Atmospheric Controls on Soil Moisture–Boundary Layer Interactions. Part II: Feedbacks within the Continental United States

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ABSTRACT

The CTP-HI_{low} framework for describing atmospheric controls on soil moisture–boundary layer interactions is described in a companion paper, Part I. In this paper, the framework is applied to the continental United States to investigate how differing atmospheric regimes influence local feedbacks between the land surface and the atmosphere. The framework was developed with a one-dimensional boundary layer model and is based on two measures of atmospheric thermodynamic properties: the convective triggering potential (CTP), a measure of the temperature lapse rate between approximately 1 and 3 km above the ground surface, and a low-level humidity index, HI_{low}. These two measures are used to distinguish between three types of early-morning atmospheric conditions: those favoring moist convection over dry soils, those favoring moist convection over wet soils, and those that will allow or prevent deep convective activity, independent of the surface flux partitioning.

Analyses of multiyear CTP-HI_{low} scatterplots from radiosonde stations across the contiguous 48 United States reveal that during the summer months (June, July, and August) positive feedbacks between soil moisture and moist convection are likely in much of the eastern half of the country. Over the western half of the country, atmospheric conditions and the likelihood of moist convection are largely determined by oceanic influences, and land surface conditions in the summer are unlikely to impact convective triggering. The only area showing a potential negative feedback is in the dryline and monsoon region of the arid Southwest. This potential arises because of the topography of this and surrounding regions. A relatively narrow band of stations lies in between the eastern and western portions of the country, in some years behaving like the stations to the west and in other years behaving like the stations to the east.

1. Introduction

a. Motivation

Feedbacks from the earth's surface to the atmosphere are an instrumental part of global climatic processes. Extensive research on the El Niño–Southern Oscillation phenomenon connects anomalous sea surface temperatures (SSTs) in the eastern Pacific Ocean with dramatic shifts in weather patterns over much of the globe. Like SSTs, vegetation cover and soil moisture content control the partitioning of energy fluxes at the earth's surface, and, like SSTs, land surface conditions in some regions yield more significant feedback influences than others. Fundamental to the determination of the potential land surface influence in a region are the predominant atmospheric conditions in that area.

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In a companion to this paper (Findell and Eltahir 2003b, hereafter Part I), the authors present a framework for assessing the structure of an early-morning atmospheric sounding and the likelihood and nature of feedbacks from the land surface, given that atmospheric structure. A one-dimensional boundary layer (BL) model was used to investigate the response to different soil moisture conditions (very wet vs very dry) in different early-morning atmospheres. The framework is based on two measures of atmospheric thermodynamic properties: the convective triggering potential (CTP), a measure of the temperature lapse rate between approximately 1 and 3 km above the ground surface, and a low-level humidity index, HI_{low}. These two measures are used to distinguish between three types of early-morning atmospheric conditions: those favoring moist convection over dry soils, those favoring moist convection over wet soils, and those that will allow or prevent deep convective activity, independent of the surface flux partitioning.

Further work with three-dimensional models and with observations is needed to asses the validity of the CTP-HI_{low} framework beyond a 1D model. That work is begun in Findell and Eltahir (2003a). This paper, however, applies the simple CTP-HI_{low} framework to the continental United States in order to determine where such future investigations might be most revealing.

b. Background

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. A negative feedback was also shown by the modeling study of Paegle et al. (1996), which focused on the low-level jet and its influence on rainfall in the Mississippi basin during the floods of 1993. Pan et al. (1996), in contrast, showed a positive feedback between soil moisture and rainfall in the United States in their model of 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, and the work of Seth and Giorgi (1998) showed that the domain size can significantly influence the model outcome.

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 First International Land Surface Climatology Project (ISLCP) Field Experiment (FIFE) experiment in Kansas, and one day from an Oklahoma summer. 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. 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,'' (Wetzel et al. 1996, 7361–7362). In the Oklahoma case, there was a very shallow nocturnal inversion, which was easily eroded. In the FIFE case, the preexisting 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 in their modeling work, claiming that under most conditions sensible heat flux plays only a secondary role in 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.

In an investigation of the role of the capping inversion on the development of hail storms observed in northeastern Colorado, Mahrt (1977) 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 explored the significance of the height of the capping inversion with their numerical model. They concluded that there is an intermediate range of inversion strengths most conducive to the development of precipitating convection. When the cap was high, entrainment was reduced because ''the depth of the initial mixed layer [was] close to that of the afternoon mixed layer,'' (Segal et al. 1995, p. 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) in a study of data from the FIFE site. 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,'' (Chen and Avissar 1994, p. 1397).

Ek and Mahrt (1994) used data from the Hydrological–Atmospheric Pilot Experiment–Modélisation du Bilan Hydrique (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 stratified. 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 CO_2 (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 that differ from these general trends. In order to fully understand the implications of these results, we need a better understanding of how interactions between soil moisture and rainfall are controlled.

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 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.

As mentioned above, the work in Part I provides a framework in which to determine the potential for and the nature of feedbacks between the land surface and rainfall, given regional atmospheric patterns and characteristics. In this paper, we use the CTP-HI_{low} framework, summarized by Fig. 1, to show how location-dependent atmospheric conditions may lead to different interactions between soil moisture and rainfall across the contiguous United States.

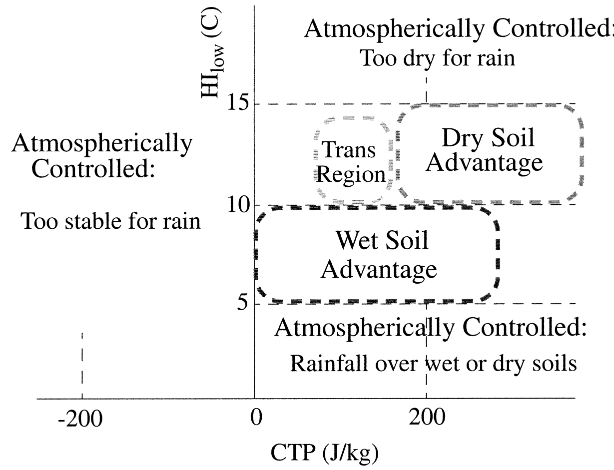

FIG. 1. The CTP-HI_{low} framework for describing atmospheric controls on soil moisture–rainfall feedbacks. Only when the early-morning atmosphere has CTP > 0 J kg^{-1} and $5 < HI_{low} < 15°C$ can flux partioning at the surface influence the triggering of convection. (Fig. reprinted from Findell and Eltahir 2003b.)

c. Outline of study

The CTP-HI_{low} framework (Fig. 1) is briefly described in section 2. Scatterplots of the CTP and HI_{low} of early-morning soundings from stations in the National Oceanic and Atmospheric Administration's (NOAA) National Virtual Data System (NVDS) were created from data from June, July, and August. At the time of the study, the database spanned the years 1957–98, and stations within the continental United States with at least 10 yr of data were included in the analyses. Regional CTP-HI_{low} characteristics quickly emerged, leading to the regions depicted in Fig. 2. The station classification protocol used in this work is described in section 3, and the regions of Fig. 2 are discussed in the four subsequent sections.

In much of the western half of the country, almost all of the days fall within the three atmospherically controlled regimes of CTP-HI_{low} space (too dry for rainfall, too stable for rainfall, rainfall expected), leaving little possibility for soil moisture conditions to impact convective triggering (section 4). One region shows a strong potential for a negative feedback (section 5), the eastern half of the country shows signs of positive feedbacks between soil moisture and rainfall (section 6), and a narrow transition region shows significant occurences of days in both the wet soil advantage regime and the

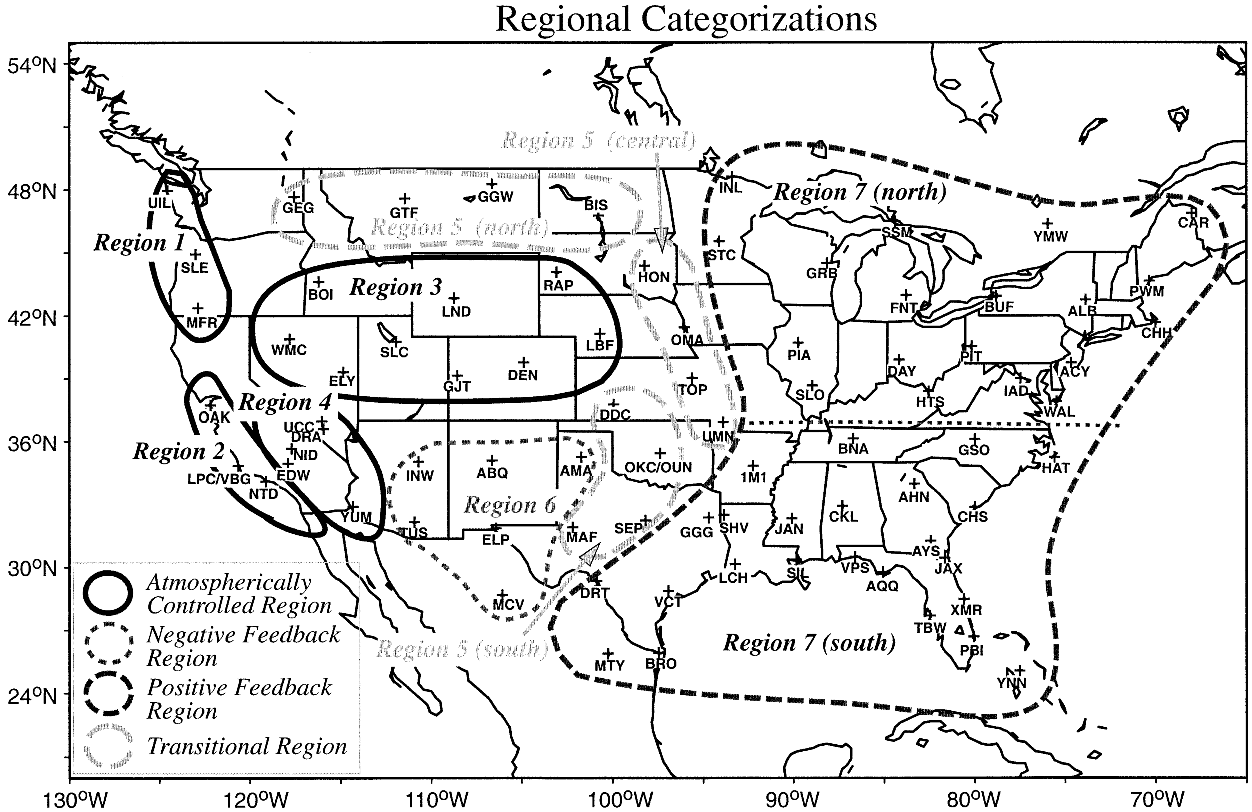

FIG. 2. Representative regions within the continental United States, based on CTP-HI_{low} scatterplots for the 89 stations with at least 10 yr of data during 1957–98 (see text for full description).

dry soil advantage regime (section 7). Section 8 has a brief discussion along with the conclusions of this work.

2. The CTP-HI_{low} framework

With the one-dimensional boundary layer modeling detailed in Part I, we established a framework for analyzing soil moisture–boundary layer interactions in different early-morning atmospheric conditions. This framework (Fig. 1) makes use of two measures of atmospheric thermodynamic properties: the convective triggering potential (CTP) and a low-level humidity index, HI_{low}. These two measures are briefly described below and fully defined in the appendix. They are used to distinguish between three types of early-morning soundings: those favoring deep convection over dry soils, those favoring deep convection over wet soils, and those whose convective potential is unaffected by the partitioning of fluxes at the surface. This last group includes days where deep, precipitating convection is very likely over either wet or dry soils, as well as days that are so dry and/or so stable that rainfall is very unlikely over any surface. A crucial third dimension of the CTP-HI_{low} framework is the vertical profile of the winds; these impacts are discussed in Findell and Eltahir (2003a).

The CTP-HI_{low} framework (Fig. 1) was developed through the use of a 1D boundary layer model initialized with wet or dry soils and with soundings taken from five different locations throughout the United States (see Part I). The CTP axis is a measure of the departure from a moist adiabatic temperature lapse rate in the region between 100 and 300 mb above the ground surface (about 1 to 3 km). When the lapse rate is close to dry adiabatic, the CTP is large and areas of high sensible heat flux were seen to have an advantage in triggering convection in the 1D model. A smaller but still positive CTP means that the lapse rate is closer to moist adiabatic, giving areas of high latent heat flux a convective advantage. Finally, a negative CTP indicates a temperature inversion that prevented deep convection over any land surface in the 1D model.

The HI_{low} axis of Fig. 1 is a measure of the humidity deficit in low-level air (the sum of the dewpoint depressions 50 and 150 mb above the surface). When the deficit was large, rainfall in the 1D model was prohibited by this atmospheric condition. When the deficit was small, the atmosphere was so close to saturation that moist convection was likely over any land surface. In between these extremes, flux partitioning at the land surface was shown to greatly influence the development of convection in the model. In drier atmospheres high sensible heat flux was the best trigger of convection, while in more humid atmospheres large contributions of humidity from the land surface was a more effective trigger. Case studies showing BL development and convective triggering in different regimes of CTP-HI_{low} space are provided in Part I.

This framework was used to divide data from early-morning soundings from stations across the United States into four separate regimes based on the expected outcome of the BL model used in Part I. The four regimes are: atmospherically controlled days, wet soil advantage days, dry soil advantage days, and transition regime days. Atmospherically controlled days can have very dry atmospheres and/or very stable atmospheres where precipitating convection cannot be forced by fluxes from the land surface in any condition, or they can have humid and conditionally unstable atmospheres where convection is likely to be triggered over any land surface. If any of these atmospheric conditions are common in an area, the land surface will have little opportunity to influence convective triggering. Therefore, particular aspects of the surface flux partitioning are not critical in this region. However, when the early-morning atmosphere is unstable and neither too humid nor too arid (Fig. 1), the land surface can and does have a significant impact on the triggering and the depth of convective rainfall.

3. Station classification protocol

NOAA's National Virtual Data System consists of stations across the continental United States with daily 1200 and 0000 UTC radiosonde launches. For each of the stations in the contiguous United States, the CTP and HI_{low} were calculated for all available 1200 UTC [0400 local time (LT) on the West Coast to 0700 LT on the East Coast–these time differences are discussed in section 8] soundings from the summers of 1957 through 1998. Stations with at least 10 yr of data were included in the results presented here. Of the 83 stations meeting this criterion, 69 had at least 20 yr of data. The shorter records were included because four of the five stations in region 4 had fewer than 20 yr of data (only DRA in Desert Rock, NV, had more) and we wanted data from more than one station to characterize this region. Florida was the only other area with a concentration of stations with records between 10 and 19 yr in length (four such stations); the other six stations with shorter records were distributed throughout the country.

At a given station and for a given summer, if 80% or more of these days fall into atmospherically controlled regimes of CTP-HI_{low} space, the station is labeled atmospherically controlled for that summer. (Sensitivity to this 80% threshold value is discussed at the end of this section.) If a station had fewer than 80% of days in atmospherically controlled regimes during a given summer, further consideration is warranted. The station is called a level-1 positive feedback station if more than half of the remaining days (the days where the land surface condition can potentially influence the triggering of convection) were in the wet soil advantage regime of CTP-HI_{low} space. Similarly, if more than half of the nonatmospherically controlled days were in the dry soil advantage regime or the transition regime, the station is a level-1 negative feedback station or a level-1 transitional station, respectively. Level-1 transitional stations were very rarely observed.

Stations that still remained unclassified were likely to show a weaker signal than the stations meeting one of the criterion listed above, and are called level-2 stations. A level-2 positive feedback station has fewer than 20% of the nonatmospherically controlled days in the dry soil advantage regime, while a level-2 negative feedback station has fewer than 20% of the nonatmospherically controlled days in the wet soil advantage regime. Level-2 transitional stations include all the remaining stations, meaning that the nonatmospherically controlled days are split fairly equally between the transition regime and the wet and dry soil advantage regimes of CTP-HI_{low} space. More specifically, there must be between 20% and 50% of these days in each of the wet and dry soil advantage regimes. Level-2 positive and negative feedback stations are rare, though level-2 transitional stations are quite common.

The regions of Fig. 2 were determined by grouping together stations with similar distributions in CTP-HI_{low} space, and with similar percentages of years classified as positive and negative (level-1 only) feedback years. Table 1 lists the average statistics for all stations within each region, and Fig. 3 shows the percentage of years labeled as either positive or negative feedback years (level-1 only). Station percentages were calculated before averaging stations within a region to account for differing record lengths.

Many stations in and west of the Rocky Mountains were atmospherically controlled almost every summer (Table 1 and Fig. 3). These four regions had between 88% and 99% of days in the atmospherically controlled regimes, suggesting an insensitivity to the cutoff value of 80%. Similarly, stations in region 7, particularly the southern part, were insensitive to the cutoff value since stations in the northern and southern parts of this region average 66% and 43% of days in the atmospherically controlled regimes, respectively. A few stations in the transitional and negative feedback regions were sensitive to the cutoff value, though varying it does not change the regional boundaries or classifications of Fig. 2 since these are based on the station distributions in CTP-HI_{low} space, as well as the percentages of positive and negative feedback years. Changing the threshold to 75% yields more atmospherically controlled years, but the ratio of positive to negative years remains similar in each region. For example, the ratio of percent positive feedback years to percent negative feedback years in the central transitional region changes from $27.7/6.2 = 4.5$ when 80% is used as the threshold value to $17.5/3.7 = 4.7$ when 75% is used. In the southern transitional

TABLE 1. Regional averages: percentages of days in each of the four general regimes of CTP-HI_{low} space for all years of data at each station that was operational for at least 10 yr between 1957 and 1998. Station averages were computed before regional averages to account for differing record lengths. The final column lists the number of stations in each region. In the dryline and moonsoon region, 1.5% of days were outside of the regions of CTP-HI_{low} space defined by the earlier analyses (pink triangles in Fig. 5b).

Region	Atmospherically controlled days (%)	Wet soil advantage days (%)	Transition regime days (%)	Dry soil advantage days (%)	Stations in region
Pacific NW	94.5	4.0	1.2	0.2	3
Pacific SW	98.7	0.4	0.6	0.3	4
Central Rockies	87.8	4.5	2.1	5.3	9
Dry intermontane	97.1	0.9	0.3	1.6	5
Transitional (north)	80.0	8.3	4.8	6.8	4
Transitional (central)	71.9	12.3	5.8	9.9	4
Transitional (south)	67.3	13.4	5.2	13.9	5
Dryline and moonsoon	75.8	8.5	1.4	12.8	6
Eastern United States (north)	65.8	17.8	7.5	8.8	19
Eastern United States (south)	43.2	35.7	10.4	10.7	24

region this ratio changes from 10.9/19.0 = 0.57 to 8.2/15.0 = 0.55. In the northern transitional region, the few years that qualify as negative feedback years with the 80% threshold (1.2%) no longer qualify when the threshold is lowered to 75%. The percent of positive feedback years changes from 15.1 to 8.2. In the dryline and monsoon region very few years qualify as positive feedback years with the 80% threshold value (2.7%) while 25.9% qualify as negative feedback years, so changing the threshold impacts the negative feedback years more significantly: the percentage falls to 17.1% and the ratio of positive to negative years increases from 0.10 to 0.16. All of the results presented in this paper were generated using a threshold of 80%.

The next four sections of this paper focus on the four types of regions shown in Fig. 2. Causes of interannual variability are briefly discussed in the final section, but a full investigation of these causes is beyond the scope of this paper.

4. Atmospherically controlled regions

Well over 90% of summer days between 1957 and 1998 were atmospherically controlled in the Pacific Northwest (region 1: Figs. 3 and 4a, Table 1). Most of the early-morning soundings were stable (CTP $<$ 0 J kg^{-1}), and a large percentage were very humid, with HI_{low} $<$ 5°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

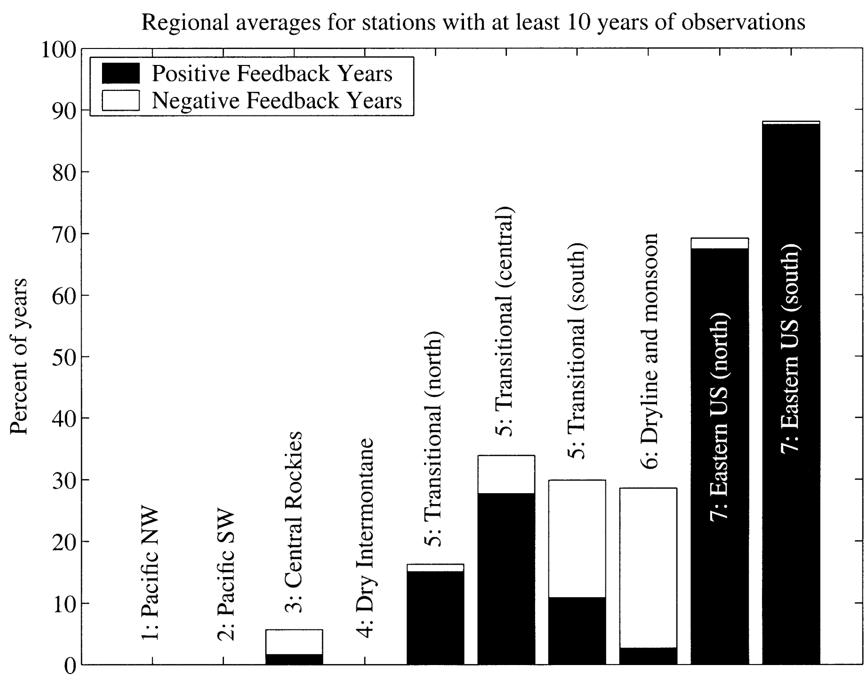

FIG. 3. Bar chart showing the average percentage of years a station in each of the regions outlined in Fig. 2 is classified as a positive or a negative feedback station.

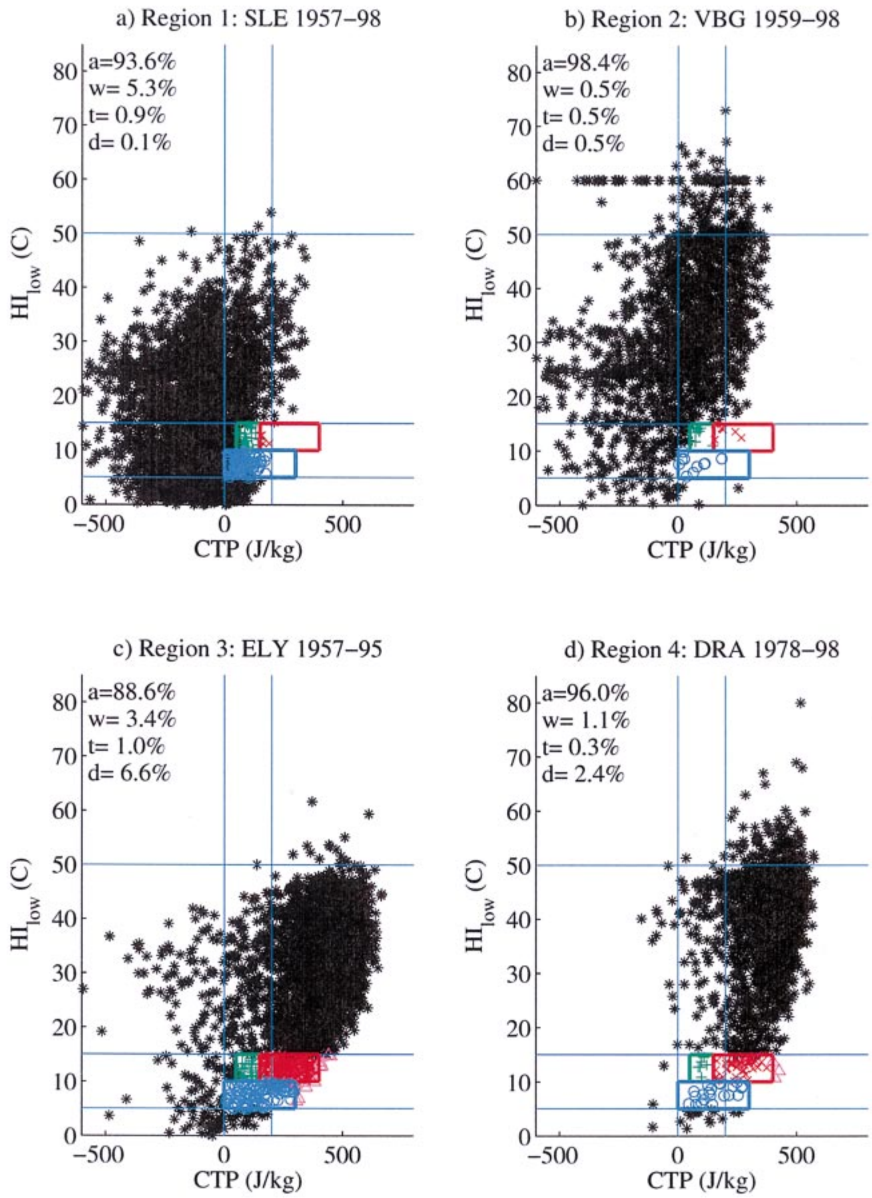

FIG. 4. Early-morning CTP and HI_{low} for all available soundings from Jun, Jul, and Aug 1957–98 from representative stations in (a) region 1, the Pacific Northwest (SLE: Salem, OR); (b) region 2, the Pacific southwest (VBG: Vandenberg, CA); (c) region 3, the central Rocky Mountains (ELY: eastern-central NV); and (d) region 4, the dry intermontane region (DRA: Desert Rock, NV). In the legends, *a* is the percentage of atmospherically controlled days (black star); *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 cross); and *d* is the percentage of days in the dry soil advantage regime (red x). Pink triangles are outside the regions of CTP-HI_{low} space defined by earlier analyses (less than 1% in each case).

Coast cool, with a nearly moist adiabatic lapse rate, and with high humidity to a considerable depth. The 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, consistent with the Bryson and Hare statement, but this mid-CTP region inversion creates strong stability (and negative CTPs, e.g., Fig. 4a), which should prohibit deepening of the preexisting low-intensity shallow clouds and/or rainfall.

Soundings from the Pacific southwest (region 2) also fall almost entirely in atmospherically controlled regions (Figs. 3 and 4b, Table 1), 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°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 this region, 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 description is consistent with the CTP-HI_{low} distribution shown in Fig. 4b.

Further inland, in the dry intermontane region (region 4), we see very different scatterplots of early-morning CTP and HI_{low} (Fig. 4d). These stations are still predominantly atmospherically controlled, with an average of 97% of days falling in these regimes (Fig. 3 and Table 1). 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 traveling over the coastal mountains and the western plateau, however, heating from the land surface raises the wet-bulb potential temperature (θ_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 θ_w is clearly evident in the increase in CTPs from the Pacific southwest to the dry intermontane region (Figs. 4b and 4d, respectively): the mean CTP at Vandenberg, California, is 15 J kg^{-1}, while that at Desert Rock, Nevada, is 363 J kg^{-1}. (The mean HI_{low} changes from 33° to 38°C.) Nearly all of the days at stations in region 4 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. Days when the land surface moisture could have an impact on the potential for rainfall are rare enough that no overriding signal of either a positive or a negative soil moisture–rainfall feedback is expected in this region (Table 1).

There is slightly more variability between stations within the central Rocky Mountain region (region 3) than in the previously discussed regions, though these stations are all still atmospherically controlled (e.g., Fig. 4c). On average, a station in the Rockies has 4.5% of days in the wet soil advantage regime, 5.3% in the dry soil advantage regime, 2.1% in the transition regime, and the remainder in atmospherically controlled regimes, suggesting little possibility of a response of rainfall to soil moisture (Fig. 3 and Table 1). As with the stations in the dry intermontane region to the south and west, the most common source region is the Pacific. The near-surface layers of this air again warm (increasing the CTP) while traveling 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. Thus, days at the Rocky Mountain stations are most typically in the atmospherically controlled regime with CTP > 0 J kg^{-1} and $HI_{low} > 15$°C, but the extremely high values of HI_{low} observed in the Pacific southwest and the dry intermontane regions (up to 85°C at some stations) are not as common at these stations.

5. Negative feedback region: The dryline and monsoon region

The stations in region 6, the dryline and monsoon region of the arid Southwest, have similar distributions in CTP-HI_{low} space, but stations on the western border (INW, Winslow, AZ, and TUS, Tucson, AZ) are influenced by the atmospherically controlled region to their west. Summers at these stations qualify as negative feedback years only 12% and 10% of the time, respectively. Stations ABQ (Albuquerque, NM) and AMA (Amarilio, TX) show the strongest signs of a potential negative feedback, with summers qualifying as negative feedback years 40% of the time. The other two stations, ELP and MCV have negative feedback summers 18% and 35% of the time, respectively. On average, these six stations had 75.8% of days in atmospherically controlled regimes, 8.5% in the wet soil advantage regime, 12.8% in the dry soil advantage regime, and 1.4% in the transition regime (Fig. 3 and Table 1). The CTP-HI_{low} distribution from Albuquerque, New Mexico, is shown in Fig. 5b.

The western half of the negative feedback region shown in Fig. 2 is the area where the North American monsoon typically extends from Mexico into the United States. As discussed in Part I in conjuction with 1D modeling results using soundings from station ABQ in Albuquerque, New Mexico, this system usually affects New Mexico and much of Arizona during July and August, sometimes beginning in late June or extending into September (Higgins and Shi 2000). The impact of the monsoonal incursion into New Mexico and Arizona is increased humidity without significant changes to the temperature profile (Wallace et al. 1999). Wallace et al.'s (1999) analysis of soundings from Tucson and Phoenix (northwest of Tucson), Arizona, found 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. 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. The temperature profiles at Tucson were essentially identical on these two types of days, but the dewpoint temperatures were 4°–6°C different from the surface extending far up the soundings. Records from

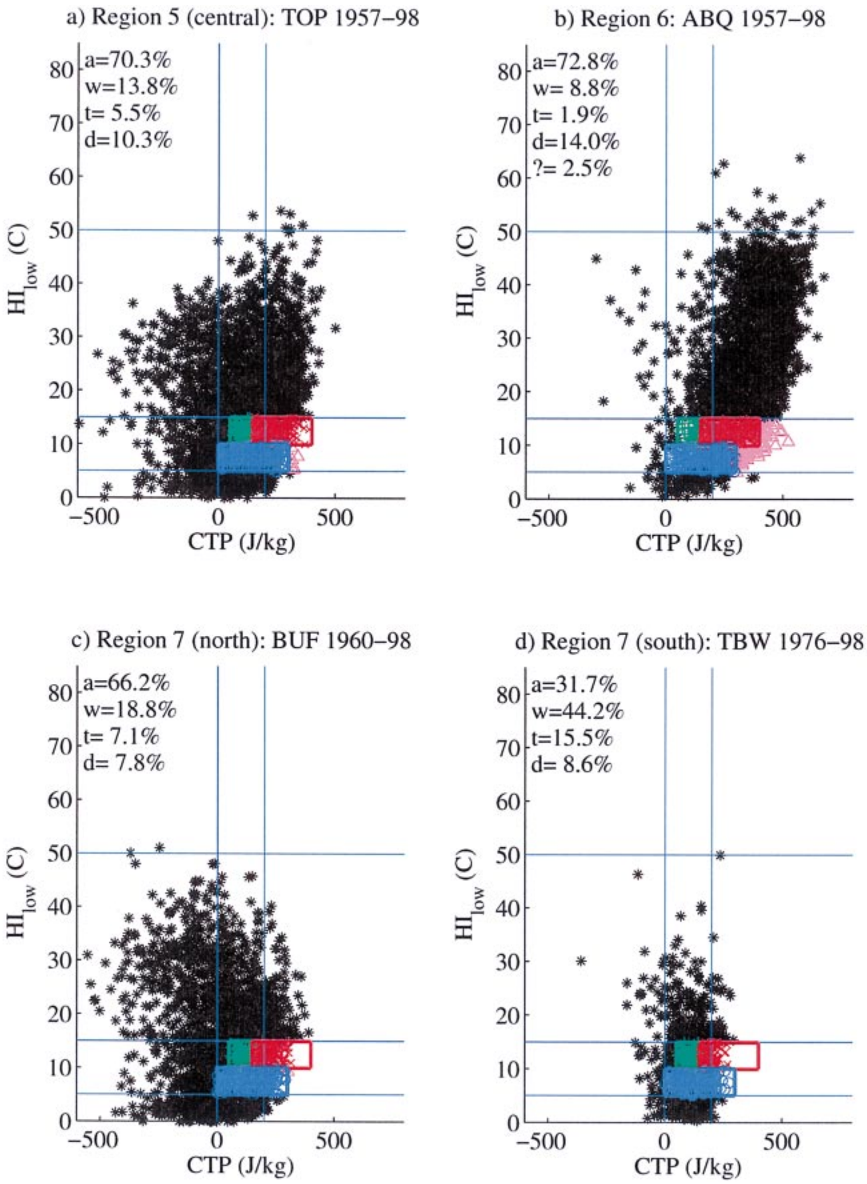

FIG. 5. As in Fig. 4 but for stations from other regions: (a) region 5 (central), one of the transitional zones (TOP: Topeka, KS); (b) region 6, the dryline and monsoon region of the arid Southwest (ABQ: Albuquerque, NM); (c) region 7 (north): Minnesota to Maine (BUF: Buffalo, NY), (d) region 7 (south): the Gulf Coast Region (TBW: Tampa Bay, FL). A question mark in the legend of the ABQ plot indicates that >1% of days fall outside the regions of CTP-HI_{low} space defined by the earlier analyses (pink triangles).

Phoenix were sparse in the NVDS database, but during the summer of 1998 there was an increased occurence of days with $HI_{low} < 15°C$ at Tucson relative to Phoenix: about 20% of days at Tucson versus only about 10% at Phoenix. This is consistent with the observations of Wallace et al. (1999).

The eastern half of the negative feedback 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 dewpoint 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 that allow a dryline to develop, and that also allow for the high CTP, moderate HI_{low} conditions necessary for a negative soil moisture–rainfall feedback.

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 U.S." (Benjamin 1986, 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 1966–68 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 Second European Stratospheric Arctic and Mid-latitude Experiment (SESAME) field experiments of 1979. These and other case studies (e.g., Crawford and Bluestein 1997; Hane et al. 1997, 1993; Benjamin and Carlson 1986; Ziegler and Rasmussen 1998; Anthes et al. 1982; Ogura et al. 1982) helped to establish a complete picture of dryline formation in the American southwest, where drylines tend to run close to north–south. 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, near the bottom 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 θ_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.

Prior to much of 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 airstreams. 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). Further research will investigate the hypothesis that these areas are also negative soil moisture–rainfall feedback regions.

Looking at the CTP-HI_{low} distribution from Albuquerque, New Mexico (Fig. 5b) and comparing it to the distribution from Desert Rock, Nevada (Fig. 4d) in the context of this other work strongly suggests that the topographic and synoptic characteristics of this dryline and monsoon region create the conditions for a potential negative feedback between soil moisture and rainfall. In both the western and eastern portions of region 6, an external source (monsoon or underrunning the dryline) brings low-level humidity into a region of typically high CTPs and high humidity deficits. The additional humidity lowers the HI_{low}s of early-morning soundings without significantly reducing the CTPs. This allows for a significant number of days to fall into the dry soil advantage regime of CTP-HI_{low} space, indicating a potential negative feedback.

6. Positive feedback regions

Most of the eastern portion of the continental United States shows the potential for a positive feedback between soil moisture and rainfall. The signal is weakest in the extreme north, where up to 75% of days are atmospherically controlled (e.g., Buffalo, NY, Fig. 5c), but gradually strengthens southward (e.g., Tampa Bay, FL, Fig. 5d). Stations in the Gulf Coast region (the southern portion of region 7-south) have about 40% of days in the wet soil advantage regime, and only about 10% in the dry soil advantage regime. 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, rendering land surface conditions less important for convection than in inland regions. In a study of summertime convective initiation in the coastal area of Mobile, Alabama, Medlin and Croft (1998) found that most soundings are humid with a deep section showing a moist adiabatic lapse rate, consistent with the CTP-HI_{low} observations for stations in this region. Medlin and Croft also find, as stated above, that most convection in this area is triggered by sea breezes.

North of these Gulf Coast stations, however, conditions are largely continental, and the land surface condition can indeed play a significant role in the development of convection. The southern half of region 7 (Gulf Coast included) shows a strong positive feedback signal, with 35.7% of days in the wet soil advantage regime, 10.7% of days in the dry soil advantage regime, and 10.4% in the transition regime (Table 1, Fig. 5d). Additionally, 88% of summers at stations in this region qualify as positive feedback years (Fig. 3). Air arriving at all stations in this region is strongly influenced by the Gulf of Mexico to the south, but early-morning soundings further inland show greater variability in both CTP and HI_{low} than stations on the coast: the humidity deficits tend to be slightly larger and the CTP is more likely to be negative, reflecting increased stability as the air moves over the land.

The northern half of region 7, which extends from Minnesota to Maine and south to the Tennessee border shows even greater variability due to the strong source region influences from both the south and the west. The positive feedback signal is weaker here than it is to the south, with 17.8% of days in the wet soil advantage regime, 8.8% of days in the dry soil advantage regime, and 7.5% in the transition regime (Table 1, Fig. 5c). Despite the lower percentage of days falling into the wet soil advantage regime of CTP-HI_{low} space, 67% of summers still qualify as positive feedback years.

7. Transitional regions

Stations classified as transition regions have significant percentages of days in each of the three nonatmospherically controlled regimes: the dry soil and wet soil advantage regimes and the transition regime. The CTP-HI_{low} distributions of these stations tend to show characteristics of stations in the surrounding regions. For example, the station at Topeka, Kansas (TOP; Fig. 5a) has a distribution that shows similarity to stations to both the east and the west. The mean CTP and HI_{low} values from TOP are between those of the Buffalo, New York, distribution well to the east (Fig. 5c) and the central Nevada distribution well to the west (Fig. 4c). (More similarity with the stations in region 3 is seen with stations closer to the eastern border of region 3.)

As in the other regions, stations in the transitional regions are grouped according to their CTP-HI_{low} distributions and their percentages of positive and negative feedback years (see Table 1). The northern transitional region has positive feedback summers about 15% of the time and negative feedback summers only about 1% of the time. The central transitional region has more frequent occurences of both conditions: 28% positive feedback summers and 6% negative feedback summers. The southern transitional region is poised between the negative feedback region of the arid southwest and the strong positive feedback region of the southeast. About 11% of the summers qualify as positive feedback summers in this region, and almost twice as many (19%) qualify as negative feedback summers. As might be predicted from their location between very different regions, interannual variability is significant in these transitional regions.

8. Discussion and conclusions

Though a full analysis of the causes leading to year-to-year variability in the observed feedback potentials at each station is beyond the scope of this paper, it is clear that interannual variability is 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 United States, which could, in turn, shift the CTP-HI_{low} distributions at many stations, particularly those in the transitional regions. This is a topic of ongoing research.

Another factor that needs to be discussed is the use of 1200 UTC soundings at all stations throughout the country. The CTP was developed in Part I for use prior to early-morning degradation of the noctural stable layer. In Illinois, 0600 LT soundings were usually well suited to this purpose. Since erosion of the noctural stable layer typically takes at least a few hours after sunrise, the CTP should be the same at 0700 LT (1200 UTC on the East Coast) as it would be at 0600 LT. The earlier soundings on the West Coast and in the mountain time zone should also have the same CTP, since continued growth of the stable layer after 0400 or 0500 LT should remain below the 100-mb above ground surface (AGS) 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.

The CTP-HI_{low} framework is a new and innovative way of determining the influence of local atmospheric conditions on the potential for feedbacks between the land surface and the atmosphere. This reference frame, developed with the one-dimensional boundary layer model described in Part I, 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 ($HI_{low} > 15°C$) or very stable ($CTP < 0$ J kg^{-1}), moist convection cannot occur, independent of flux partitioning at the surface. When the atmosphere is humid ($HI_{low} < 5°C$) and unstable ($CTP > 0$ J kg^{-1}), then rainfall should occur over both wet and dry soils, with deeper rainfall depths expected over wet soils. When the atmosphere is unstable ($CTP > 0$ J kg^{-1}) and the humidity is intermediate (HI_{low} between 5° and 15°C), then the land surface can significantly influence the likelihood of moist convection.

Though this CTP-HI_{low} framework was developed with a 1D model, it has been tested with a mesoscale model over Illinois and with data from the FIFE experiment in Kansas (Findell and Eltahir 2003a). That work shows that the behavior of the winds is a crucial third component to the framework. Other factors such as the topography throughout much of the western United States also influence convection and must be studied in the context of this framework. The nationwide analyses presented in this paper help to highlight where additional data studies might yield interesting results.

Using this framework, it was determined that much of the eastern half of the country should show a positive feedback between soil moisture and rainfall, as indicated by the positive feedback region outlined in Fig. 2. Furthermore, the arid Southwest is the only region likely to see a negative feedback. The rest of the western half of the country is unlikely to see strong feedbacks between the land surface moisture state and subsequent rainfall. A relatively narrow band of stations lie in between the eastern and western portions of the country, in some years behaving like the stations to the west and in other years behaving like the stations to the east. These results can best be understood in the context of the dominant wind patterns that affect the United States. Prevailing westerlies dictate that atmospheric conditions over the western part of the country are largely determined by the oceanic source regions of air advected in off the Pacific. As this air is pushed eastward, continental influences become increasingly important and positive feedbacks between soil moisture and moist convection become increasingly likely. Over the eastern half of the country, this air from the west mixes with moist air coming off the Gulf of Mexico and the strength of the positive feedback increases southward toward this source of humidity. The arid Southwest is an anomolous region within this larger picture: the potential for a negative feedback is created by the topography of this region and surrounding regions, particularly the Mexican plateau to the south.

Future research will include three-dimensional modeling of the Southwest to determine the areal extent, interannual persistence, and the role of winds in this potential negative feedback region. Research into the CTP-HI_{low} characteristics—and thus the nature of climate-scale feedbacks between soil moisture and rainfall—of other regions of the world is currently under way.

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APPENDIX

Definitions of CTP and HI_{low}

a. The convective triggering potential

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 of the area of integration is bounded by a constant pressure line 300 mb above the surface.

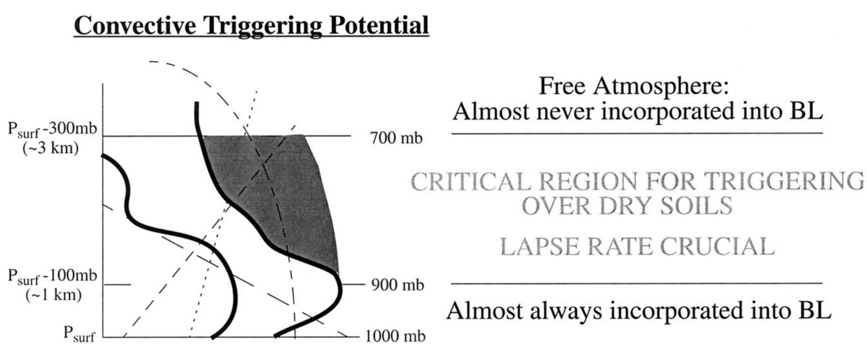

FIG. A1. A sketch of the definition of the convective triggering potential on a thermodynamic diagram. Thick solid lines are the temperature and dewpoint temperature profiles; straight long-dashed line is a dry adiabat (constant potential temperature); straight short-dashed line is constant temperature; 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. (Figure reprinted from Findell and Eltahir 2003b.)

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. A sketch of this definition is given in Fig. 1 of Part I and is included here (Fig. A1) for completeness.

b. The humidity index

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

$$\mathrm{HI} = (T_{850} - T_{d,850}) + (T_{700} - T_{d,700}) + (T_{500} - T_{d,500}), \tag{A1}$$

where T_p is the temperature at pressure level p and $T_{d,p}$ is the dewpoint temperature at pressure level p. A more useful parameter for assessing the soundings from Illinois used in Part I is the sum of the dewpoint depressions at 950 and 850 mb:

$$\mathrm{HI}_{\mathrm{low},1} = (T_{950} - T_{d,950}) + (T_{850} - T_{d,850}). \tag{A2}$$

This version of the humidity index was then generalized as the sum of the dewpoint depressions 50 and 150 mb above the ground surface. The generalized version is appropriate for use in all regions, including those with surface pressures significantly different from 1000 mb:

$$\mathrm{HI}_{\mathrm{low}} = (T_{P_{\mathrm{surf}}-50\,\mathrm{mb}} - T_{d,P_{\mathrm{surf}}-50\,\mathrm{mb}}) + (T_{P_{\mathrm{surf}}-150\,\mathrm{mb}} - T_{d,P_{\mathrm{surf}}-150\,\mathrm{mb}}). \tag{A3}$$

Lytinska et al. (1976) suggested as a threshold for rain $\mathrm{HI} \leq 30°\mathrm{C}$. The threshold for $\mathrm{HI}_{\mathrm{low}}$ is 15°C (see text).

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