire boundary layer was required to describe atmospheric controls on soil moisture-boundary layer interactions.

The goal of this work is to understand the interactions between soil moisture and the boundary layer, and to investigate how feedbacks might be dependent on the atmospheric conditions typical of a particular region. The hypothesis to be addressed contends that features of the early morning atmosphere significantly control the degree to which soil moisture can impact BL growth and development. The critical portion of the atmosphere was found to be between 100 and 300 mb above the surface. This region is the critical interface between the near-surface region, which is almost always incorporated into the growing BL, and free atmospheric air, which is almost never incorporated into the BL. A relatively high temperature lapse rate in this region suggests that the air is easy to incorporate into a growing BL. Dry soils have an advantage for triggering convection in these circumstances, because the boundary layer grows more slowly over wet soils and may not reach this easily-entrained region before the midday peak of available energy. When the lapse rate in this region is relatively small, closer to moist adiabatic, the Level of the Free Convection (LFC) is sig-
significantly reduced by small increases in the moist static energy ($\theta_E$ or $\theta_w$) of the boundary layer. In these circumstances, wet soils have an advantage, since boundary layers over wet soils tend to be shallower, more humid, and have a higher $\theta_E$ than those over dry soils.

These features of the early-morning atmosphere are captured by a new measure of the Convective Triggering Potential ($CTP$), developed from Illinois sounding data incorporated into a modified one-dimensional boundary layer model. This modeling work and the development of the $CTP$ are discussed in Chapters 2 and 3.

The results of the one-dimensional work show that using the $CTP$ and a measure of the low-level humidity deficit, $HI_{low}$, one can determine with a fairly high degree of certainty if wet soils have an advantage for the triggering of rainfall (wet soil advantage), or if dry soils have such an advantage (dry soil advantage), or if the land surface conditions are inconsequential to the likelihood of rainfall (atmospherically controlled). In Illinois, a larger proportion of early-morning soundings fall into the wet soil advantage regime than in the dry soil advantage regime, suggesting that this area is likely to see a positive soil moisture-rainfall feedback.

This suggestion is supported by the results of three-dimensional modeling work with The Fifth-Generation Penn State/NCAR Mesoscale Model (MM5) (Grell et al., 1995), which is described in Chapter 4. In Chapter 5, results from MM5 simulations are analyzed within the context of the $CTP-HI_{low}$ framework of understanding to investigate how this approach for studying the interactions between the land surface and the atmosphere holds up in a 3D setting. The significance of the vertical profile of the wind conditions will be stressed in the discussion of these results, where the 1D-based framework was found to be both valid and helpful, as long as the winds are capable of supporting convection.

This framework of understanding is used to analyze early-morning soundings from 76 daily sounding stations throughout the United States. With this analysis, regions of the country that are likely to see a positive soil moisture-rainfall feedback, a negative feedback, and no clear impact of the soil condition on rainfall were identified. This analysis is discussed in Chapter 6.

Chapter 7 presents conclusions from all three parts of this work, and briefly mentions some implications of these results, and questions to be addressed with future research.
Chapter 2

One-dimensional boundary layer model description

The model used in this work is a modified version of Kim and Entekhabi's (1998a, b) mixed-layer model of the surface energy budget and the planetary boundary layer (PBL). The heart of the model is comprised of equations for soil temperature ($T_s$), mixed-layer potential temperature ($\theta$), mixed-layer specific humidity ($q$), and the height of the PBL ($h$). In order to look at boundary layer growth on days with different early-morning atmospheric conditions, alterations to the original model were required. The model has been altered in the following ways (Figure 2.1):

1. cloud fraction is set to zero;
2. soil saturation is fixed for the duration of the model runs;
3. the growing BL entrains air from a user-input prescribed sounding, rather than from constant lapse rate profiles;
4. free convection is triggered when the growing BL reaches the level of free convection (LFC): at this point, the model assumptions of a well-mixed, cloud-free boundary layer are no longer valid and the simulation is terminated.

The last two changes are fundamental changes in the nature of the model. They allow for a melding of data analysis and model simulations. Confining the analysis to clear skies allows us to focus on the impacts of land surface conditions in the triggering of convection,
be it deep, precipitating convection, or weak convection producing shallow clouds. The model halts whenever either of these conditions occurs, since after free convection the model assumptions, including the no-cloud assumption, are no longer valid. We are considering time scales on the order of 12 hours, during which the assumption of constant soil saturation is reasonable.

The model is initiated in the early morning, preferably at or near sunrise, and proceeds until the end of the day or until free convection is triggered. Thus, there are three potential outcomes of each model run: deep convection which is likely to produce rain, shallow convection which is not likely to produce rain, or no convection. The first case will generally be referred to as “rain,” and the second case as “shallow clouds,” though it is recognized that these terms simply refer to the likelihood of rain and shallow clouds. The distinction between rain and shallow clouds depends on both the convective available potential energy (CAPE) and on the depth separating the level of free convection (LFC) from the level of neutral buoyancy (LNB). For rain to occur, it is assumed that the CAPE must be greater than 400 J/kg and that the depth of convection must be greater than 5 km. These threshold values are appropriate for the mid-latitude continental regimes studied in this thesis (Battan, 1973). Battan (1973) cites cloud-census studies to show that precipitating convection in tropical oceanic environments often occurs from much shallower clouds than in mid-latitude continental regimes.

Model results are not sensitive to changes in these threshold values when they are in the range of 3-5 km and 200-400 J/kg. All of the shallow cloud events in the model runs with Illinois data either have CAPE < 200 J/g and/or convection depths < 3 km. In
Chapter 6 the one-dimensional model described in this chapter will be used with soundings from four other stations: one in Ohio, one in South Carolina, one in Louisiana, and one in New Mexico. At three of these four stations, the same gap in the bivariate CAPE-depth distribution occurs. At the New Mexico station, three of the 86 soundings used to initialize the model resulted in convection being triggered over dry soils with CAPEs in the mid- to high-300s. No other cases were close to the 5 km and 400 J/kg thresholds.

Details of the 1D boundary layer model are discussed in the next sections, beginning with the budget equations, and continuing with the growth and collapse of the BL, the turbulent surface fluxes, entrainment, the treatment of radiation in the model, and a brief example with comparisons to observations.

### 2.1 Budget Equations

Following Kim and Entekhabi (1998a), the model has prognostic equations for soil temperature $T_s$, mixed-layer potential temperature $\theta$, and mixed-layer specific humidity $q$:

\[
\begin{align*}
    z_t C_s \frac{dT_s}{dt} &= R_s (1 - \alpha) + [R_{ad}(1 - \epsilon_a) + R_{sd}]\epsilon_s - R_{gu} - H - \lambda E \quad (2.1) \\
    \rho c_p h \frac{d\theta}{dt} &= [R_{ad} + R_{gu} + (R_{ad}(1 - \epsilon_a) + R_{sd})(1 - \epsilon_s)]\epsilon_a - R_{sd} - R_{su} + H - H_{top} \quad (2.2) \\
    \rho h \frac{dq}{dt} &= E - E_{top} \quad (2.3)
\end{align*}
\]

The soil temperature equation shows the influence of three radiative terms: incoming solar ($R_s$, reduced according to the albedo $\alpha$), downwelling longwave from within ($R_{sd}$) and above ($R_{ad}$) the boundary layer (adjusted according to the emissivities of the mixed layer ($\epsilon_a$) and the soil ($\epsilon_s$)), and upwelling longwave radiation from the ground ($R_{gu}$). In addition, the soil temperature is influenced by the fluxes of sensible and latent heat ($H$ and $\lambda E$) from the ground surface. Other parameters in Equation 2.1 include the soil thermal depth $z_t$ and the volumetric soil heat capacity $C_s$.

The mixed layer temperature is affected by multiple longwave radiative terms, including direct radiation into the BL from above ($R_{ad}$) and below ($R_{gu}$), and that reflected off of
the land surface (the term within the parentheses within the square brackets). Longwave radiation lost from the BL is also included, both from the bottom \( R_{sd} \) and from the top \( R_{su} \). Finally, sensible heat fluxes from the surface and from entrainment at the top \( (H - H_{top}) \) are also included. Density is given by \( \rho \), \( c_p \) is the specific heat of dry air, and \( h \) is the mixed layer height.

The mixed layer humidity budget equation takes a much simpler form than the other two: only latent heat fluxes from the surface and from entrainment at the top need to be considered.

### 2.2 Mixed-Layer Height Evolution

Another core equation in the model describes the growth of the boundary layer height \( h \). Both the daytime growth of the BL, which proceeds mainly in response to the virtual sensible heat flux from the surface, \( H_v \), and the collapse of the BL when the solar forcing disappears are modeled. \( H_v \) is similar to the sensible heat flux, \( H \), but it includes information about the surface evaporative flux, \( \lambda E \), to account for the impact of moisture on buoyancy. It is given by \( H_v = H + 0.61 \theta c_p E \approx H + 0.07 \lambda E \).

Following the work of Smeda (1979), the rate of change of \( h \) is given by:

\[
\frac{dh}{dt} = \frac{2(G_s - D_1 - \delta D_2)\theta}{gh\delta_\theta} + \frac{H_v}{\rho c_p \delta_\theta}.
\]

The numerator of the first term includes terms to represent the production of mechanical turbulent energy \( G_s \), the dissipation of mechanical turbulent energy in all circumstances \( D_1 \), and the excess dissipation of mechanical turbulent energy in unstable situations \( D_2 \) (\( \delta = 0 \) in stable conditions, 1 in unstable conditions). Additional variables are potential temperature \( \theta \), the inversion strength of potential temperature \( \delta_\theta \), and the acceleration of gravity \( g \). The second term, accounting for the virtual heat flux from the surface, is the primary forcing mechanism during daylight hours, which is the focus period of this work. At night, the first term becomes dominant. For complete descriptions of the turbulent terms and the usage of Smeda's (1979) formulation for the transition between daytime and nighttime regimes, see Kim and Entekhabi (1998a).
One adaptation of this formulation for the boundary layer height is discussed below with the description of the entrainment process in Section 2.4.

2.3 Surface Fluxes

2.3.1 Surface Evapotranspiration

The potential evapotranspiration rate $\lambda E_p$ is dependent on the mixed layer humidity deficit $q^*(T_s, p_s) - q$, the aerodynamic resistance $r_{aero}$, and the minimum stomatal resistance $r_{smin}$. The actual evapotranspiration rate $\lambda E$ is the potential rate multiplied by the factor $f_{res} = (r_{aero} - r_{smin})/(r_{aero} - r_s)$. The aerodynamic resistance varies only slightly, and will be discussed in greater detail in the next section. The stomatal resistance $r_s$, on the other hand, can vary significantly. It is dependent on soil moisture stress, levels of photosynthetically active radiation (PAR), and the vapor pressure deficit, according to the formulation by Kim and Verma (1991):

$$r_s = \frac{r_{smin}}{\sum_{i=1}^{3} f_i}$$

where

$$f_1 = \frac{R_{par}}{a_1 + R_{par}}$$

$$f_2 = \frac{1}{(1 + a_2)(q^*(T_s, p_s) - q)}$$

$$f_3 = \begin{cases} 
1 & \psi \leq \psi_{lim} \\
\log\psi - \log\psi_{lim} & \psi_{lim} \leq \psi \leq \psi_{wilt} \\
0 & \psi \geq \psi_{wilt}
\end{cases}$$

Following Kim and Entekhabi (1998b), $a_1$ is set to 50 $W/m^2$, $a_2$ to 0.18, $\psi_{lim}$ to 5 m, $\psi_{wilt}$ to 160 m, and $r_{smin}$ to 50 s/m. Soil suction $\psi$ is determined by the Clapp and Hornberger (1978) formulation of as a function of soil wetness, $W$:

$$\psi = \psi_{sat} W^{-B}$$

The Clapp-Hornberger values used in this work are those for loamy sand: $\psi_{sat} = 0.09 m$. 

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and $B = 4.38$. (Note that the Brooks-Corey (1966) formulation is identical, but with $m = 1/B = 0.228$.)

Another important effect of soil moisture is the impact on the albedo. With high soil moisture, the ground surface tends to be darker than with low soil moisture. This is assumed to impact net radiation $R_{net}$ through the albedo $\alpha$ according to the equation

$$\alpha = 0.20 - 0.10 \times W$$  \hspace{1cm} (2.8)

The combined effects of the stomatal resistance and the albedo dependences on soil moisture lead to the order of 60 $W/m^2$ more net radiation at mid-day in the wet soil moisture scenarios than in the dry soil moisture scenarios.

### 2.3.2 Surface Sensible Heat Flux

The sensible heat flux $H$ is forced by the temperature difference between the soil and the mixed layer:

$$H = \rho c_p (T_g - \theta) / r_{aero}$$  \hspace{1cm} (2.9)

where $T_g$ is the ground temperature, $\theta$ is the mixed layer potential temperature, $\rho$ is the density, $c_p$ is the specific heat of dry air, and $r_{aero}$ is the aerodynamic resistance, which is itself dependent on the Monin-Obukov length. The Monin-Obukov length is a function of the Richardson number, which is in turn a function of stability. However, in this model we are calculating the aerodynamic resistance near the surface (i.e., within the well-mixed boundary layer) where by definition the model is always neutral. Aerodynamic resistance then shows very little variability with time or with soil saturation. This effectively means that the sensible heat flux is dependent only on the temperature difference between the ground and the air.

### 2.3.3 Ground Heat Flux

The ground heat flux is determined to maintain surface energy balance:

$$G = R_{net} - H - \lambda E.$$  \hspace{1cm} (2.10)
2.4 Entrainment of Overlying Air

The most significant alteration made to the original version of Kim and Entekhabi's (1998a,b) PBL model is the representation of air above the boundary layer (Figure 2.1). In the original model, the boundary layer is represented as a homogeneous mixed layer with potential temperature $\theta$ and specific humidity $q$. Above this layer is an initial specified step jump of strength $\Delta \theta$ and $\Delta q$. Above this step jump, lapse rates of temperature ($\gamma_\theta$) and moisture ($\gamma_q$) are specified. The strengths of these inversions, $\Delta \theta$ and $\Delta q$, change with time according to the equations

$$\frac{d\Delta \theta}{dt} = \gamma_\theta \frac{dh}{dt} - \frac{d\theta}{dt} \quad (2.11)$$

$$\frac{d\Delta q}{dt} = \gamma_q \frac{dh}{dt} - \frac{dq}{dt}. \quad (2.12)$$

In the revised formulation of the model, an initial condition sounding is input by the user, and the growing boundary layer entrains air with the characteristics of this sounding. This enables us to see the effects of different initial atmospheres on boundary layer growth and development, and on the relative importance of soil moisture-boundary layer feedbacks in different initial atmospheres. The new formulation is applied by re-evaluating the air overlying the BL top at each time step. The values of $\Delta \theta$ and $\Delta q$ are determined by the difference between the BL $q$ and $\theta$ and the next sounding observation above the BL top. The lapse rates are determined between this sounding level and the next. (Soundings used in the model typically have data points every 15 or 20 mb.) When $\Delta \theta$ is positive, the air overlying the BL is potentially warmer and the model calculations proceed according to the equations above. When $\Delta \theta$ is negative however, the BL is more buoyant than the overlying air and the model can proceed in two ways.

First, if $\Delta \theta$ is negative because the BL top has reached the level of free convection (LFC), then free convection is triggered and the simulation is terminated. As discussed earlier, free convection can mean the likelihood of rain or the likelihood of shallow clouds, depending on the $CAPE$ and the depth of convection. $CAPE > 100 \, J$ and $(LNB - LFC) > 3 \, km$ is taken to suggest that rain is likely. If one or both of these conditions are not met, shallow clouds are assumed to result. Since the model halts at this point, there is no suggestion that rainfall
could not occur after the onset of shallow clouds. Further work without the assumption of
zero cloud cover is necessary to address this question.

A recent improvement to the model has included the capacity for turbulent overshooting
of small negative areas below the LFC. Rising air can often have large vertical velocities, such
that momentum can carry the air well above the top of the well-mixed layer. We allow for this
turbulence to overcome small negative areas (order 5 J/kg), determining "small" according
to the magnitude of the turbulence. A scaling analysis shows that turbulent kinetic energy
(TKE) is proportional to the surface buoyancy flux (BF) and the boundary layer height (h) according
to the equation

\[ TKE \propto (h \times BF)^{2/3}, \]  \hspace{1cm} (2.13)

where the BF is determined from the surface fluxes:

\[ BF = \frac{g}{T} \left( \frac{H}{c_p} + 0.608 \frac{\lambda E}{\lambda} \right). \] \hspace{1cm} (2.14)

Convection is triggered when the negative area under the LFC is less than 2 times this calculated TKE. The results are not particularly sensitive to the choice of the proportionality constant: a constant of 1.5 rather than 2 stalls the triggering for a few minutes but does not change the occurrence of convection.

The second reason that \( \Delta \theta \) can be negative is the existence of a shallow layer (on the order of 10 mb) of air that is potentially cooler than the boundary layer but still below the LFC. The model can entrain this shallow layer into the growing BL without assuming that free convection has begun. The BL top then becomes the neutrally buoyant level at the top of the shallow layer of cooler air, and the new BL \( \theta \) and \( q \) are determined by weighted mixing of this new volume with the old BL volume.

This feature of the model was a necessary addition to allow for use with real soundings. Many profiles have these shallow marginally stable layers mentioned in the previous paragraph. They are often just one observation in the vertical, and yet without this additional feature, these segments would cause the model to halt before the top of the boundary layer has reached the true LFC. When soundings are particularly complicated, with the path of a
parcel lifted from the mixed layer crossing the environmental sounding multiple times, the calculation of the LFC can be imperfect, and this mixing feature can be employed inappropriately when the model run should in fact be halted. Each model run is carefully analyzed to be sure that these cases are corrected, and free convection is assumed to have occurred when the LFC and the BL top met.

2.5 Radiative Terms

2.5.1 Short Wave Radiation

Computation of incoming short wave radiation follows the work of Liou (1980). There is dependence on the time of year, time of day, and latitude of interest. In the original formulation of the model, this calculated solar radiation, $I_{act}$, reaches the surface in cloudless conditions. The presence of clouds reduces the amount of short wave radiation reaching the surface, often quite dramatically. Rather than complicate our analysis with the impacts of clouds, we simply set the cloud fraction to zero and concern ourselves with the behavior of the boundary layer prior to the initiation of free convection, at which time cloud cover can no longer be assumed to be negligible.

We do, however, account for clear sky absorption and scattering of short wave radiation, according to the empirical method of Eagleson (1970), as reported by Bras (1990):

\[
I_{\text{clear}} = I_{\text{down}} \exp(-na_1m)
\]  

(2.15)

where $n$ is a turbidity factor that varies from about 2.0 for clear mountain air to 4 or 5 for smoggy urban air (2.5 is used for these runs dealing with agricultural areas of Illinois), $m$ is the optical air mass, given by

\[
m = [\sin \alpha + 0.1500(\alpha + 3.885)^{-1.253}]^{-1},
\]  

(2.16)

$\alpha$ is the solar altitude, and $a_1$ is the molecular scattering coefficient, given by:

\[
a = 0.128 - 0.054 \log_{10} m.
\]  

(2.17)
2.5.2 Long Wave Radiation

Computation of downwelling longwave radiation remains unchanged from the formulation of Kim and Entekhabi (1998a). It accounts for the dependence on atmospheric conditions, following the work of Brutsaert (1975) and Brubaker and Entekhabi (1995). Upwelling longwave radiation from the surface $R_{gu}$ is determined according to the Stefan-Boltzmann law:

$$R_{gu} = \epsilon_s \sigma T_s^4. \quad (2.18)$$

2.6 Comparisons with Observations

Section 3.2 discusses results of model runs initialized with data from NOAA’s National Virtual Data System, where soundings are collected daily at 12Z and 00Z (6 am and 6 pm Central time). Since we are concerned with the evolution of the boundary layer throughout the day, we would like to evaluate the performance of the model by comparing the vertical structure of the model to observations as the BL grows, not simply 12 hours after initialization. These sounding data do not allow for such validation. Data from the Flatland Boundary Layer Experiment, however, are suitable for some verification of the model’s performance.

The Flatland Boundary Layer Experiment (Angevine et al., 1998) ran during portions of the summers of 1995 and 1996 in central Illinois. On every day of these campaigns, radiosondes were released at approximately local noon. Some days included more detailed radiosonde coverage, usually with balloon releases at 9 am, 10:30 am, Noon, and 1:30 pm. Fifteen days of the Flatland experiment had the early morning soundings necessary for model initialization: these were the days used to confirm that the model predictions of boundary layer properties were within the realm of observation. There were additional days with 9 am soundings that were not suitable for model initialization because the near-surface air was already warmer than the mixed-layer. Because the model does not represent the near-surface layer, convection was immediately triggered from the warm near-surface layer. (This is not a problem with the NOAA data used in the bulk of this work, since this superadiabatic surface layer rarely develops by 6 am.) On the 15 days when this near-surface layer did
Figure 2.2: Model-calculated boundary layer height, potential temperature, and humidity (solid lines) for the model run initialized with the Flatland sounding from 9 am, 06 August 1996. Soil saturation is 20%. Crosses are observations from the Flatland BL profiler; stars are observations from the four Flatland radiosondes launched on this day.

not lead to immediate triggering, the observed mixed layer properties fell within the realm of the properties obtained from the extremely wet and extremely dry soil model runs. The example from 06 August 1996 presented in Figure 2.2 is one particularly fine example of a good match. The model run with soil saturation set at 20% matches the observations of boundary layer height, temperature and humidity quite well at the four sounding times. Observed soil saturation in the top 5 cm was approximately 20% on this day, which was eight days into a drying cycle (eight days after a rainfall event). Also shown in Figure 2.2 are the 30 minute averages of BL height measured by a BL wind profiler.

Figure 2.3 shows one of the poorer fits of model results to observations. Though the model run with 20% soil saturation matches observed BL height and potential temperature, the modeled humidity is about 1 g/kg too high. The model run with 100% soil saturation (not shown) performs more poorly: humidity is over-predicted by about 2.5 g/kg, while
the BL height and potential temperature are under-predicted. The other comparisons with Flatland data are more like to the good performance shown in Figure 2.2.

Though the model does a good job in matching the boundary layer height, temperature and humidity, as well as the net radiation (not shown), the evaporative fraction is not so easily matched. Data from other field experiments (e.g., FIFE: Betts and Ball, 1995) show some correspondence between soil moisture and Bowen ratio or evaporative fraction, but the Flatland data do not display a simple relationship. The drying and rewetting episodes at Flatland are clearly evidenced in the soil moisture data, while the evaporative fraction responds in a more complicated manner. This seems to be largely due to the fact that the soil moisture data is collected from the top 5 cm only, while evapotranspiration clearly occurs from plant roots at significantly deeper depths. The surface soil moisture is high immediately after a rainfall event, while the evaporative fraction remains low until a day or two after a rainfall event and continues to rise for at least two more days. This complicated behavior is very difficult to match with the simple boundary layer model used in this work. However, the
example presented above and the other days of intensive observation show that the model is capable of adequately representing the conditions of BL growth observed in Illinois. We did not tune the model to replicate observed days since our intent is not to use this model as a predictive tool. Rather, given the adequate representation of BL height, temperature, and humidity seen on days of varying soil moisture levels, we proceed with our investigation of the role of soil moisture in BL growth and development in different atmospheric settings. The insights gained from this work will be taken into further work with a more realistic three-dimensional model.
Chapter 3

One-dimensional boundary layer modeling work

The one-dimensional boundary layer model described in Chapter 2 was used with three summers worth of data (June-August, 1997-99) from a NOAA radiosonde station located in Lincoln, Illinois. This station, ILX, is not far from the Flatland experimental site mentioned in Section 2.6. NOAA’s National Virtual Data System (NVDS) consists of 76 stations across the continental US, with daily 12Z and 00Z (6 am and 6 pm Illinois time) radiosonde launches. Station ILX is the only station in Illinois.

Of the 273 days during the three summers, 225 were available for use for model initialization. The 48 other days were either missing from the NVDS database, or already showed rain or heavy cloud cover at 6 am. Each of the 225 valid cases was used to initialize two model runs: one with very dry soils (soil saturation set to 20%) and one with very wet soils (soil saturation set to 100%). Given the three possible model outcomes explained in Chapter 2, the results were divided into four main categories: rain over both wet and dry soils, shallow clouds over both, no convection over either, and cases where different outcomes resulted over different soil conditions. The first three categories are all situations where the partitioning of fluxes at the land surface did not influence the convective potential of the atmosphere: these are called atmospherically controlled cases. Cases in the fourth category are non-atmospherically controlled: these are the cases where the land surface moisture
condition has the potential to determine whether or not convection is triggered.

In this chapter, we attempt to find common properties of the early-morning atmospheres within each of the four outcome categories. More importantly, within the non-atmospherically controlled cases, we strive to find a way to distinguish days where rainfall is more likely to occur over wet soils from those where rainfall is more likely to occur over dry soils. Early work with many commonly used measures of atmospheric stability, humidity, and energy content proved less effective than the newly developed measure of the Convective Triggering Potential, \( CTP \). The \( CTP \) is defined in the next section. Section 3.2 describes results of the modeling work using Illinois soundings, focusing first on atmospherically controlled cases, then on the cases where soil moisture influences the model outcome. Additionally, the performance of the \( CTP \) and other measures will be assessed in terms of their ability to distinguish days according to the influence that soil moisture can have on convection. Next, we will give a brief presentation of results of some sensitivity experiments with altered radiative conditions. Following a discussion of why the \( CTP \) is a strong indicator of the influence of soil moisture in different atmospheric settings, we conclude this 1D work, setting the stage for bringing the \( CTP-HI_{low} \) framework to the results of 3D work presented in the following chapters.

### 3.1 The Convective Triggering Potential

The hypothesis that prompted this work was that certain atmospheric conditions favor rainfall triggering over wet soils (positive soil moisture-rainfall feedback; Findell and Eltahir, 1999, 1997), while other atmospheric conditions favor rainfall triggering over dry soils (negative feedback). Our intent was to determine the differences between these initial atmospheric settings and their frequency of occurrence. The properties of the early morning soundings used to initialize the boundary layer model were analyzed to determine the interplay between atmospheric and soil moisture initial conditions. A number of stability indices have been in use for many years in thunderstorm and weather prediction. As Mueller et al. (1993) report and this work confirms, these traditional stability indices are helpful in ruling out the possibility of rain in very stable atmospheric conditions, but when instability is indicated,
they give no further clues of where and when—or even if—convection might be triggered. Definitions of the many traditional indices and parameters used in this work are given in Appendix A.

We will now define the Convective Triggering Potential (CTP), then discuss the CTP results with the Illinois soundings, showing that when coupled with a measure of low-level humidity, the CTP can be used to assess the potential for soil moisture to influence convection. Finally, we will discuss the implications of the CTP on our understanding of soil moisture-rainfall feedbacks.

3.1.1 Definition of the CTP

An early-morning atmospheric profile can be broken down into three basic zones (Figure 3.1):

- the near-surface zone which is sure to be incorporated into the day’s boundary layer (order 75-100 mb, or 1 km),
- the free atmosphere which is sure to be untouched by the day’s BL (beginning about 300 mb or 3 km above the surface),
- the zone between these two layers: its incorporation into the growing BL depends on both the surface fluxes and the temperature lapse rate of the profile in this region.

As shown in Figure 3.1, the CTP focuses on this middle zone. The value of the CTP is determined by integrating the area between the environmental temperature profile and a moist adiabat drawn upward from the observed temperature 100 mb above the surface to a point 300 mb above the surface. (Since surface pressure in Illinois is commonly close to 1000 mb [usually in the 990’s], we will often present this critical CTP region as between 900 and 700 mb, as noted in the figure.) As previously mentioned, many of the traditional measures do a good job of identifying atmospheric conditions that are highly stable, but they are not so effective in identifying conditions likely to rain when the index suggests instability. This is demonstrated for the Showalter stability index (SI, Showalter, 1953) in Figure 3.2. The SI is obtained by lifting a parcel dry adiabatically from 850 mb until it reaches its Lifting Condensation Level (LCL), and then moist adiabatically until 500 mb (see Appendix A). Since the stability index only includes information from 850 and 500
Convective Triggering Potential

Free Atmosphere:
Almost never incorporated into BL

CRITICAL REGION FOR TRIGGERING OVER DRY SOILS!!! LAPSE RATE CRUCIAL!

Almost always incorporated into BL

Figure 3.1: A sketch of the definition of the Convective Triggering Potential on a thermodynamic diagram. Thick solid lines are the temperature and dew point temperature profiles; straight long-dashed line is a dry adiabat (constant potential temperature); straight short-dashed line is constant temperature; straight dotted line is constant mixing ratio; curved short-dashed line is a moist adiabat (constant equivalent potential temperature). The CTP is determined by integrating the area between the observed temperature sounding and a moist adiabat originating at the observed temperature 100 mb above the surface. The top is bounded by a constant pressure line 300 mb above the surface.

mb, it cannot distinguish between soundings with identical $T_{500}$, $T_{850}$, and $T_{d,850}$ but different temperature characteristics between these two pressure levels. The CTP, on the other hand, is a physically-based measure of the temperature profile throughout its definition region, not just at the top and bottom (Figure 3.3).

Note that the CTP can be negative if the temperature of the moist adiabat originating from the $P_{surf} - 100$ mb level is less than the observed temperatures. Also, the CTP will be zero if the observed profile is moist adiabatic above the point of origin.

The CTP includes information about the temperature profile throughout the critical CTP region, but it tells us nothing about the humidity in the troposphere. Humidity is also a crucial component of convective potential, however, and must be assessed in order to determine the potential for rainfall. Lytinska et al. (1976) defined a humidity index which is designed to distinguish atmospheres that were too dry for rainfall, from those where rainfall was possible. The original definition of the humidity index is the sum of the dew point depressions at 850 mb, 700 mb, and 500 mb:

$$HI = (T_{850} - T_{d,850}) + (T_{700} - T_{d,700}) + (T_{500} - T_{d,500}),$$

(3.1)
Figure 3.2: Two soundings that would have identical values of the Showalter stability index ($SI$), despite their radically different temperature characteristics between 850 and 500 mb.

Figure 3.3: Schematic of the $CTP$ for the two soundings with identical values of the stability index from Figure 3.2. The $CTP$ is the area between the environmental temperature profile and a moist adiabat drawn upward from the observed temperature 100 mb above the surface to 300 mb above the surface.
where $T_p$ is the temperature at pressure level $p$ and $T_{d,p}$ is the dew point temperature at pressure level $p$. Though this index was indeed somewhat helpful in distinguishing between very dry and very humid atmospheres, the 500 mb information included in this index is generally beyond the reach of typical boundary layer growth, and is therefore not relevant for this work. Other combinations of dew-point depressions at levels below 500 mb all prove to be helpful in assessing the convective potential of Illinois soundings. The most effective is the sum of the dew point depressions at 950 mb and 850 mb:

$$HI_{low} = (T_{950} - T_{d,950}) + (T_{850} - T_{d,850}).$$

(3.2)

$HI_{low}$ will be used extensively throughout the rest of this thesis. Lytinska et al. (1976) suggested as threshold for rain $HI \leq 30^\circ C$. The threshold for $HI_{low}$ is $15^\circ C$ for the Illinois data.

In the next section, we isolate the results according to the response of the model to the two different prescribed soil moisture conditions, and show that the $CTP$ and $HI_{low}$ are more effective at separating cases according to model response to soil moisture than any pairs of the traditional indices.

### 3.2 Results from Illinois Soundings

As stated above, the model was initialized with soundings from all of the available days from the summers of 1997-99 at station ILX in Lincoln, Illinois. For each available sounding, a wet soil and a dry soil simulation were performed, using soil saturations of 100% and 20%, respectively. The final tally of model outcomes over these two different land surface conditions is shown in Figure 3.4. The 225 days are those from June, July, and August of the three years studied for which data was available, rainfall was not already occurring, and cloud cover was not present at the time of the sounding. In general, convection is more likely over wet soils: 39% of initial soundings lead to convection over wet soils, as opposed to 27% over dry soils. Furthermore, rain is likely 22% of the time over wet soils, but only 13% over dry soils. The rest of the days (60% over wet soils, 72% over dry) result in the
model reaching the end of the day with no triggering of convection.

Our primary concern is determining how soil moisture-boundary layer interactions behave in different atmospheric settings. We wish to discover which initial atmospheres lead to different results over dry soils than over wet soils. Figure 3.5 divides the results into four possible combinations of outcomes. Both soil conditions lead to the same outcome 72% of the time (11% both rain, 6% both have shallow clouds, 55% neither convect), and different outcomes 28% of the time. In Section 3.2.1 we will briefly discuss the atmospheric conditions that predominate on days when the model results are the same over wet and dry soils. The focus of this work, however, lies in Section 3.2.2, where we discuss the cases where the soil moisture condition changes the final outcome of the model.

### 3.2.1 Atmospherically Controlled Outcome

In this section we discuss the cases where the model outcome was unaffected by the land surface condition. As a first approximation, boundary layer dynamics and the potential for rainfall on these days are assumed to be atmospherically controlled. The $CTP$ and the modified humidity index $HI_{low}$ together do an excellent job of stratifying the atmospherically controlled cases (Figure 3.6). Figure 3.7 shows results plotted according to the traditional stability indices discussed earlier and defined in Appendix A.

In Figure 3.8 and Table 3.1 we attempt to show the relative ability of each of nine different combinations of measures (e.g., $CTP$ and $HI_{low}$) to separate the three clusters of data (rainy...
Figure 3.5: The outcome of about three-quarters of the 225 available days is unaffected by the soil condition; 25% have different results over the dry and the wet soils.

Figure 3.6: Values of the $CTP$ and $HI_{low}$ for days when outcomes of dry soil and wet soil model runs are the same (rain over both, shallow clouds over both, no convection over either).
Figure 3.7: Values of some traditional stability indicies when outcomes of dry soil and wet soil model runs are the same (rain over both, shallow clouds over both, no convection over either). Labels as in Figure 3.6.

days, R; shallow cloud days, SC; days with no convection, N). The data in each of these plots have been normalized according to the full-sample population means and standard deviations for each measure. For example, the upper right plot of Figure 3.8 is the same data as that shown in Figure 3.6, but the $CTP$ and $HI_{low}$ data have each been normalized according to

$$CTP_{Normalized} = \frac{CTP_{orig} - \mu_{CTP}}{\sigma_{CTP}}.$$  \hfill (3.3)

The mean ($\mu_{CTP}$) and standard deviation ($\sigma_{CTP}$) have been determined from the full samples (all 225 days of valid data), such that the both the atmospherically and non-atmospherically
controlled cases are normalized with the same values. The normalization procedure is the same for all of the variables used in Figure 3.8. For each of the three clusters (R, SC, N), the centroid of the cluster distribution is determined, as are the standard deviations in each of the x and y directions. These are indicated on Figure 3.8 with red lines and arrows from each group centroid. Greater separation distances indicates greater separation between group centroids. The sum of the three separation distances (from group R to group N, from group R to group SC, and from group SC to group N) are given in column 2 of Table 3.1. This value is largest (ranked 1) for the CAPE – CIN-HIlow combination of measures, but is closely followed by the deep convective index with HIlow, the K index with HIlow, and then the CTP with HIlow. There is a larger gap between these measures and the five pairs that follow.

In addition to the spread between cluster centers, we also wish to have a way of quantifying within-cluster spread, and the amount of cluster overlap. The standard deviations of the distances between all the elements within a cluster is one measure of within-cluster spread (this is related to the Mahanalobis $D^2$ [Kendall and Stuart, 1968]), as is the standard deviation of distances between each element of a cluster and its own centroid. Columns 3 and 4 of Table 3.1 list the three-cluster sums of these measures for each of the nine pairs.

Significant within-cluster spread, however, does not necessarily indicate poor performance of a measure. For example, the large spread of humidity deficits in cases with no rain is simply an indication that an $HI_{low}$ of 40°C is just as effective as an $HI_{low}$ of 60°C at prohibiting convection. Of greater interest as an indicator of the performance of these measures is the frequency of mis-categorized days. This, however, requires defining threshold values separating the groups (rain expected, shallow clouds expected, no convection expected) for each of the ten measures considered. A simpler technique is to calculate the percentage of cases falling closer to the centroid of another cluster than to the centroid of their own cluster, as in column 5 of Table 3.1. The pair with the fewest mis-categorized days, the $K$ index with $HI_{low}$, receives top rank by this measure of cluster overlap. Note that although CAPE–CIN showed the greatest centroid separation, it ranks tied for sixth by this measure of variability.

The final column of Table 3.1 sums the ranks of the two relevant measures, with lower ranks indicative of better performance. It shows that $K$, CTP, DCI, and CAPE – CIN,
Atmospherically Controlled Cases

<table>
<thead>
<tr>
<th></th>
<th>Sum of centroid separations (rank)</th>
<th>Sum of ( \sigma_{\text{Mahan}} )</th>
<th>Sum of ( \sigma_{\text{dist centroid}} )</th>
<th>% closer to another centroid (rank)</th>
<th>Sum of ranks columns 1+4</th>
</tr>
</thead>
<tbody>
<tr>
<td>( K )</td>
<td>4.04 (3)</td>
<td>2.70</td>
<td>1.32</td>
<td>31.1% (1)</td>
<td>4</td>
</tr>
<tr>
<td>( CTP-HI_{\text{low}} )</td>
<td>3.96 (4)</td>
<td>2.96</td>
<td>1.58</td>
<td>32.5% (2)</td>
<td>6</td>
</tr>
<tr>
<td>( DCI )</td>
<td>4.23 (2)</td>
<td>2.73</td>
<td>1.42</td>
<td>35.8% (4.5)</td>
<td>6.5</td>
</tr>
<tr>
<td>( CAPECIN )</td>
<td>4.24 (1)</td>
<td>2.89</td>
<td>1.52</td>
<td>36.4% (6.5)</td>
<td>7.5</td>
</tr>
<tr>
<td>( SI )</td>
<td>3.60 (7)</td>
<td>2.89</td>
<td>1.57</td>
<td>35.8% (4.5)</td>
<td>11.5</td>
</tr>
<tr>
<td>( TT )</td>
<td>3.74 (5)</td>
<td>2.83</td>
<td>1.53</td>
<td>36.4% (6.5)</td>
<td>11.5</td>
</tr>
<tr>
<td>( SLI )</td>
<td>3.47 (9)</td>
<td>2.73</td>
<td>1.55</td>
<td>35.1% (3)</td>
<td>12</td>
</tr>
<tr>
<td>( LI )</td>
<td>3.66 (6)</td>
<td>2.72</td>
<td>1.45</td>
<td>40.4% (8)</td>
<td>14</td>
</tr>
<tr>
<td>( CTP-HI_{\text{orig}} )</td>
<td>3.53 (8)</td>
<td>2.77</td>
<td>1.65</td>
<td>41.7% (9)</td>
<td>17</td>
</tr>
</tbody>
</table>

Table 3.1: Measures of group separation of atmospherically controlled data. The three groups are rain over both wet and dry soils, shallow clouds over both, and no convection over either. Except for the \( CTP-HI_{\text{orig}} \) pair, all listed atmospheric measures are paired with \( HI_{\text{low}} \), as in Figure 3.8. Greater sum of centroid separations (column 1) indicates better separation of the three groups. The rank of 1 is given to the pair with the greatest value in this column. Conversely, minimum overlap between clusters is indicative of better performance, and the rank of 1 is given to the pair with the smallest percentage in column 5. The best performance in terms of both maximum cluster separation and minimum overlap is indicated by the smallest ranks in column 6.

when each is coupled with \( HI_{\text{low}} \), perform quite similarly, and out-perform the other five combinations. The non-atmospherically controlled cases presented in Section 3.2.2, however, will demonstrate that \( CTP-HI_{\text{low}} \) greatly out-perform the other eight combinations, giving a net superiority to this combination.

Both wet and dry soils lead to rain

There are 25 cases during the summers of 1997-1999 at the Lincoln, Illinois station when both extremely wet and extremely dry soils result in deep convection in the model. These days are labelled atmospherically controlled because deep convection results over either soil condition. Note that the phase “atmospherically controlled is used in two different contexts. When describing model results, it is used to indicate the cases where the model outcome is the same over both wet and dry soils (e.g., for a given initial sounding, rainfall occurs in the model runs over both wet and dry soils). Later, this phrase will be used in a predictive
Figure 3.8: Measures of cluster separation for nine combinations of variables when outcomes of dry soil and wet soil model runs are the same. Black x: no convection over either soil type; green diamond: shallow clouds over both; blue star: rain over both. Each group centroid and ± one standard deviation in each direction are indicated in red. Variables are normalized using the mean and standard deviation of all cases (atmospherically and non-atmospherically controlled), such that the normalization is identical for this figure and for Figure 3.13.
Figure 3.9: Average ± one standard deviation of (a) free convection triggering time, (b) dew point depression, (c) precipitable water, (d) $CAPE$, (e) $\theta_E$, and (f) depth of convection, for the 31 instances when deep convection (likely to rain) is triggered in both the saturated and the dry soil runs.

It is not entirely clear how to describe atmospheric conditions where the land surface fluxes cannot impact the potential for convection.

Despite the atmospherically controlled label applied to the 25 cases where rainfall is triggered over both wet and dry soils, Figure 3.9 shows that the properties of the boundary layer at the time of convective triggering are significantly different over soils of different moisture content.

The anticipated result of higher soil moisture leading to higher boundary layer $\theta_E$ is indeed noticeable, with a 5.4°C difference being significant at the $\alpha = 0.0375$ level. Accompanying these higher $\theta_E$ values come larger $CAPE$s (an 850 J/kg difference), deeper convection depths (a 1.18 km difference), and smaller dew point depressions (a 4.2°C difference).
ference), all significant at the $\alpha = 0.0015$ level. (It should be noted that the convection depths are sometimes underestimated, as the level of neutral buoyancy sometimes exceeds the sounding top, particularly over wet soils. Correction for this underestimate would make the difference between the mean convection depths even more highly significant.) Each of these differences in the mean properties is a direct result of the higher evaporative fraction (lower Bowen ratio) over wet soils leading to lower boundary layer temperatures, higher specific humidities, lower boundary layer heights, and less entrainment. The differences between the mean triggering times and the mean precipitable water in the entire column are not statistically significant.

From these results, we conclude that even when the occurrence of rainfall is atmospherically controlled, the land surface moisture condition can indeed impact the depth of rain. This is supported by the studies of Williams and Renno (1993) and Eltahir and Pal (1996): Williams and Renno (1993) demonstrated that $CAPE$ tends to be linear and close to zero below some threshold temperature value, while above this threshold there is a $\sim 1000 \text{ J/kg}^\circ\text{C}$ slope of increasing $CAPE$ with increasing $\theta_w$ (wet bulb potential temperature: a measure of moist static energy, like $\theta_e$). Eltahir and Pal (1996) also found this threshold behavior, and further showed that above this threshold, $CAPE$ is linearly correlated with rainfall depth. This suggests a positive feedback mechanism between soil moisture and the depth of rainfall. This result is consistent with the work of Findell and Eltahir (1997), who showed that late spring/early summer large-scale moisture conditions are positively correlated with the total rainfall depth over the course of the summer in Illinois.

Of the 25 days that lead to rainfall over both soil moisture conditions, three types of initial soundings occur:

- soundings that are close to saturation from the surface up to high levels and have little to no surface inversion;

- soundings that have a surface inversion to overcome, but are then able to freely convect; and

- soundings that are dominated by a warm air mass near the surface which must be overcome before convection can occur. They may or may not have a small surface inversion leading into this warm air mass.
All of the 25 soundings have a $CTP$ between 0 and \( \sim 230 \text{ J/kg} \) (only three have $CTP > 200 \text{ J/kg}$, and only three have $CTP < 80 \text{ J/kg}$), and all but two have $HI_{low} < 10^\circ \text{C}$.

**Both wet and dry soils lead to shallow clouds**

Thirteen of the 225 cases explored from the summers of 1997-1999 lead to the formation of shallow clouds over both wet and dry soils. Of these 13 cases, eight are initial soundings with a warm and dry air mass at upper levels that prevents deep convection. The other five are cases where the initial sounding is nearly moist adiabatic essentially all the way up from near the ground surface. In these cases, significant $CAPE$ cannot form before free convection is triggered.

Seven of the 13 cases have $CTP < 0 \text{ J/kg}$ and $HI_{low} < 10^\circ \text{C}$. Five of the remaining six cases are in what appears to be a transition zone: $0 < CTP < 200 \text{ J/kg}$ and $9 < HI_{low} < 15^\circ \text{C}$. There is one outlier with an $HI_{low}$ of almost $20^\circ \text{C}$ and a $CTP$ slightly under $-200 \text{ J/kg}$.

**No convection over wet or dry soils**

A total of 124 of the cases investigated led to no convection over either dry or saturated soils. About a third of these are initial soundings that are warm and dry all the way up from the surface. They typically have a small surface inversion that is capped by air that is close to moist adiabatic up to very high levels. About a sixth of the cases are initial soundings with a warm and dry air mass that intrudes at low levels (around 900 mb), but is distinct from the near-surface air. As with the cases that are warm and dry from the surface, this intruding mass prevents both the buildup of any $CAPE$ or the triggering of free convection. In both categories, the LFC is often undefined, since the path of a parcel from the BL is always less buoyant than the warm air of the overlying mass. The other half of the cases are energetically controlled: large $CAPE$s can build up (particularly over moist soils), but a barrier of warm air prevents convection in either soil moisture scenario.

The modified humidity index does an excellent job of screening out the cases where convection is limited by excessive aridity. When $HI_{low} > 15^\circ \text{C}$, there is not enough low-level humidity to allow for rainfall or shallow clouds. An $HI_{low}$ between 10 and 15$^\circ \text{C}$ is in a
transition region where any of the three outcomes is possible, but no convection is likely over either wet or dry soils. Cases with a sounding is too stable for rainfall to occur are well classified by either a $CTP < 0 \text{ J/kg}$ or an $SI > 0 \text{ K}$.

### 3.2.2 Soil Moisture Determines Outcome

Figure 3.10 shows the occurrences of the possible pairs of model outcomes when different soil moisture conditions led to different model results. (Note that there are no instances of shallow clouds forming over wet soils with rain over dry.) Figure 3.11 shows that these data are fairly well stratified into three groups in $CTP-HI_{low}$ space: a dry soil advantage regime (rain or shallow clouds only over dry soils), a wet soil advantage regime (rain only over wet soils), and a group with shallow clouds only over wet soils. The wet soil advantage cases (blue asterisk) all have $CTP \leq 180 \text{ J/kg}$ and most have $HI_{low} \leq 12 \text{ K}$. All but one of the cases with rain over dry soils but not over wet (red asterisk) has $CTP \geq 180 \text{ J/kg}$, and all but one has $HI_{low} \geq 12 \text{ K}$. In Figure 3.12 we see the performance of some of the other combinations of measures, while Figure 3.13 and Table 3.2 show that these other combinations of indices do not do as good a job of separating out the different responses to soil moisture conditions.

The format of Figure 3.13 and Table 3.2 are the same as Figure 3.8 and Table 3.1, but for the three clusters in these non-atmospherically controlled cases. As mentioned, the $CTP$ is
Dry soil advantage:
- Rain over dry soils
- SC over dry soils

Wet soil advantage:
- Rain over wet soils
- SC over wet soils

Figure 3.11: Values of the CTP and $H_{I_{low}}$ with model outcomes when different soil conditions led to different results. (One shallow cloud over wet soils outlier is removed: $CTP \approx -540 \ J/kg$, $H_{I_{low}} \approx 50 \ K$.) The red oval encloses the area where dry soils have an advantage for convective triggering; the blue oval encloses the area where wet soils have the advantage.

Non-atmospherically Controlled Cases

<table>
<thead>
<tr>
<th>Non-atmospherically Controlled Variables</th>
<th>Sum of centroid separations (rank)</th>
<th>Sum of $\sigma_{Mahalanobis}$</th>
<th>Sum of $\sigma_{dist centres}$</th>
<th>$%$ closer to another centroid (rank)</th>
<th>Sum of ranks columns 1+4</th>
</tr>
</thead>
<tbody>
<tr>
<td>$CTP-H_{I_{low}}$</td>
<td>3.55 (1)</td>
<td>2.80</td>
<td>1.22</td>
<td>25.5% (1)</td>
<td>2</td>
</tr>
<tr>
<td>$CTP-H_{I_{orig}}$</td>
<td>3.46 (2)</td>
<td>2.65</td>
<td>1.72</td>
<td>32.7% (2.5)</td>
<td>4.5</td>
</tr>
<tr>
<td>$DCI$</td>
<td>2.43 (3)</td>
<td>3.10</td>
<td>1.80</td>
<td>40.0% (5.5)</td>
<td>8.5</td>
</tr>
<tr>
<td>$CAPECIN$</td>
<td>1.72 (5)</td>
<td>2.95</td>
<td>1.30</td>
<td>38.2% (4)</td>
<td>9</td>
</tr>
<tr>
<td>$TT$</td>
<td>1.66 (8)</td>
<td>2.98</td>
<td>1.40</td>
<td>32.7% (2.5)</td>
<td>10.5</td>
</tr>
<tr>
<td>$SI$</td>
<td>1.77 (4)</td>
<td>2.93</td>
<td>1.25</td>
<td>41.8% (7)</td>
<td>11</td>
</tr>
<tr>
<td>$SLI$</td>
<td>1.66 (7)</td>
<td>2.99</td>
<td>1.39</td>
<td>40.0% (5.5)</td>
<td>12.5</td>
</tr>
<tr>
<td>$LI$</td>
<td>1.68 (6)</td>
<td>2.85</td>
<td>1.29</td>
<td>49.1% (9)</td>
<td>15</td>
</tr>
<tr>
<td>$K$</td>
<td>1.56 (9)</td>
<td>3.07</td>
<td>1.80</td>
<td>43.6% (8)</td>
<td>17</td>
</tr>
</tbody>
</table>

Table 3.2: Measures of group separation of non-atmospherically controlled data. The three groups are dry soil advantage, wet soil advantage (rain over wet soils), and shallow clouds over wet soils. Column details as in Table 3.1.
Figure 3.12: Values of some traditional stability indices with outcomes of dry soil and wet soil model runs. One shallow cloud over wet soils outlier is removed (\(CAPE - CIN \approx -6100 \text{ J/kg}\), \(HI_{low} \approx 50 \text{ K}\), \(SI \approx 26 \text{ K}\), \(LI \approx 2 \text{ K}\), \(K \approx 250 \text{ K}\), \(TT \approx 2 \text{ K}\), \(DCI \approx 532 \text{ K}\)) Labels as in Figure 3.11.

far more effective at separating these three clusters than any of the other measures. Both \(HI_{low}\) and \(HI_{orig}\) are effective humidity measures to couple with the \(CTP\), but \(HI_{low}\) is the better of the two. This is because the original humidity index was defined as the sum of the dew-point depressions at 850 mb, 700 mb, and 500 mb. The 500 mb information included in the definition is not particularly relevant for the BL growth considerations emphasized in this work. Other variations on the original humidity index were also considered, and all were effective when coupled with the \(CTP\), but \(HI_{low}\) was the most consistent of the good performers.
Figure 3.13: Measures of cluster separation for the six best combinations of variables when outcomes of dry soil and wet soil model runs are the same. Blue stars indicate rainfall over wet soils (either shallow clouds or nothing over dry soils); red stars indicate dry soil advantage (either rain over dry with less convection over wet soils); green circles indicate shallow clouds over wet soils (nothing over dry). Each group centroid and ± one standard deviation in each direction are indicated in black. Normalization as in Figure 3.8.
The measure that follows the $CTP$ in Table 3.2 is $CAPE - CIN$, but the drop from the top ranking $CTP-HI_{low}$, with the minimum possible rank of 2, to the rank of 7 given to this pair is quite large. Given the close performance of the top few measures in Table 3.1 and the clearly superior performance of the $CTP$ in Table 3.2, the $CTP$ and $HI_{low}$ are deemed the best indicators of both the potential for soil moisture to influence convection, and the nature of that influence.

**Case studies highlighting the relevance of the $CTP$**

In both the real world and the model world, convection is triggered when the level of free convection (LFC) and the boundary layer top meet. In simplified terms, this can occur when the LFC remains constant and the BL grows up to the LFC, or when the BL height remains constant and the LFC drops to the top of the BL. Obviously many combinations of BL growth and LFC descent can also bring these two levels together for convection. The extremes, however, describe the characteristic manner in which convection is triggered over very dry and very wet soils, respectively. We will now present two case studies highlighting these different methods for triggering convection.

Figure 3.14 shows two initial 6 AM soundings with very different $CTP$ values. These soundings are indicative of the types of initial atmospheric conditions which lead to rain over dry but not over wet soils (Figure 3.14a: 23 July 1999, $CTP = 254$ J/kg), and those which lead to rain over wet but not over dry soils (Figure 3.14b: 03 July 1999, $CTP = 88$ J/kg).

In many model runs, the boundary layer height over wet soils grows slowly but steadily until noon or 2 PM, and then remains relatively constant. The $\theta_E$ continues to grow due to the continued input of moisture from the land surface. Over dry soils, on the other hand, the behavior of these two variables is often reversed: the BL height grows steadily and more rapidly throughout the day, but the $\theta_E$ plateaus or even drops in the afternoon. In the dry soil case, the BL top and the LFC will meet only if the BL grows high enough to reach the LFC; in the wet soil case, this will occur only if the $\theta_E$ grows large enough to bring the LFC down to the BL top.

Consider, for example, the case of 03 July 1999 (Figure 3.14b). The boundary layer
Figure 3.14: Profile of initial conditions for (a) 23 July 1999: $CTP = 254 \text{ J}$, rainfall occurs only over dry soils, and (b) 03 July 1999: $CTP = 88 \text{ J}$, rainfall occurs only over wet soils.
height, the level of free convection (LFC), and $\theta_E$ values for the wet and dry model runs on this day are shown in Figure 3.15, and the soundings from 1:00 pm are shown in Figure 3.16. The dry soil boundary layer continues to grow higher after this point, but the $\theta_E$ remains constant. For this process to trigger convection, the BL must grow from 890 mb to 685 mb (the point where the parcel path crosses the environmental temperature line in Figure 3.16a). In contrast, the wet soil boundary layer grows more gradually than that over the dry soil, but the $\theta_E$ is also increasing. The pseudoadiabats on Figure 3.16b indicate that the wet-bulb potential temperature $\theta_w$ must increase by only $\sim 1^\circ C$ ($\sim 4^\circ C$ in $\theta_E$) in order to bring the LFC down from 855 to 940 mb. Given the steep lapse rate in this particular sounding (the quality which leads to an intermediate CTP), a small increase in $\theta_E$ leads to a large decrease in the LFC. Indeed, when convection is triggered just over 1 hour later, the $\theta_E$ has risen $\sim 2^\circ C$ and the LFC has fallen from 855 to 920 mb.

Over the dry soils, the boundary layer grows only to 840 mb by the end of the day: still 135 mb below the LFC (about 1.5 km in Figure 3.15. The CTP calculated from the initial 6 am profile on this day was 88 J/kg. Recall from Figure 3.11 that the intermediate range of the CTP where convection is limited by the energy (rather than the height) of the boundary layer is $\sim 0-180$ J/kg. Indeed, as anticipated by the CTP, the behavior on this day is representative of an energy-limited BL.

Now consider the case of 23 July 1999 (Figure 3.14a), when the opposite circumstances occur. Figure 3.17 shows that the growth of the BL is much slower over the wet soils than over the dry. The rapid growth over dry soils between 11 am and noon was due to the easy entrainment of neutrally buoyant air between 950 and 840 mb (see initial sounding, Figure 3.14a). It is the presence of this dry adiabatic portion of the sounding within the CTP region that yields a high CTP. At about noon the $\theta_E$ in the dry soil case levels out. For convection to be triggered over the dry soils by growth of the BL at constant $\theta_E$, the BL must rise from 850 to 770 mb (Figure 3.18a). The BL does continue to grow after 12:30 pm, and convection is triggered at 1:00 pm.

By 12:30 pm, 601 J/kg of CAPE are already trapped in the moist boundary layer shown in Figure 3.18b. For convection to be triggered over the wet soils by increasing $\theta_E$ at constant BL height, the $\theta_E$ must increase by $\sim 10$ K ($\sim 2.5$ K in $\theta_w$). At 6 pm, just before the BL
collapses at the end of the day, the $\theta_E$ has increased by almost 4 K and the CAPE has increased to over 3700 J/kg. This very large amount of energy cannot be released, for the BL is still a few degrees shy of the $\theta_E$ necessary for triggering convection in this scenario. The CTP calculated from the 6 am initial sounding on 23 July 1999 was 254 J/kg, well above the 180 J/kg threshold seen in the data of Figure 3.11.

These two cases are indicative of circumstances leading to rain over different land surfaces; shallow clouds, however, can form when the $CTP > 0$ J/kg, and often form when the $CTP < 0$ J/kg. When a warm and often dry air mass intrudes into a sounding, typically ~200 to 300 mb above the surface, the $CTP$ is greatly reduced, often becoming negative. In many cases, convection is triggered but then blocked by these masses, allowing only for
Figure 3.16: Profile of model runs at 1:00 pm in (a) the dry soil run, and (b) the wet soil run on 03 July 1999.
the formation of shallow clouds. This is more likely over wet soils than over dry (green circles in Figure 3.11) because the BL over dry soils will often grow right up to the base of the intruding air mass prior to any convective triggering. The shallow BLs over wet soils, on the other hand, often allow for the triggering of convection below the intruding mass, such that shallow clouds form between the triggering level and the base of the intrusion. In cases where rainfall occurs over wet soils, but shallow clouds form over dry (blue asterix with yellow circles in Figure 3.11), there is often a slightly warm air mass protruding only far enough to block the freely convecting air from the boundary layer over the dry soils (and to keep the $CTP$ below $\sim 180$ J/kg). The $\theta_E$ over the wet soils is large enough at the time that convection is triggered that it is not blocked by this relatively warm air mass.
Figure 3.18: Profile of model runs at 12:30 pm in (a) the dry soil run, and (b) the wet soil run on 23 July 1999.
Though the \( CTP \) provides us with a great deal of information, large humidity deficits, particularly in the \( \sim 200 \) mb closest to the surface, can prevent convection of any kind. When \( HI_{low} > 15 \text{ K} \), no convection is likely to occur, regardless of the \( CTP \) value.

### 3.3 Sensitivity of Results to Radiative Conditions

In order to extend the \( CTP-HI_{low} \) framework to stations beyond Illinois, it is necessary to verify that the results are not highly sensitive to radiative conditions. Two sets of experiments using the 1D model were run in order to address this issue. In both sets, the same Illinois soundings (though only from the summer of 1997) were used, but the radiative conditions were changed from the default latitude of 40.0°N and the default mid-summer Julian Day 210 (July 29th). In the first set of experiments, the Julian Day is kept at 210, but the latitude is changed to 26.0°N: just above to the southern-most tips of Florida and Texas. In the second set of experiments, the latitude is again set to 40.0°N, but the Julian Day is set to 258 (September 15th). The effect of the latitude change is to increase both the incoming short wave and the net radiation by approximately 10% at noontime. The effect of the change of seasons is to reduce both the incoming short wave and the net radiation by approximately 20% at noontime.

The increased radiation caused by changing the location used in conjunction with the Illinois data yields only minor changes in the 1D model results. Indeed, of 67 days of available and non-rainy 6 AM soundings from the summer of 1997, there are no additional cases of deep convection over either the wet or the dry soils. There was a reduction in shallow cloud events with increased radiation: 15 versus 13 times over wet soils, and 13 versus 12 times over dry soils. As previously mentioned, many of the shallow cloud events—particularly over wet soils—occur below an inversion. With increased net radiation, the boundary layer could grow up to the base of the inversion before reaching the level of free convection. Overall, the significant change of 14° latitude yielded only minor changes in model outcome.

The second set of experiments, changing the season from mid-summer to early-fall, led to a 20% decrease in noontime net radiation for all of the model runs. In this case, there were significant changes in the number of convective events, but a favoring of convection over
wet soils was evident. While 21 cases led to rain over wet soils in the control runs for the summer of 1997, only 11 did so in these reduced-radiation runs. Over dry soils, the number was reduced from 18 to 6. There were a few more incidences of shallow clouds over wet soils in these runs as compared to the control runs: 18 versus 15. This is caused by the same mechanism which led to fewer shallow cloud events in the increased radiation experiments: reduced radiation allows from more likely triggering of shallow clouds below inversions. Over dry soils, however, there were 3 fewer shallow cloud events than in the control runs (10 versus 13). These three shallow cloud cases in the control runs were the style with nearly moist adiabatic profiles above the triggering level, leading to $CAPE < 400 \text{ J/kg}$ and/or convection depths $(\text{LNB-LFC}) < 5 \text{ km}$. In the reduced radiation runs, the growing boundary layers are not able to reach these same triggering conditions.

Only three runs with $HI_{low} > 9^\circ \text{C}$, and only two with $CTP > 200 \text{ J/kg}$ produce any rain at all. For the higher $CTP$ and/or $HI_{low}$ cases, more available energy is needed to trigger rainfall. Though the number of rainy events falls by about 50% in these early-fall runs, the higher percentage of convective events over wet soils (15% deep convection, 25% shallow clouds) as opposed to dry soils (8% deep convection, 14% shallow clouds) suggests that the small but significant positive feedback mechanism is active even at these reduced radiation levels.

### 3.4 Discussion

The cases presented in Section 3.2 highlight the significance of the Convective Triggering Potential. They also allude to the many pieces of information about the temperature profile that are included in the $CTP$. The discussion above focused primarily on the ability of the BL over dry soils to grow from levels around 900 mb to levels between 700 and 800 mb when the surface pressure was around 1000 mb. As shown in Figure 3.3, the temperature lapse rate between 100 and 300 mb above the surface is the central aspect of the $CTP$. We will refer to this portion of the sounding as the critical $CTP$ region. This region is important for the $CTP$ because the bottom boundary ($P_{surf} - 100 \text{ mb}$, or $\sim 900 \text{ mb}$ in Illinois: about 1 km, in general), separates the levels from which wet and dry soils typically convect (wet
soils from levels near or below this, dry soils from levels above). Within this critical region
the temperature lapse rate is important for determining the ease with which entrainment,
and therefore BL growth, can occur. A great deal of information about the likelihood for
deep convection is provided by this region.

A high lapse rate—close to dry adiabatic—in the critical $CTP$ region, yields a high $CTP$.
When the lapse rate is close to dry adiabatic, air is neutrally buoyant and therefore easy to
entrain, suggesting that the BL and LFC could easily be brought together in areas of high
sensible heat flux. Since the BL over moist soils rarely grows deeper than 100 or 150 mb,
a dry adiabatic lapse rate in the $CTP$ region is advantageous only for the high boundary
layers over dry soils. When the lapse rate in this region is intermediate, the $CTP$ is also
intermediate. Entrainment is more difficult than with a neutrally buoyant atmosphere, so
the dry soils no longer have a great advantage. Additionally, a small increase in $\theta_E$ can
produce a large decrease in the LFC height when the lapse rate is close to moist adiabatic.
Thus, areas of high latent heat flux have an advantage for triggering convection in these
circumstances. When the $CTP$ is near zero, no energy is contained in the sounding and
if convection is triggered, it will not be deep. And finally, a negative lapse rate yields a
negative $CTP$, which indicates the intrusion of a warm air mass that will serve as a barrier
to deep convection.

The 1D analytical work of Haiden (1997) and of Ek and Mahrt (1994) both support these
findings. Haiden found that in cases of moderate to high stability, cumulus onset is favored
in low Bowen ratio (wet soil) environments, while in less stable environments, cumulus onset
is favored in high Bowen ratio (dry soil) regimes. In unstable environments, the onset time
is very sensitive to the sensible heat flux because of rapid BL growth. Additionally, more
rapid growth means that entrainment is more important than the surface latent heat flux in
the BL moisture budget. Thus, the impact of reduced latent heat flux is not crucial in the
triggering of convection.

Stable environments, on the other hand, Haiden (1997) found to be more conducive to
rain over wet soils because of the rapid rise of the lifting condensation level (and fall of
$\theta_E$) that accompanies rapid BL growth. When the BL growth rate is small and the flux of
moisture from the surface is large, then the rise of the LCL accompanying the BL growth is
overpowered by the fall of the LCL accompanying the BL moistening and the $\theta_E$ increase. The work presented here extends the results of Haiden (1997) to real data, identifies the portion of a sounding that is most critical, and gives physical justification for why we see evidence of a positive soil moisture–rainfall feedback in Illinois.

Ek and Mahrt (1994) presented a strong case regarding the importance of the structure of the atmosphere in response to different land surface conditions. Using data from HAPEX-MOBILHY and a 1D model of the soil and boundary layer, they looked at the relative humidity at the top of the BL (because of the control this has on the development of BL clouds) in response to variations in soil moisture, large-scale vertical motion, and the moisture and temperature stratification above the BL. They show very clearly that in their model, “The influence of soil moisture on relative humidity [at the top of the BL] varies dramatically according to initial atmospheric conditions and the prescribed mean subsidence” (pg. 2718). When stratification above the BL is weak, then BL growth dominates the relative humidity tendency equation, and dry soils lead to higher relative humidities, and presumably greater incidence of clouds. When air above the BL is strongly stratified or quite dry, on the other hand, then the moistening terms dominate the relative humidity tendency equation, and wet soils are more likely to lead to BL clouds. These two scenarios are consistent with a high $CTP$ case where a negative feedback is expected between soil moisture and rainfall, and a negative $CTP$ case where wet soils are more likely to lead to shallow clouds, as long as the low-level humidity deficit is not too large.

Betts and Ball (1995) found similar evidence for positive soil moisture–rainfall feedbacks at the FIFE site in Kansas. They found that increased soil moisture led to an increased diurnal $\theta_E$ range, and was accompanied by a decrease in the peak depth to the LCL from 140 mb over wet soils to 240 mb over dry soils. Note that the 100 mb between these two LCL depths is captured by the critical $CTP$ region. This difference in $\theta_E$ behavior over soils of different moisture content is important, Betts and Ball determine: “If soils are moist enough over large enough horizontal scales, then the associated higher equilibrium $\theta_E$ and the lower cloud-base can be expected to organize mesoscale convective systems, just as warmer sea surface temperatures do over the ocean” (pg 25,692).

Since the Convective Triggering Potential only measures properties of the temperature
profile, it is logical that this measure will yield the most information when coupled with a measure of the humidity. The lowest levels of the atmosphere proved to be the most important for humidity considerations with these soundings from Illinois, as shown by the stratification of the data with $HI_{low}$ and as discussed in Section 3.2.2. This index is the sum of the dew point depressions at 950 mb and 850 mb. Figure 3.19 summarizes the predictive capability gained from use of these two measures of the early morning atmospheric setting. As shown in this figure, when $HI_{low}$ is less than about 5 K or greater than about 15 K, the model outcome is atmospherically controlled:

- $HI_{low} < 5^\circ$C:
  - $CTP > 0$ J/kg: rain will occur over any soil condition;
  - $CTP < 0$ J/kg: shallow clouds will result over any soil condition;
- $HI_{low} > 15^\circ$C
  - Any $CTP$: no convection will result over any soil condition.

Also shown in Figure 3.19, when $HI_{low}$ is between 5 and 15 K, the land surface moisture condition can significantly impact the likelihood of rain, and the $CTP$ can help to determine what that impact will be:

- $5^\circ$C < $HI_{low}$ < $10^\circ$C:
  - $CTP < 0$ J/kg: Shallow clouds over wet soils. No convection over dry soils.
  - $CTP > 0$ J/kg: Wet soils favored! Rain over wet soils, rain likely (but not certain) over dry soils.
- $10^\circ$C < $HI_{low}$ < $15^\circ$C:
  - $CTP < 50$ J/kg: Shallow clouds likely (but not certain) over wet soils. No convection over dry soils.
  - $50$ J/kg < $CTP < 180$ J/kg: Transition zone: Any outcome possible. Convection of either kind is more likely over dry than wet soils, but no convection is highly likely over either.
  - $CTP > 180$ J/kg: Dry soils favored! No convection over wet soils, rain or shallow clouds possible over dry.
Figure 3.19: Anticipated 1D model outcomes given early-morning values of the $CTP$ and $HI_{low}$ over wet and dry soils.
Figure 3.20: The $CTP-HI_{low}$ framework for describing atmospheric controls on soil moisture-rainfall feedbacks. Only when the early morning atmosphere has $CTP > 0 \text{ J/kg}$ and $5 < HI_{low} < 15^\circ \text{C}$ can flux partitioning at the surface influence the triggering of convection. In the wet soil advantage regime, the advantage manifests primarily through enhanced CAPEs, and therefore enhanced rainfall depths, and secondarily through increased triggering of convection. In the dry soil advantage regime, the advantage manifests through increased triggering of convective events. The Atmospherically Controlled region where rainfall is expected also shows enhanced CAPEs over wet soils as compared to dry soils.

While Figure 3.19 separately shows the expected responses to wet or dry soils in different $CTP-HI_{low}$ regimes, Figure 3.20 draws these two figures together into the full $CTP-HI_{low}$ frame of reference. This reference frame shows that the land surface moisture or vegetative condition can influence the potential for rainfall only in a limited range of early-morning atmospheric conditions. When the atmosphere is very dry ($HI_{low} > 15^\circ \text{C}$) or very stable ($CTP < 0 \text{ J/kg}$), rainfall cannot occur, independent of flux partitioning at the surface. When the $HI_{low}$ is $< 5^\circ \text{C}$ and the $CTP$ is $> 0 \text{ J/kg}$, then rainfall should occur over both wet and dry soils. When the $HI_{low}$ is $< 15^\circ \text{C}$ and $> 5^\circ \text{C}$, and the $CTP$ is $> 0 \text{ J/kg}$, then the land surface can significantly influence the likelihood of rainfall, as described in detail above.

De Ridder (1997) considered half of the problem studied here. With a 1D PBL box
model, he calculated the dependence of $\theta_E$ on the evaporative fraction, $\alpha$, and determined that the potential for moist convection increases with $\alpha$, except in very dry atmospheres. Haiden (1997), however, found that "static stability and temperature determine the sign of the Bowen ratio effect, with atmospheric humidity merely affecting its magnitude." Our results indicate that within a particular range of humidity, Haiden’s assessment holds for our results as well, but when the humidity deficit is sufficiently large or sufficiently small, the stability and temperature characteristics do not determine the sign of the Bowen ratio effect. In fact, when the humidity deficit is sufficiently large or sufficiently small, the likelihood for convection is independent of the land surface fluxes.

### 3.5 Conclusions

A one-dimensional model of the planetary boundary layer (BL) and surface energy budget has been modified to allow the growing BL to entrain air from an observed atmospheric sounding. The model is used to analyze the impact of soil saturation on BL development and the triggering of convection in different atmospheric settings.

Sounding data from the summers of 1997-99 from a station in central Illinois are used to initialize the model. For each sounding, the model is run once with saturated soils and once with very dry soils. The BL development is observed from initialization at 6 am until the triggering of free convection or until the end of the day. Results show that rainfall is more likely to occur over wet soils than over dry soils (22% versus 16% of the time). When deep convection is triggered over both dry and wet soils, the land surface moisture condition can impact rainfall depth. These impacts are seen in the following ways:

- Soil moisture conditions do not significantly impact the time of triggering of convection.
- At the time of triggering of free convection, the $\theta_E$ is significantly greater in high soil moisture conditions than in low soil moisture conditions.
- In addition to a higher $\theta_E$, the high soil moisture conditions lead to a significantly larger $CAPE$ and deeper convection depth at the time of triggering.
- Since $CAPE$ has been shown to be positively correlated with rainfall quantities (Eltahir and Pal, 1996), we conclude that higher soil moisture conditions lead to higher rainfall efficiency and higher rainfall amounts.
Thus there is evidence of a positive feedback between soil moisture and total rainfall amount.

A new measure of the Convective Triggering Potential (CTP) provides an excellent indication of the likelihood of rainfall or shallow clouds over soils of differing moisture states, particularly when considered in conjunction with a measure of the humidity deficit near the surface (Figure 3.19). The initial soundings which favor rain over dry soils are those with a high CTP (greater than ~180 J/kg) and an $HI_{low}$ between 10 and 15°C. In general, they are dominated by a near-surface layer that can only be penetrated by deep mixing, and not by the build up of energy in the boundary layer. They are characterized by a large surface inversion topped by an extensive zone (typically up to at least ~200 mb above the surface) with a lapse rate close to the dry adiabatic lapse rate. The high sensible heat flux over dry soils is able to more quickly penetrate above the surface inversion. After entraining the surface inversion, further entrainment and BL growth is rapid, due to the near-neutral state of the atmosphere in the critical CTP region (between 100 and 300 mb above the surface). Because of low sensible heat flux, the boundary layer only slowly entrains the surface inversion over wet soils, potentially never fully breaking through, despite the buildup of large amounts of energy (CAPEs well above 3000 J/kg). Dry soils have the advantage in these circumstances because they are more likely to reach the near-neutral region by or before the mid-day peak of radiation. Deep convection is possible over both wet and dry soils when $5°C < HI_{low} < 15°C$, but is favored over dry soils when $HI_{low}$ is on the high end of this range.

The initial soundings which favor rain over wet soils have an $5°C < HI_{low} < 12°C$, and an intermediate CTP, between 0 and ~180 J/kg, by the measure of the non-atmospherically controlled cases, and between 0 and ~200 J/kg by the measure of the atmospherically controlled cases. In contrast to the high CTP soundings, they are more easily overcome by the buildup of energy in the boundary layer than by deep BL growth with a low $\theta_E$. Like the high CTP soundings, ones with an intermediate CTP are also often characterized by warm air near the surface, but the air above this mass differs from the above-described soundings in a critical way. Rather than an extended region above the surface inversion with a lapse rate close to dry adiabatic, the lapse rate in this region is much closer to moist adiabatic.
Wet soils lead to much higher values of moist static energy in the BL than dry soils do, with the BL $\theta_E$ differing between wet and dry soil runs by well over 10 or 12°C in many cases. It is this $\theta_E$ difference that makes convection possible over wet soils but not over dry soils in these intermediate CTP circumstances. The greater stability found in these nearly-moist adiabatic soundings means that entrainment is more difficult than in high CTP soundings, and that an increase in the $\theta_E$ leads to a greater decrease in the level of free convection than in neutral or unstable soundings. Both of these facts lead to a favoring of convective triggering over wet soils when the CTP is in this intermediate range.

When CTP is negative, either the temperature profile is nearly moist adiabatic and has limited CAPE, or an overlying warm air mass is present. In both cases, deep precipitating convection is not possible, though shallow clouds are likely over wet soils when the CTP is less than 0 J/kg and $HI_{low} < 10^\circ C$.

This work offers considerable insight into the process of convective triggering, and the interplay between atmospheric and land surface conditions. In the next two chapters, we take the CTP-$HI_{low}$ framework developed here into the three-dimensional world of MM5 to determine if the insights gained here hold up when 3D effects are included. The conclusions of the MM5 work indicate that the framework is indeed valid and helpful, paving the way for the nationwide analysis of the potential for soil moisture-rainfall feedbacks presented in Chapter 6.
Chapter 4

Description of The Fifth-Generation Penn State/NCAR Mesoscale Model (MM5)

4.1 Introduction

Through the one-dimensional boundary layer modeling work of Chapters 2 and 3, the CTP-$H_I_{low}$ framework for describing atmospheric controls on soil moisture-rainfall feedbacks was developed. In these next two chapters, we apply this framework to three-dimensional modeling work, to see if the 1D concepts apply to a 3D setting. This 3D analysis is performed with the Fifth-Generation Penn State/NCAR Mesoscale Model (MM5), using a small domain centered in Illinois, modeling days from the summers of 1996-99.

MM5 is a product of modeling work that began in the 1970's and was originally documented by Anthes and Warner (1978). The current incarnation includes multiple-nest capability, nonhydrostatic dynamics, and a four-dimensional data assimilation capability (Grell et al., 1995). The vertical dimension is described by the terrain-following $\sigma$-coordinate, which is always equal to 1 at the land surface and 0 at the top of the model domain. Model calculations are performed with a finite difference algorithm with velocity variables staggered with respect to other fields. For many physical processes, multiple options are available, and
more options are being added with each new release. The model version used and described in this thesis is Version 2.12.

The following sections describe the grid and domain used in these simulations, the initial and boundary condition data, and the representations of boundary layer processes, radiation process and cloud microphysics.

4.2 Model Domain Configuration

The model runs detailed in this thesis were all run with the same single-nested domain configuration. Though MM5 is capable of running nested domains, initial multi-nested runs with outer nest grid spacing ranging from 6 km to 18 km on a side indicated that results were dependent on the convection scheme used. This is consistent with the work of Pal (1997). In order to remove the dependence on convection schemes, simulations were run on a 200 km by 200 km domain with 2 km grid resolution. At this level of detail, we can rely on MM5's capacity to explicitly resolve vertical velocity and convective motion.

In the vertical, there are twenty-three levels between the 100 mb top and the surface. The half-sigma levels include 1.00 (surface), 0.99, 0.97, 0.94, 0.91, 0.87, 0.83, 0.77, 0.73, and 0.67 at and below the critical $CTP$ region. (The 0.67 level is approximately at the $P_{surf} - 300$mb top of this critical region.) Sigma intervals of approximately 0.05 separate the levels between 0.67 and the top at 0.03.

The domain was centered over Illinois, near the Flatland site at 40.0N, 88.3W (Figure 4.1). Initial and boundary conditions were provided by Eta Model Assimilated Data at 40 km resolution (Rogers et al., 1995; Black, 1994). Detailed analysis and comparisons with observations are performed on the central 64 km by 64 km portion, in order to be sufficiently far from any potential boundary effects. Simulations were initialized at 6 am using Eta Data for several days during the summers of 1996-1999. The questions to be addressed by these simulations concern the interactions between the early-morning atmosphere and fluxes from the land surface and how these conditions might impact the triggering and the amount of rainfall on a given day. Therefore, nighttime conditions were not relevant and simulations were halted at 9 pm, after 15 simulated hours.
4.3 Initial and boundary conditions: Eta model data

Initial and boundary condition data are provided from Eta Model output. The boundary data are treated with a relaxation condition whereby model-predicted variables are “relaxed” or “nudged” toward the large-scale Eta Model analysis. This treatment impacts the four grid points closest to the boundary, decreasing with distance from the boundary.

The Eta Model is one of the National Centers for Environmental Prediction's (NCEP) operational forecast models. Characteristics of the model will be summarized here; for further details, see e.g., Rogers et al. (1995), Black (1994), Janjic (1994), Mesinger and Black (1992). It is run twice daily, at 00Z and 12Z, yielding analysis products for the time of
the run and forecast products covering 3 hour intervals up to 48 hours after each run. The horizontal resolution of the grid has changed a number of times since the original 80 km grid of the early 1990s. The Eta data used in these MM5 simulations was from a 40 km output grid on a Lambert Conformal projection (NCEP Grid No. 212, known as AWIPS Grid 212). Twenty-five constant pressure surfaces cover the vertical domain from 1000 mb to 25 mb, at intervals of 50 mb, with 25 mb intervals added in lowest 2.0 km and at the tropopause jet level.

The model terrain is represented with the eta coordinate (see e.g., Black [1994] for a full description), where terrain surfaces are only allowed on a vertical eta level. This creates a step-like pattern of topography throughout the domain. This coordinate was chosen to reduce errors in sigma-coordinate models known to occur along steeply sloping terrain (Mesinger and Black, 1992).

The original Eta Model created its initial and boundary conditions through optimal interpolation (OI) of data from the Global Data Assimilation System. The mesoscale version of the model which became operational in 1994 used a more detailed system called the Eta Data Assimilation System (EDAS). This system uses the OI to establish a first guess field 12 hours before the forecast time. This OI field is used to initialize the Eta Model, which is run for 3 hours and then stopped to allow for assimilation of new observations. Three more cycles of 3 hour runs followed by assimilation of new data are performed until the actual forecast time is reached. This final EDAS field is the initial field for the Eta Model forecasts. (See Black [1994] for a description of EDAS and Rogers et al. [1995] for information on the improvements derived from the implementation of this system.)

The fundamental prognostic variables in the Eta Model are temperature, specific humidity, horizontal wind components, surface pressure, turbulent kinetic energy and cloud water. Model variables are distributed over the semi-staggered Arakawa E grid, with mass variables (e.g., temperature and moisture) in the middle of each grid box, surrounded by four corner points where the winds are defined.

The physics of the Eta model include both grid scale and convective precipitation. Grid scale precipitation occurs whenever the relative humidity in a given grid box exceeds 95%. The convective parameterization is determined by the Betts (1986) convective adjustment
scheme, plus treatment of explicit liquid water. The radiation scheme comes from the Geo­
physical Fluid Dynamics Laboratory, with shortwave treatment detailed by Lacis and Hansen
(1974), and longwave by Fels and Schwarztkopf, (1975). Boundary layer processes are treated
with the Mellor-Yamada turbulent kinetic energy formulation level 2.0 for the surface layer,
and level 2.5 formulation above the surface layer (Mellor and Yamada 1974, 1982).

4.4 Surface Energy Budget

The ground surface temperature, \( T_g \), is calculated in MM5 with a force-restore method, based
on that developed by Blackadar (Zhang and Anthes 1982). The surface budget equation is
given by

\[
C_g \frac{\partial T_g}{\partial t} = R_n - G - H_s - \lambda E_s
\]

where \( C_g \) is the thermal capacity of the slab per unit area, \( R_n \) is the net radiation, \( G \) is the
ground heat flux, \( H_s \) is the sensible heat flux into the atmosphere, and \( \lambda E_s \) is the latent heat
flux into the atmosphere, with \( \lambda \) being the latent heat of vaporization and \( E_s \) the surface
moisture flux. The following sections will address each of the terms on the right side of
Equation 4.1, following a discussion of the treatment of cloud microphysics, since clouds are
an important aspect of the radiation calculations.

4.5 MM5 Cloud Microphysics

Rainfall occurs through many mechanisms which operate on many spatial and temporal
scales. No numerical representation of precipitation can describe all of these mechanisms in
one manner, and the complex spatial heterogeneity of rainfall requires different parameteri­
zation techniques for different scales of averaging or simulating. To allow for a wide range of
both spatial grid scales and rainfall mechanisms, MM5 includes both explicit (grid-resolved)
and implicit (parameterized) precipitation schemes. Implicit schemes are used when the grid
spacing is larger than a few kilometers, when detailed resolution of vertical velocity is not
possible. Explicit schemes, on the other hand, treat resolved precipitation processes, and
can be used with or without an accompanying implicit scheme.

The work of Pal (1997) and Pan et. al (1996) illustrate the influence of implicit convection schemes on modeling results. Both studies show that the sensitivity of rainfall to soil moisture conditions is dependent on the convection scheme. In order to avoid this dependence, these MM5 simulations are performed using a small enough grid spacing to allow for fully explicit resolution of cloud microphysics. According to Frank (1983), models with horizontal scales equal to or smaller than scales typical of convective updrafts and downdrafts can explicitly resolve convection. These convective element scales are on the order of 0.1-10 km (Frank, 1983). The work of Zhang et. al (1988) indicated that at grid scales of 12.5 km, both explicit and implicit schemes were required to fully reproduce convective precipitation responsible for two storms leading to severe flooding in Jonestown, Pennsylvania in 1977. Similarly, Wang et. al (2000) state that MM5's capability to explicitly resolve moist convection is appropriate with grid sizes <~10 km. The 2 km grid scale used in this work is well below scales appropriate for the use of implicit convection schemes, so moist convection is assumed to be fully represented by an explicit moisture scheme.

MM5 has many different schemes for explicitly resolving precipitation, all of which are activated whenever grid-scale saturation (or a specified fraction of grid-scale saturation) is reached. The most simple approach involves only two steps: remove super-saturation as precipitation, and add the latent heat to the thermodynamic equation. Many other schemes are much more sophisticated, accounting for cloud ice and snow and even graupel. The seven available options are dry, stable, warm rain, simple ice, mixed phase, and two different graupel schemes. The simulations run for this study all used the mixed phase explicit moisture scheme, which is built on Dudhia's simple ice scheme (1989). The differences and additional processes included in the mixed phase scheme will be discussed after a general description of the simple ice scheme.

4.5.1 The Explicit Moisture Scheme

In the tropical oceanic environments where Dudhia's (1989) simple ice scheme was first developed, in-cloud processes most frequently occur at below-freezing temperatures. Many
explicit moisture schemes account for this by simply treating cloud water as ice, and rain as snow when temperatures are below 0°C. This is generally computationally efficient, since only three moisture arrays are needed: one accounting for water vapor, one accounting for cloud water and cloud ice, and one accounting for rain and snow. Indeed, this is the general approach of the simple ice schemes of both Hsie et al (1984) and Dudhia (1989), and it provides the foundation for many more advanced schemes, including the one used in these simulations.

In Dudhia's simple ice representation of cloud microphysics, it is assumed that there is no supercooled water or superwarmed snow or ice of any kind. Melting is assumed to take place within one model level of the freezing point, such that cloud ice and snow melt immediately upon descending past the 0°C level, and cloud water and rain immediately freeze upon rising above this level. This assumption is good for slowly falling particles, but doesn't capture the true history of heavy particles or particles subjected to intense vertical motion. Particles such as graupel and hail that are formed through much mixed phase growth are neglected, since it is not possible to adequately represent them with the assumptions mentioned above.

Though the presence of supercooled water is neglected in the simple ice scheme, supersaturation with respect to vapor is not. This differential treatment is justified given that condensation occurs very quickly, while deposition of vapor onto ice crystals can be slow due to limited nuclei (Rogers and Yau, 1989). Observations from the GATE (GARP Atlantic Tropical Experiment) experiment (Rutledge, 1986) and from the WMONEX (Winter Monsoon Experiment) experiment (Dudhia, 1989; Churchill and Houze, 1984) were used to develop Dudhia's simple ice scheme. In those mesoscale cloudy settings away from convective updrafts, there was little to no evidence of mixed phase ice particle growth, suggesting that this simple ice parameterization could describe cloud growth in these settings. One reason Dudhia (1989) cites for this adequate representation is that at subfreezing temperatures water droplet build-up is unlikely because of the Bergeron-Findeisen process: vapor is preferentially consumed by ice particles rather than water droplets, and the water droplets evaporate.

Behavior of the ice processes represented in the model depend heavily on the assumed number concentration of ice crystals, which Dudhia calls the least certain assumption in the
model. This concentration is strongly dependent on the temperature, increasing by a factor of about 400 for each 10°C drop (Dudhia, 1989).

In addition to the processes represented by the simple ice scheme, the mixed phase scheme used in this thesis also allows for snow and ice to exist at temperatures above zero. This allows for falling particles to partially melt, be lifted again, and develop the riming structures observed in graupel and hail. Computationally, this means that separate arrays are used to store vapor, cloud, rain, cloud ice, and snow (Grell et. al, 1995). Physically, the mixed phase scheme is a more appropriate representation of the continental convection studied in this thesis than the simple ice schemes, which are most appropriate for tropical oceanic convection.

4.6 The Modified CCM2 Radiation Scheme

From Hack et. al (1993), we find that the primary steps in the CCM2 radiation scheme include calculation of (1) cloud water properties (including cloud fraction, cloud liquid water content and path, emissivity and effective cloud cover), (2) latitudinally dependent radiation parameters, (3) solar radiation, and (4) longwave radiation. Both Kiehl et. al (1994) and Hack (1998) found the need for improvements in the treatment of clouds and their radiative properties in the CCM2 parameterization. CCM2 had some systematic biases, two of which are particularly important to investigations of mid-latitude summers. First, there was a substantial over-prediction of July surface temperatures over most of the Northern Hemisphere. Second, precipitation maxima were also consistently overpredicted, particularly over warm land areas. These biases were determined to be associated with deficiencies in the cloud optical properties, namely, cloud liquid water path and the cloud effective radius.

Two main changes in the treatment of clouds were undertaken to attempt to correct these biases, and Hack (1998) shows that these improvements in the cloud liquid water path and the cloud drop effective radius lead to substantial improvements in CCM2 performance. These two items, discussed in detail below, are the primary differences between the CCM2 and the CCM3 cloud parameterizations. They have also both been included in the version of the radiation scheme used in this thesis, making it effectively closer to the CCM3 scheme.
than the CCM2 scheme.

Cloud Liquid Water Path  In CCM2, the in-cloud liquid water path, $CWP$, is dependent only on latitude, and not on calculated liquid, vapor, ice or snow mixing ratios. The grid-based area-averaged liquid water path does involve some dependence on local conditions, through the cloud fraction:

$$LWP = A_c CWP.$$  \hspace{1cm} (4.2)

The total cloud fraction, $A_c$, is determined by both convective cloud cover and non-convective clouds.

The in-cloud liquid water path is evaluated according to the function

$$CWP = \int \rho_l dz$$  \hspace{1cm} (4.3)

where

$$\rho_l(z) = \rho_l^0 e^{-z/h_l}$$  \hspace{1cm} (4.4)

and $\rho_l^0 = 0.18 \text{ g/m}^3$. The liquid water scale height, $h_l$, is the element which changes dramatically between CCM2 and CCM3. In CCM2, $h_l$ is time-independent and meridionally varying according to

$$h_l = A + B \cos^2 \phi$$  \hspace{1cm} (4.5)

where $A = 1080 \text{ m}$ and $B = 2000 \text{ m}$. Thus, $h_l$ varies only according to latitude: it is large in the tropics and small at high latitudes. At the latitude of Illinois (approximately $38^\circ$N to $42^\circ$N), $h_l$ ranges from 2180 m to 2320 m, by this definition. This scale height, in turn, forces the three-dimensional distribution of in-cloud liquid water density profile, $\rho_l$, to be independent of the calculated liquid water content. It is fully prescribed for a given latitude.

CCM3, on the other hand, calculates $h_l$ with dependence on the vertically integrated water vapor:

$$h_l = a \ln \left( 1.0 + \frac{b}{g} \int_{P_c}^{P_e} q dp \right),$$  \hspace{1cm} (4.6)

where $a = 700 \text{ m}$ and $b = 1 \text{ m}^2/\text{kg}$. This allows $h_l$ to vary anywhere from 0 to 4300 m and more, depending on the humidity of the model layer. For constant cloudy sky fraction, this change leads to higher liquid water paths for a column with greater precipitable water, and lower liquid water paths for a column with less precipitable water. Hack (1998) is careful to
include a caveat stating that this method is still empirical, and "does not address the issue of whether an exponentially decaying profile for in-cloud liquid water concentration is an appropriate approximation" (pg. 1500). Comparisons of simulations with this formulation for $h_i$ versus the original CCM2 formulation significantly reduced the high temperature bias seen throughout Northern Hemisphere summers. When this change was made in the radiation scheme used in this work, this amounted to a temperature reduction of approximately 0.5 K in Illinois.

**Cloud Water Droplet Effective Radius** In CCM2, this value was set as a constant of 10 $\mu$m. In CCM3 and in this work, a formula allows it to vary from 5 to 10 $\mu$m. Hack (1998) follows the observations of Kiehl (1994) which showed that this formula more adequately describes over-land drop size distributions.

\[
T > -10^\circ C : r_e = 5\mu m
\]

\[
-30^\circ C \leq T \leq -10^\circ C : r_e = 5 - 5 \left( \frac{T + 10}{20} \right) \mu m
\]

\[
T < -30^\circ C : r_e = 10\mu m
\]

### 4.7 Ground Heat Flux

The flow of heat into the ground due to molecular conduction is given by

\[
G = K_mC_g(T_g - T_m)
\]

where $K_m$ is the heat transfer coefficient, given by $K_m = 1.18\Omega$, $\Omega$ is the angular velocity of the earth, $T_m$ is the substrate temperature, and the other variables have the same meaning as in Equation 4.1. The value of $T_m$ is currently taken as the mean of the surface-air temperature over the period of the simulation.
4.8 Boundary Layer Parameterization: Sensible and Latent Heat Fluxes

The sensible and latent heat fluxes, $H_s$ and $\lambda E_s$ in Equation 4.1, are determined by the boundary layer parametrization. The Blackadar planetary boundary layer (PBL) scheme used in MM5 is well documented in Zhang and Anthes (1982). Blackadar (1979) made a strong argument for the need for a PBL scheme with high vertical resolution in order to adequately model the transition from well-mixed daytime conditions to stratified nighttime conditions, which are often characterized by strong gradients of temperature, wind and moisture. The MRF scheme (originally used in NCAR's Medium-Range Forecast model; Hong and Pan, 1996, Troen and Mahrt, 1986) is quite similar to the Blackadar scheme, except in its treatment of countergradient fluxes during free convection. Early sensitivity studies showed that these MM5 experiments were not sensitive to changes between these two schemes. The results presented in Chapter 5 are for simulations with the MRF BL scheme. We will now describe the Blackadar scheme, summarizing the more complete descriptions in Grell et al (1995) and Zhang and Anthes (1982), and then discuss the differences between this and the MRF scheme used in this work.

4.8.1 Sensible and Latent Heat Fluxes

The surface heat flux $H_s$ is given by

$$H_s = -C_{pm} \rho_a k u_* T_*, \quad (4.11)$$

where $C_{pm}$ is the specific heat at constant pressure for moist air, $\rho_a$ is the air density at the lowest model level, $k$ is the von Karman constant, $u_*$ is the friction velocity, given by

$$u_* = \text{MAX} \left( \frac{kV}{\ln \frac{z_a}{z_0}} - \psi_m, u_{*0} \right), \quad (4.12)$$

$T_*$ is

$$T_* = \frac{\theta_a - \theta_g}{\ln \frac{z_a}{z_0} - \psi_n}, \quad (4.13)$$
\( V \) is velocity, \( z_a \) is the height of the lowest \( \sigma \)-level, \( z_0 \) is the roughness parameter (specified by land-use category; 15 cm in summer over agricultural lands), \( u_* \) is a background value (0.1 m/s over land, 0 over water), and \( \psi_m \) and \( \psi_h \) are nondimensional stability parameters that are a function of the bulk Richardson number, which will be discussed below. These stability parameters are defined by different equations in the four behavioral regimes used to represent turbulent mixing in different settings. The determination of the appropriate regime and the stability parameter equations for each regime will be discussed in the next section.

The surface moisture flux equation is also dependent on the stability parameters:

\[
E_s = M \rho_a I^{-1}(q_{oa}(T_g) - q_{va}),
\]

where

\[
I^{-1} = k u_* [\ln (\frac{k u_* z_a}{K_a}) + \frac{z_a}{z_l} - \psi_h]^{-1},
\]

\( M \) is the moisture availability, \( z_l \) is the depth of the molecular layer (0.01 m over land, \( z_0 \) over water), and \( K_a \) is a background molecular diffusivity equal to \( 2.4 \times 10^{-5} \) m\(^2\)/s.

### 4.8.2 Determination of Behavioral Regime

The Blackadar PBL parameterization has two primary modules to represent turbulent mixing: a stable, nocturnal model and a free-convection module. The module used at a given time step is determined by the Richardson number, by the ratio of the mixed layer height to the Monin-Obukhov length, and by the sign of the temperature gradient in the layer closest to the surface.

**On the Monin-Obukhov length** The Monin-Obukhov length, \( L \), is a measure of the relative strength of the velocity scales associated with mechanical turbulence to those associated with buoyant turbulence (Emanuel, 1994). Considering a dimensional analysis of a simplified case of a semi-infinite fluid with both convective and mechanical turbulence, a length scale of the form \( L = -M^{1/2}/Q_0 \), arises, where \( Q_0 \) and \( M \) are the integrated buoyancy and momentum sinks, respectively. The Monin-Obukhov length is conventionally defined as
negative in unstable conditions when the surface buoyancy flux, $Q_0$, is positive.

Representing $Q_0$ and $M$ with surface fluxes, we can use $Q_0 = \overline{w'\theta'}$ and $M = \overline{w'u'}$ to compute

$$L = -\frac{(\overline{w'u'})^{3/2}}{\overline{w'\theta'}}$$

(4.16)

In the model, $L$ is calculated by

$$L = -\frac{C_{pm} \rho_a \theta_a v_s^3}{kg H_a}$$

(4.17)

Mechanical turbulence is dominant when $z < -L$, and convective turbulence is dominant for $z > -L$. Note that a purely mechanical boundary layer would have $-L \to \infty$, while a purely convective boundary layer would have $-L \to 0$. A typical daily cycle has $L$ changing from within the positive range of 5 to over 200 during the night, with a relatively abrupt transition to negative (unstable regime) values of less than -150 around sunrise. $L$ typically stays negative but gets much smaller in magnitude throughout most of the day, until another negative peak near sunset leading into an abrupt transition back to the positive range of nighttime conditions (Stull, 1988, pg. 181). The Zhang and Anthes (1982) PBL scheme enters the free-convection regime only when $|z_h/L| > 1.5$, where $z_h$ represents the height of the boundary layer, and the Bulk Richardson number $< 0$.

**On The Richardson Number** The Richardson number is a measure of the ratio of buoyant production of turbulent kinetic energy (TKE) to mechanical production of TKE. Since wind and temperature data are provided at discrete model levels, the bulk Richardson number is used in the Blackadar scheme. The model definition of the bulk Richardson number for the lowest model layer, denoted by the subscript $a$, is given by

$$R_B = \frac{g z_a (\theta_a - \theta_g)}{\theta_a (V_a)^2}$$

(4.18)

where $V_a$ is the speed of the surface wind.

In Zhang and Anthes' (1982) PBL scheme, the stability regime is, in part, determined by whether the bulk Richardson number is less than or greater than a critical value of $R_B$. Zhang and Anthes use 0.2 as this critical value demarking the transition between laminar and turbulent flow.
Regime 1: \( R_B \geq 0.2 \) means nighttime stable conditions
Regime 2: \( 0.0 < R_B < 0.2 \) means damped mechanical turbulence
Regime 3: \( R_B < 0.0 \) and \( \left| \frac{z_h}{L} \right| < 1.5 \) means forced convection
Regime 4: \( R_B < 0.0 \) and \( \left| \frac{z_h}{L} \right| > 1.5 \) means free convection

4.8.3 The Nocturnal Module

The nocturnal module is comprised of the first three regimes: stable conditions, damped mechanical turbulence, and forced convection. These three regimes are treated in the same way in both the Blackadar and the MRF PBL schemes. In each of these regimes, local-K theory is used, where the eddy transfer coefficient \( K \) is a function of the gradient Richardson number, \( Ri \). The primary assumption in this theory is that turbulent correlations are proportional to their vertical lapse rates.

Regime 1: Stable Conditions When \( R_B \) is greater than the critical value of 0.2, the surface fluxes are determined using \( u_* = u_{*o} \) and

\[
\psi_m = \psi_h = \frac{z_a}{z_o}. \tag{4.19}
\]

Additionally, \( H_s \) is has a lower limit of -250 W/m².

Regime 2: Damped Mechanical Turbulence When \( 0.0 < R_B < 0.2 \), surface scaling parameters are given by

\[
\frac{z_a}{L} = \frac{R_B}{1 - 5R_B} \frac{\ln z_a}{z_0} \tag{4.20}
\]

and

\[
\psi_h = \psi_m = -5z_a/L. \tag{4.21}
\]

Regime 3: Forced Convection This regime is used for marginally unstable states when \( R_B < 0.0 \) and \( \left| \frac{z_h}{L} \right| < 1.5 \). This latter conditions implies that mechanically-generated turbulence is stronger than buoyancy-generated turbulence. Rather than use time consuming iterative procedure to determine the scaling parameters for these conditions, a “quasi-neutral” assumption is made, whereby the scaling parameters can be approximated by

\[
\frac{z_a}{L} = R_B \ln \frac{z_a}{z_0} \tag{4.22}
\]
The Free Convection Module

This fourth regime, the free convection module, is used when \( R_B < 0.0 \) and \( |z_h/L| > 1.5 \).
This occurs when buoyancy turbulence is dominant.

\[
\psi_h = -3.23 \frac{z_a}{L} - 1.99 \left( \frac{z_a}{L} \right)^2 - 0.474 \left( \frac{z_a}{L} \right)^3, \tag{4.24}
\]

\[
\psi_m = -1.86 \frac{z_a}{L} - 1.07 \left( \frac{z_a}{L} \right)^2 - 0.249 \left( \frac{z_a}{L} \right)^3, \tag{4.25}
\]

where \( \frac{z_a}{L} \) is restricted by a lower limit of -2.0.

In general, \( \frac{z_a}{L} \) is a function of \( \psi_m \), and an iterative approach is required to solve the equations. In the model, however, a time-saving approximation is made:

\[
\frac{z_a}{L} = R_B \ln \frac{z_a}{z_0}. \tag{4.26}
\]

With this four-part scheme, \( \psi_m \) is continuous for all values of \( R_B \).

This regime captures behavior when heating of the land surface is particularly strong, and a super-adiabatic layer forms near the land surface. Buoyant plumes can lift this air to levels far above the surface. The Blackadar boundary layer scheme models this type action by allowing vertical mixing between the lowest layer and each layer in the mixed layer. In local-K theory, mixing occurs only between adjacent layers, and low-level clouds tend to be oversimulated due to inefficient moisture transport (Hong and Pan, 1996; Holtslag and Boville, 1993). Accounting for these so-called countergradient terms (Deardorff, 1972) is consistent with the observation that most of the transport of mass and momentum within the PBL is accomplished by the largest eddies (Hong and Pan, 1996). This formulation allowed for more efficient transport of all quantities when compared to schemes using K-theory, but it was most noticable in the transport of water vapor.

The MRF boundary layer scheme used in the work shown here is quite similar to the Zhang and Anthes (1982) version of the Blackadar scheme described above. The main
difference comes through the properties attributed to eddies. In the Zhang and Anthes version, eddies are given properties of surface air, as described above. In the MRF scheme (Hong and Pan, 1996; Troen and Mahrt, 1986), eddies have the bulk properties of the PBL, rather than local properties. As mentioned above, early sensitivity tests between these two schemes showed few changes in the model outcome. Of the 24 simulations performed with both boundary layer schemes, the occurrence of rainfall or clouds was different only one time, and this was a borderline case. The results presented in this thesis are all from simulations with the MRF scheme.

4.8.5 Treatment of the land surface moisture

The Version 2.x series of MM5 releases all treat soil moisture with a moisture availability term that is dependent on vegetation type and season. The moisture availability does not change with evaporation or precipitation: it is constant for the vegetation type throughout the course of a simulation. The landuse type over the entire experimental domain is agriculture. The default moisture availability for this vegetation class is 30%. Results presented here show model runs with wet conditions simulated using a moisture availability of 80%, and dry conditions simulated using a value of 10%. Though this treatment is crude, it is suitable to the task at hand for a number of reasons. First, our primary concern is the response of the growing boundary layer to different fluxes from the land surface. A more intricate land surface scheme would add many unnecessary layers of complexity to the calculation of evapotranspiration. Second, on the time scale of the 15 hours it is not unreasonable to assume that the soil moisture changes little, except in the event of rainfall over dry soils.

4.9 Comparisons of Rainfall Results with Observations

A thorough presentation of simulation results is presented in Chapter 5. However, we will now briefly present comparisons of modeled and observed rainfall to validate our usage of the model in the described configuration. Though we initialize each model run with atmospheric data from a particular summer day in Illinois, no tuning of the model has been performed to
try to match the results to observations. The results presented in Figure 4.2 are comparisons of domain-average (domain refers to the focus region depicted in Figure 4.1) rainfall with the average of hourly raingauge data from nine stations within 0.45° latitude or longitude of the focus region. Raingauge data is taken from the EarthInfo, Inc. database, which is a subset of the National Climatic Data Center's (NCDC's) TD-3240 file. Only three raingauge stations were within the focus region boundaries: six additional stations were added by the slight expansion of the region of observation.

The model simulations were initialized at 6 am and terminated at 9 pm, so observed rainfall measurements plotted in Figure 4.2 are cumulative hourly totals between 6 am and 9 pm. Of the 68 simulations from the summer of 1996, the rainfall in 62 cases was simulated reasonably well by at least one of the two simulations for the day of interest. Removal of the six worst cases greatly improves the goodness-of-fit of the modeled-to-observed rainfall, particularly for the wet soil runs: the $r^2$ for these 62 cases is 72.8% for the wet soil runs and 41.4% for the dry soil runs. Four of the six poor-performers were model under-estimates, and two were model over-estimates. The model under-estimate cases were all very high rainfall events, and were the only events in this range of rainfall observation. This suggests that the model is unable to adequately model extreme events. Specifics of these cases are discussed in the next section. The upper right plot in Figure 4.2, however, suggests that except for extreme events, the model is capable of capturing and adequately modeling non-extreme rainfall events. Comparison of the upper and lower right plots of this figure further suggest that the model is also sensitive to soil moisture. These two suggestions allow us to go forward using the model to address questions of atmospheric controls on soil moisture-rainfall feedbacks.

4.9.1 Description of Outliers: Model Overestimates

The two cases where the model severely overestimated rainfall were Run V73 (08 June 1996), the most extreme outlier, with observed rainfall averaging about 2.5 cm over the 65-by-65 km focus region, and Run V98 (26 August 1996), with observed rainfall of about 1.6 cm. The initial sounding for Run V73 (Figure 4.3) shows that a super-adiabatic layer was present
at the surface in the 6 am initial conditions. This is also true for Run 98, though this superadiabatic layer of about 25 mb depth is capped by a stable layer extending up to to 950 mb. The presence of these super-adiabatic layers indicates that early-morning heating of the surface was substantial even before the 6 am conditions captured by these Eta Model-derived initial conditions. This condition also occurs on some of the other simulated days, but on most days in Illinois, the nocturnal inversion is not removed until mid-morning. Without such an inversion, the boundary layers in Runs V73 and V98 grew and developed more quickly than in other runs with similar conditions above the anomalous lower ~30 mb.
Figure 4.3: Domain average initial sounding for run V73 ($CTP = -61$ J/kg, $H_{low} = 5.7^\circ$C.) Note that the surface air is already positively buoyant at 6 am. Rainfall in both the model runs begins at about 4 pm. Rainfall is observed at 2 of 9 nearby stations, also beginning at about 4 pm.

Modelled rainfall over both the wet and dry soils began at about 4 pm in both cases: without this early-morning advantage, rainfall may not have occurred. Similarly, two of the nine raingauge stations also had rainfall beginning at about 4 pm on both June 8 and August 26, though clearly not at high enough intensities to produce quantities close to the simulated domain-average values. One possible explanation for the model's poor performance is that the domain-average initial conditions given by the Eta Model are in fact only representative of a small portion of the domain in these two cases.

4.9.2 Description of Outliers: Model Underestimates

In four of the six outliers, rainfall was substantially underestimated by the model. As previously mentioned, the 3D wind configuration has the potential of enhancing or diminishing convective potential. Mechanisms controlling these processes will be discussed in
Section 5.4.1. At this stage, we will only mention that Runs V10 (17 June 1996) and V18 (21 July 1996) have $CTP-HI_{low}$ combinations which suggest that no rainfall should occur: V10's $HI_{low}$ of 15.9°C is greater than 15°C, placing it in the Too Dry region, and V18’s $CTP$ of -18 J/kg is less than 0 J/kg, placing it in the Too Stable region. Both of these also had veering winds in the lowest 300 mb which greatly enhanced the convective potential and allowed some rainfall to occur, though not enough to match observations. Indeed, no rainfall occurred in the reduced-winds simulations of either of these two cases. As will be discussed in the next chapter, other model runs with veering winds but with values in $CTP-HI_{low}$ space falling in regions where rain is likely do simulate rainfall amounts close to observations. In these two cases, however, the buoyancy generation of veering winds was much stronger in the real world than in the model.

The other two cases where the model underestimates rainfall are Runs V11 (01 June 1996) and V13 (18 August 1996). On 18 August, rainfall was already occurring at the raingauges at the time of model initialization, and a domain average of 1.0 cm had already fallen by 10 am. Late afternoon rainfall at three of the nine observation stations increased the final observed average to about 1.6 cm. July 21, the day simulated by Run V18 mentioned above, also had about 2/3 of observed rainfall occur before 10 am. The only other day with rainfall observed in the morning was 17 August (Run V116). The wet soil model run for this day matches the final observed rainfall of about 1 cm quite well, but this 1 cm is modelled as late afternoon rainfall rather than morning rainfall. These three cases suggest that the model cannot appropriately simulate convection immediately after initialization. Time series of temperature, humidity, and surface fluxes are typically smooth from the beginning of the simulations. Sometimes spin-up effects are seen in abrupt jumps from the initial values to the 7 am values, but they never extend into the second hour of the simulation. This suggests that poor simulation of morning rain in these three cases is not due to spin-up problems. Other explanations are being investigated, along with explanations for the lack of modelled rain in Run V11.
4.9.3 Conclusions Regarding Model Performance

By the metric of simulated versus observed rainfall, the version and configuration of MM5 used in this thesis works well in the majority of circumstances common to Illinois summers. Exceptions to this statement—circumstances where the model performs poorly—include

- anomalous early-morning conditions:
  - rainfall occurring at initialization time, or
  - a well-developed super-adiabatic surface layer present at initialization time; and,
- rainfall events on the high end of observations for this Illinois setting (rainfall depths in excess of 1.5 cm, averaged over very large areas; individual station depths might greatly exceed this), particularly with strongly veering winds.

Rainfall in other circumstances is simulated well by the model. Note that not all veering wind scenarios are simulated poorly. This will be apparent in the details of the simulation results presented in the next chapter.
Chapter 5

MM5 Results

This chapter describes results of three-dimensional modeling work using the Fifth-Generation Penn State/NCAR Mesoscale Model (MM5), described in Chapter 4. Based on the one-dimensional boundary layer modeling detailed in Chapters 2 and 3, expectations for three-dimensional modeling results were formed according to Figure 3.20. As the figure depicts, atmospheric conditions were expected to control the outcome whenever the $CTP < 0$ J/kg (no rainfall), or the $HI_{low} > 15^\circ C$ (no rainfall), or the $HI_{low} < 5^\circ C$ (rainfall, as long as the $CTP > 0$ J/kg). In the region where $CTP > 0$ J/kg and $5 < HI_{low} < 15^\circ C$, the soil moisture condition was expected to significantly impact the outcome of the simulation, with higher values of the two measures favoring convection over dry soils, and lower values favoring convection over wet soils. Figure 5.1 shows the distribution of domain averages of initial values of $CTP$ and $HI_{low}$ for the entire suite of MM5 simulations. (Domain averages refer to the interior focus region of the full domain described in the previous chapter, Figure 4.1.) As the following discussion will explain, expectations based on the 1D results were met, conditioned on the favorable configuration of the winds.

The distribution of domain averages of initial $CTP$ and $HI_{low}$ from the model runs (Figure 5.1) is not the same as the observed early-morning $CTP-HI_{low}$ distribution from Illinois. Sixty-eight cases are from the summer of 1996, nine are from 1997, 14 are from 1998, and seven are from 1999. Almost all days with data available from the summer of 1996 were simulated, both to cover the range of observed $CTP-HI_{low}$ combinations and to
Figure 5.1: Question marks are used to indicate domain average (for focus region shown in Figure 4.1) initial $CTP$ and $HI_{low}$ values for each day in the entire suite of MM5 simulations. Expected results, based on 1D results, are indicated by the various encircled regions. Actual results are presented in the separate sections of this chapter. Sixty-eight cases are days from the summer of 1996, 9 from 1997, 14 from 1998, and 7 from 1999.

provide ample data for the comparisons with observed rainfall presented in Section 4.9. Days from other summers were specifically selected for their $CTP-HI_{low}$ characteristics, in order to better understand the behavior in each of the regimes. This led to a greater frequency of days in the dry soil advantage regime in the model runs (Figure 5.1) than would typically be observed in a given summer in Illinois.

The most significant change between the 3D and 1D models used in this work was the addition of wind effects in the 3D model. Advection of wind from areas unrepresentative of the local land surface conditions can skew the interpretation of local effects if not properly accounted for. Over the relatively flat and homogeneous terrain of the Illinois simulation region, the upstream land surface conditions are quite similar to those in the center of the simulation domain, so the impact of these effects are assumed to be minimal. This is par-
particularly true in this version of MM5 (version 2.12) where soil moisture is represented by a moisture availability at the surface which is dependent on vegetation type. The only vegetation type in the modeled domain is agriculture, which has a default moisture availability of 30%. Many simulations were run at this value, but the simulations which will be discussed here have moisture availabilities set at 80% in the wet soil runs and 10% in the dry soil runs. For each set of initial and boundary conditions, the fluxes, temperature, humidity, and rainfall amounts of the 30% runs were consistently in between the results of the 10% and 80% runs.

A very important impact of the 3D winds is the potential for backing winds or unidirectional winds with great shear to suppress convective potential. Due to this suppression of convection in certain wind conditions, far fewer simulations produced rain than would be anticipated based solely on the 1D framework of understanding. However, when the winds allowed, convection occurred in a manner consistent with the 1D-based expectations. Generally speaking, in the region where dry soils were expected to have an advantage, convection was triggered over dry soils more often than over wet; in the region where wet soils were expected to have an advantage, convection was more frequently triggered over wet soils than over dry. Additionally, when triggering was not restricted, rainfall depths were typically greater over wet soils.

An additional set of 3D simulations were run to highlight the impact of 3D wind effects. A subset of the full suite of simulations were run with the observed initial and boundary winds reduced to 10% of their original values. This greatly reduces the most important effect that is absent from the 1D simulations, suggesting that the results should be similar to the 1D-based expectations, without the caveats based on wind configurations. Rainfall is triggered much more frequently in these simulations than in those with observed winds, and rainfall depths are most often larger than in the counterpart normal wind runs.

This chapter is organized with separate sections focusing on each of the expected outcome regions shown in Figure 5.1.
5.1 Atmospherically controlled: Too dry for rain

When the early-morning atmosphere is quite dry, rainfall cannot be triggered regardless of the flux partitioning at the surface. The $HI_{low}$ cutoff value determined from the 1D work presented earlier is 15°C. Most of the days that fell into this region occurred when the domain was under a strong high pressure system. Such a system would typically be accompanied by subsidence, bringing dry, cold air from aloft down to lower levels. Indeed, this would lead to the high humidity deficits exhibited in each of these cases.

The results presented in Figure 5.2 show that 19 of the 22 simulations in this region produced no convective activity. One run had shallow clouds over dry soils only, and two runs close to the 15°C cutoff had rainfall over both wet and dry soils. One of these rainy runs, V10 (domain average $CTP = 261$ J/kg, $HI_{low} = 15.9°C$), was one of the six outliers discussed in Section 4.9 with veering winds which increased buoyancy and allowed rainfall when it otherwise would not have occurred. Indeed, no rainfall occurred in the reduced-wind run of case V10. The other of these rainy runs, V116 (domain average $CTP = 105$ J/kg, $HI_{low} = 17.5°C$), was also mentioned in Section 4.9. This run was one of the three scenarios with observed rainfall at the time of model initialization. Each of these three cases were poorly simulated by the model.

5.2 Atmospherically controlled: Too stable for rain

When the early-morning atmosphere is very stable, usually as a result of an upper-air inversion (frequently with inversion base between 800 and 700 mb), then, as in the very dry atmospheric conditions, rainfall cannot be triggered regardless of the flux partitioning at the surface. The $CTP$ cutoff value determined from the 1D work presented earlier is 0 J/kg. The results presented in Figure 5.3 show that 12 of the 15 simulations in this region produced no clouds or rainfall. (Note that some simulations fall into both the too stable and the too dry regions.)

The three runs that did show some convective activity all had $HI_{low}$s less than 7°C where, according to the one-dimensional results, shallow clouds were likely to result over wet soils.
Rain and cloud triggering given initial CTP and HIlow, observed winds

Atmosphere too dry for rain

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Run V18 (average CTP = -18 J/kg, average HI_low = 6.6°C) and Run V73 (average CTP = -61 J/kg, average HI_low = 5.7°C), were two of the six outliers discussed in Section 4.9. Run V18 was one of the cases where strongly veering winds greatly enhanced convection in the observations, but only weakly enhanced convection in the model. This is also one of the three cases where the observed rainfall occurred in the morning: conditions that the model simulated poorly in all three circumstances. It is interesting to note that the modelled rainfall in Run V18 occurred only over wet soils, and only in the portion of the domain where the CTP was greater than zero, consistent with the CTP-HI_low framework.

Run V73, on the other hand, was one of the model over-estimates: a day with only minimal rainfall at two of nine nearby raingauges. Observations on this day, then, are consistent with the CTP-HI_low framework, though the simulations are not. See Section 4.9 for specifics of this case.

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Figure 5.2: Outcome of the MM5 simulations with initial conditions falling in the atmospherically controlled region that is too dry for rainfall to occur (HI_low > 15°C).
Rain and cloud triggering given initial CTP and Hlow, observed winds

Atmosphere too stable for rain

Rain over both
Clouds over both
Nothing over either
... over wet soils only
... over dry soils only

Figure 5.3: Outcome of the MM5 simulations with initial conditions falling in the atmospherically controlled region that is too stable for rainfall to occur \((CTP < 0 \text{ J/kg})\).

5.3 Atmospherically controlled: Rainfall expected

Rainfall is expected over both wet and dry soils when the early-morning atmosphere is close to saturation (yielding an \(HI_{low} < 5^\circ \text{C}\)) and exhibits some degree of instability \((CTP > 0 \text{ J/kg})\). Figure 5.4 shows that the MM5 results do not fit the 1D-based expectations as closely as anticipated. Despite domain-average instability and very low humidity deficits, two of six cases show no rain over either wet or dry soils, and one rains only over wet soils. Note that the total rainfall depth was greater over wet soils than over dry soils in three of the four cases with rain. On average, the domain-average rainfall was 0.65 cm over wet soils, but only 0.48 cm over dry soils. In the cases where rainfall occurred in at least one of the soil moisture cases, rain over the wet soils was 0.98 cm, but only 0.71 cm over the dry soils. This is consistent with the results from the 1D work showing higher \(CAPEs\) over wet soils (Section 3.2.1), since higher \(CAPEs\) are typically associated with higher rainfall depths.
In the six scenarios that fall in this Rainfall Expected Region, the main differences between the runs with rain and those without is in the vertical profile of the winds. The next section, highlights two cases to show that strongly sheared winds can suppress convection. These two cases have similar $CTP$ and $HI_{low}$ values, but markedly different wind profiles. The discussion of the Wet Soil Advantage Regime results (Section 5.4) includes specifics of cases which demonstrate that strongly backing winds can also suppress convection by decreasing the buoyancy of rising air parcels.

5.3.1 The effects of strongly sheared winds

Both runs V42 and V22 fall well within the interior of the Rainfall Expected Region of our $CTP-HI_{low}$ space, yet one ends with rainfall over much of the domain (Figure 5.5) and one does not. The rainy scenario, run V42 ($CTP = 113$ J/kg, $HI_{low} = 4.0^\circ$C, Figure 5.6), has gentle veering and shearing of the winds in the lowest 300 mb (Figure 5.8). Run V22 ($CTP = 61$ J/kg, $HI_{low} = 4.5^\circ$C, Figure 5.7), on the other hand, has very strongly sheared unidirectional winds (Figure 5.9) and fails to produce any rainfall over either land surface condition.

The most striking contrast between the soundings of runs V42 and V22 is evident in the trace of the hodographs: V22 is so strongly sheared in this region that deep convection is not able to develop. Ziegler and Rasmussen (1998) observed this in their analysis of data from the COPS (Central Oklahoma Profiler Studies project) and VORTEX (Verification of the Origins of Rotation in Tornadoes Experiment) field experiments. They found many cases where the convective initiation energy (CIN) went to zero, but convection was not triggered because of excessive wind shear. Since highly sheared winds enhance mixing between updrafts from low levels and typically drier air from higher levels, shearing tends to be accompanied by drying of the updraft air. This, in turn, will elevate the Lifted Condensation Level (LCL) and the Level of Free Convection (LFC), making convection more difficult to trigger. Barnes and Newton (1986) also note that though the slantwise organization of convection caused by pronounced wind shear creates an efficient thermodynamical-mechanical process, the precipitation efficiency of squall lines and large thunderstorms actually decreases with an increase
Rain and cloud triggering given initial CTP and HIlow, observed winds

Rain expected over wet and dry soils

Rainfall depths over wet (blue) and dry (red) soils, observed winds

Dry soil ave = 0.71 cm
Wet soil ave = 0.98 cm
(Only when one > 0.1 cm)

Dry soil ave = 0.48 cm
Wet soil ave = 0.65 cm
(All runs, including rain-free)

Figure 5.4: Outcome of the MM5 simulations with initial conditions falling in the rainfall expected region ($HI_{low} < 5^\circ C, CTP > 0$ J/kg).
Figure 5.5: Total rainfall depths over dry soils (left figure) and over wet soils (right figure) in run V42 (initial conditions shown in Figures 5.6 and 5.8).
Figure 5.6: Initial $CTP$ (left figure) and $HI_{low}$ (right figure) for run V42. Rainfall is modelled over both wet and dry soils.
Figure 5.7: Initial $CTP$ (left figure) and $H_{low}$ (right figure) for run V22. No rain occurs in model runs with wet or dry soils.
Figure 5.8: Initial domain average sounding for run V42. The solid red line marks the trajectory of a lifted surface parcel. The hodograph tracks the winds in the lowest 300 mb.

Figure 5.9: As in Figure 5.8, but for run V22.
of vertical shear. Ziegler and Rasmussen (1998) determined that “moist boundary layer air parcels must be lifted to their lifted condensation level and level of free convection prior to leaving the mesoscale updraft to form deep convection” (p. 1106). Furthermore, they found that “initiation of forced or active cumulus convection requires that the magnitude of the horizontal flux of dry air ... be locally negligible in relation to the vertical flux of moist air in the mesoscale updraft below the LCL or LFC, respectively” (p. 1126).

In the one-dimensional boundary layer modeling used to develop the $CTP-HI_{low}$-based expectations, the assumed trigger for convection was $CIN=0$. (Actually, triggering could even occur when $CIN$ was slightly positive (order $\lesssim 5$ J/kg), since turbulence can often overcome small amounts of $CIN$. See the 1D model development chapter for more details.) Given the above observations from field studies, it is not surprising that there would be fewer rainy cases in the full three-dimensional simulations than predicted by this assumption. In order to study these wind effects in more detail, another set of MM5 experiments were performed where the boundary and initial winds were reduced to 10% of their actual values. Model-calculated winds within the domain were not altered from their calculated values: only the forcing winds were reduced. These reduced-winds runs should more closely mimic the 1D simulations, since the most important 3D effect has been severely minimized.

Figures 5.10 and 5.11 show the resultant rainfall for the reduced-winds runs for V42 and V22, respectively. Reducing the highly sheared winds in run V22 unleashed torrential (and probably not particularly realistic) downpours in both the wet and dry soil runs. In sharp contrast, the total rainfall in scenario V42 is actually less in the reduced winds runs than in the normal winds runs. This is because some degree of shear is helpful for rainfall production since mild shearing allows the convective downdraft to develop downwind of the updraft, rather than directly on top of it (Barnes and Newton [1986]). In addition, veering winds impart additional buoyancy to rising air. (The buoyancy effect of veering winds can be explained through the thermal wind equation, and will be discussed in more detail in the next section.) When the forcing winds were reduced, these influences were removed.

Figure 5.12 shows that all 5 of the cases from the Rainfall Expected Regime run with reduced winds produced significant amounts of rainfall. This includes the 3 cases that did not rain in both soil conditions with the normal winds, and 2 of the 3 cases that produced rain.
over both soil moisture states. One case was not run because of the demand for computing time.

### 5.4 Wet Soil Advantage Regime

Results of MM5 simulations falling in the Wet Soil Advantage Regime are summarized in Figure 5.13. The most striking feature of this figure is the lack of convection in this region. Based on the 1D expectations, rainfall should definitely occur over wet soils, and is likely to occur over dry soils in all of these cases. In stark contrast to these expectations, rain occurs over dry soils in only 3 of 29 cases, and over wet soils in only 6 of these 29 cases. However, as in the Rainfall Expected Region, when rain does occur, rainfall depths are consistently larger over wet soils than over dry soils: 0.62 cm versus 0.36 cm in the runs with rain over at least one of the soil moisture conditions, and 0.16 cm versus 0.09 cm when all cases are averaged. Again, this is consistent with the 1D results and with a small but significant positive soil moisture-rainfall feedback in Illinois.

As mentioned in the previous section, the suppression of convection is due to the influence of the 3D wind profiles. We have already provided an example and an explanation about the impact of excessive unidirectional shear on convection; we will now discuss the thermal wind equation and the impact of backing winds on the buoyancy of rising air. One or both of these wind conditions occurred in each of the non-convective cases in the Wet Soil Advantage Regime.

#### 5.4.1 Effects of Thermal Wind on Convection

The thermal wind equation relates the vertical shear of the geostrophic wind to the horizontal temperature gradient. To discuss this equation, we must first begin by defining the geostrophic wind. (For more complete descriptions, see e.g., Rogers and Yau [1989] or Wallace and Hobbs [1977].) When an air parcel moves, three primary forces act on it and control the direction of the air (and thus the winds): a pressure gradient force, the Coriolis force, and friction. The pressure gradient force accelerates the parcel from high to low pressure.
Figure 5.10: Total rainfall depths over dry soils (left figure) and over wet soils (right figure) in run V42 with winds reduced to 10% of observed. (Compare with rainfall in run V42 with observed winds: Figure 5.5.)
Figure 5.11: Total rainfall depths over dry soils (left figure) and over wet soils (right figure) in run V22 with winds reduced to 10% of observed. No rainfall occurred in the simulations with observed winds. Reducing the winds removed the effects of strong unidirectional wind shear and allowed large amounts of rain to fall.
Rain and cloud triggering given initial CTP and Hllow, 10% of observed winds

Rain expected over wet and dry soils

Rainfall depths over wet (blue) and dry (red) soils, 10% of observed winds

Dry soil ave = 2.40 cm
Wet soil ave = 2.61 cm
(All are > 0.1 cm)

Figure 5.12: As in Figure 5.4, but for runs with reduced winds. Only 5 of the original 6 simulations were performed.
Rain and cloud triggering given initial CTP and Hllow, observed winds

Rainfall depths over wet (blue) and dry (red) soils, observed winds

Figure 5.13: Outcome of the MM5 simulations with initial conditions falling in the Wet Soil Advantage Region. Symbols in upper figure as in Figure 5.2.
The Coriolis force deflects the parcel to the right in the northern hemisphere (to the left in the southern hemisphere), and is dependent on the wind speed. When the Coriolis force and the pressure gradient force are of equal magnitude but opposite directions, geostrophic balance is achieved and the flow is parallel to isobars, with low pressure on the left (in the northern hemisphere). Frictional effects act to slow the winds, which in turn reduces the Coriolis force so that it is not as large as the pressure gradient force. This leads to some component of the flow across the isobars, down the pressure gradient towards the low pressure center, leading to convergence above the low. Geostrophic balance occurs when frictional effects are negligible and the pressure gradient force and the Coriolis force are in balance. It is described by the following equations:

\[
\begin{align*}
-\frac{1}{\rho} \frac{\partial p}{\partial x} &= - f v_g \\
-\frac{1}{\rho} \frac{\partial p}{\partial y} &= + f u_g
\end{align*}
\]

The pressure gradient terms are on the left, and the Coriolis terms are on the right, with \( u_g \) and \( v_g \) representing the geostrophic wind components and \( f = 2\Omega \sin(\text{latitude}) \) the Coriolis parameter.

Since the geostrophic wind is related to the horizontal gradient of pressure, any vertical variation in this pressure gradient will lead to a vertical variation in the geostrophic wind. Furthermore, since gradients of pressure are related to gradients in temperature, horizontal gradients of temperature are related to vertical shear in the geostrophic winds. The thermal wind equation describes this relationship, and is obtained by differentiating Equation 5.2 with respect to \( z \), and making use of the hydrostatic equation and the definition of potential temperature:

\[
\begin{align*}
-f \frac{\partial v_g}{\partial z} &= \frac{\partial}{\partial z} \left( -\frac{1}{\rho} \frac{\partial p}{\partial x} \right) = - \frac{g}{\theta} \frac{\partial \theta}{\partial x} \\
f \frac{\partial u_g}{\partial z} &= \frac{\partial}{\partial z} \left( -\frac{1}{\rho} \frac{\partial p}{\partial y} \right) = - \frac{g}{\theta} \frac{\partial \theta}{\partial y}
\end{align*}
\]

This equation tells us that the geostrophic wind is constant with height only when the
potential temperature is uniform in the horizontal. If there is a horizontal temperature gradient, then the geostrophic wind will vary with height. The vector difference between the geostrophic wind at two levels is termed the thermal wind.

The change of the wind direction with height is characterized by the terms veering and backing. Backing of the winds occurs when the geostrophic wind vector turns in the same sense as the planetary rotation. Thus, in the northern hemisphere, backing winds are when the geostrophic wind vector rotates in a counterclockwise direction with increasing height. Veering, on the other hand, is when the geostrophic wind vector rotates in the opposite sense as the planetary rotation with increasing height (clockwise in the northern hemisphere). As discussed extensively in Wallace and Hobbs (1979, pp. 387-390) and Barnes and Newton (1986), for example, the differential advection of temperature in different layers of the atmosphere caused by this thermal wind effect can alter atmospheric stability. Backing winds lead to advection of air from the colder portion of the region defined by thermal gradients into the warmer portion. Conversely, veering winds lead to warm advection. Thus, veering winds lend additional buoyancy to lifting air parcels, while backing winds decrease the buoyancy. This conclusion has significant implications for the interpretation of the results of the MM5 simulations.

Figure 5.14 shows the 300 mb hodographs for two cases in the wet soil advantage regime where reduction of the initial and boundary winds increased the convective activity (Runs V38 and V58). Note that in both cases, the winds back with height. Since backing winds are associated with cold air advection and a decrease of buoyancy, removing the winds allows rising parcels to maintain their surface buoyancy, thereby increasing the convective activity. Both runs yielded no convection with these observed winds, while rain developed over both wet and dry soils in the reduced-wind runs. The $CTP$ and $HI_{low}$ of Run V38 were nearly constant throughout the domain, at 93 J/kg and 7.6°C, respectively (not shown). The rainfall distributions over wet and dry soils for the reduced winds runs of case V38 are shown in Figure 5.15.

It is important to note that though backing winds reduce the buoyancy of rising air, they do not necessarily fully suppress convection. Run V95 (Figures 5.16 and 5.17) is an example of a day with backing winds where rainfall still developed over both dry and wet
soils (Figure 5.18). However, as in the other cases with backing winds, removing the negative effect of the winds allows even more rain (and perhaps unrealistic amounts of rain) to develop (Figure 5.19).

Many other cases in this wet soil advantage regime are also limited in their production of rainfall by the observed winds. Figure 5.20 shows the results of all the reduced-wind simulations from this regime. Rainfall frequency and depth are both significantly increased by removing the winds, and rainfall depths remain greater over wet soils than over dry soils (1.22 cm vs. 1.07 cm).

5.5 Transition Region

MM5 results for runs in transition region determined from the 1D work are shown in Figure 5.21. This region can be effectively eliminated from consideration by extending the bounds of the Dry Soil Advantage Region from a low CTP value of 200 J/kg to 150 J/kg. This is done in the next section.
Figure 5.15: Total rainfall depths over dry soils (left figure) and over wet soils (right figure) in run V38 with winds reduced to 10% of observed. Reduction of the winds removed the buoyancy-reduction effects of backing winds and allowed rainfall to occur.
5.6 Dry Soil Advantage Regime

The results of the observed winds simulations for cases in the dry soil advantage regime are presented in Figure 5.22. This figure shows that rainfall is triggered more frequently over dry soils than over wet, as anticipated (eleven times versus seven times). Additionally, the average rainfall depths no longer favor wet soils: they are now essentially equal at 0.26 cm over wet soils and 0.24 cm over dry soils. Five of the seven cases where rain occurs over both soil types have more rainfall over wet soils, but in two cases the rainfall depth is greater over dry soils, and in four additional cases rainfall only occurs over dry soils. As predicted by the 1D modeling work, triggering can potentially occur over both wet and dry soils in this regime, but boundary layers over dry soils are more likely to reach the neutrally bouyant layers which yield the high $CTP$. In the five cases where rainfall is greater over wet soils, the boundary layer over both soil conditions was able to tap into this neutrally bouyant layer. In the six cases with more rainfall over dry soils, the boundary layer over the wet soils did...
Figure 5.17: Initial $CTP$ (left figure) and $HI_{low}$ (right figure) for run V95.
Figure 5.18: Total rainfall depths over dry soils (left figure) and over wet soils (right figure) in run V95 with observed winds.
Figure 5.19: Total rainfall depths over dry soils (left figure) and over wet soils (right figure) in run V95 with winds reduced to 10% of observed. Reduction of backing winds allowed more rainfall to occur than in the observed winds runs (Figure 5.18).
Rain and cloud triggering given initial CTP and H\textsubscript{low}, 10\% of observed winds

Wet soil advantage region

Rainfall depths over wet (blue) and dry (red) soils, 10\% of observed winds

Dry soil ave = 2.12 cm
Wet soil ave = 2.43 cm
(Only when one > 0.1 cm)

Dry soil ave = 1.07 cm
Wet soil ave = 1.22 cm
(All runs, inc. rain–free)

Figure 5.20: As in Figure 5.13, but for reduced winds.
Rain and cloud triggering given initial CTP and Hllow, observed winds

Figure 5.21: Outcome of the MM5 simulations with initial conditions falling in the 1D-based Transition Region. The half of this region with CTP > 150 J/kg will be added to the Dry Soil Advantage Regime. Symbols as in Figure 5.2.

not grow high enough early enough in the day to benefit from the high CTP zone, though in two of these cases there were small pockets of rain over wet soils.

Run V90 is a good example of the advantage that boundary layers growing over dry soils have in these high CTP environments. The domain-average initial sounding (Figure 5.23) shows an extensive zone between 945 mb and 710 mb with a lapse rate that is nearly dry adiabatic. The CTP and H llow in this case are 282 J/kg and 13.7°C, respectively (Figure 5.24). Figure 5.25 shows that six hours into the run (local noon), the boundary layer has grown to 3.5 km over dry soils, but only to 2 km over wet soils. This allows for convection to occur over the dry soils, but not over the wet, despite the 7°C difference in θE between the two simulations at the time of triggering. Figure 5.26 shows noontime soundings at a point where rain occurs over dry soils but not over wet. It is clear that clouds have already developed at this time over dry soils, and surface parcels can freely convect. Over wet soils, however, surface parcels cannot reach their level of free convection at noontime, and Figure 5.25 shows that the boundary layer does not grow any deeper and θE does not increase.
Rain and cloud triggering given initial CTP and Hllow, observed winds

Dry soil advantage region

Rainfall depths over wet (blue) and dry (red) soils, observed winds

Dry soil advantage region

Dry soil ave = 0.43 cm
Wet soil ave = 0.48 cm
(Only when one > 0.1 cm)

Dry soil ave = 0.24 cm
Wet soil ave = 0.26 cm
(All runs, including rain-free)

Figure 5.22: Outcome of the MM5 simulations with initial conditions falling in the Dry Soil Advantage Region. Symbols in upper figure as in Figure 5.2.
from the noontime values. Thus, the higher boundary layer growth over dry soils has allowed for convective triggering, while the high moist static energy in the BL over wet soils is not large enough to trigger convection in this high $CTP$ environment.

As in the wet soil advantage regime, wind effects play an important role in determining the production of rainfall in the dry soil advantage regime. Figure 5.27 shows the results of 12 reduced-winds simulations from this region. In three of the runs, convective activity decreased when the winds were reduced to 10% of observations. In each of these runs, the original winds veered with height, as shown in the hodographs of Figure 5.28. In three other runs, convective activity increased when the winds were removed. Figure 5.29 shows that none of these original wind profiles veer enough to bring additional buoyancy to rising air in the normal winds runs: removing these winds allows for deeper convection to occur.

5.7 Conclusions from MM5 simulations

Conclusions drawn from this work are threefold:
Figure 5.24: Initial \( CTP \) (left figure) and \( H_{low} \) (right figure) for run V90.
Figure 5.25: Time series of fractional coverage of rainfall (upper left), cumulative rainfall depth (in cm, upper right), planetary boundary layer height (in m, lower left), and $\theta_E$ (in K, lower right) in the wet soil (solid black lines) and dry soil (dashed green lines) simulations for case V90.

- The Convective Triggering Potential ($CTP$) offers significant information regarding the likely impact of the land surface condition on the potential for the triggering of convection, particularly when coupled with a measure of the humidity in the lowest levels of the atmosphere (e.g., $H_{I_{low}}$).

- The land surface condition can impact the potential for convection only when the atmosphere is not already predisposed to convect or not to convect. This atmospheric predisposition can be determined by analyzing the $CTP$, the $H_{I_{low}}$, and the vertical profile of the winds.

- Areas such as Illinois exhibit a small but significant positive feedback between soil moisture and rainfall because the frequency of days falling in the wet soil advantage
Figure 5.26: Noontime soundings (6 hours into the simulations) for run V90 over dry soils (left figure) and over wet soils (right figure). Rainfall at this point occurs over dry soils but not over wet.
Rain and cloud triggering given initial CTP and Hllow, 10% of observed winds

Rainfall depths over wet (blue) and dry (red) soils, 10% of observed winds

Dry soil advantage region

Dry soil ave = 0.61 cm
Wet soil ave = 0.51 cm
(Only when one > 0.1 cm)

Dry soil ave = 0.32 cm
Wet soil ave = 0.25 cm
(All runs, including rain-free)

Figure 5.27: As in Figure 5.22, but for reduced winds.
Figure 5.28: Hodographs to 300 mb above ground surface (AGS) for the three cases in the dry soil advantage regime where reduction of the initial and boundary winds reduced the convective activity. Note that in all three cases, the winds veer with height (run V20 veers only to 200 mb AGS). Run V9 rained over both wet and dry soils in the normal wind runs, but only over dry soils in the 10% wind runs. Runs V20 and V25 rained only over dry soils in the normal wind runs, but V20 only produced shallow clouds over dry soils and V25 showed no convective activity in the 10% wind runs.

Figure 5.29: Hodographs to 300 mb above ground surface for the three cases in the dry soil advantage regime where reduction of the initial and boundary winds increased the convective activity. All three runs rained over both wet and dry soils in the 10% wind runs, while in the normal wind runs runs V33 and V44 had shallow clouds over dry soils only and run V34 rained over dry soils only.
regime of \( CTP - HI_{low} \) space exceeds the frequency of days falling in the dry soil advantage regime.

As a follow-up to the third point, it is important to recall that the distribution of days simulated with MM5 is not fully representative of the observed distribution in Illinois. Most of the days simulated (68) were from the summer of 1996; other days were chosen from the summers of 1997 (9 days), 1998 (14 days) and 1999 (7 days), specifically to find conditions in the regions of interest, particularly the dry soil advantage regime. Therefore, the distribution of initial conditions for the MM5 runs has a larger percentage of dry soil advantage days than would normally be observed in Illinois.

It is relevant to note the relationship between these results and the work that originally inspired this investigation of atmospheric controls on soil moisture-rainfall interactions. Findell and Eltahir (1997) found a small but significant positive feedback between soil moisture and rainfall in Illinois. Expanding on this work, Findell and Eltahir (1999) used near-surface atmospheric data and found a significant correlation between soil moisture and wet-bulb depression, \( T_{dpr} \), and then between \( T_{dpr} \) and subsequent rainfall. They did not, however, find a significant correlation between soil moisture and wet-bulb temperature, \( T_w \), or between \( T_w \) and subsequent rainfall.

The current results seem to be consistent with these findings. \( HI_{low} \) should be closely correlated with \( T_{dpr} \), since it considers the dew point depression at relatively low levels (specifically 950 mb and 850 mb). Given the importance of \( HI_{low} \) in the current results, it is not surprising that the surface wet-bulb depression is also a helpful indicator of the link between the land and the atmosphere. The wet-bulb temperature, on the other hand, is a measure of the surface energy, much like \( \theta_E \). The current work shows that the surface energy alone is not enough to determine either the potential for rainfall or the impact of the surface moisture on this potential. The \( CTP \) is helpful in both of these determinations because it considers the temperature profile well above the surface, and because it focuses on the portion of the atmosphere that is between the region that is almost always incorporated into the growing boundary layer and the portion of the free atmosphere that is almost never incorporated into the growing BL.
The nature of the atmospheric structure in this critical region of the troposphere, about 1 to 3 km above the ground surface, determines the manner in which soil moisture can impact rainfall. A positive feedback is likely when the temperature profile in this region is close to moist adiabatic. In these circumstances, convection is most easily triggered by increasing boundary layer moist static energy (MSE) because this greatly reduces the level of free convection. The high latent heat flux over wet soils increases the BL MSE more than the smaller latent heat flux over dry soils. A negative feedback is likely when this region has a temperature profile close to dry adiabatic. In these circumstances, convection is most easily triggered by increasing the height of the BL: a process requiring a high sensible heat flux like that seen over dry soils.

In the next chapter, we look at radiosonde data from stations throughout the contiguous 48 United States to find regions where these different circumstances are commonly observed.
Chapter 6

Nationwide Application of the CTP as an Indicator of the Potential for Soil Moisture-Rainfall Feedbacks

6.1 Introduction

With the one-dimensional boundary layer modeling described in the early chapters of this thesis, it was discovered that the critical region of the atmosphere which determines the potential of the land surface to influence the triggering of convection lies between 100 and 300 mb above the ground surface (AGS). Important characteristics of this critical region were captured by the Convective Triggering Potential (CTP). When coupled with a measure of the low-level humidity deficit, the CTP provided an effective means of determining when a positive or a negative feedback between the soil moisture condition and the development of rainfall might occur in Illinois. The full analysis included one-dimensional boundary layer modeling of 12Z soundings from three summers at a station in Illinois. Plotting the initial CTP and $HI_{low}$ from these 12Z (6 am in Illinois) soundings on a scatter plot like that of Figure 3.20 revealed that more days fell in the wet soil advantage regime than in the dry soil advantage regime, suggesting that there should be a small but significant positive soil moisture-rainfall feedback in Illinois. This is consistent with the analyses of Findell and

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Eltahir (1999, 1997), and with the 3D work described in Chapter 5.

Similar $CTP-HI_{low}$ scatter plots have been developed for 76 radiosonde stations throughout the contiguous 48 United States using data from the summer (June, July, August) of 1998. The $CTP$ was initially developed in a region with surface pressures close to 1000 mb (typically in the 990's). Typical surface pressures at the other 75 radiosonde stations, however, are often far from 1000 mb. The $CTP$ values shown in all the plots in this thesis are calculated relative to the surface pressure, with the critical region defined as $P_{surf} - 100$ mb to $P_{surf} - 300$ mb. In Illinois, this generally meant ~900 mb to ~700 mb. The $HI_{low}$ values reported from Illinois, on the other hand, were defined at specific pressure levels, namely 950 mb and 850 mb. In order to extend the $HI_{low}$ concept to other regions, the $HI_{low}$ is re-defined for all the plots of this chapter as the sum of the dew point depressions 50 and 150 mb above the surface.

These scatter plots reveal ten region across the US (Figure 6.1): within each of these regions, the $CTP-HI_{low}$ characteristics are relatively uniform. In five of these regions, primarily in the western half of the country, almost all of the days fall within the three atmospherically controlled regimes (too dry for rainfall, too stable for rainfall, rainfall expected), leaving little possibility for soil moisture conditions to impact convective triggering. In the eastern half of the country, two regions, including one encompassing Illinois, show signs of positive feedbacks between soil moisture and rainfall. One region shows a strong potential for a negative feedback, and two regions show significant occurrences of days in both the wet soil advantage regime and the dry soil advantage regime. Each of these ten regions will now be discussed. Following these regional discussions, four stations will be analyzed with the 1D boundary layer model described in Chapter 2, following the techniques used in Chapter 3 with soundings from Illinois. These additional results confirm that the upper and lower bounds on the critical region of the atmosphere are independent of mid-latitude location, and that the $CTP-HI_{low}$ framework is a robust indicator of soil moisture-rainfall feedbacks.
Figure 6.1: Representative regions, based on $CTP-H_{1low}$ scatter plots from 76 radiosonde stations.
6.2 Atmospherically controlled regions

Each of the five regions encircled in black in Figure 6.1 showed few, if any, days in either the wet soil advantage regime or the dry soil advantage regime. Almost 100% of days from the summer of 1998 were atmospherically controlled in the Pacific Northwest (Figure 6.2). Most of the early morning soundings were stable ($CTP < 0 \text{ J/kg}$), and a very large percentage were very humid, with $HI_{low} < 5^\circ C$. Bryson and Hare (1974), in their review of the climatic patterns of North America, state that westerlies off of the north Pacific arrive on the west coast cool, with a nearly moist adiabatic lapse rate, and with high humidity to a considerable depth. The 1998 summer soundings from Salem, Oregon (SLE) agree with this general description, but they also frequently exhibit a strong inversion around 850 mb with saturation or close to saturation below this level, indicating the existence of rain or clouds at the time of the sounding. Above the 850 mb inversion, the lapse rate is commonly moist adiabatic, as anticipated by the Bryson and Hare statement, but this mid-$CTP$ region inversion creates strong stability which should prohibit deepening of the pre-existing low intensity shallow clouds and/or rainfall.

Soundings from the Pacific Southwest also fall almost entirely in atmospherically controlled regions (Figure 6.3), though the distribution in $CTP$-$HI_{low}$-space is very different than at the coastal stations to the north. Most of these soundings are very dry, with many $HI_{low}$ values greatly exceeding $15^\circ C$. These dry atmospheres are no doubt a result of the influence of the anticyclonic system that resides over the Pacific (Bryson and Hare, 1974). In most of the western states, thunderstorms are suppressed by subsidence on the eastern side of this oceanic anticyclone. Air coming off the Pacific is further inhibited by the low moist static energy resulting from the cool waters of the California current (Barnes and Newton, 1986). The airstream emerging from the Pacific anticyclone travels south, paralleling the coast, with increasing subsidence through its southward course (Bryson and Hare, 1974). This is apparent in the higher $CTP$ and $HI_{low}$ values at Station NKX, the southern-most of the three Pacific Southwest stations.

Further inland, in the Dry Intermontane region, we see very different scatter plots of early morning $CTP$ and $HI_{low}$. These stations are still predominantly atmospherically controlled,
Figure 6.2: Early-morning (12Z, 4 am local time) $CTP$ and $HI_{low}$ for all available days from June, July, and August of 1998 from the stations in the Pacific Northwest. In the legend, $d$ is the percentage of days in the dry soil advantage regime (would be plotted with a red square, if there were any such days); $w$ is the percentage of days in the wet soil advantage regime (blue circle); $t$ is the percentage of days in the transition region (green triangle); and $a$ is for atmospherically controlled days (black star).

Figure 6.3: As in Figure 6.2, but for stations from the Pacific Southwest region.
with between 82% and 96% falling in these regimes. The original source of air arriving in this region is the same dry air off of the Pacific anticyclone that strongly impacts the Pacific Southwest region. After travelling over the coastal mountains and the western plateau, however, heating from the land surface raises the wet-bulb potential temperature ($\theta_w$) in the lower 1-2 km to values comparable to that of maritime tropical air (Barnes and Newton, 1986). This increase in low-level $\theta_w$ is clearly evident in the increase in $CTP$ from the Pacific Southwest to the Dry Intermontane region. Nearly all of the days at these stations are characterized by positive $CTP$ values, indicating some degree of convective potential. This potential, however, is effectively removed by the extreme aridity of the air: almost all of the atmospherically controlled days are too dry to produce rainfall. The remaining 4 to 18% of days at the different stations are days when the land surface moisture could have an impact on the potential for rainfall. Still, these days are rare enough that no over-riding signal of either a positive or a negative soil moisture-rainfall feedback is expected.

It is interesting to compare the $CTP-HI_{low}$ plots from Phoenix (PHX) and Tucson (TUS), Arizona with the observations presented in Wallace et al. (1999). These authors determined that there were generally two types of soundings observed at Tucson, indicating two very different types of days: dry days and monsoon days. Monsoon-style days were not observed in Wallace et al.'s Phoenix data. The temperature profiles at Tucson were essentially identical on these two types of days, but the dew point temperatures were 4-6°C different far up the soundings. On the dry days, the air derived from the eastern Pacific, with deep westerlies up to the tropopause. The monsoon days, on the other hand, showed southeasterly midtropospheric flow, indicating an influence from the Mexican monsoon to the south (Wallace et al., 1999).

Relating their conclusions to the Phoenix and Tucson plots in Figure 6.4, this influence of the Mexican monsoon could potentially bring moist air at low enough levels to affect the $HI_{low}$. This would be consistent with the increased occurrence of days with $HI_{low} < 15°C$ at Tucson relative to Phoenix: about 20% of days at Tucson versus only about 10% at Phoenix during the summer of 1998. This monsoonal influence is more prominent in the data from the station at Albuquerque, New Mexico (ABQ). This will be discussed in Sections 6.3 and 6.7.1.

There is slightly more variability between stations in the Rocky Mountain region (Fig-
Figure 6.4: As in Figure 6.2, but for stations from the Dry Intermontane region.

Figure 6.5) than in the previously discussed regions. Atmospherically controlled conditions prevail about 81% of the time at the six stations in the central and southern portion of the domain (LKN, SLC, GJT, BOI, DEN, UNR), but only about 68% of the time at the three stations nearest the Canadian border (OTX, TFX, GGW). As with the stations in the Dry Intermontane region to the south and west, the most common source region is air off of the Pacific, the near-surface layers of which again warm (increasing the $CTP$) as it travels inland. Though this air is also coming off the Pacific anticyclone, and therefore shows signs of subsidence, it is not as dry as the air in the regions to the south because the Pacific source is not as far south. As mention earlier, the air traveling parallel to the coast gets increasingly dry as it moves southward (Bryson and Hare, 1974). Thus, days at the Rocky Mountain stations are most typically in the atmospherically controlled regime with $CTP > 0$ J/kg and
$HI_{low} > 15^\circ$C, but the extremely high values of $HI_{low}$ observed in the Pacific Southwest and the Dry Intermontane regions (up to $85^\circ$C) are not as common at these stations. Indeed, the stations which are further north and east (DEN, OTX, TFX, GGW, and UNR) have maximum $HI_{low}$ values only in the 40's, and also have the highest percentage of days with $HI_{low} < 15^\circ$C in the Rocky Mountain Region.

All but one of the nine stations in the Rockies region has at least 10% of days in the dry soil advantage regime, and the three northern stations have closer to 15% of days in this regime. These three stations, however, also have close to 10% of days in the wet soil advantage regime, suggesting that the two influences would balance each other out and lead to a neutral response of rainfall to soil moisture.

The final region with no clear potential for either a positive or negative feedback between soil moisture and rainfall is the Plains States region (Figure 6.6). Of the five stations in this region, Stations LBF, OAX, and TOP have atmospherically controlled conditions about 75% of the time, with the rest of the days being approximately equally divided between the wet soil advantage regime, the dry soil advantage regime, and the transition region. The other two stations (BIS and ABR), which are both in the northern part of the Plains States region, have atmospherically controlled days about 65% of the time, wet soil and dry soil advantage days about 10% of the time, and transition region days about 15% of the time. Like the Rocky Mountain region, the influences of these various regions are expected to balance each other out and lead to a neutral response of rainfall to soil moisture.

### 6.3 Negative Feedback Region: The Dryline and Southwest Monsoon Region

Only seven of the 76 stations distributed throughout the contiguous 48 United States show evidence of a potential negative feedback between soil moisture and rainfall. $CTP-HI_{low}$ scatter plots for these stations are given in Figure 6.7. The western half of the negative feedback region shown in Figure 6.7 is the area of the Southwest American Monsoon, which typically affects New Mexico and much of Arizona during July and August, sometimes beginning in
Figure 6.5: As in Figure 6.2, but for stations from the Rocky Mountain region.
late June or extending into September (Higgins and Shi, 2000). This area will be discussed in detail in Section 6.6 in conjunction with 1D modeling results using soundings from Station ABQ in Albuquerque, New Mexico.

The eastern half of this region corresponds to an area that is frequently characterized by a dryline: a sharp gradient of surface moisture over a very short horizontal distance, commonly with dew-point temperature changes on the order of 15°C in just 2 km (Schaefer et al., 1986). The topographic and synoptic setting of this region create very specific conditions which allow a dryline to develop, and which also allow for the high \( CTP \), moderate \( HI_{low} \) conditions necessary for a negative soil moisture-rainfall feedback to develop.

Carlson and Ludlam (1968) developed a conceptual framework to explain why drylines are often associated with outbreaks of severe storms, particularly in the American southwest.
The critical component in dryline formation is a warm, elevated mixed layer moving over cooler near-surface air at lower elevations, forming a lid with a capping inversion. The Mexican plateau serves as the source region for the lid that frequently forms over much of central and eastern Texas. Indeed, Benjamin (1986) notes that “the time of strongest differential heating between the Mexican plateau and the region to its east coincides with the spring severe storm maximum in the south central US” (p 331). This time is generally between April and June. Schaefer et al. (1986) cite a 1973 study where the same authors looked at all days in April, May and June from 1966-1968 and found drylines present over the Great Plains on more than 41% of the days.

Carlson et al. (1983) refined the conceptual model of Carlson and Ludlam (1968), and
modeled and analyzed three case studies from the SESAME field experiments of 1979. These and other case studies [e.g., Crawford and Bluestein (1997), Hane et al. (1997), Hane et al. (1993), Benjamin and Carlson (1986), Ziegler and Rasmussen (1983), Anthes et al. (1982), Ogura et al. (1982)] helped to establish a complete picture of dryline formation in the American southwest. As stated above, a dryline is an area exhibiting a sharp gradient in surface moisture over a very short horizontal distance. In the Texas region, drylines tend to run close to north-south. On the eastern side of the dryline, surface winds carry moist air northward from the Gulf of Mexico into central and eastern Texas, (more generally, they are easterly or northeasterly into southern Texas, and then turn to the north), while winds from the south and west carry very dry air into western Texas. Air to the west of the dryline is very dry with a nearly adiabatic lapse rate. Over the moist air on the east side of the dryline, a capping inversion is created by air moving off the elevated Mexican plateau. Because the plateau is so much higher than most of Texas, the base of this air mass tends to be located at about 850 mb: in the middle of the CTP region. This lid prevents convection over much of Texas, despite the buildup of moisture and energy within the shallow, capped boundary layer. The dryline represents the edge of a capping inversion, so vertical differential advection can cause moist BL air to flow out from beneath the lid, leading to rapid destabilization and explosive storm development (Schaefer et al., 1986). This process of underrunning brings high $\theta_E$ air to a region of high sensible heat flux, providing the lifting mechanism necessary to raise the air past its level of free convection.

Ziegler and Rasmussen (1983) and Ziegler et al. (1997) studied the initiation of convection at the dryline. The earlier study found that the dryline is a favored zone for cumulus formation, with the peak in cumulus cloud formation occurring about 15 km east of the dryline due to above-surface westerlies carrying the lifted surface air away from the surface dryline location. The later study confirmed that the long, narrow band of moisture convergence is indeed coincident with the dryline and with maximum thermal gradients. As discussed with the results of the MM5 simulations, however, both studies showed that the shear created by the westerlies cannot be too great, or convective initiation will be shut down. Ziegler and Rasmussen (1983) concluded that boundary layer air must be lifted to its lifting condensation level and level of free convection prior to leaving the updraft for clouds
and/or deep convection to form. This need for favorable wind shear was also noted many years earlier by Carlson and Ludlam (1968) in their original formulation of the conditions for the occurrence of severe storms.

Prior to the work mentioned above, the capping inversions frequently seen over Texas were assumed to be caused by subsidence. Carlson and Farrell (1983) discuss the differences between an elevated mixed layer lid and a lid created by a subsidence inversion, revealing that the primary difference is seen in the relative humidity. Above an elevated mixed layer lid, the relative humidity tends to increase with height above the lid base. Additionally, the extreme variation of potential temperature and specific humidity across the lid suggests that the air above and below the lid base are from two completely different air streams. Other locations where such elevated mixed-layer lids are known to occur include France (lid formation over the elevated regions of northern Spain), tropical West Africa (lid formation over the Sahara, according to Carlson and Ludlam [1968]; Schaefer et al. [1986] suggest that the Intertropical Convergence Zone often acts like a dryline), and India during the monsoon (lid formation over Arabia).

6.4 Positive Feedback Regions

Stations in the Gulf Coast region of Figure 6.1 have between 20 and 44% of days in the wet soil advantage regime, but only between 0 and 13% in the dry soil advantage regime (Figure 6.8). Though these statistics suggest the potential for a significant positive feedback between the land surface soil moisture and rainfall, convection in this region is largely controlled by effects of the land-sea border. In a study of summertime convective initiation in the coastal area of Mobile, Alabama, Medlin and Croft (1998) find that most soundings are humid with a deep section showing a moist-adiabatic lapse rate, consistent with the $CTP-HI_{low}$ observations in Figure 6.8 (Stations VPS and SIL are closest to Mobile). Medlin and Croft also find, as stated above, that most convection in this area is triggered by sea breezes.

The other positive feedback region, however, is largely continental, and the land surface condition can indeed play a significant role in the development of convection. The largest of all the 10 regions, the Great Lakes and Northeast region includes 15 stations. Twelve of these
Figure 6.8: As in Figure 6.2, but for stations from the Gulf Coast region. Note that figure axes have changed slightly.
meet strong requirements for being labelled indicative of a positive feedback (Figure 6.9),
while three do not (Figure 6.10).

The twelve stations with a strong positive feedback signal (Figure 6.9) all have at least
twice as many days in the wet soil advantage regime as days in the dry soil advantage regime.
Seven of the 12 stations (INL, MPX, ILX, PIT, RNK, GSO, CHH) have between 20 and 27% of
days in the wet soil advantage regime, and a maximum of 10% in the dry soil advantage
regime. Three stations (ILN, IAD, ALY) have 17% of days in the wet soil advantage and
5-7% in the dry soil advantage regime. The final two stations (DVN and DTX) have only
12 and 14% in the wet soil advantage regime, but they also have only 2% of days in the dry
soil advantage regime. Each of these stations has a CTP-HIlow distribution that suggests
the potential for a small but significant positive feedback between soil moisture and rainfall,
similar to that seen in Illinois.

Two of the other three stations included in the positive feedback region (Figure 6.10)
are on the Great Lakes (GRB, BUF) and the third is on Long Island (OKX), so land-water
contrasts may be non-negligible at these stations. These three stations have between 14 and
18% of days in the wet soil advantage regime, but they also have between 8 and 13% of days
in the dry soil advantage regime. These stations, then, may not show a clear positive soil
moisture-rainfall feedback.

The other three stations shown in Figure 6.10–APX in northern Michigan, and GYX and
CAR in Maine–are not included in any of the 10 regions. Their characteristics are closest
to the Plains States stations, with about 70% of days atmospherically controlled, and the
other 30% close to evenly divided between the three other regimes. This division of days is
not expected to show a strong rainfall response to the land surface moisture condition.

### 6.5 Transition Regions

Stations classified as transition regions have significant percentages of days in each of the
three non-atmospherically controlled regimes: the dry soil and wet soil advantage regimes,
and the transition region. The Inland Southeast region (Figure 6.11) tends to have about
15% of days in each of these three regimes. One exception is Station FFC which could be
Figure 6.9: As in Figure 6.2, but for stations from the Great Lakes and Northeast region. Note that figure axes have changed slightly.
classified in a positive feedback region, with 22% of days in the wet soil advantage regime, and only 11% in the dry soil advantage regime. However, due to its location between the rest of the Inland Southeast region and the Bermuda High Impact Region, it was left in the transition region.

The Bermuda High Impact Region is quite different from the Inland Southeast. These three stations (CHS, MHX, and WAL, Figure 6.12), have more dry soil advantage days than wet soil advantage days (18-29% vs. 15-19%, respectively), because of the impact of the high pressure system that is typically centered over Bermuda. As with the coastal stations in the Gulf Coast Region, however, convection is largely controlled by land-sea contrasts. Monthly mean surface wind streamlines presented in Bryson and Hare (1974) show that in July these
three stations are the only ones on the east coast to receive air directly from the Atlantic. Further north, the streamlines show air coming from the west after traveling north from the Gulf of Mexico. Florida, too, is strongly influenced by Gulf air. The Bermuda High Impact region is likely to receive air that has a high $CTP$ because of subsidence associated with the anticyclone typically centered near Bermuda, but this air also has a moderate to low $HI_{low}$ because of the high evaporation rates over warm Gulf Stream waters.

Figure 6.11: As in Figure 6.2, but for stations from the Inland Southeast region. Note that figure axes have changed slightly.
Figure 6.12: As in Figure 6.2, but for stations from the Bermuda High Impact region. Note that figure axes have changed slightly.

6.6 One-dimensional BL results from other stations

In order to determine if the $CTP-HI_{low}$ approach used to classify positive and negative feedback regions is valid outside of the region for which it was developed, the methodology of Chapter 3 is applied to another station within the same positive feedback region that Illinois is in, to stations in each of the two transition zones, and to one station from the Dryline and Monsoon Region. For each of these four additional stations, 1D model runs were performed for each day from the summer of 1998 (the summer used to create the $CTP-HI_{low}$ scatter plots detailed throughout this chapter), with radiative conditions determined for the actual latitude of the station on Julian day 210 (July 29). Results show consistency with the $CTP-HI_{low}$ framework developed from Illinois soundings, suggesting that the framework is applicable to a wide range of atmospheric and geographic settings.

6.6.1 Another Positive Feedback Region Station

Station ILN

Like the Illinois sounding station, Station ILN is in the Great Lakes and Northeast Region. At latitude 39.4°N, this station is in Wilmington, Ohio. The 1998 data at this station are in atmospherically controlled regimes 63% of the time, in the transition region 14% of the
time, in the wet soil advantage region 17% of the time, and in the dry soil advantage region only 6% of the time (Figure 6.9). This distribution suggests an over-all wet soil advantage.

Indeed, signs of a positive feedback between soil moisture and rainfall are revealed in Figure 6.13, which shows the results of the 1D model run with the soundings making up the ILN scatter plot of Figure 6.9. Of a total of 69 available and non-rainy soundings, 50 lead to the same model outcome over both wet and dry soils (both rain in 16 cases, both have shallow clouds in six cases, and 28 have no convection over either soil condition), while 19 have different responses to the different soil conditions: one has rain over dry soils only, five have rain over wet soils only, four have rain over wet and shallow clouds over dry, and nine yield shallow clouds over wet soils only. Thus, 25 cases lead to rain over wet soils, but only 17 lead to rain over dry soils (36% versus 25%). It is important to note that one summer’s worth of soundings does not provide enough data points, particularly in the non-atmospherically controlled plot, to draw solid conclusions about the existence of a positive feedback in Ohio, but the results are consistent with this hypothesis.

The two cases in Figure 6.13 with rain over wet soils only and very high $HI_{low}$ are both cases with very humid near-surface layers and a sharp humidity drop below 850 mb (one of the two levels included in the $HI_{low}$ value), but above the level at which convection is triggered. The more extreme of the two cases, with an $HI_{low}$ of about 24°C, has a specific humidity drop of 6 g/kg between 860 and 840 mb (8 to 2 g/kg) while convection is triggered at about 910 mb. Thus, the value at 850 mb is not representative of the conditions in the mixed layer at the time of triggering.

One concern about extending the $CTP-HI_{low}$ framework developed with data from one location to a broad geographical region was the possible location-to-location variability of values marking a transition from one response regime to another. Figure 6.13 shows that these data fit within the expected regimes established with data from Illinois. This is not surprising, given the relative proximity of Ohio and Illinois, and the fact that the two stations have similar distributions in $CTP-HI_{low}$ space, suggesting similar atmospheric conditions. Nevertheless, it is encouraging to verify that the framework is not solely dependent on the Illinois location. The following analyses at stations in different regions of the country will further verify this conclusion.
6.7 Transition Region Stations

Station CHS

Station CHS is at latitude 32.9°N in Charleston, South Carolina, within the Bermuda High Impact Region. Of a total of 73 available and non-rainy soundings, 57 lead to the same model outcome over both wet and dry soils (both rain in 24 cases, both have shallow clouds in one case, and 32 have no convection over either soil condition), while 16 have different responses to the different soil conditions: nine have rain over dry soils only, one has rain over wet soils only, one has rain over wet and shallow clouds over dry, one has shallow clouds over wet soils only, and four have shallow clouds over dry soils only (Figure 6.14). Thus, there is a greater incidence of triggering of convection over dry soils than over wet (26 versus 33 times,
or 36% versus 45%), but due to the likelihood of deeper rainfall depths when convection is triggered over wet soils as opposed to dry, this is not necessarily indicative of a negative feedback between soil moisture and rainfall. It is, however, in line with expectations from a transition region, which has about equal numbers of days in the wet soil and the dry soil advantage regions. At this station, the dry soil advantage manifests in more events begun (which was also expected, given the 29% of days in the dry soil advantage region but only 19% in the wet soil advantage region [Figure 6.12]), but the 24 cases with rain over both soil types are in the wet soil advantage region, where rainfall depths are likely to be greater over wet soils than over dry.
Station SHV

Station SHV, in the Inland Southeast Transition Region, is located in Louisiana at the Shreveport Regional Airport. The latitude of this station is 32.5°N. Of a total of 75 available and non-rainy soundings, 65 lead to the same model outcome over both wet and dry soils (both rain in 12 cases, both have shallow clouds in two cases, and 51 have no convection over either soil condition), while ten have different responses to the different soil conditions: three have rain over dry soils only, two have rain over wet soils only, three have shallow clouds over wet soils only, and two yield shallow clouds over dry soils only (Figure 6.15). Like Station CHS, these results conform to expectations for a station in a transition region: no strong advantage is seen by either wet soils or dry soils. Note, too, however, that the two transition stations are in very different regions: the Bermuda High Impact Region is a significantly more humid environment than the Inland Southeast Region. This is reflected in the distribution of $H_{low}$ seen in Figures 6.14 and 6.15, and in the percentage of days likely to rain: between 36 and 45% at Station CHS, but only between 19 and 20% at Station SHV.

6.7.1 Negative Feedback Region Station

Model Adjustments for Use in the Dryline and Monsoon Region

Evapotranspiration is calculated in the 1D boundary layer model according to a Penman-Monteith equation (see Chapter 2). Soil properties and stomatal resistance values appropriate for the eastern half of the United States are entirely inappropriate for the desert-like setting of the Dryline and Monsoon Region. Modifications were made to the minimum and maximum stomatal resistances to produce realistic evapotranspiration values for this region.

Analysis of both 12Z soundings (those used for model initialization) and 00Z soundings from Albuquerque showed that the boundary layer typically grows up to between 650 and 550 mb by 00Z, starting from a surface pressure of about 840 mb. (Note that this is consistent with the description of the critical region described in the definition of the CTP (Figure 3.1, which states that the boundary layer rarely grows higher than 300 mb AGS. The desert-like conditions of New Mexico provide good confirmation that this upper-limit is indeed rarely
Figure 6.15: Results from the 1D model initialized with soundings from Station SHV in the Inland Southeast Region.

surpassed.) The specific humidity at the surface most commonly changes little during the day, and is frequently around 9 g/kg, though it can be much smaller.

Model stomatal resistance parameters were selected by two criterion: that this commonly observed behavior of relatively constant specific humidity at the surface occur for soil saturations of 20%, and that the Bowen ratio remain above 1.0 at soil saturations of 50%. These two conditions were satisfied by using a minimum stomatal resistance of 400 s/m (contrasted with 50 s/m for Illinois and the other three stations modeled). With normal conditions tuned to 20% of soil saturation, 50% was used for wet soil runs, and 10% for dry soil runs. The mid-day Bowen ratio computed from this $r_{smin}$ (with the values of the other parameters detailed in Section 2.3.1 unchanged) is approximately 1.5 for a soil saturation of 50%. For the dry soil case, however, the stomatal resistance is determined by the value of $r_{smax}$: this parameter was fixed at 15,000 s/m, to yield a mid-day Bowen ratio of about
25 for a soil saturation of 10%. (Note that the computed stomatal resistance at the other four stations was always well below the prescribed maximum value of 1000 s/m.) Wallace and Hobbs (1977) note that Bowen ratios over extremely dry land surfaces can be in the many hundreds. With these two simple modifications, the 1D model behavior appropriately simulated conditions in the arid southwest.

Station ABQ

Station ABQ, in Albuquerque, New Mexico, is at latitude 35.0°N. The elevation of this station is 1.6 km, yielding surface pressures of about 840 mb—well under the ~1000 mb values in Illinois. The index $HI_{stillow}$ plotted in Figure 6.16 is equivalent to the version of $HI_{low}$ used in the scatter plots of the earlier parts of this chapter: it is the sum of the dew-point depressions at $P_{surf} - 50$ mb and $P_{surf} - 150$ mb. (The plots of 1D results from the other stations use the original fixed-pressure-level definition.)

Of a total of 86 available and non-rainy soundings from Albuquerque, 73 lead to the same model outcome over both wet and dry soils (both rain in ten cases, both have shallow clouds in one case, and 62 have no convection over either soil condition), while 13 have different responses to the different soil conditions: seven have rain over dry soils only, one has rain over wet and shallow clouds over dry, and five have shallow clouds over dry soils only (Figure 6.16). Thus, 11 cases lead to rain over wet soils, and 17 lead to rain over dry soils. This 13% versus 20% difference in triggering of rainfall is a more substantial contrast than that observed at the two transition stations analyzed above, and is evidence that a negative feedback between soil moisture and rainfall may be present in New Mexico. Further analysis is needed, however, to increase the statistical validity of this claim and to determine the impact of the expected higher rainfall depths over wet soils in the cases where rainfall occurs over both soil conditions. Future research will include MM5 investigations of this region similar to those of Illinois described in Chapter 5.

All of the days with rainfall developing in these one-dimensional simulations of Albuquerque occur in July and August, after the onset of the southwest monsoon. The mean and median onset days between 1948 and 1996 were July 7 and 9, respectively (Higgins and
1D Modeling results from Station ABQ

Atmospherically controlled cases

Non-atmospherically controlled cases

Figure 6.16: Results from the 1D model initialized with soundings from Station ABQ in the Southwest Monsoon Region.

Shi, 2000). The 1998 soundings used in this analysis showed very low relative humidities (high $HI_{low}$) throughout June and in the first three days of July. A sudden onset of more humid conditions begins on July 4, 1998 and lasts through July 11. Model simulations initialized with eight soundings from this period led to rainfall over both wet and dry soils four times, and to rainfall over dry soils only four times. A similar humid period occurred between July 23 and July 29, 1998. With soundings from this week, the model developed rainfall over both soil conditions four times, and over dry soils twice (one sounding was removed from the analysis because rain was occurring at the sounding time). It is likely that these two humid periods were times when the summer monsoon, which is consistently present south of Albuquerque in Mexico, extended far enough north to be observed at Station ABQ.
6.7.2 Composites of Results From All Four Stations

Since one summer's worth of analyses did not yield a substantial amount of data at each individual station, the results of all four additional stations are composited in Figure 6.17. The humidity index used in these figures, in contrast to the earlier figures from Stations ILN, CHS, and SHV, is $HI_{stilllow}$, since this is well-defined for all four stations. Since these three stations all have surface pressures close to 1000 mb, $HI_{low}$ and $HI_{stilllow}$ are usually quite similar, though a few cases with sharp humidity gradients around 850 mb may be substantially different.

Figure 6.17 shows that the $CTP-HI_{low}$ framework is valid for a wide range of locations and atmospheric settings. It suggests that the $CTP$ and $HI_{low}$ values marking the transition from wet soil to dry soil advantage regimes are independent of location (within the range of circumstances studied here). More significantly, it suggests that, for matters of convective triggering and response to land surface conditions, the critical portion of the atmosphere is independent of location: approximately 1 to 3 km above the ground surface.

6.8 Discussion and Conclusions

Though this analysis is complete in its coverage of the continental US, it is incomplete in its temporal coverage: these data are from only one summer. Inter-annual variability may be significant. For example, Court (1974) points out that the location of the anticyclone associated with the Bermuda High impacts the rainfall distribution from New Mexico to the southern Atlantic coast. When the anticyclone is west of its normal position in summer, Texas receives more moist air and more rain showers, while the southeast gets descending air bringing little rain. An eastward shift of the Bermuda anticyclone allows for more hot, dry air from Mexico than normal to extend into New Mexico, Texas and beyond, while the southeast receives more moisture and more rainfall than normal. Clearly, other factors can alter the summertime atmospheric patterns over the US, which could, in turn, shift the locations of the ten regions in Figure 6.1. More research is needed to determine accurate $CTP-HI_{low}$ climatologies throughout the country, and to determine how large-scale conditions,
Figure 6.17: Composites of the 1D model results from the four additional stations described in the text and in the previous four figures. The wet soil advantage region defined with analysis of Illinois data is approximately bound by the blue ellipse, and the dry soil advantage region by the red ellipse. Conditions with $CTP > 400 \text{ J/kg}$ did not occur in Illinois. The results from these four stations are consistent with the framework developed with data from Illinois.

such as the location of the Bermuda High or the presence or absence of an El Nino, might lead to deviations from this long-term mean.

Another factor that needs to be discussed is the use of 12Z soundings at all stations throughout the country. On the west coast, 12Z is 4 am local time, while on the east coast, it is 7 am. The $CTP$ was developed for use prior to early morning degradation of the nocturnal stable layer. In Illinois 6 am soundings were typically well-suited for this purpose. (Though two anomalous conditions with super-adiabatic near-surface layers already present at 6 am led to MM5 outputs inconsistent with observations, as discussed in Chapter 4.) Since erosion of the nocturnal stable layer typically takes at least a few hours after sunrise, the $CTP$ should be the same at 7 am as it would be at 6 am. The earlier soundings should
also have the same CTP, since continued growth of the stable layer after 4 or 5 am should remain below the 100 mb-high base of the critical CTP region. The HI_{low} may be affected by these time changes, but it is assumed that these effects are small.

Despite these two caveats, the analyses in this chapter strongly support the conclusion that the CTP-HI_{low} framework is valid for locations far removed from Illinois. Using this framework, it was determined that much of the eastern half of the country should show a small but significant positive feedback between soil moisture and rainfall, as indicated by the two positive feedback regions outlined in Figure 6.1. Furthermore, the arid southwest is the only region likely to see a negative feedback, as indicated by the Dryline and Monsoon Region in Figure 6.1. Indications of a negative feedback at Albuquerque, New Mexico were seen with one-dimensional boundary layer modeling work described above, although, as previously mentioned, only one summer of data was used. Future research will include three-dimensional modeling of this area.

The most important conclusion that can be drawn from this chapter, however, is that, for matters of convective triggering and response to land surface conditions in mid-latitudes, the critical portion of the atmosphere is independent of location: approximately 1 to 3 km above the ground surface. As long as the low levels of the troposphere are relatively humid, when the temperature lapse rate in this region is dry adiabatic, a negative feedback is likely; when it is moist adiabatic, a positive feedback is likely; and when there is a temperature inversion in this region, deep convection cannot occur.
Chapter 7

Conclusions

This thesis has addressed the question of how the early morning atmospheric thermodynamic structure affects the interactions between fluxes from the land surface (and thus the soil moisture state) and the growth and development of the boundary layer (BL) and the triggering of convection. It is concluded that in mid-latitudes, the atmospheric structure in the region between 100 and 300 mb (approximately 1 and 3 km) above the ground surface (AGS) is the critical factor for determining the nature of the influence that soil moisture has on the triggering of convection. The great influence of this region results from its location between the lowest ~1 km, which is almost always incorporated into the boundary layer, and the free atmospheric air above ~3 km, which is almost never incorporated into the BL. The pressure levels of these bounds on the critical region were confirmed at five different locations in four different synoptic settings within the United States.

As long as the low levels of the troposphere are relatively humid ($HI_{low} < 15°C$), but not extremely close to saturation ($HI_{low} > 5°C$), the atmospheric response to soil moisture is largely determined by the temperature profile within this critical region.

- When a temperature inversion is present within this critical region, convection is not triggered, regardless of the flux partitioning at the land surface.

- When this region has a temperature lapse rate close to dry adiabatic, dry soils are more likely to lead to rainfall (negative feedback), because high sensible heat fluxes are the most efficient mechanism for bringing together the boundary layer height and the
Level of Free Convection (LFC).

- When this region has a temperature lapse rate close to moist adiabatic, wet soils are more likely to lead to rainfall (positive feedback), because high latent heat fluxes are the most efficient mechanism for bringing together the boundary layer height and the LFC.

In the negative feedback case, the primary mechanism for convective triggering is BL growth, while the LFC tends to change little after rapid decline in the morning; BL growth occurs more readily in areas of high sensible heat flux. In the positive feedback case, the primary mechanism for convective triggering is an increase in BL moist static energy, which leads to a rapid fall of the LFC, while the BL height tends to change little after the typical rapid growth phase of the early morning; BL moistening occurs more readily in areas of high latent heat flux.

These features of the atmospheric thermodynamic structure are all captured by the newly developed Convective Triggering Potential (CTP). In the first case, when a temperature inversion is present within this region, the CTP is negative. In the negative feedback case, when the temperature lapse rate is close to dry adiabatic, the CTP is above 150-200 J/kg. In the positive feedback case, when the temperature lapse rate is close to moist adiabatic, the CTP is between 0 and 200-250 J/kg. To distinguish between feedback implications in the overlap region, information about the humidity of the lower troposphere is required.

Many measures of the humidity of the lower troposphere were effective at distinguishing between days where convection could or could not occur, and days which favored wet soils from those which favored dry soils. Most effective was $HI_{low}$, a variation on the humidity index (Lytinska et al., 1976) which summed the dew-point depressions at two levels; one below the critical CTP region and one in the lower portion of this region. Other height combinations in humidity index variations were also effective.

The CTP-$HI_{low}$ framework for assessing the influence of the land surface on the potential for rainfall is summarized by Figure 3.20. For cases with positive CTPs, threshold values of $HI_{low}$ are 5, 10 and 15°C. Below 5°C, rainfall is likely over both wet and dry soils. Between 5 and 10°C, rainfall is favored over wet soils. Between 10 and 15°C, rainfall is favored over dry
soils. In the dry soil advantage regime, the advantage manifests in an increased likelihood of convective triggering. In the wet soil advantage regime, the advantage manifests primarily in enhanced rainfall depths over wet soils, and to a lesser degree in a greater number of events triggered. Above 15°C, the atmosphere is too dry for convection to develop. These threshold values between positive and negative feedback regimes were not significantly different at the five different stations closely investigated with a one-dimensional boundary layer model.

The $CTP-HI_{low}$ framework captures critical details about the temperature and humidity structure of the atmosphere, but, as modeling results with MM5 reveal, the vertical profile of the winds is also crucial. These results show that backing or strongly sheared unidirectional winds can suppress convective activity. Conditions most favorable for convective development have veering winds with little to moderate shear. Thus, it is concluded that the land surface condition can impact the potential for convection only when the atmosphere is not already predisposed to convect or not to convect. This atmospheric predisposition can be determined from the $CTP$, $HI_{low}$, and the vertical profile of the winds.

Illinois was found to exhibit a small but significant positive feedback between soil moisture and rainfall here and also in Findell and Eltahir (1997) because the frequency of days falling in the wet soil advantage regime of $CTP-HI_{low}$-space exceeds the frequency of days falling in the dry soil advantage regime (e.g., 22% versus 8% during the summer of 1998, Figure 6.9). Analyses of radiosonde data from the summer of 1998 shows that stations in other regions of the country have significantly different distributions in $CTP-HI_{low}$-space. In much of the eastern half of the United States, a positive feedback like that seen in Illinois is likely. In the Dryline and Monsoon Region of the arid southwest, on the other hand, frequent high-$CTP$, intermediate-$HI_{low}$ days suggest the potential for a negative feedback. The rest of the western half of the US shows little potential for land surface impacts to influence convective potential, which in these regions is almost entirely controlled by the early morning atmospheric thermodynamic structure.

More research is needed to determine accurate $CTP-HI_{low}$ climatologies throughout the country, and to determine how large-scale conditions in a given year, such as the location of the Bermuda High or the presence or absence of an El Nino, might lead to deviations from this climatology. Investigations of inter-decadal variability may reveal information about
large-scale responses to climatic change, and to remote physical features with far-reaching impacts, such as the El Nino-Southern Oscillation or the Pacific-North American pattern.

Other future areas of research include 3D mesoscale modeling over the Dryline and Monsoon Region of the arid southwest, to look for further evidence of a negative feedback between soil moisture and rainfall. Additionally, the nationwide analysis of Chapter 6 should be extended to continental scales. Convection in the tropics is quite different than at mid-latitudes; it would be interesting to determine if the critical region is 1-3 km there, as well as in mid-latitudes. A positive feedback is anticipated for the tropics, since early-morning atmospheric profiles are typically close to moist adiabatic in tropical regimes.

These conclusions have strong implications regarding the importance of high resolution of data and model levels throughout the critical $CTP$ region. In fact, early attempts at a nationwide analysis with NCEP reanalysis data were quickly abandoned because the two data levels within this region (at 850 and 700 mb) did not adequately resolve key details of the atmospheric structure. As discussed in the introductory chapter, previous observational and modeling studies have shown evidence of both positive and negative feedbacks. This could be a result of the individual study locations, since the nationwide analysis revealed highly variable $CTP-HI_{low}$ characteristics throughout the United States. These differing results, however, could also result from different model and/or forcing-data resolution in the critical $CTP$ region. Modeling studies investigating interactions between the land surface and the atmosphere are no doubt influenced by the vertical resolution of both the model and the forcing data within this critical zone of the atmosphere. Further research is needed to determine the vertical resolution required to adequately represent this region and its control on land surface-boundary layer interactions.
References


Appendix A

Appendix: Definitions of Various Stability Indicies

In order to discuss the distinguishing characteristics of initial atmospheric profiles, we employ a number of meteorological indicies and thermodynamic parameters. Many of these indicies have been in use for rainfall and thunderstorm forecasting for many years. We find Showalter's Stability Index (1953) and the Humidity Index (Lytinska et al., 1976) to be particularly useful and important. Many other well-used indicies were also investigated, including many described and cited in a comprehensive assessment by Peppler and Lamb (1989). Some of these include the K-index (K), the Lifted Index (LI), Total Totals (TT), the convective available potential energy (CAPE), the convective initiation energy (CIN), and the deep convective index (DCI), to name a few.

The Showalter Stability Index  The value of the Showalter Stability Index (SI) is given by

\[ SI = T_{500} - T_{pcl,850to500}, \]  \hspace{1cm} (A.1)

where \( T_{500} \) is the observed temperature at 500 mb, and \( T_{pcl,850to500} \) is obtained by lifting a parcel dry adiabatically from 850 mb until it reaches its Lifting Condensation Level (LCL), and then moist adiabatically until 500 mb (Showalter 1953), as sketched in Figure 3.2. A negative SI means that the parcel temperature is warmer than the environmental tempera-
ture at 500 mb, indicating instability; conversely, a positive SI indicates stability. Showalter (1953) gives the following prediction guidelines:

- showers are likely when the index is less than +3°C,
- thunderstorms are increasingly likely as the index falls from +1 to -2°C,
- severe storms may occur when the index is less than -3°C, and
- tornadoes are possible when the index is less than -6°C.

The Humidity Index  

Lytinska et al.'s (1976) original definition of the humidity index is the sum of the dew point depressions at 850 mb, 700 mb, and 500 mb:

$$HI = (T_{850} - T_{d,850}) + (T_{700} - T_{d,700}) + (T_{500} - T_{d,500}),$$  \hspace{1cm} (A.2)

where $T_p$ is the temperature at pressure level $p$ and $T_{d,p}$ is the dew point temperature at pressure level $p$. A more useful parameter for assessing this group of soundings from Illinois is the sum of the dew point depressions at 950 mb and 850 mb:

$$HI_{low} = (T_{950} - T_{d,950}) + (T_{850} - T_{d,850}).$$  \hspace{1cm} (A.3)

Lytinska et al. (1976) suggested as threshold for rain $HI \leq 30°C$. The threshold for $HI_{low}$ is 15°C for the Illinois data.

The K Index  

Originally defined by George (1960), the K index combines information about the thermal lapse rate between 850 and 500 mb with the dew point at 850 mb and the dew point depression at 700 mb:

$$K = (T_{850} - T_{500}) + T_{d,850} + (T_{700} - T_{d,700}).$$  \hspace{1cm} (A.4)

The K index is particularly useful for non-severe convection. Threshold levels vary with season, location, and synoptic setting. In 1976, Anthes wrote, "The leading predictor [of thunderstorms] is the K index, which is a measure of the stability of the lower half of the atmosphere and the amount of water vapor present in the lower levels" (pg. 426).
The Lifted Index  The Lifted Index was defined by Galway in 1956, a few years after the Showalter Stability Index was first described. It is similar to the SI, except that the lifted parcel is given the mean properties of a specified surface layer and is lifted from the midpoint of the layer. Galway used the 3000 feet closest to the ground as the near-surface layer, while Peppler and Lamb (1989) use the lowest 50 mb:

\[
LI = T_{500} - T_{p_{cl},SFCLAYERto500}.
\]  

(A.5)

The Total Totals  The Total Totals (TT) is the sum of two other indices: the Cross Totals (CT) and the Vertical Totals (VT). The VT assesses the vertical temperature lapse rate: \( VT = T_{850} - T_{500} \), while the CT assesses the low-level moisture: \( CT = T_{d,850} - T_{500} \). Thus,

\[
TT = (T_{850} - T_{500}) + (T_{d,850} - T_{500}).
\]  

(A.6)

The Deep Convective Index  The Deep Convective Index is given by

\[
DCI = T_{850} + T_{d,850} - SLI.
\]  

(A.7)

The Surface Lifted Index (SLI) is a modified version of the LI, where the lifted parcel is taken directly from the surface, rather than from a mixed near-surface layer. The DCI provides information about the stability of the profile through the SLI, and also provides information about the total energy at 850 mb through the temperature and dewpoint at this level.

The Convective Available Potential Energy  In contrast to the previously described indices, the CAPE is a thermodynamic measure with a very physically-based definition: it is the energy difference between a parcel lifted from the surface and the observed profile.

\[
CAPE = \int_{LFC}^{LNB} R_d(T_{pp} - T_{p_{env}})dlnp
\]  

(A.8)
where $T_{pp}$ is the density temperature of the lifted parcel, $T_{penv}$ is the density temperature of the environmental air, LFC is the level of free convection and LNB is the level of neutral buoyancy.

The Convective Inhibition The CAPE is defined between the LFC and the LNB. Soundings can contain a great deal of energy, but this energy can only be released if there is a lifting mechanism to raise surface parcels to the LFC. The CIN measures the amount of work that must be performed to lift surface air to the LFC. Just as the CAPE is the positive area between a parcel path and the environmental temperature profile, the CIN is the negative area between these two vertical temperature profiles.