Atmospheric Controls on Soil Moisture–Boundary Layer Interactions. Part I: Framework Development

KIRSTEN L. FINDELL

Geophysical Fluid Dynamics Laboratory, Princeton University, Princeton, New Jersey

ELFATIH A. B. ELTAHIR

Parsons Laboratory, Department of Civil and Environmental Engineering, Massachusetts Institute of Technology, Cambridge, Massachusetts

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ABSTRACT

This paper investigates the influence of soil moisture on the development and triggering of convection in different early-morning atmospheric conditions. A one-dimensional model of the atmospheric boundary layer (BL) is initialized with atmospheric sounding data from Illinois and with the soil moisture set to either extremely wet (saturated) or extremely dry (20% of saturation) conditions. Two measures are developed to assess the low-level temperature and humidity structure of the early-morning atmosphere. These two measures are used to distinguish between four types of soundings, based on the likely outcome of the model: 1) those soundings favoring deep convection over dry soils, 2) those favoring deep convection over wet soils, 3) those unlikely to convect over any land surface, and 4) those likely to convect over any land surface. Examples of the first two cases are presented in detail.

The early-morning atmosphere is characterized in this work by the newly developed convective triggering potential (CTP) and a low-level humidity index, HI_{low}. The CTP measures the departure from a moist adiabatic temperature lapse rate in the region between 100 and 300 mb (about 1–3 km) above the ground surface (AGS). 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. Together, these two measures form the CTP-HI_{low} framework for analyzing atmospheric controls on soil moisture–boundary layer interactions.

Results show that in Illinois deep convection is trigged in the model 22% of the time over wet soils and only 13% of the time over dry soils. Additional testing varying the radiative conditions in Illinois and also using the 1D model with soundings from four additional stations confirm that the CTP-HI_{low} framework is valid for regions far removed from Illinois.

1. Introduction

Feedbacks between the land surface and the atmosphere have been the focus of much recent inquiry into questions ranging from the maintenance of extreme drought or flood conditions, to the influence of deforestation on rainfall, to responses to increases in atmospheric concentrations of greenhouse gases. Many studies of the midwestern U.S. drought of 1988 and flood of 1993, for example, suggest that the soil moisture condition in these cases helped to sustain the extreme circumstances throughout the summer (Trenberth and Guillemont 1996; Trenberth et al. 1988; Atlas et al. 1993). Others suggest that there is actually a negative feedback between soil moisture and drought (flood) conditions (Giorgi et al. 1996; Paegle et al. 1996). Ek and Mahrt (1994) caution against extending the results of these studies to all locations and synoptic settings. With a one-dimensional model initialized with data from the Hydrological–Atmospheric Pilot Experiment–Modélisation du Bilan Hydrique (HAPEX–MOBILHY) experiment, they show that the influence of the land surface on the development of boundary layer (BL) clouds is highly dependent on the initial (early morning) condition of the atmosphere. Furthermore, modeling studies are also dependent on factors such as the convection scheme (Pan et al. 1996; Pal 1997), the domain size (Seth and Giorgi 1998), or the vertical resolution of the model and/or the forcing data.

This work is an effort to define the physical mechanisms controlling the interactions between the land surface and the atmospheric boundary layer (BL). Specifically, this paper addresses the question of how the ear-

Corresponding author address: Dr. Kirsten L. Findell, Geophysical Fluid Dynamics Lab, Princeton University, P.O. Box 308, Princeton, NJ 08542.
E-mail: kirsten@alum.mit.edu

ly-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 BL, leading to the triggering of convection.

There are three main characteristics of the early-morning atmospheric structure that significantly influence the nature and evolution of the boundary layer during the course of the coming day:

- the properties of the residual layer, since this air will quite likely be incorporated into the BL (e.g., Chen and Avissar 1994; Rabin et al. 1990; Cutrim et al. 1995; Rabin and Martin 1996);
- the depth of the nocturnal stable layer (e.g., Wetzel et al. 1996; Segal et al. 1995), since this will determine the ability of the growing BL to reach beyond the air of this near-surface stable layer and the time at which it does so; and,
- the height and strength of the inversion separating the mixed layer from the overlying free atmosphere (e.g., Betts et al. 1996; Ek and Mahrt 1994; Mahrt 1997; Mahrt and Pierce 1980; Segal et al. 1995), 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). 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 vegetation and/or soil moisture to influence the development of convection. We describe such a measure in this paper.

The convective triggering potential (CTP) focuses on the temperature lapse rate between 100 and 300 mb (about 1–3 km) above the ground surface (AGS). The CTP was developed from work with a one-dimensional boundary layer model initialized with sounding data from Illinois. It is coupled with a low-level humidity index, HI_{low}, to better describe early-morning atmospheric conditions and help diagnose the likelihood for deep convection during that day.

The model is briefly described in section 2. In section 3 we define the CTP and the HI_{low}, and in section 4 we present two case studies highlighting the advantage of dry soils in high CTP environments and wet soils in intermediate CTP environments. In section 5, we present the full suite of model results generated using initial soundings from the summers of 1997–99 at a station in central Illinois. These results show that deep convection is more likely to occur in the model over wet soils than over dry soils, given the atmospheric environment of central Illinois. This is consistent with the small but significant positive soil moisture–rainfall feedback previously reported for Illinois (Findell and Eltahir 1997, 1999).

The results in section 5 are presented in the context of the CTP-HI_{low} framework. The framework is used to distinguish between four different atmospheric regimes: one where it is easier for high sensible heat flux regions to trigger deep convection, another where it is easier for high latent heat flux regions to trigger deep convection, a third where the atmosphere is so dry and/or stable that deep convection is unlikely over any surface, and a fourth where the atmosphere is very humid and marginally unstable so that convection is very likely over any surface. Note that in both the third and fourth conditions, the likelihood of deep convective activity is independent of the surface flux partitioning.

To provide further support for the extension of this framework beyond the original development location, section 6 includes analyses of 1D results from four additional stations. These additional results confirm that the CTP-HI_{low} framework is a robust indicator of soil moisture–rainfall feedbacks. Sections 7 and 8 include a brief discussion followed by the conclusions of this work.

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

The first 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 12–15 h, during which the assumption of constant soil saturation is reasonable.

The model is initiated in the early morning, preferably at or near sunrise (0600 local time soundings are used for the bulk of the results presented here), 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 that is likely to produce rain, shallow convection that 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^{-1} and that the depth of convection must be greater than 5 km. Model results are not sensitive to changes in these threshold values within a given range. The results are bimodal in nature, with a gap between 3 and 5 km and 200 and 400 J kg^{-1}, suggesting that this distinction between shallow clouds and deep clouds in the model is appropriate. These threshold values are appropriate for the midlatitude continental regimes studied in this work (Battan 1973).

The reader is referred to Kim and Entekhabi (1998a,b) for details of the basic structure of the original model, and to Findell (2001) for details of the procedures for entrainment of overlying air and the triggering of convection. Convection is triggered when the convective inhibition energy (CIN) is zero or slightly positive (on order of 5 J kg^{-1}, to allow for turbulent overshooting of small negative areas below the LFC). Noted differences from the original model formulation are that the Clapp–Hornberger values used to determine stomatal resistance in this work are those for loamy sand (ψ_{sat} = 0.09 m and $B = 4.38$), and that the radiative conditions for all runs were the same (incoming solar calculated for 29 July conditions), such that the land surface influences were not masked by different solar conditions on different days of the year. Calculations of incoming solar radiation are dependent on the latitude of the location of interest.

As extreme examples of land surface influences, the model was run twice for each sounding: once with soil saturation set to 100%, and once with it set to 20%. The soil moisture value, W, comes into play in the model in two ways: through the stomatal resistance and through the albedo (see Kim and Entekhabi 1998a,b; Findell 2001). With high soil moisture, the ground surface tends to be darker than with low soil moisture. This is assumed to impact net radiation R_n through the albedo α according to the equation

$$\alpha = 0.20 - 0.10W. \tag{1}$$

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 midday in the wet soil moisture scenarios than in the dry soil moisture scenarios. Betts and Ball (1995) report a 22 W m^{-2} difference in radiation available at noon at the First International Satellite Land Surface Climatology Project (ISLCP) Field Experiment (FIFE) site in Kansas between days with soil moisture below 14% (average $R_n - G = 505$ W m^{-2}, where G is ground heat flux) and days with soil moisture above 20% (average $R_n - G =$ 527 W m^{-2}). Given the much larger range in soil moisture quantities used in this study, the 60 W m^{-2} difference appears reasonable.

Model runs with soil saturation set to other values (0%, 30%, 50%, and 80%) were also performed for some of the days presented here. The BL height and the moist static energy content of the intermediate runs was consistently between that of the 20% and 100% runs, and the 0% runs always had the fastest BL growth with the lowest moist static energy, as measured by the equivalent potential temperature, θ_E. The value θ_E is a measure of both the temperature and humidity content of the air and is conserved in dry adiabatic or pseudoadiabatic processes. [See Bolton (1980) for a highly accurate empirical formula used to calculate θ_E.] Triggering of deep convection in the 80% run occurred through a similar increased θ_E mechanism as seen in the fully saturated runs, though not quite as frequently. Triggering in the 20% and 0% runs occurred through similar BL growth mechanisms, at about the same frequency. These mechanisms of convective triggering will be discussed in detail in section 4. The BL properties (height, θ_E, LFC) in the intermediate soil moisture runs were always between those of the extremely wet and extremely dry soil moisture runs. When convection was triggered with 50% soil saturation, it was also triggered in either the very wet or the very dry case. The 20% and 100% runs were chosen as clear examples of the two distinct means of convective triggering: rapid BL growth and strong BL moistening, respectively. Convective triggering through a combination of these mechanisms clearly deserves more detailed investigation, but is beyond the scope of this paper. Only the 20% and the 100% cases are presented here.

The midday Bowen ratio (ratio of sensible to latent heat) in the model experiments changed from a range of 1.4–1.6 in the dry soil cases to a range of 0.3–0.4 in the saturated cases. Observations of this quantity during the Flatland Boundary Layer Experiment in central Illinois during the summers of 1995 and 1996 (Angevine et al. 1998) ranged from about 0.3 to 1.1. The Flatland data were used to verify the model's performance. Predawn soundings were not available, but 15 days of the experiment included 0900 LT soundings that were suitable for model initialization, plus subsequent soundings at 1030, 1200, and 1330, along with continuous boundary layer heights obtained from BL wind profilers, and observations of soil moisture in the top 5 cm. These were the days used to confirm that the model predictions

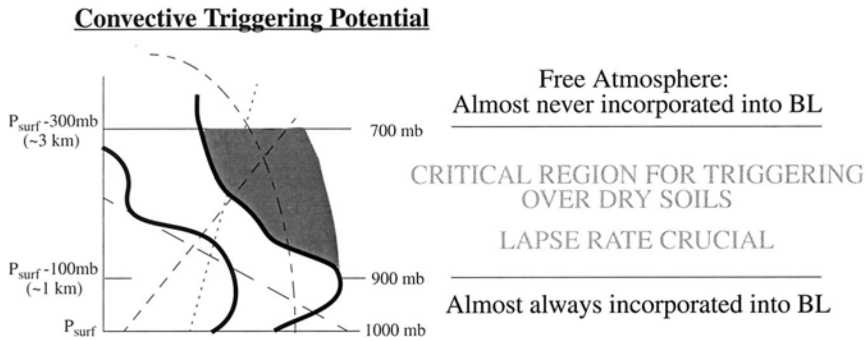

FIG. 1. A sketch of the definition of the convective triggering potential on a thermodynamic diagram. Thick solid lines are the temperature and dewpoint temperature profiles; straight long-dashed line is a dry adiabat (constant potential temperature); straight short-dashed line is constant temperature; 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. Note that the CTP can be negative if the value of the moist adiabat originating from the $P_{surf} - 100$ mb level is less than the observed equivalent potential temperatures at higher levels. Also, the CTP will be zero if the observed profile is moist adiabatic above the point of origin.

of boundary layer properties were within the realm of observation.

The comparisons of model results with Flatland observations (not shown, see Findell 2001) 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 observations for a given day since our intent was not to use this model as a predictive tool. Rather, given the adequate representation of BL height, temperature, humidity, and net radiation 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.

3. The convective triggering potential

An early-morning atmospheric profile can be broken down into three basic zones (Figure 1):

- the near-surface zone, which is sure to be incorporated into the day's boundary layer (order of 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 Fig. 1, the newly developed convective triggering potential 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 many regions is close to 1000 mb we will often present this critical CTP region as between 900 and 700 mb, as noted in the figure.) When the temperature profile in this critical region is close to dry adiabatic the CTP is large and convective triggering is favored over dry soils where the boundary layer height grows more quickly than over wet soils. When the temperature profile in the critical region is close to moist adiabatic the CTP is intermediate (still positive but smaller) and convective triggering is favored over wet soils where the boundary layer moist static energy grows more quickly than over dry soils (e.g., Betts and Ball 1995; Eltahir 1998). When there is a temperature inversion in this region, the CTP is negative, the atmosphere is stable, and convection will not occur, independent of the land surface condition. The qualitative results of this study were not sensitive to variations of the CTP bounds of up to 25 mb on the lower boundary and 50 mb on the upper boundary. Results were clearest when the measure was defined by the 100 and 300 mb AGS levels.

The hypothesis that prompted this work was that certain atmospheric conditions favor rainfall triggering over wet soils [positive soil moisture–rainfall feedback, e.g., Findell and Eltahir (1999, 1997)], while other atmospheric conditions favor rainfall triggering over dry soils [negative feedback, e.g., Giorgi et al. (1996)]. Our intent was to determine the differences between these initial atmospheric settings and their frequency of occurrence in different geographic regions. 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 the results of this work confirm, 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. Similarly, traditional humidity indices are helpful in ruling out days where the atmosphere is too dry for rainfall to develop, but are less helpful in more humid situations. The CTP is a measure of atmospheric stability, ruling out convection in stable conditions, as many traditional stability indices do (e.g., Showalter index, Showalter 1953). However, the CTP is also—and perhaps more importantly—a measure of the influence of surface flux partitioning on the likelihood of convection in unstable situations. In the next section, we present two case studies that demonstrate how the CTP effectively discriminates between atmospheric conditions favoring the development of rainfall over wet soils from those favoring the development of rainfall over dry soils.

In subsequent sections we will couple the CTP with HI_{low} to improve on this ability to discriminate between differing atmospheric conditions. The HI_{low} is a variation on the humidity index of Lytinska et al. (1976), which was defined as the sum of the dewpoint depressions at 850, 700, and 500 mb:

$$\mathrm{HI} = (T_{850} - T_{d,850}) + (T_{700} - T_{d,700}) + (T_{500} - T_{d,500}), \qquad (2)$$

where T_p is the temperature at pressure level p and $T_{d,p}$ is the dewpoint 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 dewpoint depressions at levels below 500 mb all prove to be helpful in assessing the convective potential of Illinois soundings. The most effective was the sum of the dewpoint depressions at 950 and 850 mb:

$$\mathrm{HI}_{low} = (T_{950} - T_{d,950}) + (T_{850} - T_{d,850}). \qquad (3)$$

Defined more generally, HI_{low} is the sum of the dewpoint depressions 50 and 100 mb above the ground surface. This is the definition that will be used throughout this work.

4. Case studies highlighting the relevance of the CTP

Convection is triggered in the model when the level of free convection 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. 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 2 shows two initial 0600 LT soundings with very different CTP values. These soundings are indicative of the types of initial atmospheric conditions that lead to rain over wet but not over dry soils (Fig. 2a: 3 July 1999, CTP = 88 J kg^{-1}), and those that lead to rain over dry but not over wet soils (Fig. 2b: 23 July 1999, CTP = 254 J kg^{-1}). The boundary layer height, the level of free convection, and the moist static energy (as described by θ_E) values for the wet and dry soil model runs for the first day (3 July 1999) are shown in Fig. 3. Model profiles at 1300 LT are shown in Fig. 4. Similar plots for the second day (23 July 1999) are shown in Figs. 5 and 6.

In many model runs, the boundary layer height over wet soils grows slowly but steadily until 1200 or 1400 LT, and then remains relatively constant (Fig. 5a). The θ_E continues to grow due to the continued input of moisture from the land surface (Fig. 5c). Over dry soils, on the other hand, the behavior of these two variables is often reversed (Figs. 3b,d): the BL height grows steadily and more rapidly throughout the day, but the θ_E plateaus or even drops in the afternoon, primarily because of increased entrainment of dry air from above the BL and limited moisture flux contributions from the land surface, which are then spread out over a deep BL. In the dry soil case, the BL top and the LFC will meet only if the BL grows high enough to reach the LFC. The critical factors influencing the BL growth are the sensible heat flux (determined by the land surface soil moisture and/or vegetation) and the temperature lapse rate of the air being entrained. In the wet soil case, the BL top and the LFC will meet only if the θ_E grows large enough to bring the LFC down to the BL top. The critical factor influencing the fall of the LFC are the BL θ_E and the temperature lapse rate of the air through which the LFC drops. The BL θ_E is determined by the latent heat flux (determined by the land surface soil moisture and/or vegetation) and the initial low-level humidity (assesed by a humidity index such as HI_{low}). Note that the temperature lapse rate in the critical CTP region is a central factor in both styles of convective triggering.

Consider, for example, the case of 3 July 1999 (initial sounding Fig. 2a). The θ_E in the BL over dry soils peaks just after 1000 LT, and levels off at 1300 LT (Fig. 3d). The BL height, however (Fig. 3b), continues to increase until almost 1600 LT. For BL deepening to trigger convection with no accompanying increase in θ_E, the BL must grow from 890 to 685 mb (the point where the parcel path crosses the environmental temperature line in Fig. 4b). In contrast, the wet soil boundary layer grows more gradually than that over the dry soil, but the θ_E is also increasing. The pseudoadiabats in Fig. 4a

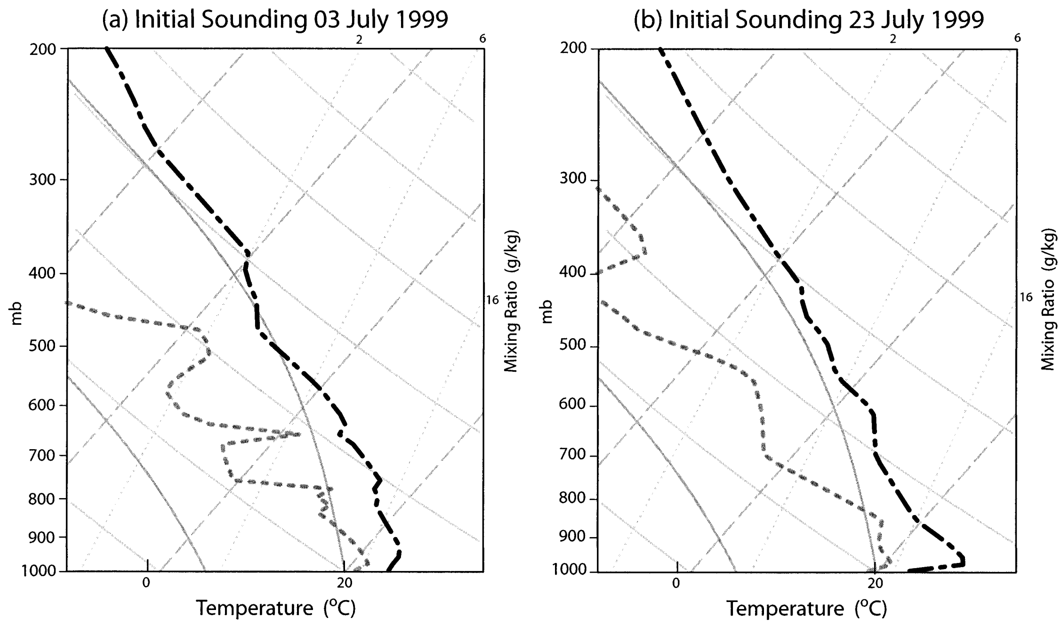

FIG. 2. Profile of initial conditions for (a) 3 Jul 1999: CTP = 87 J kg^{-1}, HI_{low} = 10.6°C, rainfall occurs only over wet soils; and (b) 23 Jul 1999: CTP = 254 J kg^{-1}, HI_{low} = 11.6°C, rainfall occurs only over dry soils. Thick dashed-dotted line is profile temperature; thick dashed line is profile dewpoint temperature; light solid lines are dry adiabats; long-dashed lines are constant temperature; dotted lines are constant mixing ratio.

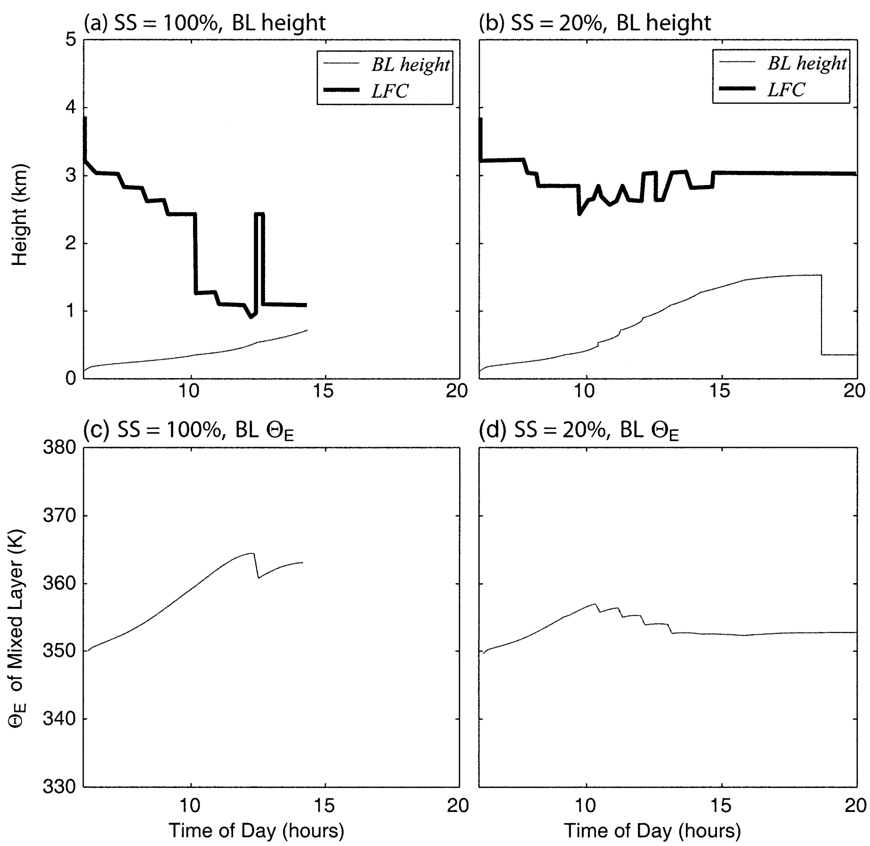

FIG. 3. The boundary layer height and the level of free convection in the (a) wet soil and (b) dry soil model runs for 3 Jul 1999; θ_E for these same model runs: (c) wet soil case, and (d) dry soil case.

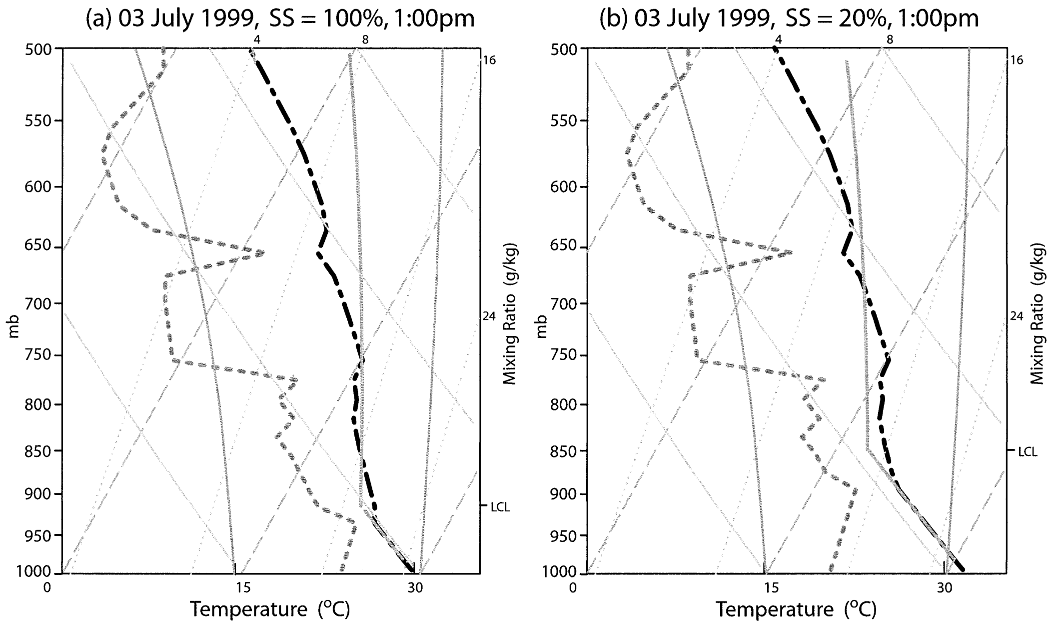

FIG. 4. Profile of model conditions at 1300 LT in (a) the wet soil run and (b) the dry soil run on 3 Jul 1999. Lines as in Fig. 2. Additional line shows the path of a surface parcel: it follows a dry adiabat until reaching its lifting condensation level, then it follows a moist adiabat until reaching the level of neutral buoyancy (off plot).

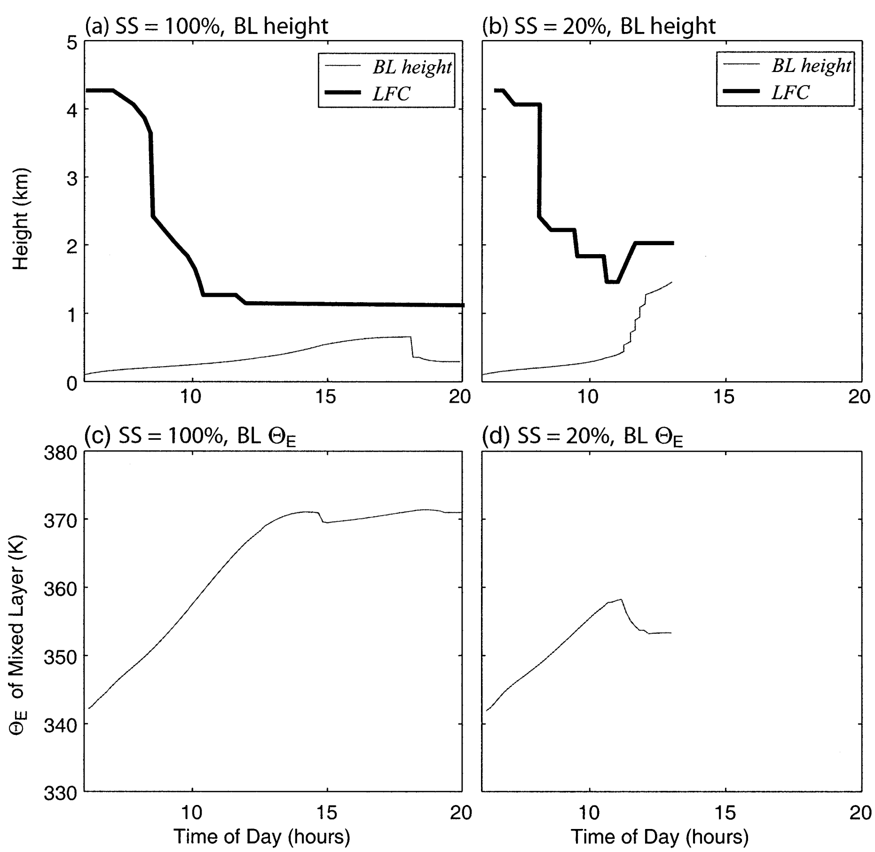

FIG. 5. As in Fig. 3 but for 23 Jul 1999.

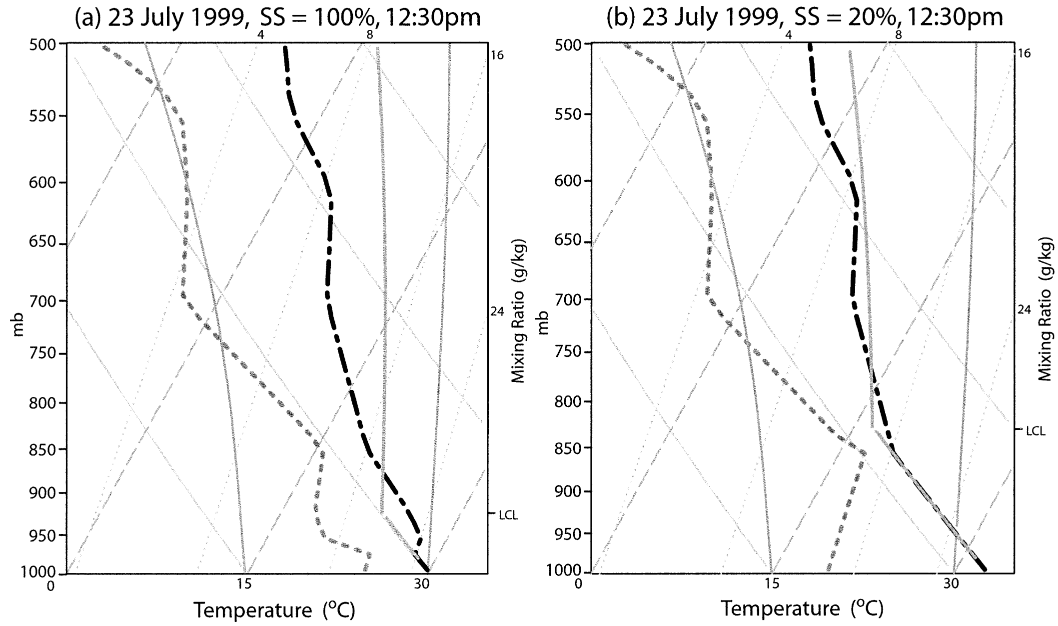

FIG. 6. As in Fig. 4 but for 1230 LT 23 Jul 1999.

indicate that the wet-bulb potential temperature θ_w (another measure of moist static energy) must increase by only ~1°C (~4°C in θ_E) in order to bring the LFC down from 855 to 940 mb. Given the steep lapse rate in this particular sounding (the quality that leads to an intermediate CTP), a small increase in θ_E leads to a large decrease in the LFC. Indeed, when convection is triggered just over 1 h later, the θ_E has risen ~2°C and the LFC has fallen from 855 to 920 mb (~1 km to ~0.8 km, Fig. 3).

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 Fig. 3b). Thus, with this early-morning sounding, convection occurs over wet soils but not over dry. The behavior on this day is representative of an energy-limited BL, in which convection is more easily triggered by a buildup of moist static energy in the BL than by deep BL growth.

Now consider the case of 23 July 1999 (initial sounding in Fig. 2b), when the opposite circumstances occur. Figure 5 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 1100 and 1200 LT was due to the easy entrainment of neutrally buoyant air between 950 and 840 mb (see initial sounding, Fig. 2b). It is the presence of this dry adiabatic portion of the sounding within the CTP region that yields a high CTP. At about 1200 LT the θ_E in the dry soil case levels out. For convection to be triggered over the dry soils by growth of the BL at constant θ_E, the BL must rise from 850 to 770 mb (Fig. 6b). The BL does continue to grow after 1230 LT, and convection is triggered at 1300 LT.

By 1230 LT, 601 J kg^{-1} of CAPE are already trapped in the moist boundary layer shown in Fig. 6a. For convection to be triggered over the wet soils by increasing θ_E at constant BL height, the θ_E must increase by ~10 K (~2.5 K in θ_w). At 1800 LT, just before the BL collapses at the end of the day, the θ_E has increased by almost 4 K and the CAPE has increased to over 3700 J kg^{-1}. This very large amount of energy cannot be released, for the BL is still a few degrees shy of the θ_E necessary for triggering convection in this scenario.

5. Results from Illinois

The one-dimensional boundary layer model described earlier was used with three summers worth of data (June–August 1997–99) from a National Oceanic and Atmospheric Administration (NOAA) radiosonde station located in Lincoln, Illinois (station ILX). NOAA's National Virtual Data System (NVDS) consists of 70 to 80 stations across the continental United States, with daily 1200 and 0000 UTC (0600 and 1800 Illinois local time) radiosonde launches. Station ILX was the only station in Illinois operational during the late 1990s.

Of the 273 days from the three summers, 1200 UTC soundings from 225 days were available for model initialization. The 48 other days were either missing from the NVDS database, or already showed rain or heavy cloud cover at 0600 LT. 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%). As explained in section 2, the BL in the model grows until one of three things happens: deep convection is triggered with rainfall likely, shallow convection is triggered with shallow clouds likely, or the day ends with no convective triggering. In general, convection was triggered in the mod-

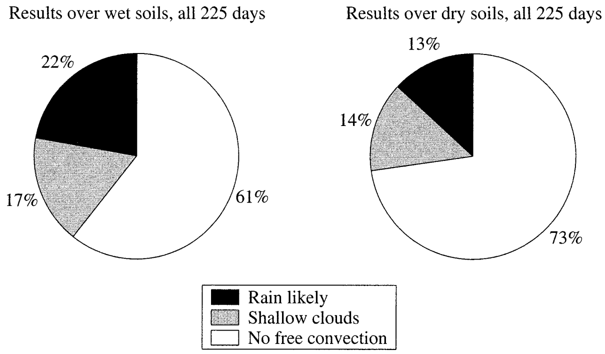

FIG. 7. For all of the 225 available days at Lincoln, IL, during the summers of 1997–99, convection in the model is more likely to be triggered over wet soils than over dry soils.

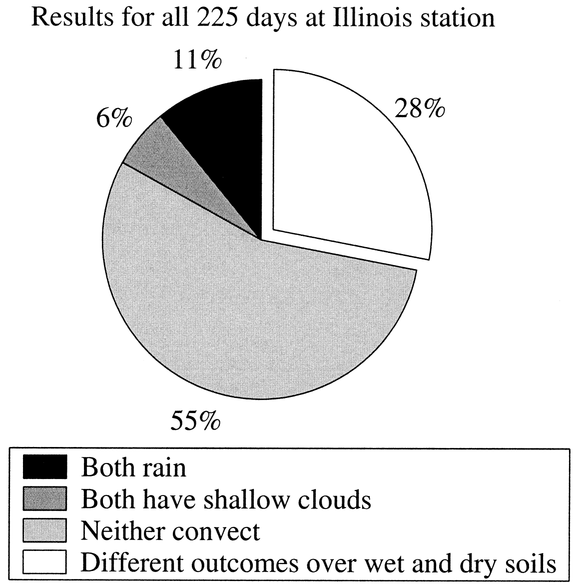

FIG. 8. The outcome of about three-quarters of the 225 available days is unaffected by the soil condition: 28% have different results over the dry and the wet soils.

el more often over wet soils than over dry (Fig. 7): rain was likely 22% of the time over wet soils, but only 13% over dry soils. Shallow clouds were likely 17% of the time over wet soils, and 14% over dry soils. The rest of the model runs (61% over wet soils, 73% over dry) ended with no triggering of convection.

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 occured over different soil conditions. The first three categories are all situations where the partitioning of fluxes at the land surface was not the critical factor determining the convective potential of the system: these are called atmospherically controlled cases. Cases in the fourth category are called nonatmospherically controlled: these are the cases where the land surface moisture condition has the potential to determine whether or not convection is triggered. Both soil conditions led 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 (Fig. 8).

We will now briefly discuss the predominant atmospheric conditions on days when the model results are the same over wet and dry soils. In section 5b, we discuss in greater detail the cases where the soil moisture condition changes the final outcome of the model. These are the cases where the newly developed convective triggering potential, when coupled with a measure of low-level humidity such as HI_{low}, distinguishes days where rainfall is more likely to occur over wet soils from those where rainfall is more likely to occur over dry soils. Comparisons of the ability of these two measures to separate clusters of rainy, cloudy, and nonconvective days with other pairs of commonly used meteorological measures and indices indicate that the CTP and HI_{low} outperform all other pairs (not shown, see Findell 2001).

a. Atmospherically controlled outcome

In this section we discuss the days where convective triggering 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 controlled by the condition of the early-morning atmospheric profile. Figure 9 shows the values of the CTP and the modified humidity index HI_{low} with symbols to indicate whether rainfall, shallow clouds, or no convection occured.

A total of 124 of the cases investigated led to no convection over either dry or saturated soils (Fig. 9). Almost all of these days were too dry for convection to be triggered in the model, and many were also too stable. The modified humidity index does an excellent job of screening out cases where convection is limited by excessive aridity. When $HI_{low} > 15°C$, there is not enough low-level humidity to allow for rainfall or shallow clouds to develop in the model, regardless of surface flux contributions: about 3/4 of these 124 cases had an $HI_{low} > 15°C$. Cases with a sounding that is too stable for rainfall to occur are well classified by a CTP < 0 J kg^{-1}: about 1/3 of the 124 cases met this condition. Many of the traditional atmospheric measures and indices also filter out many of these very stable cases.

Thirteen of the 225 cases explored from the summers of 1997–99 led to the formation of shallow clouds over both wet and dry soils (Fig. 9). Of these 13 cases, 8 were initial soundings with a warm and dry air mass at upper levels that prevented deep convection but allowed shallow clouds to form beneath this inversion. The other five were cases where the initial sounding was nearly moist adiabatic essentially all the way from near the ground surface to the top of the sounding (higher than 200 mb). In these cases, significant CAPE could not form before free convection was triggered.

There were 25 cases during the summers of 1997–99

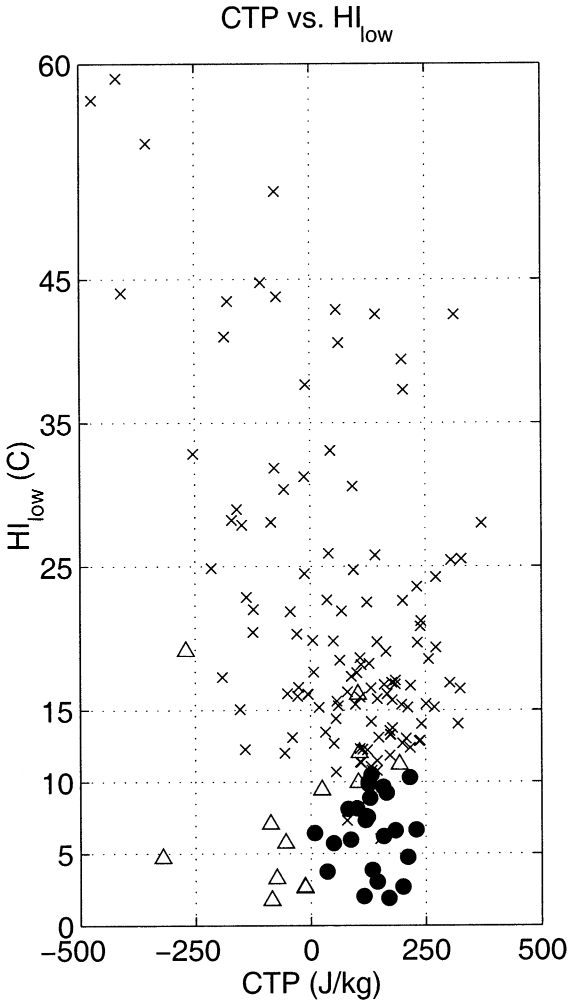

FIG. 9. Values of the CTP and HI_{low} for days when outcomes of dry soil and wet soil model runs are the same. Filled circles indicates rain over both soil states; triangles: shallow clouds over both; x: no convection over either.

at the Lincoln, Illinois, station when both extremely wet and extremely dry soils result in deep convection in the model (Fig. 9). On these days, the early-morning atmosphere was very humid ($HI_{low} \leq 10.5$°C in all 25 cases) and contained no inversions to block deep convection (CTP > 0 J kg^{-1}). Despite the atmospherically controlled label applied to these 25 cases, Fig. 10 shows that the properties of the boundary layer at the time of convective triggering were significantly different over soils of different moisture content. The anticipated result of higher soil moisture leading to higher boundary layer θ_E (Betts et al. 1996; Eltahir 1998) is indeed noticable, with a 5.4°C difference being significant at the α = 0.0375 level. Accompanying these higher θ_E values come larger CAPEs (an 850 J kg^{-1} difference), deeper convection depths (a 1.18-km difference), and smaller dewpoint depressions (a 4.2°C difference), all significant at the α = 0.0015 level. 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 wet-bulb potential temperature value (θ_w), while above this threshold there is a ~1000 (J kg^{-1}) °C^{-1} slope of increasing CAPE with increasing θ_w. Eltahir and Pal (1996) also found this threshold behavior, and further showed that above this threshold, CAPE is linearly correlated with rainfall depth in the Amazon. This suggests a positive feedback mechanism between soil moisture and the depth of rainfall in Illinois. 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.

b. Soil moisture affects outcome

Figure 11 divides the 63 cases when different soil moisture conditions led to different model results into wet soil advantage and dry soil advantage days. Rain occurs over wet soils but not over dry 41% of the time, while the reverse occurs only 8% of the time. Similarly, shallow clouds occur over wet soils but not dry 40% of the time, but only 11% of the time are there shallow clouds over dry but not over wet soils. Figure 12 shows that these data are fairly well stratified in CTP-HI_{low} space. The wet soil advantage cases with rain (filled circles) all have $0 \leq$ CTP ≤ 200 J kg^{-1} and 18 of the 24 have $5 \leq HI_{low} \leq 12$ K. All but one of the cases with rain over dry soils but not over wet (filled triangles) have CTP ≥ 200 J kg^{-1} and $HI_{low} \geq 11$ K. The case studies presented in section 2 describe behavior typical of the wet soil and dry soil advantage days.

c. Summary of Illinois results

Figure 13 is a composite sketch of the information provided in Figs. 9 and 12, separated into responses in the model runs with wet soils and those with dry soils. This figure summarizes the predictive capability gained from use of the CTP and HI_{low} as measures of the early-morning atmospheric setting, according to this 1D modeling study using data from Illinois. As shown in this figure, in very dry or very humid atmospheres, the model outcome is determined by the atmosphere alone:

- $HI_{low} > 15$°C:

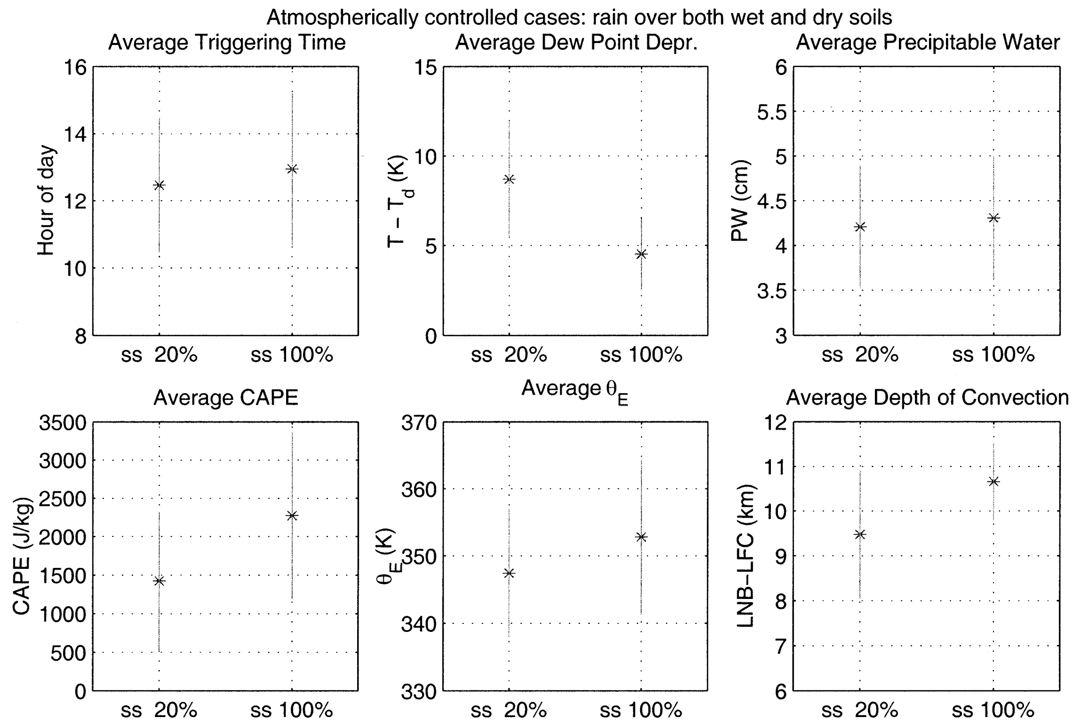

FIG. 10. Average ± one standard deviation of (a) free convection triggering time, (b) dewpoint depression, (c) precipitable water, (d) CAPE, (e) θ_E, and (f) depth of convection, for the 25 instances when deep convection (likely to rain) is triggered in both the saturated and the dry soil runs. Differences between the wet soil and dry soil values of the dewpoint depression, CAPE, θ_E, and depth of convection are highly statistically significant (see text).

  - Any CTP: no convection will result over any soil condition.
- $HI_{low} < 5°C$:
  - $CTP > 0$ J kg^{-1}: rain will occur over any soil condition;
  - $CTP < 0$ J kg^{-1}: shallow clouds will result over any soil condition;

At intermediate humidity levels, the land surface moisture condition can significantly impact the likelihood of rain, and the CTP can help to determine what that impact will be:

- $5°C < HI_{low} < 10°C$:
  - $CTP < 0$ J kg^{-1}: Shallow clouds over wet soils. No convection over dry soils.
  - $CTP > 0$ J kg^{-1}: Wet soils favored. Rain over wet soils, rain likely (but not certain) over dry soils.
- $10°C < HI_{low} < 15°C$:
  - $CTP < 50$ J kg^{-1}: Shallow clouds likely (but not certain) over wet soils. No convection over dry soils.
  - 50 J kg^{-1} $< CTP <$ 200 J kg^{-1}: Transition zone: Any outcome possible. Convection of either kind is more likely over dry than wet soils, but no convection is the most likely result over either.
  - $CTP > 200$ J kg^{-1}: Dry soils favored. No convection over wet soils, rain or shallow clouds possible over dry.

6. One-dimensional BL results from other stations

In order to determine if the CTP-HI_{low} approach used to classify atmospheric conditions yielding a wet soil or a dry soil convective advantage was valid outside the region of Illinois, the methodology of section 5 was applied to four additional stations from other parts of the United States. For each of these four additional stations, [Wilmington, OH (station ILN, latitude 39.4°N), Shreveport, LA (station SHV, latitude 32.5°N), Charleston, SC (station CHS, latitude 32.9°N), and Albuquerque, NM (station ABQ, latitude 35.0°N)], 1D model runs were performed for each day from the summer of 1998 with radiative conditions determined for the actual latitude of the station on 29 July. At the four stations, 70, 75, 73, and 86 days, respectively, were used for model initialization. Composite plots of the results from these four stations are given in Fig. 14. These results show consistency with the CTP-HI_{low} framework developed from Illinois soundings, suggesting that the framework is applicable in a wide range of atmospheric and geographic settings. Similarly, sensitivity tests with varing radiative conditions at the Illinois station (not shown) showed expected changes with more (less) deep convection with increased (decreased) solar radiation, but a continued agreement with the wet and dry soil advantages described by the CTP-HI_{low} framework. The results of Fig. 14 also add information to a portion of

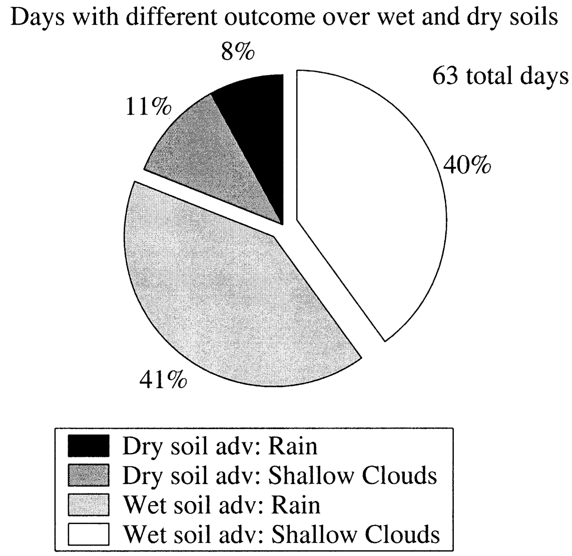

FIG. 11. Division of the outcome combinations for the 63 days in which the model resulted in different outcomes over wet and dry soils.

CTP-HI_{low} space that was not covered by data from Illinois. In the HI_{low} range of 5°–10°C, there were no Illinois soundings with CTP $>$ 225 J kg^{-1}. Data from these additional stations show that the wet soil advantage regime can be extended to CTPs of up to 300 J kg^{-1} in this HI_{low} range (Fig. 14). This emphasizes the point that both the temperature structure and the humidity of the low-level air are critical factors determining the nature of interactions between the land surface and the boundary layer.

Experiments with various modifications on the pressure levels included in the definition of HI_{low} indicated that most low-level humidity measures could be used to create a robust CTP-HI_{low} framework. The value HI_{low} as defined in this work (the sum of the dewpoint depressions 50 and 150 mb above the ground surface, usually near 950 and 850 mb) was the best performer, but all had a few outliers caused by sharp humidity drops just below one of the levels included in the definition. This happens in two cases in Fig. 14. Both show rain over wet soils only with intermediate CTPs and very high HI_{low}s. Both are cases from station ILN 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^{-1} between 860 and 840 mb (8–2 g kg^{-1}) while convection was triggered in the model 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. Future work will include the development of an index that describes the humidity throughout the lower troposphere, rather than at a few distinct levels. This should not be as sensitive to extreme humidity changes.

Table 1 separates the data points of Fig. 14 into the outcomes at the four individual stations analyzed. Within the framework of this model, the numbers suggest that during the summer of 1998 there was the potential for a positive feedback between soil moisture and deep convection at station ILN (Wilmington, OH), negative feedbacks at stations CHS (Charleston, SC) and ABQ (Albuquerque, NM), and a neutral response at station SHV (Shreveport, LA). The high frequency of modeled rainfall events and the likelihood of deeper rainfall depths when convection is triggered over wet soils as opposed to dry soils is expected to dampen the negative feedback signals. This dampening effect should be stronger at station CHS than at station ABQ because of the high percentage of days with rainfall expected over both soil types at CHS (32.9%; only 11.6% at ABQ). These feedback signals are discussed in detail in Findell and Eltahir (2003, hereafter Part II), with particular emphasis placed on the negative feedback signals seen in some years in the southwest (the region influenced by the North American monsoon system and the dryline region of the Texas and Oklahoma panhandles).

Figure 14 shows that, as predicted by the model, 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, although the range of circumstances must be further expanded to fully cover CTP-HI_{low} space. More significantly, it suggests that for matters of convective triggering and response to land surface conditions, the degree of departure from moist adiabatic conditions between approximately 1 and 3 km AGS is important in all the locations studied. These results, coupled with those from Illinois summarized by Fig. 13, were used to generate the full CTP-HI_{low} framework for analyzing soil moisture–rainfall feedbacks presented in Fig. 15.

7. Discussion

The case studies presented in section 4 highlight the significance of the convective triggering potential within the context of a one-dimensional boundary layer model. In Findell and Eltahir (2003a) we apply the framework developed here to three-dimensional simulations with the fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5; Grell et al. 1995) and to an analysis of observations of soil moisture, rainfall, and BL properties from the FIFE experiment in Kansas (Sellers et al. 1992). The work in Findell and Eltahir (2003a) highlights the importance of the vertical profile of the winds in influencing the triggering of deep convection; it shows that the winds form a crucial third dimension to the CTP-HI_{low} framework. This study, however, focuses on the results of the one-dimensional BL model and the physical reasons behind the differing behavior seen in

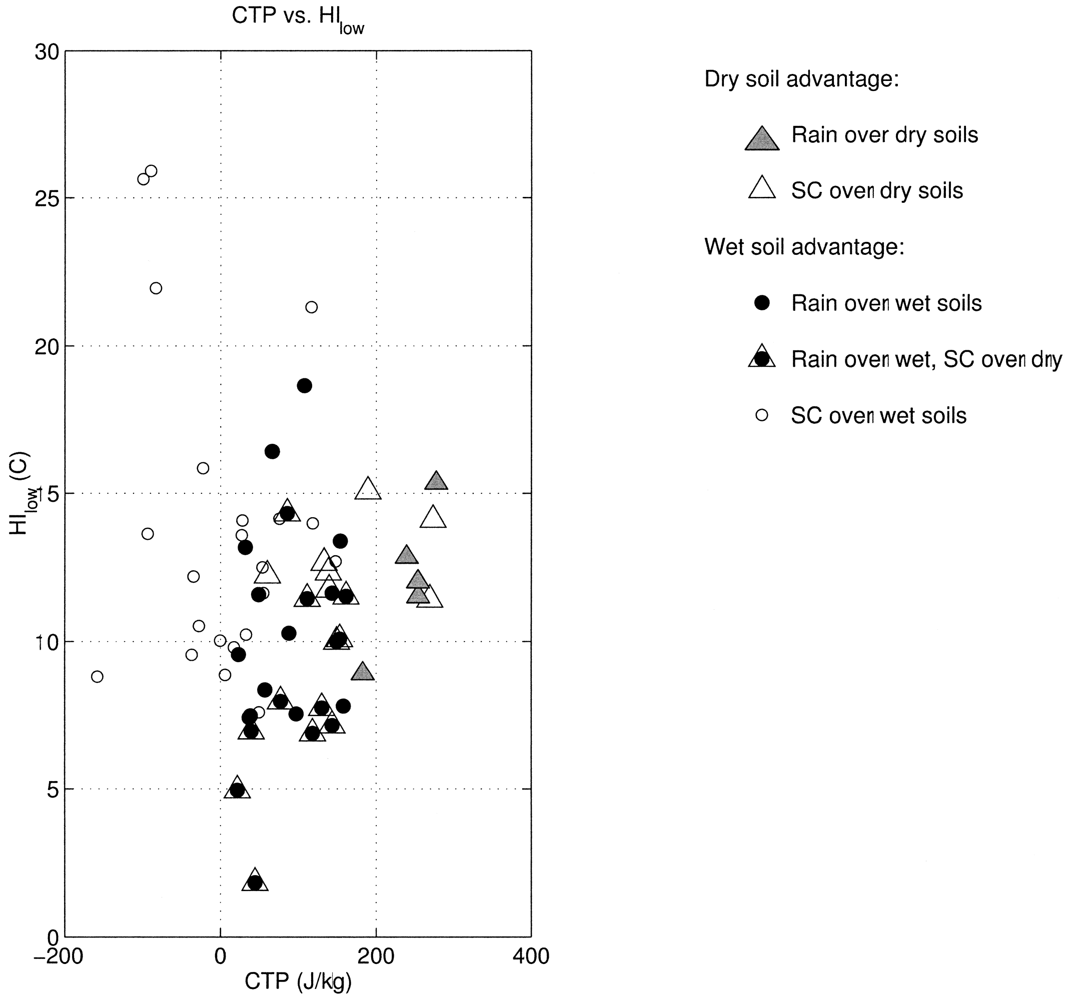

FIG. 12. Values of the CTP and HI_{low} with model outcomes when different soil conditions led to different results. Triangles indicate dry soil advantage, circles indicate wet soil advantage; filled symbols indicate rain, open symbols indicate shallow clouds. (One shallow cloud over wet soils outlier is removed: CTP ≈ -540 J kg^{-1}, $HI_{low} \approx 55$ K.)

model runs initialized with different CTP-HI_{low} environments.

Within the critical CTP region the temperature lapse rate is important for determining the ease with which entrainment, and therefore boundary layer growth, can occur. Similarly, the temperature lapse rate in this region determines the rate that the LFC will drop with increasing BL θ_E. This explains why the CTP assesses more than just the stability of a sounding: the CTP also provides critical information about the boundary layer response to surface fluxes in a given atmospheric setting.

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 θ_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, little 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.

Though the temperature lapse rate is a crucial element of atmospheric predisposition to convection, so too is the low-level humidity. Just as many stability indices have been used to distinguish very stable soundings from those with some convective potential (e.g., Show-

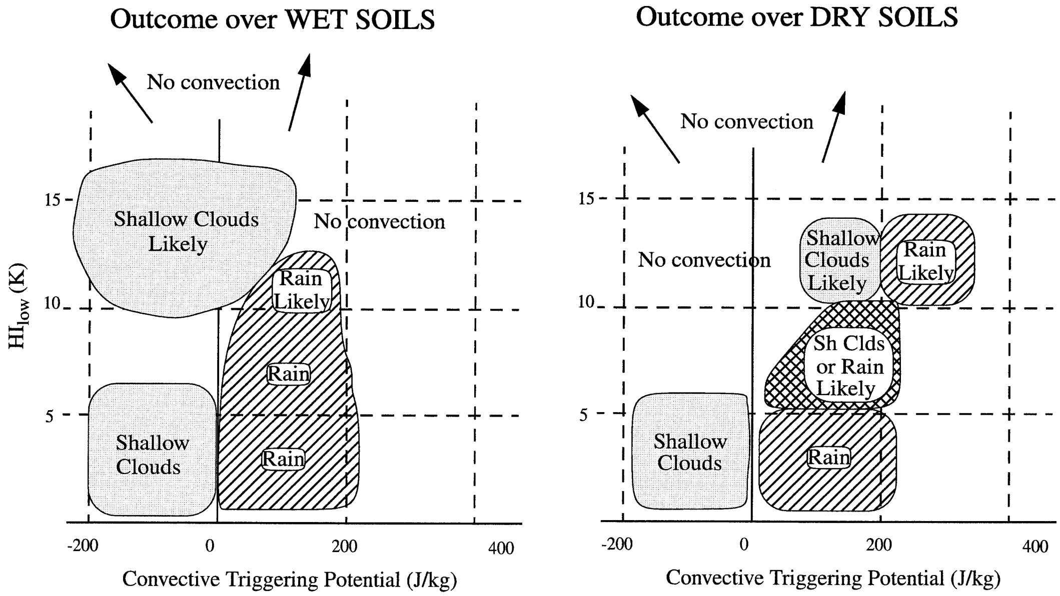

FIG. 13. Summary of 1D model outcomes over wet and dry soils, given early-morning values of the CTP and HI_{low} from Illinois.

alter index, Showalter 1953), many humidity indices have been used to separate soundings that are too dry for rainfall from those with enough humidity to allow for rainfall (e.g., humidity index, Lytinska et al. 1976). The limitation of these indices is that additional information is needed when the sounding shows sufficient instability or relative humidity: they do not provide enough information to determine where or even if convection will occur in these circumstances. When the CTP is coupled with a variation on the humidity index, HI_{low}, the two measures create a framework that greatly improves the ability to distinguish early-morning soundings likely to lead to convection in the model from those unlikely to lead to convection. More importantly, the CTP-HI_{low} framework (Fig. 15) allows one to distinguish soundings predisposed to convection in regions of high latent heat flux from those predisposed to convection in regions of high sensible heat flux.

The 1D analytical work of Haiden (1997) and of Ek and Mahrt (1994) support these findings. Haiden (1997) 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 in his 1D model because of the rapid rise of the lifting condensation level (and fall of θ_E) that accompanies rapid BL growth over dry soils. For wet soils, 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 θ_E increase. The work presented here extends the results of Haiden (1997) to initialization with soundings rather than with idealized potential temperature and humidity lapse rates. This work also shows that the lowest 300 mb of a sounding are critical in processes related to feedbacks from the land surface.

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 both 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,’’ (Ek and Mahrt 1994, p. 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

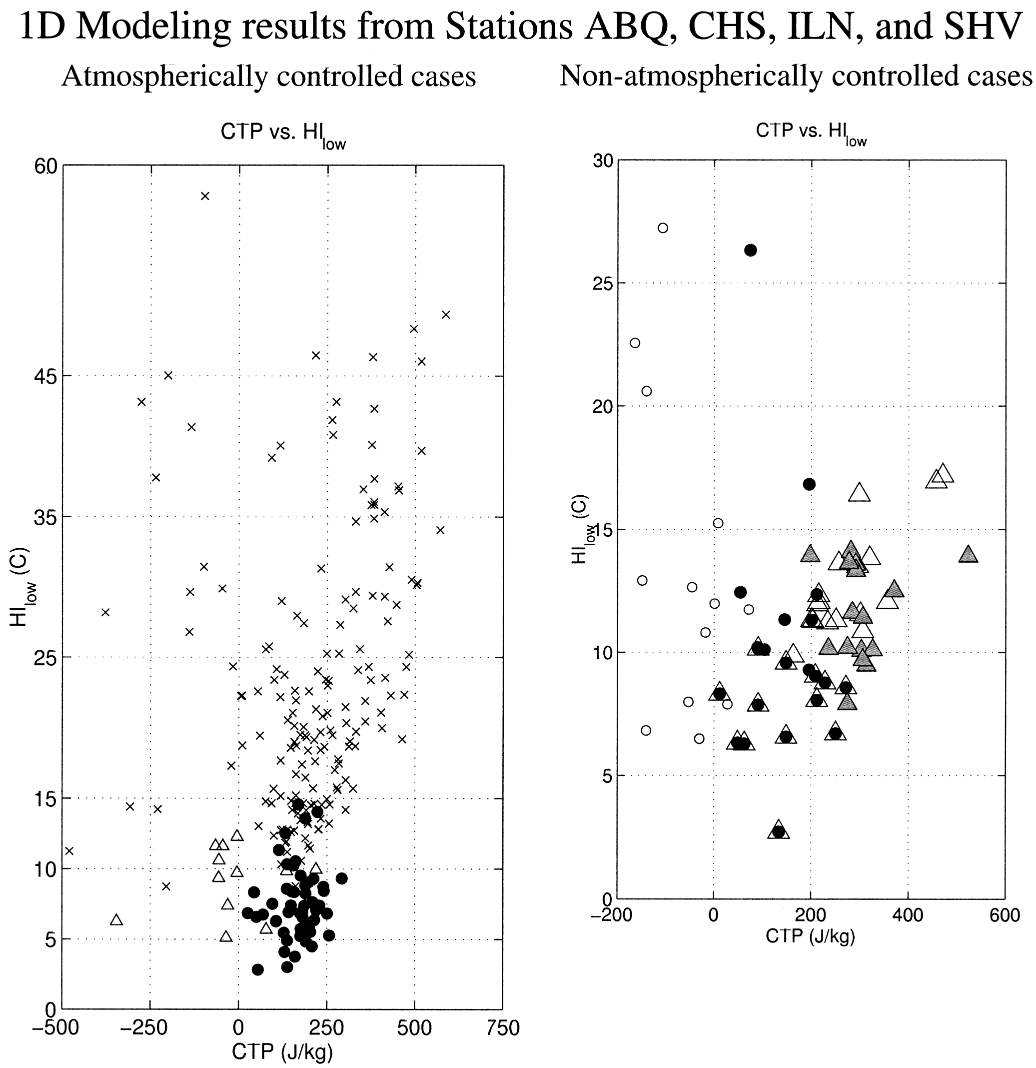

FIG. 14. Composites of the 1D model results from the four additional stations described in the text and in the previous four figures. The wet soil advantage region is approximately bound by the blue ellipse, and the dry soil advantage region by the red ellipse. The results from these four stations are consistent with the framework developed with data from Illinois, and add additional information in areas unpopulated by data from Illinois. Symbols for the left plot as is Fig. 9; those for the right plot as in Fig. 12.

TABLE 1. Results of 1D model runs from four additional stations (occurences in %). The numbers suggest that during the summer of 1998 there was a positive feedback between soil moisture and rainfall at Station ILN (Wilmington, OH), negative feedbacks at stations CHS (Charleston, SC), and ABQ (Albuquerque, NM), and a neutral response at station SHV (Shreveport, LA).

	Atmospherically controlled cases (%)			Nonatmospherically controlled cases (%)			
Station	Both rain	Both SC	Neither convective	Wet Ad: Rain	Wet Ad: SC	Dry Ad: Rain	Dry Ad: SC
ILN	23.2	8.7	40.6	**13.0**	**13.0**	1.4	0.0
SHV	16.0	2.7	68.0	2.7	4.0	4.0	2.7
CHS	32.9	1.4	43.8	2.7	1.4	**12.3**	**5.5**
ABQ	11.6	1.2	72.1	1.2	0.0	**8.1**	**5.8**

SC = shallow clouds; Ad = advantage.

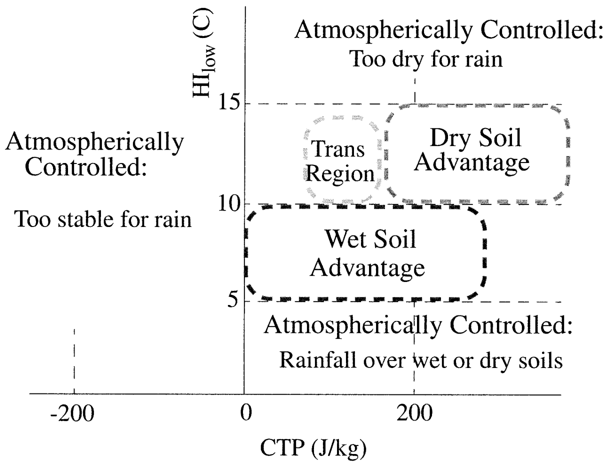

FIG. 15. The CTP-HI$_{low}$ framework for describing atmospheric controls on soil moisture–rainfall feedbacks. Only when the early-morning atmosphere has CTP > 0 J kg^{-1} and $5 < \mathrm{HI}_{low} < 15$°C can flux partioning at the surface influence the triggering of convection.

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 in data from the FIFE site in Kansas. They found that increased soil moisture led to an increased diurnal θ_E range, and was accompanied by a decrease in the peak depth to the LCL from $\sim$230 mb over dry soils to $\sim$130 mb over wet soils. Note that the 100 mb between these two LCL depths is captured by the critical CTP region. This difference in θ_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 θ_E and the lower cloud-base can be expected to organize mesoscale convective systems, just as warmer sea surface temperatures do over the ocean,'' (Betts and Ball 1995, p. 25 692).

With a 1D PBL box model, de Ridder (1997) calculated the dependence of θ_E on the evaporative fraction, α, and determined that the potential for moist convection increases with α, except in very dry atmospheres. This is consistent with the results here regarding the lack of convection above a threshold humidity decificit, independent of land surface conditions. 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, 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, in these circumstances, the likelihood for convection is independent of the land surface fluxes.

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, though further work with observations is necessary. The value HI$_{low}$ should be closely correlated with T_{dpr}, since it considers the dewpoint depression at relatively low levels. 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 θ_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.

8. Conclusions and future work

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, rather than from profiles produced by idealized potential temperature and humidity lapse rates. The model is used to analyze the impact of soil saturation on BL development and the triggering of convection in different atmospheric settings. Using early-morning atmospheric soundings from Illinois to initialize the model, a small positive feedback was seen between soil moisture and rainfall from three summers worth of data from central Illinois, consistent with the work of Findell and Eltahir (1997, 1999).

The newly developed convective triggering potential (CTP) is a measure of the early-morning atmospheric thermodynamic structure in the region between 100 and 300 mb (approximately 1 and 3 km) above the ground surface (AGS). The great influence of this region results from its location between the lowest $\sim$1 km, which is almost always incorporated into the boundary layer, and the free atmospheric air above $\sim$3 km, which is almost never incorporated into the BL.

The CTP is coupled with a low-level humidity index, HI_{low}, to help distinguish between different types of early-morning soundings based on model response to these differing initial states. The sounding classes are those favoring rainfall over dry soils, those favoring rainfall over wet soils, and those whose convective potential is unaffected by surface fluxes. Together, these two measures form the CTP-HI_{low} framework for analyzing atmospheric controls on soil moisture–boundary layer interactions (Fig. 15). This framework was initially developed initializing the model with data from Illinois, but additional testing using soundings from Ohio, Louisiana, South Carolina, and New Mexico suggest that it is valid for locations far removed from Illinois.

This work demonstrates that the early-morning temperature and humidity structure must be considered in order to determine how the growing boundary layer will respond to fluxes from the land surface. It shows that within the 1D model, 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 ($HI_{low} > 15°C$) or very stable ($CTP < 0$ J kg^{-1}), rainfall cannot occur, independent of flux partitioning at the surface. When the atmosphere is humid and unstable ($HI_{low} < 5°C$ and $CTP > 0$ J kg^{-1}), then rainfall should occur over both wet and dry soils, with deeper rainfall depths expected over wet soils. In the remaining circumstances (HI_{low} between 5° and 15°C, and $CTP > 0$ J kg^{-1}), then the land surface can significantly influence the likelihood of rainfall, with dry soils more likely to trigger rainfall in the high CTP–high HI_{low} section of this range, and wet soils more likely in the low CTP–low HI_{low} section.

The power of this framework lies in the ability to determine from a simple analysis of early-morning soundings whether a geographical region is likely to see climate-scale feedbacks between soil moisture and rainfall, and what the nature of those feedbacks are likely to be. In Part II, the framework is used with data from all of the contiguous 48 United States to locate regions of potential positive and negative feedbacks between the land surface and rainfall, and regions where the land surface conditions cannot play a large role in triggering convection. Research into the CTP-HI_{low} characteristics of other parts of the world is currently underway.

A third paper (Findell and Eltahir 2003a) highlights the effects of the vertical profile of the winds through experiments with the three-dimensional, fifth-generation Mesoscale Model (MM5; Grell et al. 1995) and an analysis of data from the FIFE experiment in Kansas (Sellers et al. 1992) in the context of the CTP-HI_{low} framework. These analyses show that the winds form a crucial third dimension to the CTP-HI_{low} framework. The identification of potentially coherent feedback regions (e.g., Part II) will help determine where to locate future observational missions to test the validity of the CTP-HI_{low} framework beyond a 1D model and to improve understanding of soil moisture–boundary layer interactions. Given the rarity of study sites with a statistically significant number of days with soil moisture observations, precipitation observations, and early-morning radiosondes, it is useful to gleen as much from models and currently available data as is possible. More work with available observational datasets is underway and will continue in the future.

The work presented here has strong implications regarding the importance of high-resolution data and model levels throughout the critical CTP region: it is critical that the vertical resolution in this region be sufficient to distinguish between a moist adiabatic and a dry adiabatic temperature lapse rate. As discussed in the introduction, 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 five stations presented in this paper and the nationwide analysis of Findell and Eltahir (2003b, Part II) reveal highly variable CTP-HI_{low} characteristics throughout the United States. However, this could also result from different model and/or forcing-data resolution in the critical CTP region. Further research is needed to determine the vertical resolution required to adequately represent this region and its control on land surface–boundary layer interactions.

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