

## CAPE and Convective Events in the Southwest during the North American Monsoon

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### ABSTRACT

The relationship between atmospheric stability, measured as CAPE, and deep precipitating convection has been widely studied but is not definitive. In the maritime tropics, CAPE and precipitation are usually inversely correlated. In continental convection (i.e., midlatitude and tropical), no consistent relationship has been found. In this study of the semiarid Southwest, a moderate positive correlation exists, approaching 0.6. Correlations based on radiosonde data are found to be sensitive to the parcel level of origin. The strongest correlations are found by modifying the preconvective morning sounding with the maximum reported surface temperature, assuming well-mixed adiabatic layers to the level of free convection with pseudoadiabatic ascent. These results show that the upper bounds on parcel instability correlate best with precipitation. Furthermore, the CAPE–precipitation relationship is argued to depend on the convective regime being considered. The North American monsoon convective regime requires essentially only moisture advection interacting with the strong surface sensible heating over complex topography. Elimination of strong convective inhibition through intense surface sensible heating in the presence of sufficient water vapor leads to the positive CAPE–precipitation relationship on diurnal time scales. These results are discussed in light of contradictory results from other continental and maritime regions, which demonstrate negative correlations.

### 1. Introduction

Understanding the relationship between the energy available for atmospheric convection and convective activity has engendered a great deal of research in the atmospheric sciences. One common measure of convective energy used in many studies is the convective available potential energy (CAPE), the vertically integrated parcel buoyant energy (Moncrieff and Miller 1976). CAPE is most frequently used as a forecasting tool for gauging severe thunderstorm likelihood since it provides a rough estimate of vertical updraft magnitude (Doswell and Rasmussen 1994). CAPE has also been used as an indicator variable of global climate change (Ye et al. 1998; Gettelman et al. 2002), as a predictor of electrification/lightning intensity in deep tropical convection (Williams et al. 1992, 2002), as an estimate of kinetic energy produced in steady-state tropical convec-

tion (Rennó and Ingersoll 1996; Adams and Rennó 2003), and to estimate precipitation (McBride and Frank 1999; Yano et al. 2005).

The calculation of CAPE is somewhat open to debate as it entails various assumptions about the parcel level of origin, parcel moisture and temperature characteristics, the thermodynamic path taken during ascent, water loading effects on density, and the presence or absence of the ice phase (Williams and Rennó 1993; Doswell and Rasmussen 1994; Emanuel 1994; Craven et al. 2002). The results obtained in any given study may be sensitive to these assumptions. For example, the magnitude of CAPE for tropical deep convection in “quasi-equilibrium” with large-scale forcing is sensitive to the parcel level of origin and to the inclusion of the ice phase (Xu and Emanuel 1989; Williams and Rennó 1993; Emanuel 1994). Nevertheless, in the broadest sense, the underlying assumption is that CAPE is related to measures of convective activity (e.g., diabatic heating, cloud-top temperatures, or simply precipitation).

How strongly CAPE actually relates to convective activity is still debated and is far from resolved. Corre-

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lations between parcel energetics and convective activity have given contradictory results with correlations not only varying in magnitude, but also in sign (Zawadzki et al. 1981; Pepler and Lamb 1989; McBride and Frank 1999; Tompkins 2001; Yano et al. 2005). Many observational and modeling studies focusing on deep convection in the tropics have found negative correlations between CAPE and convection. In a detailed study of stability and convection during the Australian monsoon, McBride and Frank (1999) found negative relationships between CAPE and areal measures of precipitation with large sensitivity to parcel level of origin. Near-surface parcels were shown to correlate negatively with convective activity, while parcels only 100 hPa above surface level were uncorrelated with convection. Over the tropical oceans, Sobel et al. (2004) and Yano et al. (2005) found a weak negative correlation between convection and rainfall rate. Yano et al. (2005), however, did find very weak positive relationships over land. The inverse relationship between CAPE and convection noted in these studies lies in that deep precipitating convection warms the upper troposphere and, more importantly, stabilizes the boundary layer through the influx of low equivalent potential temperature air,  $\theta_e$ , due to downdrafts originating in the midlevel troposphere (Raymond 1995). The cooler and drier boundary layer reduces CAPE values during convective precipitation outbreaks, leading to a negative relationship. Sherwood (1999), examining predictor variables for disturbed conditions in the tropical western Pacific, found that sizeable CAPE is always present in the large-scale mean. This omnipresence of CAPE implies that it is not a limiting factor in convective activity and therefore correlates poorly with convective outbreaks. However, Tompkins (2001) considering results from a cloud-resolving model study, argues that the CAPE–convection relationship becomes positive even in the deep maritime tropics. He notes that on larger scales (hundreds of kilometers) CAPE–convection correlations are positive and ultimately depend upon the spatial and temporal scales considered. McBride and Frank (1999) also point out that soundings taken in convective wakes will not be representative of the large-scale undisturbed atmosphere where sizeable CAPE may exist even during widespread convective outbreaks. CAPE–precipitation correlations may therefore be dependent upon properly identifying undisturbed conditions if the sounding is to truly represent the “large-scale environment.”

Outside of the deep tropics, in midlatitude continental regions, the inverse CAPE–convection relationship is not apparent. In the central United States, Pepler and Lamb (1989) found that measures of parcel buoy-

ant energy had no clear relationship with precipitation amounts. Zhang (2002), for the same general area, also indicates there is essentially no relationship between CAPE and precipitation. However, Zawadzki et al. (1981), for a continental region of eastern Canada, found large positive correlations between areal rainfall rates and parcel energetics. So the relationship between CAPE and precipitation is even less clear for continental convective regimes.

In addition to its theoretical implications for the relationship between thermodynamic energy and convection, the CAPE–precipitation relationship also has bearing on convective parameterization in large-scale numerical models. In many quasi-equilibrium convective parameterizations, CAPE functions as both trigger and mass flux closure for convection (e.g., Zhang and McFarlane 1995, among many others). The closure is based on the principle that cumulus convection acts to stabilize an atmosphere (consumption of CAPE) for which large-scale forcing is acting to destabilize it (production of CAPE). Ignoring the details of model precipitation microphysics, a positive CAPE–precipitation relationship is implied by these schemes. As noted by Xie and Zhang (2000) and Sobel et al. (2004), negative CAPE–convection correlations are problematic in the context of CAPE-based closures. In the CAPE-based, quasi-equilibrium framework, the quantity and timing of upper-tropospheric heating and the resulting dynamical and cloud–radiation effects are critically related to the presence of CAPE. Since the presence of CAPE acts as a trigger for convection, numerous studies have shown that for continental regions, these parameterizations begin convection too early, essentially mimicking the diurnal cycle in surface forcing (Xie and Zhang 2000; Chaboureau et al. 2004; Grabowski et al. 2006). This overactive diurnal cycle of convection leads to unrealistic tropospheric temperature profiles (Xie and Zhang 2000). In general, for diurnally forced “triggered” convection, quasi-equilibrium between large-scale tropospheric forcing and the convective response would not seem to apply at small time scales (i.e., order of several hours; Donner and Phillips 2003; Zhang 2002). On the subdiurnal scale, particularly over land surfaces, CAPE values are not correlated with precipitation because frequently substantial convective inhibition energy (CIN) exists to stifle convection despite large values of CAPE. Only when CIN has been diminished considerably because of, for example, strong surface sensible heating, can CAPE be released. Hence, the correlations obtained between convection and CAPE are strongly impacted by *when* the thermodynamic profile is sampled (i.e., by the “large-scale convective environment” the sounding represents).

### *Motivations for study*

From a critical reading of the above studies, it becomes apparent that the CAPE–convection relationship is sensitive to parcel origin (McBride and Frank 1999), the nature of the convective forcing (Donner and Phillips 2003; Zhang 2002), and the space–time scales (i.e., phase relationship) for which the relationship is observed (Sherwood 1999; Tompkins 2001; Sobel et al. 2004). These studies also highlight what seems to be a regional or regime-dependent relationship between large-scale forcing, the local thermodynamic environment, and the phase relationship between instability and convective activity. In this study, we investigate the nature of the CAPE–precipitation relationship for the convective regime along the periphery of the North American monsoon (NAM). The motivations for this study lie in understanding the relationship between CAPE, the thermodynamic vertical structure, and deep convection in a regime of strong boundary layer forcing and weak large-scale forcing.

Several of the above-cited studies of the deep tropics have utilized large-scale networks of radiosondes, in situ, and satellite precipitation measurement, along with other observation during intensive field campaigns (McBride and Frank 1999; Sobel et al. 2004). At these fine temporal resolutions, the thermodynamic structure and variability of the convecting atmosphere and the lag time between convection and its “large-scale” effects on temperature and moisture can be gauged (McBride and Frank 1999; Sobel et al. 2004). In the NAM region no such small resolution, frequently sampled upper-air data exist and other data must be utilized. In this study, long-term (multiyear) precipitation and radiosonde data during the convective season are used to characterize the thermodynamic profiles during deep convective events. With these data, we investigate the relationship between convective instability and convective activity measured as areal rainfall. We argue that the positive correlations observed in this study for the NAM are due to the relative “simplicity” of this convective regime. The main aim of this study is to understand what leads to the positive CAPE–precipitation relationship in light of results from tropical maritime and midlatitude continental convective regimes. With this in mind, we specifically examine the following:

- 1) The relationship between the parcel level of origin and thermodynamic characteristics and their influence on the correlation between CAPE and convective precipitation.
- 2) Which large-scale environment represented by the soundings gives the strongest CAPE–precipitation

correlations. Specifically, this is done by considering both the morning and evening soundings and “modified” uncontaminated morning soundings, which produce convective environments of maximum instability.

The first one will help indicate where the strongest forcing occurs in this convective environment, while the second sheds light on the importance of when the sounding is taken in determining the strength of the CAPE–convection relationship; that is, which large-scale environment most strongly correlates with convection.

In the sections that follow, we first present an overview of our study area within the NAM region and the nature of its convective regime with respect to the deep tropics and midlatitude continental regimes. This is followed by a description of the sounding and precipitation records utilized. The methodology for calculating CAPE as a function of parcel origin and how the morning soundings are modified to give maximum instability are then presented. Correlation results then follow and are discussed in the context of the thermodynamic environment during convectively active “burst” and less active “break” periods of the NAM. And, finally, the implications of the CAPE–precipitation relationship ascertained in this study are discussed.

## **2. Study region, data, and methodology**

### *a. The NAM convective regime*

The NAM system results in relatively frequent widespread convective precipitation events during mid- to late summer in northwestern Mexico (Douglas et al. 1993; Adams and Comrie 1997). The Southwest, lying on the periphery of the monsoon, likewise experiences convective outbreaks resulting from invasions of tropical maritime air on weak synoptic flow from northwestern Mexico. This moist air interacts with intense surface sensible heating and complex orography resulting in deep precipitating convection. The NAM regime of the desert Southwest differs from other deep tropical maritime regions (e.g., equatorial western Pacific) where water vapor is abundant and forcing, typically weak at the surface, is associated with various atmospheric modes of oscillation (Sobel et al. 2004). In addition, unlike the NAM, in the deep tropics complex feedbacks between convection, radiation, surface energy fluxes, and waves can act to further enhance or suppress convection. The NAM also differs from the continental midlatitude regime (e.g., the central United States) in that baroclinic zones or drylines are not generally necessary for triggering deep convection. Furthermore,

convective inhibition due to a strong low-level capping inversion, commonplace during spring/summer in the central United States (Colby 1984), is typically absent. Most of the convective inhibition present in the Southwest, resulting principally from nocturnal radiational cooling of the boundary layer, is eliminated by intense surface sensible heating during the day. Elevated mixed layers that evolve over the elevated terrain may act as lids to convection, but not nearly as extensively as in the central United States (Bright and Mullen 2002). As a result, monsoon convection in the Southwest may be viewed as “simple” compared to the deep tropics or central United States. To a first order, it depends essentially on the presence of lower-tropospheric moisture with intense insolation and elevated topography acting as convective triggers. Nevertheless, NAM convection also has many inherent difficulties in modeling (Bright and Mullen 2002; Gochis et al. 2002; Collier and Zhang 2006).

The role of topography in convective initiation and organization cannot be overemphasized in the Southwest (Raymond and Wilkening 1980; Raymond and Blyth 1989; Zehnder et al. 2006; Damiani et al. 2008). Mountains acting as elevated heat sources induce low-level convergence and deepening of the boundary layer to the level of free convection, typically during the morning hours (Damiani et al. 2008). Given conducive ambient thermodynamic and local wind shear conditions, shallow cumulus towers grow and precondition their environment through mixing. Furthermore, dynamics resulting from the cumulus towers can act to induce larger mountain-scale circulations the time scale of which determines the transition to deep convection (Damiani et al. 2008). Oftentimes, elevated convection becomes organized and propagates into the surrounding valleys or it may produce outflow boundaries, which further initiates convection in the lower elevation regions (Smith and Gall 1989; Maddox et al. 1995; Wallace et al. 1999). Organized, long-lived mesoscale convective systems, including tropical squall lines, can actually contribute sizeable amounts of precipitation to seasonal totals (Smith and Gall 1989; Maddox et al. 1995).

It is also important to emphasize that synoptic-scale dynamic lifting and strong low-level moisture convergence (e.g., both important for convection in the central United States) are much less important in driving deep convection than are orography and convective outflows (R. A. Maddox 2008, personal communication). Moisture advection is typically associated with synoptic-scale wind reversal (e.g., see Fig. 3) from dry southwesterlies (break periods) to light southeasterly winds (burst periods; Adams and Comrie 1997). Synoptic-

scale transients, nevertheless, can be important for low-level moisture fluxes and destabilization of the boundary layer (Adang and Gall 1989; Fuller and Stensrud 2000). But for the most part, “large-scale forcing” of convective activity due to low-level moisture convergence is very weak, particularly compared to the above mentioned lifting mechanisms. However, the *presence* of water vapor advected by synoptic-scale winds, and not low-level moisture convergence, is critical for the thermodynamic destabilization of boundary layer air and convective activity. In the context of the large-scale forcing term in a quasi-equilibrium convective parameterization, the advection term would be quite small, but advection of moisture into the region is absolutely necessary for deep moist convection; this differentiation is subtle, but important.

#### b. Data

Unlike other studies of the deep convective environment (McBride and Frank 1999; Zhang 2002; Sobel et al. 2004), which employ data from intensive field campaigns, we use long term radiosonde data or a “climatology” of the thermodynamic environment in periods of strong and weak convective activity. Specifically, the long-term records for July and August of two radiosonde sites and numerous surrounding precipitation stations located in the Southwest are examined. During these months, periods of convective activity, *bursts*, alternate with periods of quiescence, *breaks*. Soundings from Desert Rock, Nevada (DRA; 36.6°N, 116.0°W), and Tucson, Arizona (TUS; 32.1°N, 110.9°W), provide the thermodynamic profiles from which CAPE is calculated. These two stations are comparable in many respects, both lying in broad valleys at similar elevations surrounded by a complex elevated terrain. DRA lies on the extreme fringes of the NAM experiencing relatively infrequent intrusions of tropical moisture compared to TUS; TUS being much closer to the principle monsoon region of northwestern Mexico (see Fig. 1). [The radiosonde data was obtained from the National Climatic Data Center (NCDC)/Forecast Systems Laboratory CD-ROM archive online at [http://raob.fsl.noaa.gov/General\\_Information.html](http://raob.fsl.noaa.gov/General_Information.html).]

Given that precipitation during the NAM is essentially convective, it serves as an excellent proxy for convective activity. In general, the relationship between CAPE and precipitation at any given station is weak because of the extreme spatial and temporal variability of convective rainfall (Peppler and Lamb 1989). To draw out any sort of relationship, an areal total of precipitation was calculated (Peppler and Lamb 1989; Williams et al. 1992; McBride and Frank 1999). For both DRA and TUS, daily precipitation from stations con-

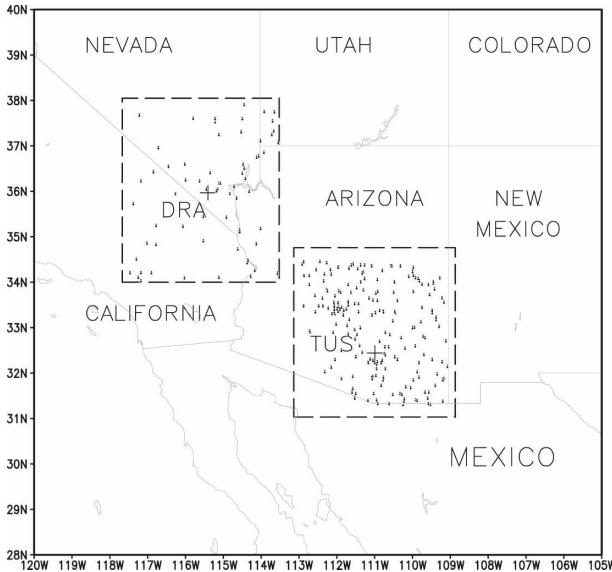


FIG. 1. Map of the study region. DRA and TUS are radiosonde sites and the small dots represent precipitation stations. The  $4^\circ \times 4^\circ$  grid box surrounding DRA contains 66 stations, while TUS contains 210. Many of the precipitation station are within several kilometers of each other and therefore are not resolved at this map scale.

tained within a  $4^\circ \times 4^\circ$  grid box, approximately centered on the radiosonde site, were summed (see Fig. 1). The area, although somewhat arbitrary in extent, could be thought of as representative of a grid box in a large-scale numerical model or as the upper-air region represented by the radiosonde in the U.S. network. Daily rainfall totals from NCDC “Cooperative Summary of the Day” were derived from 66 stations surrounding DRA and 210 stations surrounding TUS. We used 14 yr of data for DRA and 23 yr for TUS in the correlation calculations. In each case, CAPE is calculated from each radiosonde station and is correlated with the daily total precipitation derived from the surrounding stations in each  $4^\circ \times 4^\circ$  grid box.

*c. Methodology*

One focus of this study lies in determining CAPE–precipitation correlations as a function of the large-scale environment of which the radiosonde is assumed to be representative. Morning and afternoon soundings reveal decidedly different instability characteristics, which should be expected to influence the CAPE–precipitation relationship. Moreover, convective activity can contaminate soundings rendering them unrepresentative of the large-scale surroundings. With this in mind, we attempted to isolate uncontaminated soundings taken before the onset of convective activity.

These “preconvective” soundings were then modified, described below, to provide the most convectively unstable environment for a given day. The correlations derived from these modified, uncontaminated soundings are compared with those derived from the original morning and the afternoon soundings.

1) PARCEL ORIGIN AND CAPE CALCULATION

Correlations are calculated in terms of parcel level of origin and the layer depth over which parcel mixing ratio is averaged. There is no agreed-upon standard that determines which parcel to lift or how to characterize its temperature and moisture properties (Williams and Rennó 1993; Doswell and Rasmussen 1994; Craven et al. 2002). Nevertheless, lifting near-surface parcels and mixing thermodynamic properties over near-surface layers is the norm. Doswell and Rasmussen (1994) and Craven et al. (2002) have argued that lifting the most unstable parcel in the lower atmosphere has the strongest relationship with convective activity. In this study, two methods are employed in calculating lifted parcel characteristics in order to gauge the sensitivity of the results to parcel origin. First, the lifted parcels are chosen at regular pressure levels. In this case, moisture and temperature values are averaged over a 25-hPa layer, centered on that pressure level. The only exception is the “surface” layer, which is averaged over the lowest 10 hPa above the surface. The second methodology is to lift only the surface parcel, but to vary the depth of the layer over which its mixing ratio is averaged. The depth of this averaged layer is in 50-hPa increments from the surface up to 300 hPa. The only exception is what we call the “surface layer” for which the mixing ratio is averaged only over the lowest 10 hPa.

It is assumed that all parcels ascend pseudoadiabatically. For calculation in this study, CAPE is defined as follows:

$$CAPE = \int_{p_{LNB}}^{p_{LFC}} R_d(T_{vp} - T_v) d \ln p. \quad (1)$$

Here  $R_d$  is the dry air gas constant,  $p$  is pressure, and  $T_{vp}$  and  $T_v$  represent the parcel and sounding virtual temperatures, respectively. The difference between parcel and environmental virtual temperatures is integrated upward from the level of free convection,  $p_{LFC}$ , to the level of neutral buoyancy,  $p_{LNB}$ .

2) MODIFICATION OF THE PRECONVECTIVE SOUNDING

To properly identify the relationship between the convective environment and precipitation, the sound-

TABLE 1. Correlation of CAPE vs rainfall for modified 1200 UTC sounding for DRA and TUS. Depth represents the layer in hectopascals of pressure (measured from the surface) over which the mixing ratio is averaged. Boldface values are significant at  $\alpha = 0.01$ .

Depth	Correlation coef	
	DRA	TUS
Surface	<b>0.59</b>	<b>0.37</b>
25 hPa	<b>0.56</b>	<b>0.37</b>
50 hPa	<b>0.50</b>	<b>0.39</b>
100 hPa	<b>0.47</b>	<b>0.40</b>
150 hPa	<b>0.50</b>	<b>0.38</b>
200 hPa	<b>0.42</b>	<b>0.37</b>
250 hPa	<b>0.38</b>	<b>0.37</b>
300 hPa	<b>0.33</b>	<b>0.36</b>

ing must be representative of the large-scale; in this study, the  $4 \times 4$  grid surrounding each sounding site. The 1200 UTC morning sounding, launched at approximately 0500 LST at both TUS and DRA, was considered the most representative large-scale preconvective environment. The 0000 UTC (1700 LST) radiosonde is often launched during or after convective activity and hence, contaminated, at both DRA and TUS (Zehnder et al. 2006) and were, therefore, not included in the preconvective environment soundings. On occasion, however, nocturnal thunderstorms or the previous day's convection can contaminate the morning sounding through the presence of debris clouds, remnant outflow boundaries, and mesoscale downdrafts rendering the sounding entirely unrepresentative of the large-scale atmosphere (Wallace et al. 1999). The 1200 UTC soundings that were obviously contaminated were eliminated; 11% of the TUS and 2% of the DRA soundings. The greater number of contaminated soundings at Tucson results from more frequent convective activity and its proximity to the core monsoon region in northwestern Mexico.

Clearly, the radiationally cooled nocturnal boundary layer present at 1200 UTC is not representative of the unstable late afternoon convective environment. On most days, a deep, well-mixed adiabatic layer develops and convective temperatures at the surface are reached (i.e., a surface parcel rises unimpeded to its convective condensation level). This environment represents the most unstable one for any given monsoon day. To replicate this environment, uncontaminated 1200 UTC soundings were modified by replacing the soundings surface temperature with the maximum temperature reported at the sounding site for that day. A well-mixed, adiabatic layer (constant potential temperature,  $\theta$ , and constant mixing ratio,  $r$ ) is then assumed to exist from the surface up to the lifting condensation level

TABLE 2. Correlation of CAPE vs rainfall for modified 1200 UTC sounding for DRA and TUS as a function of the pressure level of parcel origin. Boldface values are significant at  $\alpha = 0.01$ .

Level of parcel origin	DRA	TUS
Surface	<b>0.59</b>	<b>0.37</b>
900 hPa	—	<b>0.41</b>
850 hPa	<b>0.42</b>	<b>0.34</b>
800 hPa	<b>0.41</b>	<b>0.32</b>
750 hPa	<b>0.31</b>	<b>0.27</b>
700 hPa	<b>0.18</b>	<b>0.08</b>
650 hPa	<b>0.11</b>	0.01
600 hPa	0.01	<b>-0.10</b>

(LCL). Assuming constant  $r$  within the adiabatic layer may be an overestimate of the actual mixing ratio as oftentimes drier air from the free atmosphere mixes down during the day with the growth of the convective boundary layer (Bright and Mullen 2002). In making the above assumptions about the convective environment, the parcels will represent the upper bounds on CAPE. These correlations from the modified soundings are compared with unmodified 1200 and 0000 UTC soundings.

### 3. Results and discussion

#### a. CAPE–precipitation correlations

Table 1 contains the correlations between precipitation and CAPE for modified morning soundings. The “depth” is the extent of the layer in hectopascals measured from the surface, over which the mixing ratio was averaged. The surface pressure is approximately 895 hPa at DRA and 920 hPa at TUS. The overall results show small-to-moderate positive correlations for both DRA and TUS, with DRA showing somewhat larger correlations most notably at the near-surface level. These correlations drop off when the depth of the

TABLE 3. Correlation of CAPE vs rainfall for unmodified 0000 and 1200 UTC sounding for DRA and TUS as a function of the pressure level of parcel origin. Boldface values are significant at  $\alpha = 0.01$ .

Level of parcel origin	DRA		TUS	
	1200 UTC	0000 UTC	1200 UTC	0000 UTC
Surface	<b>0.44</b>	<b>0.23</b>	<b>0.30</b>	<b>0.24</b>
900 hPa	—	—	<b>0.29</b>	<b>0.13</b>
850 hPa	<b>0.28</b>	<b>0.10</b>	<b>0.28</b>	<b>0.24</b>
800 hPa	<b>0.28</b>	<b>0.14</b>	<b>0.24</b>	<b>0.23</b>
750 hPa	<b>0.30</b>	<b>0.20</b>	<b>0.20</b>	<b>0.15</b>
700 hPa	<b>0.12</b>	<b>0.13</b>	0.06	0.05
650 hPa	0.03	<b>0.10</b>	0.01	-0.01
600 hPa	-0.03	0.00	0.01	0.01

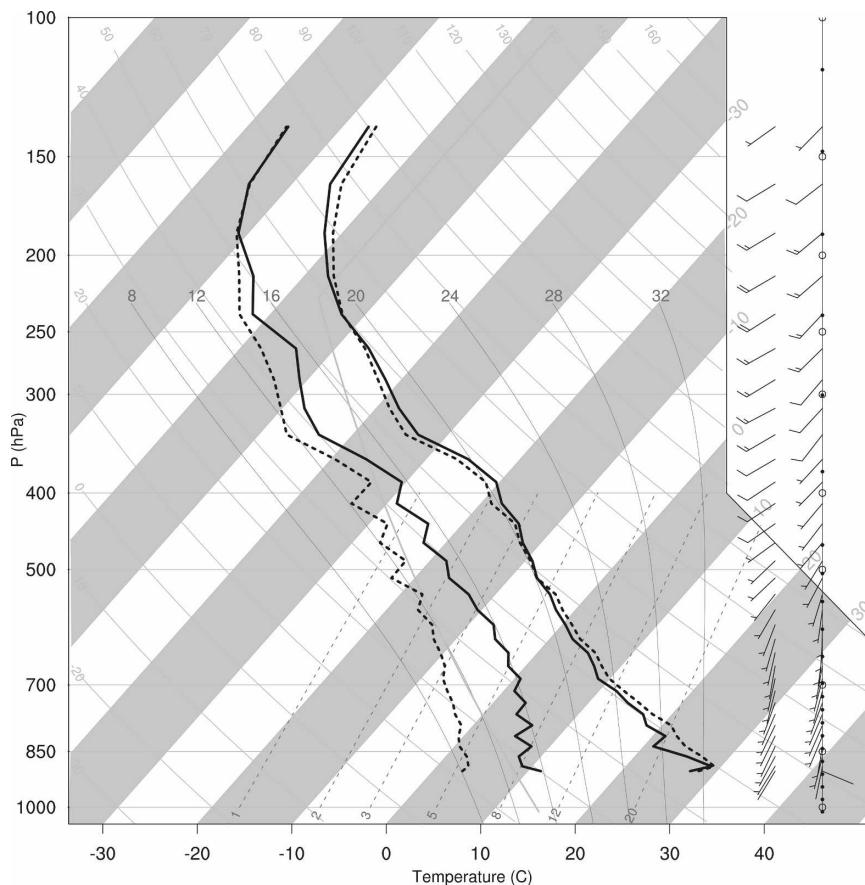


FIG. 2. Burst and break averaged soundings, 0000 and 1200 UTC combined, over the period of record for DRA. The dashed line represents the break sounding and the solid line represents the burst sounding. Wind barbs at left are for the break sounding and wind barbs at the right are for the burst sounding.

modified mixed layer is greater than 200 hPa. TUS, on the other hand, has small positive correlations regardless of the depth through which the mixing ratio is averaged; the maximum being found at a depth of 100 hPa. Although the depth of the averaging layer is somewhat arbitrary, it appears that averaging the lowest 100 hPa results in the strongest relationship. Layer averages over the lowest 100 hPa has also been recommended by Craven et al. (2002) in order to diminish the effects of unrealistically high surface relative humidities.

As mentioned previously, many studies have examined the sensitivity of CAPE to parcel level of origin. McBride and Frank (1999) found that over the tropical ocean the CAPE–precipitation correlation (negative in this case) as a function of parcel origin drops off rapidly within 100 hPa above the ocean surface. This is indicative of the moist neutrality (i.e., CAPE ~0) of the tropical maritime atmosphere to parcels lifted from above 900 hPa. Table 2 contains correlations for DRA and TUS as a function of parcel origin for modified morning

soundings. Small-to-moderate positive correlations are found at both DRA and TUS below the 650- and 700-hPa levels, respectively. Above these levels precipitation and CAPE are unrelated. The unmodified 1200 UTC morning sounding (Table 3) displays similar behavior albeit with smaller correlations for both sites. The 0000 UTC unmodified soundings (Table 3) have the smallest correlations with slightly more irregularity as a function of height. Overall, Tables 1–3 show that unlike the maritime tropics, parcels throughout a deep layer (approximately 200 hPa) exhibit a positive CAPE–precipitation relationship. This is most likely reflective of the great depth of the well-mixed layer resulting from strong surface forcing on typical monsoon days. Mixed layer depths to greater than 700 hPa are frequent (Zehnder et al. 2006). This being the case, parcels through a much deeper boundary layer (compared with other convective regimes) are candidates for deep convection. However, these results also confirm that the most unstable parcels (i.e., near-surface parcels) are

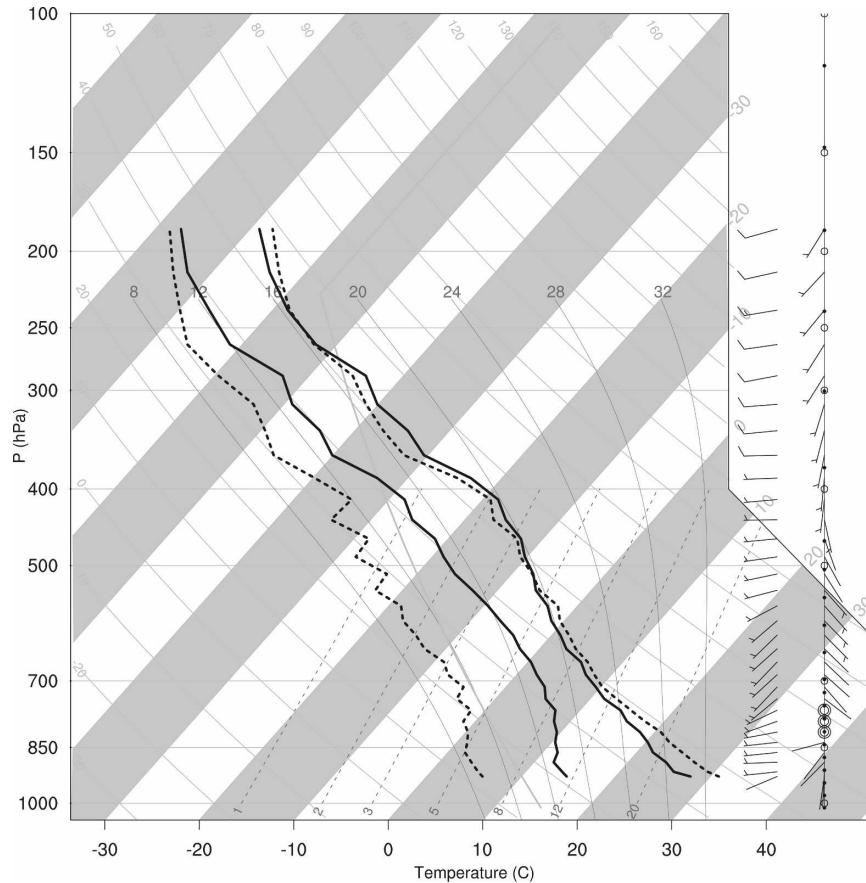


FIG. 3. As in Fig. 2, but for TUS.

those that best correlate with convective activity (Doswell and Rasmussen 1994; Williams and Rennó 1993; Rennó and Williams 1995; Eltahir and Pal 1996). Furthermore, the 1200 UTC modified sounding, which represents maximum possible instability for the day, gives the strongest correlations.

The question remains as to why positive correlations exist between CAPE and precipitation relative to other monsoon regions, the deep tropics, or central United States. In the Southwest, it is nearly entirely surface sensible heating in the presence of sufficiently deep moisture that is responsible for convective outbreaks. Intense solar radiation on most days leads to large surface sensible heat fluxes. Convective temperatures are reached eliminating any convective inhibition that may have been present for the surface parcel. As long as a sufficiently deep layer of low-level moisture exists to overcome entrainment of drier midtropospheric air, deep convection is likely (Zehnder et al. 2006). These conditions may be substantially different in the deep tropics or monsoonal regions where other factors such as the passage of easterly waves, squall lines, monsoon

troughs, and forcing due to large-scale modes of oscillation exist (Wheeler and Kiladis 1999). In tropical continental regions, large-scale propagating convective systems, entirely out of phase with local surface sensible heating (and maximum CAPE values), can modulate the diurnal cycle of convective precipitation (Rickenbach 2004) resulting in low CAPE values during convective events. Furthermore, unlike the central United States, strong capping inversions over the Southwest are much less frequent. Convection in the Southwest is much less dependent on frontal boundaries or other lifting mechanisms. These factors, which may tend to confound the CAPE–convection relationship in the deep tropics or the central United States, are essentially absent in the Southwest during the monsoon.

#### b. Thermodynamic environment

To gain insight into the CAPE–precipitation relationship, the monsoon’s thermodynamic environment for active convective burst and quiescent break periods was examined. This will help us understand what the

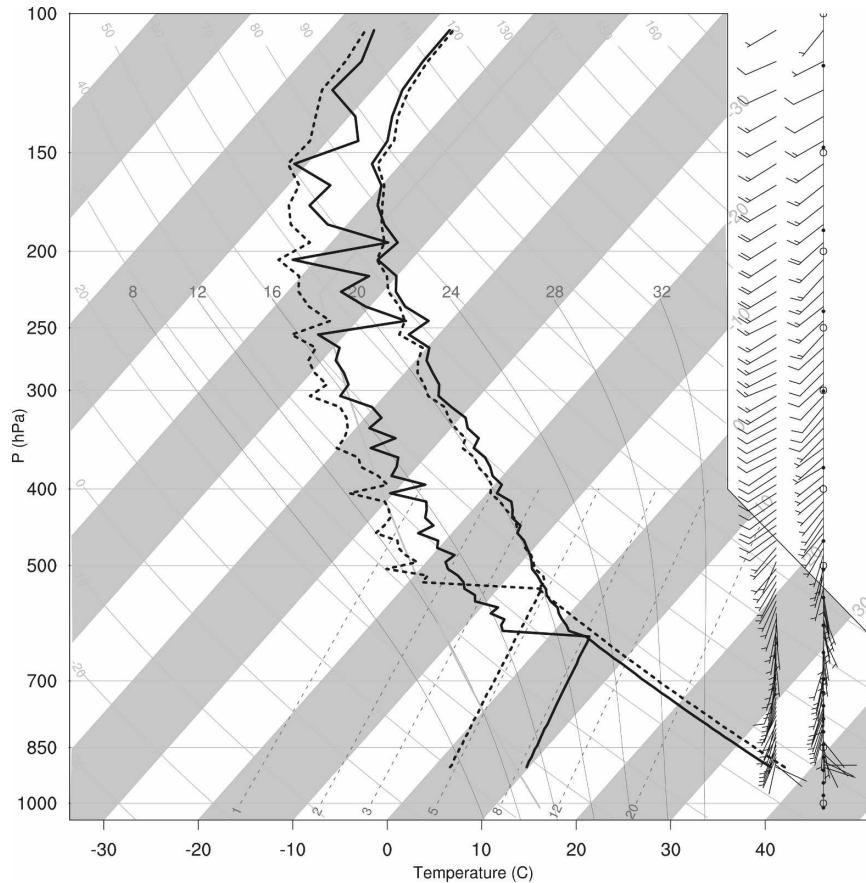


FIG. 4. As in Fig. 2, but for the 1200 UTC sounding over the period of record for DRA.

average conditions are, thermodynamically speaking, that lead to larger CAPE for convective events and less CAPE for drier periods. Burst events were classified as those days reporting greater than  $10 \text{ mm day}^{-1}$  of rainfall rate across either the TUS or DRA areas, breaks are less than  $10 \text{ mm day}^{-1}$ . The division is arbitrary, representing about 20% of nonzero precipitation events at both sites, but it separates clearly active periods from dry periods when convection was suppressed or moisture was absent. In terms of the upper-tropospheric conditions, little change in the temperature profile between convectively active and quiescent periods is observed. To show this, all unmodified soundings (0000 and 1200 UTC) were averaged for convectively active burst periods and separately for break periods (see Figs. 2 and 3). For both DRA and TUS, the soundings show evidence of large-scale moist convective overturning (slight cooling below 600 hPa and slight warming from 600 to 250 hPa) relative to break soundings. The average upper-tropospheric warming between convectively active and inactive periods is slightly greater than 1 K for both sites (virtual tempera-

ture effects were small and, therefore, neglected). However, it should be noted that for both DRA and TUS the average standard deviation in the upper-atmospheric temperature profile during both burst and break conditions is approximately 2 K reflecting a large degree of variability as might be expected for the subtropical latitudes of this study. McBride and Frank (1999) found very similar results for their study of the Australian monsoon as did Frank (1980) for the Global Atmospheric Research Program (GARP) Atlantic Tropical Experiment (GATE) studies, although the warming was less than 1 K. More recently, it has also been shown that over midlatitudinal continental regions, free troposphere temperature changes are small during convective activity (Donner and Phillips 2003; Zhang 2002), which is consistent with what has been reported for other convectively active areas (Ye et al. 1998). Even though these tropospheric temperature changes are small between burst and break conditions, they would tend to decrease CAPE, all else being equal. This would imply decreasing CAPE with increasing precipitation (i.e., an inverse relationship), not the

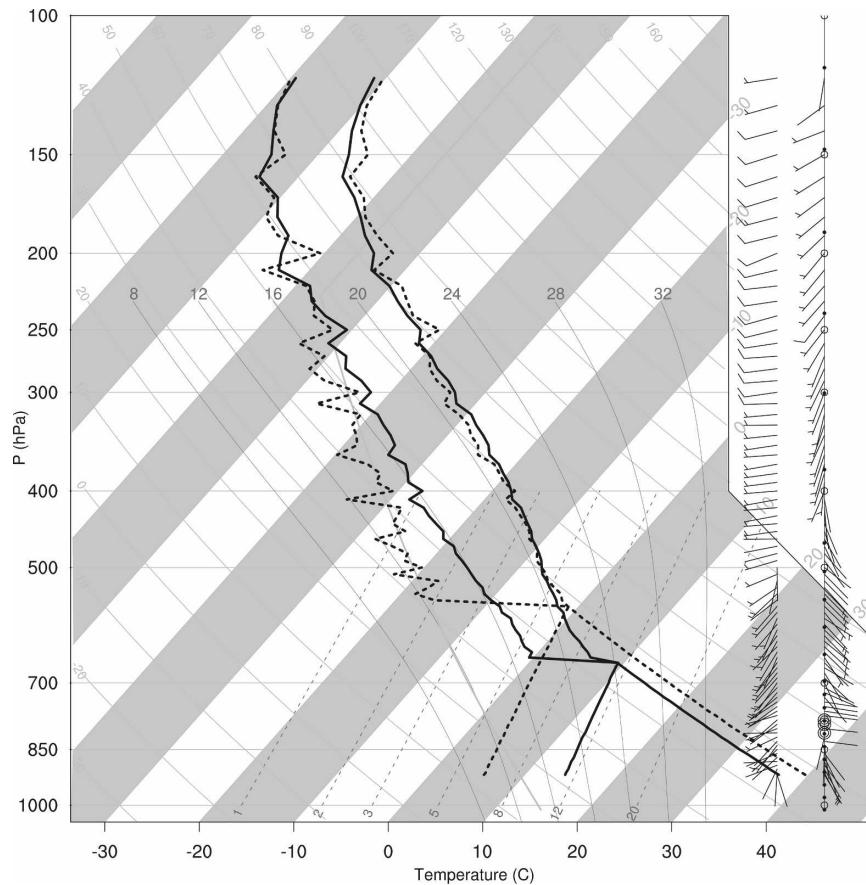


FIG. 5. As in Fig. 2, but for the 1200 UTC sounding over the period of record for TUS.

positive one observed. Relative to changes in the tropospheric temperature profile, the change in the mixing ratio profile between bursts and breaks is large. This behavior can be seen in the modified 1200 UTC soundings for burst and breaks (Figs. 4 and 5). Average CAPE for convectively active conditions are nearly 4 times greater than those for break conditions (the lifted parcel averaged over the lowest 75 hPa) (450 versus  $1636 \text{ J kg}^{-1}$  for TUS and 307 versus  $1173 \text{ J kg}^{-1}$  for DRA). The average upper-tropospheric temperature contribution of approximately 1 K would tend to decrease the burst CAPE about  $300 \text{ J kg}^{-1}$  relative to break CAPE. However, the low-level increase in mixing ratio more than offsets this decrease in CAPE due to warmer tropospheric conditions. The low-level moisture lowers the LCL, essentially shifting the parcel moist adiabat to the right. A  $4 \text{ g kg}^{-1}$  increase in mixing ratio at the surface in the burst soundings raises the parcel CAPE by over  $1000 \text{ J kg}^{-1}$ . The CAPE dependence on low-level moisture is further demonstrated in Figs. 6 and 7, which contain scatterplots of the average mixing ratio versus CAPE, where the mixing ratio has

been averaged over the lowest 75 hPa. The correlation between CAPE and mixing ratio is 0.87 for DRA and 0.83 for TUS. The surface temperature is not statistically significant in accounting for variations in CAPE values (figure not shown), which is consistent with the argument that surface heat fluxes are not a limiting factor in the Southwest. These results are consistent with those of Williams (1991), Williams and Rennó (1993), and Eltahir and Pal (1996), which have demonstrated the importance of near-surface moisture in accounting for CAPE variations.

Considering the above results and given the limitations in the dataset, the NAM regime of the Southwest, variations in large-scale forcing of convection may be attributed principally to the advection of moisture over the region. Convection does act to stabilize the temperature profile as seen in Figs. 2–5, but its overall contribution to decreasing CAPE is small compared with its increase due to water vapor. Large values of tropospheric moisture give larger values of CAPE and also precipitable water. Since surface heat fluxes do not tend to be a limiting factor, higher lower-tropospheric

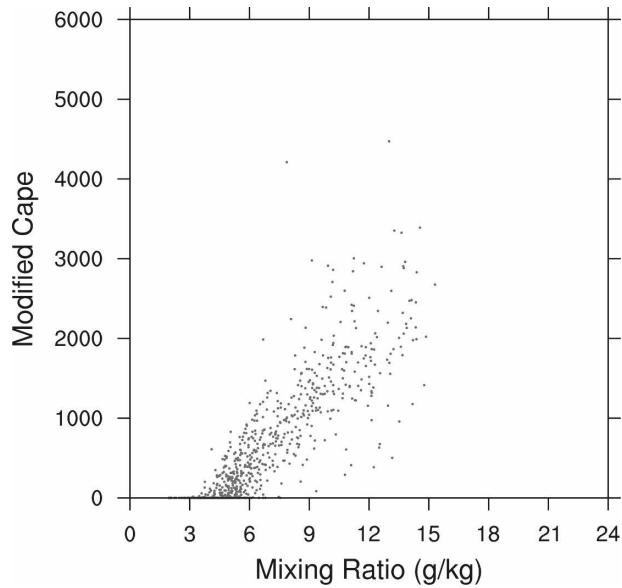


FIG. 6. Scatterplot of modified CAPE as a function of the average mixing ratio for the period of record (DRA). The parcel mixing ratio was averaged over the lowest 75 hPa.

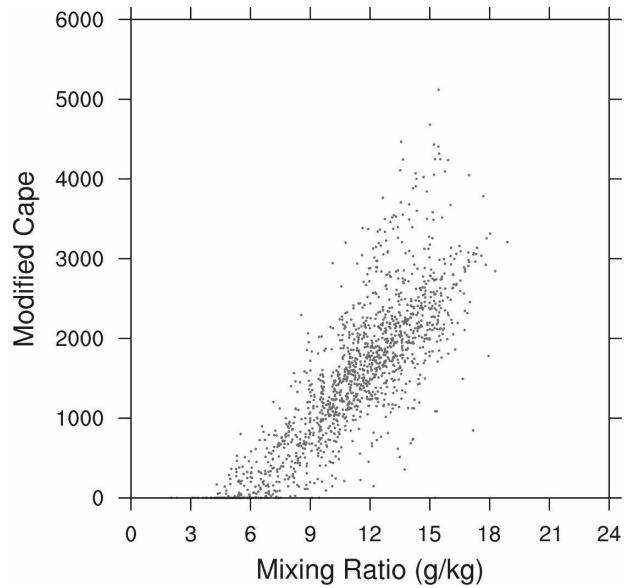


FIG. 7. As in Fig. 6, but for TUS.

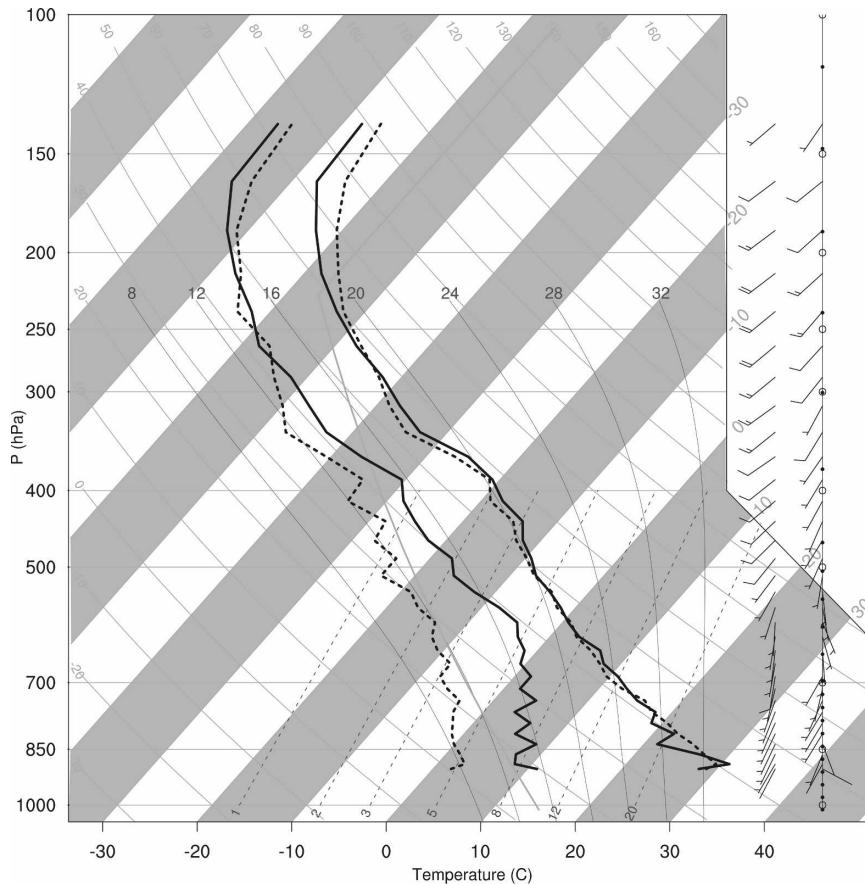


FIG. 8. As in Fig. 2, but for the top 20% of soundings from the linear fit.

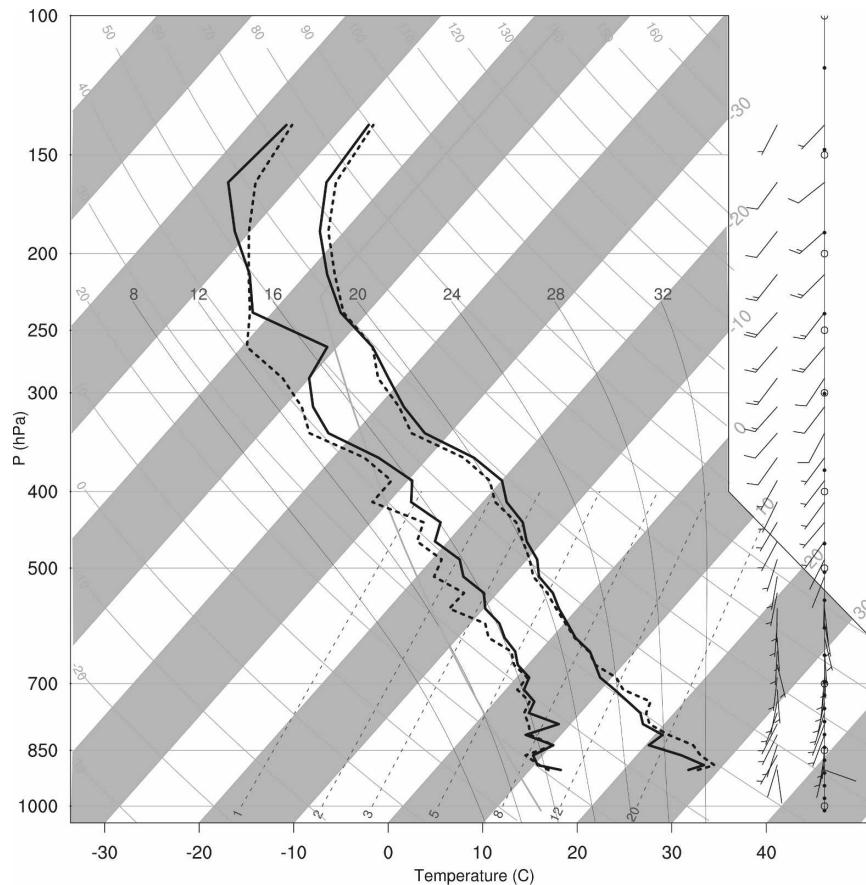


FIG. 9. As in Fig. 2, but for the bottom 20% of soundings from the linear fit.

humidities result in higher CAPE and greater areal precipitation; hence, the positive correlation.

To verify the role of low-level moisture in the positive correlations between CAPE and rainfall, soundings which deviate least and most from a linear fit between CAPE and rainfall were examined (calculated from the largest observed correlation at each site). Composites of the top and bottom 20% of soundings were made for both sites in order to characterize the moisture structures during burst and breaks for the extreme ends of the distribution. This allows for identification of the conditions under which CAPE and precipitation correlate the most and the least. Figures 8–11 are the top and bottom 20% soundings for DRA and TUS, respectively. The top 20% of soundings at DRA and TUS show, not surprisingly, nearly identical moisture and temperature structures to those in Figs. 2 and 3, respectively. The bottom 20% (i.e., least correlated) show distinctly different moisture structures from Figs. 2 and 3. In this case, the bottom 20% break cases have very large low-level moisture values, approaching those of the average burst conditions at each site. This implies

large values of low-level atmospheric water vapor even though total rainfall quantities were small and, hence, the break categorization. Analysis of stability characteristics of the bottom 20% break composite for both sites shows that large CAPE values were present ( $1200 \text{ J kg}^{-1}$  at DRA and  $2900 \text{ J kg}^{-1}$  at TUS) in the unmodified soundings. CIN, though very large at DRA ( $250 \text{ J kg}^{-1}$ ) would be easily overcome by modified maximum surface temperature. At TUC, CIN was quite low ( $10 \text{ J kg}^{-1}$ ). Sizeable instability, therefore, was possible at both sites, but rainfall was a minimum. One possible explanation for this apparent contradiction may be that convective cloudiness formed early in the day given the instability present, thereby, reducing solar insolation and, therefore, critical surface forcing to drive deep convection and precipitation.

These results suggest that factors other than just the presence of sizeable precipitable water values determine the outbreak of convective activity particularly for extreme CAPE and/or precipitation values. The stability–rainfall relationship is more complex. For example, isolating the largest 20% of rainfall events in the TUS

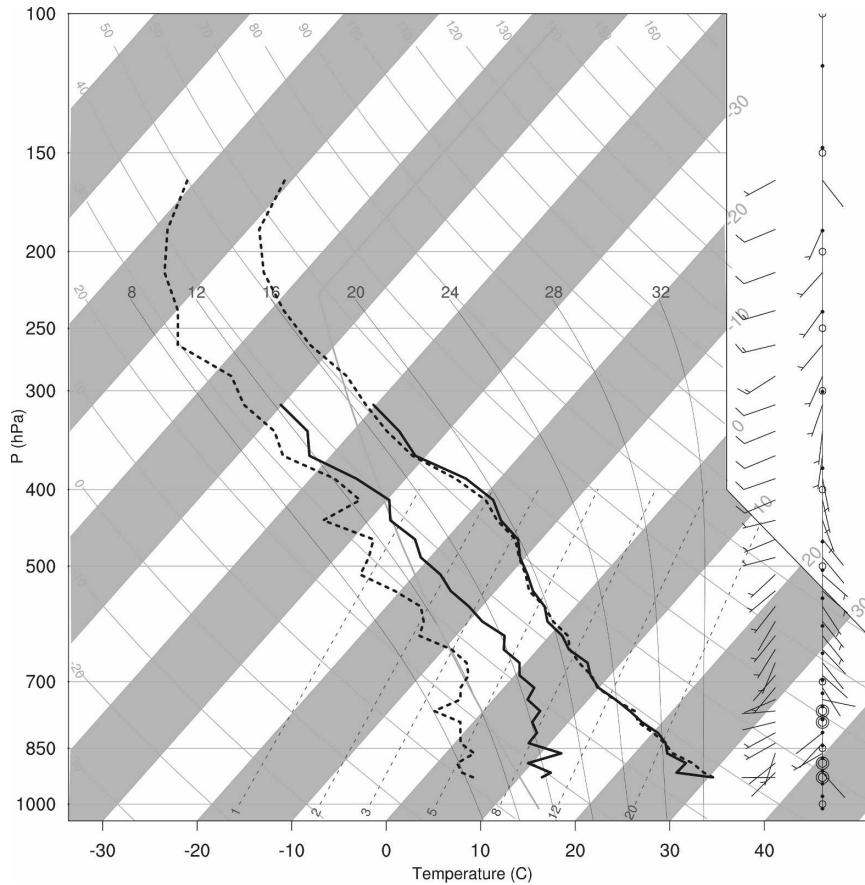


FIG. 10. As in Fig. 3, but for the top 20% of soundings from the linear fit.

region shows that a weak (not statistically significant) inverse relationship between CAPE and rainfall exists. At DRA, the correlations for the upper 20% of rainfall events are still positive, but much weaker (although this may reflect the decrease in sample size). Thus, other elements associated with synoptic-scale transients modifying convective instability and/or inhibition most likely play a role in the determination of extreme rainfall events.

The CAPE–precipitation correlations in this study represent the diurnal cycle, temporally. Numerous studies that have examined the CAPE–precipitation relationship, either observationally or modeling, have looked at forcing, convection, and precipitation on the subdiurnal scale. At the subdiurnal scale, particularly over continental regions, the relationship between CAPE and precipitation is complicated by the presence of CIN, as mentioned above. On the diurnal time scale, relevant for this study, CIN is, on any given monsoon day, very likely to be diminished or eliminated entirely given the strong surface sensible heating. This is consistent with the positive correlations found at lower lev-

els for all of the soundings, modified or unmodified. The reason is that, on average, the presence of CAPE in the sounding is reflective of the presence of low-level moisture. Given the elimination of CIN, CAPE is released and deep convection and precipitation occur.

To investigate this further, average values for CIN were calculated for both unmodified and modified 1200 UTC soundings for burst and break conditions and shown in Table 4. CIN values are quite large for the morning sounding (approximately  $200 \text{ J kg}^{-1}$  for both sites during burst periods). However, modification of these soundings using maximum surface temperature essentially eliminates CIN at both DRA and TUS. For modified break soundings CIN is small but not negligible, being slightly larger than burst conditions, entirely consistent with the idea of break conditions. Granted, the modification of the 1200 UTC sounding is crude, nevertheless, the results based on the modified sounding give the strongest correlations. Therefore, considering maximum daily instability for calculating convective–precipitation correlations has its merits if prediction of convective precipitation is the goal. Stud-

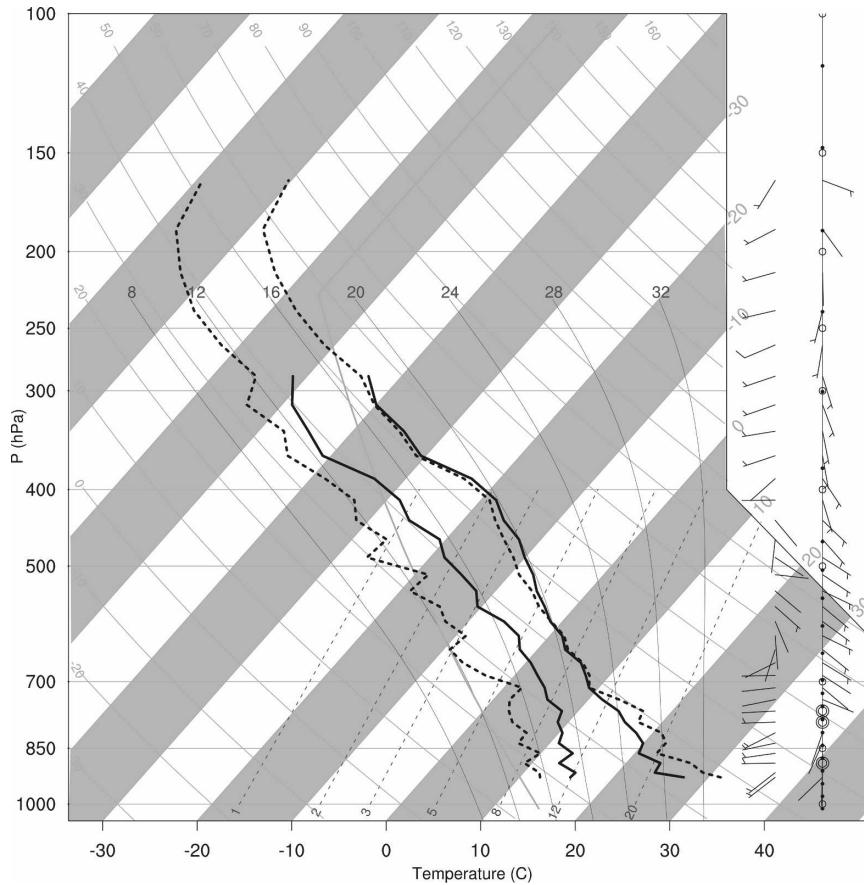


FIG. 11. As in Fig. 3, but for the bottom 20% of soundings from the linear fit.

ies at the subdiurnal scale show small or negative correlations due to most probably to the lag relationship, which reflects the complex physical relationship between convection, CAPE, and CIN found in the deep tropics and the central United States.

**4. Conclusions**

To summarize the principal results of this study, correlations between CAPE and precipitation appear to possess a strong convective regime or, regional dependence. It has been argued that the desert Southwest summertime convective regime experiences positive correlations due to its relatively stagnant conditions in terms of dynamic forcing and relatively strong surface forcing. Given the very deep boundary layers, unlike the deep tropics, positive correlations are found up to 700 hPa, though the strongest correlations are found in the lowest layers. These layers represent the most unstable parcels and correlate best with precipitation.

At its most basic, deep precipitating atmospheric convection is driven by the heating of the surface, which

destabilizes (i.e., increases CAPE) the atmosphere. All else being equal, this thermodynamic control implies a positive relationship between convection and precipitation. The correlations calculated in this and the studies cited could, in some sense, be viewed as the degree to which the observed convective regime deviates from this ideal. Even in a “simple” convective regime, a phase lag exists between measured instability and the actual occurrence of rainfall. As was shown in this study, unmodified morning soundings show weaker

TABLE 4. CIN values ( $J\ kg^{-1}$ ) for unmodified and modified 1200 UTC soundings for burst and break conditions at DRA and TUS. The parcel mixing ratio was averaged over the lowest 75 hPa.

	Unmodified	Modified
DRA		
Burst	188.0	6.38
Break	44.6	9.6
TUS		
Burst	224.0	1.24
Break	84.9	10.24

correlations than modified soundings and “postconvective” afternoon soundings show even weaker ones. Physically, the lag effect in continental convective regimes has been attributed to weak capping CIN and/or mixing of incipient cumulus with drier air above the boundary layer (Williams and Rennó 1993; Chaboureau et al. 2004). Zehnder et al. (2006) and Damiani et al. (2008) have proposed similar mechanisms for delaying the onset of convection in the Southwest. In utilizing the maximum potential convective instability each day, we have effectively eliminated the lag effect on the subdiurnal time scale (6–12 h). Tompkins (2001) also points out that the CAPE–convection relationship is a function of the sampling frequency and spatial scale examined, with positive CAPE–precipitation relations being found at longer time and larger space scales. This finding is consistent with the argument that substantial CAPE is always present in the deep convecting tropical atmosphere (Williams and Rennó 1993; Rennó and Ingersoll 1996). Likewise, an inverse CAPE–precipitation relationship from the studies of McBride and Frank (1999) and Sobel et al. (2004), appears to result, at least in part, from the local sampling of postconvective stabilized boundary layers. Considering the results from our study, elimination of these “contaminated” soundings and calculation of maximum daily instability may, in fact, modify the negative correlations observed in these deep maritime tropical regimes. Regardless, it can be said with confidence that *when* the thermodynamic profile is examined is critical in determining the CAPE–precipitation relationship.

It is also important to consider here the nature of the data utilized in this study with respect to previous studies. In several of the studies cited here, data was gathered during intensive observation periods associated with various field campaigns (McBride and Frank 1999; Sobel et al. 2004). These data provide the possibility to gauge indirectly large-scale forcing (through convergence–divergence calculations) in addition to the “immediate” atmosphere large-scale response or signal due to convective activity. Admittedly, point data (individual radiosonde sites) used as the present dataset cannot capture these details and are not ideal for investigating large-scale thermodynamic conditions on daily time scales for a large region. Nevertheless, the use of multiyear records captures the essential large-scale thermodynamic behavior and is sufficient to differentiate, in terms of average conditions, between active convective “burst” and less active “breaks.” Ideally, it would be of interest to examine the convective environment with a much denser network of radiosondes or remote sensing products to improve the analysis of atmospheric moisture transport in the re-

gion, particularly northwestern Mexico, which acts as the moisture source for the desert Southwest. The North American Monsoon Experiment (Higgen et al. 2006) utilized intensive observation networks, but the focus was understanding the hydrological cycle on a regional/continental space scales and on longer seasonal-to-interannual time scales.

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