The Influence of Soil Moisture on the Planetary Boundary Layer and on Cumulus Convection Over an Isolated Mountain. Part I: Observations

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The Influence of Soil Moisture on the Planetary Boundary Layer and on Cumulus Convection over an Isolated Mountain. Part I: Observations

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ABSTRACT

Data collected around the Santa Catalina Mountains in Arizona as part of the Cumulus Photogrammetric, In Situ and Doppler Observations (CuPIDO) experiment during the 2006 summer monsoon season are used to investigate the effect of soil moisture on the surface energy balance, boundary layer (BL) characteristics, thermally forced orographic circulations, and orographic cumulus convection. An unusual wet spell allows separation of the two-month campaign in a wet and a dry soil period. Days in the wet soil period tend to have a higher surface latent heat flux, lower soil and air temperatures, a more stable and shallower BL, and weaker solenoidal forcing resulting in weaker anabatic flow, in comparison with days in the dry soil period. The wet soil period is also characterized by higher humidity and moist static energy in the BL, implying a lower cumulus cloud base and higher convective available potential energy. Therefore, this period witnesses rather early growth of orographic cumulus convection, growing rapidly to the cumulonimbus stage, often before noon, and producing precipitation rather efficiently, with relatively little lightning. Data alone do not allow discrimination between soil moisture and advected airmass characteristics in explaining these differences. Hence, the need for a numerical sensitivity experiment, in Part II of this study.

1. Introduction

Warm-season deep convection in the western United States is strongly diurnally modulated, with cumulus convection initiating close to solar noon and a maximum in lightning and precipitation in the midafternoon (Watson et al. 1994b; Nesbitt and Zipser 2003; Carbone et al. 2002; Demko et al. 2009). Satellite imagery shows how this convection almost invariably initiates over mountain ranges and not the surrounding lower terrain (Banta and Schaaf 1987). Clearly the convection and mountain-scale circulations that control this spatial distribution are tightly coupled to the land surface, especially the surface sensible heat flux that controls the daily evolution of the convective boundary layer (CBL). The sensible heat flux mainly depends on the magnitude of the surface net radiation, and on soil moisture, which is the primary control in the partitioning of the available energy between sensible and latent heat fluxes (e.g., Sellers et al. 1992). Thus, soil moisture is a critical parameter determining CBL characteristics (e.g., Zhang and Anthes 1982). However, the impact of soil moisture on mountain-scale circulations and on the growth of orographic convection is less obvious. That is the topic of the present study. This paper (Part I) explores observations. A companion paper (Part II) uses numerical simulations.

This study should be framed in the broader question about the relationship between soil moisture and warm-season precipitation. Many numerical experiments have been designed to address this question (e.g., Sun and Ogura 1979; Carlson et al. 1983; Rowntree and Bolton 1983; Atlas et al. 1993; Koster et al. 2004). Observational studies on this topic are relatively scarce, especially in complex terrain. The simulated soil moisture feedback on precipitation over flat terrain can be positive or negative, mainly depending on scale, but also depending on numerical factors such as the surface and boundary layer schemes and the cumulus parameterization method (Pan et al. 1996; Gallus and Segal 2000; Hohenegger et al. 2009).

The main positive feedback mechanism is that moist soil enriches BL air with moist static energy, which
reduces convective inhibition (CIN) and increases convective available potential energy (CAPE; Betts et al. 1996; Eltahir 1998; Schär et al. 1999; Pal and Eltahir 2001; Findell and Eltahir 2003a,b; Wood 1997; Hohenegger et al. 2009). Another positive feedback is that increased soil moisture decreases surface albedo (Idso et al. 1975) and thus increases net radiation [and boundary layer (BL) moist static energy] under clear skies (Eltahir 1998). Boundary layer clouds may complicate this response: the addition of water vapor into the CBL may increase low-level cloudiness and stability (Hohenegger et al. 2009), but the resulting surface net shortwave radiation reduction may be compensated by a reduced net longwave radiation loss, unless the clouds are confined to the daytime (Schär et al. 1999; Pal and Eltahir 2001).

The main negative feedback mechanism is that moist soil reduces the sensible heat flux, and thus the daytime maximum temperature, the CBL depth, and the ability of air parcels to overcome CIN (Findell and Eltahir 2003a,b; Ek and Holtslag 2004; Hohenegger et al. 2009). Another negative feedback relates to spatial heterogeneity: in a flat landscape with mesoscale soil moisture variations, the wetter soil regions will be relatively cool, resulting in mesoscale subsidence as BL air diverges toward the drier, warmer patches, where the CBL becomes deeper and deep convection may erupt (Walker and Rowntree 1977; Sun and Ogura 1979; Carlson et al. 1983; Taylor et al. 2007; Nair et al. 2011). The flow toward warmer patches is density driven (solenoidal). The minimum scale supporting significant solenoidal BL circulations is at least an order of magnitude larger than the CBL eddies [i.e., at least $O(10 \text{ km})$], depending mainly on surface contrast, ambient wind, and BL depth (Chen and Avisar 1994).

These and other factors need to be considered in complex terrain. Elevated surface heating leads to significant pressure perturbations (Geerts et al. 2008) resulting in mountain-scale solenoidal BL circulations (e.g., Banta 1986) especially when soils are dry (Ookouchi et al. 1984). The initiation of cumulus convection over or near the high terrain is due to a combination of terrain-induced convergent flow (Banta 1984; Kottmeier et al. 2008; Barthlott et al. 2011) and elevated heating and early elimination of CIN (Demko et al. 2009; Demko and Geerts 2010a,b). Clearly orographic convection would be most effective if the surrounding valley soil is moist (with lush vegetation) and the mountain soil dry, since the solenoidal circulations due to the terrain and due to soil moisture would have the same sign, but in reality a history of precipitation mainly over mountains typically leads to an opposite soil moisture (and vegetation) distribution. The relation between soil moisture and orographic precipitation is complicated also because deep convection produces outflow boundaries that can trigger secondary convection (Demko and Geerts 2010b), and because the complex terrain may aid or stifle mesoscale convective organization (Nesbitt et al. 2008; Hauck et al. 2011).

To make progress on the topic of soil moisture impact on warm-season precipitation in complex terrain, this study focuses on the rather simpler component questions of how soil moisture affects CBL growth, mountain-scale convergence, and initial cumulus growth. Part I of this study is entirely observational. Its primary data source is a network of surface weather and flux stations deployed for two months (July–August 2006) around a rather isolated mountain in southern Arizona.

A 5-day period of heavy rainfall occurred across southeast Arizona in late July 2006, yielding widespread near-saturated soil conditions. About 118 mm of rain fell between 27–31 July around the Santa Catalina Range (SCR; Fig. 1), mostly at night, causing some rare local flooding (Damiani et al. 2008). This event was driven by a synoptic-scale moisture surge, which is not uncommon in Arizona in summer (e.g., Adams and Comrie 1997). This spell was unusual because of a series of nocturnal organized convective systems, aided by high surface humidity and a weak but persistent
upper-tropospheric cyclone. During this spell the air mass was as moist tropical as commonly observed farther south along the Sierra Madre Occidental in summer (e.g., Higgins et al. 2006). The National Weather Service (NWS) forecast office at Tucson, Arizona (KTUS), recorded a total of 214 mm of rain in July and August 2012, which is 193% of normal. While this unusual wet spell makes it difficult to understand the more typical nature of land–atmosphere coupling in this region (Small 2001), it does provide a strong regional-scale signal to examine soil moisture impact. Clearly this study does not examine the role of soil moisture in this flooding event—that event was largely driven by large-scale moisture advection. Instead, it examines how the changes in soil moisture resulting from this event affect diurnal changes at the surface and in the boundary layer, and the resulting thermally driven orographic circulations and orographic convection.

Observations alone do not allow discrimination between soil moisture impact and changes due to large-scale advection. A model sensitivity study, in which soil moisture is controlled, does allow a quantification of the soil moisture impact, but only if the model is realistic. This can be checked by validating that the regional model, driven by observed initial and boundary conditions, accurately simulates atmospheric and land surface conditions over the two-month period. Part II of this study does exactly that.

Section 2 describes the data sources. These data are analyzed in section 3. The influence of soil moisture is summarized in section 4. Conclusions are given in section 5.

2. Data sources

The Cumulus Photogrammetric, In Situ, and Doppler Observations (CuPIDO; http://www.eol.ucar.edu/projects/cupido/) campaign was conducted around the SCR near Tucson in southeast Arizona between 1 July and 29 August 2006 (Damiani et al. 2008). This region is relatively isolated with a horizontal scale of ~30 km and a maximum height of ~2000 m above the surrounding plains. These plains are part of the Sonoran Desert, with sparse vegetation. Ten Integrated Surface Flux Facility (ISFF) stations were positioned around the SCR (Fig. 1). All stations collected key meteorological data such as 10-m wind and 2-m temperature and humidity. These data were averaged to a 5-min time resolution. Four ISFF stations measured soil moisture and temperature at 5 cm below the surface, and all surface energy balance terms, including the surface heat fluxes. High-rate temperature, humidity, and 3D wind data were measured at these stations by means of fast-response temperature and humidity sensors and 3D sonic anemometers at a height of 10 m above ground level (AGL). The latent and sensible heat (LH and SH) fluxes are estimated using the eddy correlation technique at 5-min intervals based on 30-min running averages. The SH flux is proportional to the buoyancy flux; that is, it includes the virtual temperature correction (wθv), where w is vertical velocity and θv virtual potential temperature. The soil heat flux is computed using the gradient method. We also use corresponding data at 10 m AGL from a flux tower located on Mt. Bigelow, near the top of the SCR. This station is at an elevation 1432 m higher than that of the surrounding 10 ISFF stations.

The evolution of orographic cumuli was captured by two pairs of cameras, one located to the northwest, the other to the southwest of the mountain (Damiani et al. 2008). The pair located 30 km southwest of Mt. Lemmon (Fig. 1) is used herein, because of the prevailing southerly flow during CuPIDO. Each camera pair was optimally spaced for stereo-photogrammetry over the SCR, and thus cloud edges can be geolocated in three dimensions at high temporal resolution (Zehrnder et al. 2007). This method is used to obtain the height of the highest cumulus cloud top over the SCR.

The National Lightning Detection Network (NLNDN) data are used to examine lightning patterns over the SCR. Upper-air data come from the KTUS radiosondes at 0000 UTC [4.5 h after local solar noon (LSN)] and 1200 UTC (shortly before sunrise). These radiosondes were released from the valley floor ~40 km southwest of the highest point in the SCR (Fig. 1). On 16 days, Mobile GPS Advanced Upper-Air Sounding System (MGAUS) soundings were launched at 45–90-min intervals from a location just upstream of the mountain top [see Table 1 in Damiani et al. (2008)]. We also use the KTUS level-III radar data (http://www.ncdc.noaa.gov/nexradinv/) for the detection of precipitation echoes (radar fine lines marking outflow boundaries were rarely seen), the Climate Prediction Center’s (CPC) station-based 0.25° × 0.25° gridded daily precipitation data (http://www.esrl.noaa.gov/psd/) and the 25 km × 25 km gridded Advanced Microwave Scanning Radiometer for Earth Observing System (AMSR-E)/Aqua daily L3 soil moisture product (http://nsidc.org/).

3. Results

a. Bifurcation based on soil moisture

Data from the 10 ISFF stations within 2 h of LSN (1930 UTC ± 2 h) are examined for July–August 2006 (Fig. 2). A 4-h period around noon is chosen because the net radiation and surface heat fluxes peak in that period.
The very wet period of 27–31 July is evident in Fig. 2c, but there are three wet days in early July, and several days with more than 5 mm of rain in August. The time series of precipitation at the ISFF stations corresponds well with the fully independent CPC U.S. daily gridded dataset (Fig. 3a). The ~25-km CPC data suggest that the SCR is the wettest in the region (Figs. 3b,c), with a peak value located just southeast of Mt. Lemmon.

Soil moisture (Fig. 2c) jumps up in each wet spell, and then tapers off gradually over several days in the following dry spell, mainly due to evapotranspiration (Grayson et al. 1997). Assuming that this decay is exponential, the data suggest a soil moisture “half life” of 1.7 days, at a depth of 5 cm, for the four-station average around the SCR. The moistest soils are found on the southeast side of the SCR (station south-southeast in Fig. 1), and the driest soils on the other side (station north-northwest), consistent with the CPC precipitation (Figs. 3b,c) and with the AMSR-E/Aqua soil moisture product (Figs. 3f,g). That product is based on the spaceborne measurement of upwelling microwave radiation at several frequencies (Njoku 2004). It indicates a strong west–east soil moisture gradient across southern Arizona, with an ~3% higher volumetric soil moisture to the east. The east–west gradient is broadly consistent with CPC precipitation data, although the AMSR-E soil moisture data fail to capture the widespread anomalously high soil moisture from late July into early August 2006 across the domain.

To examine the impact of soil moisture in a statistically significant way, we divide the two-month period into two parts: the wet soil period (WSP; 38 days) and dry soil period (DSP; 22 days) (Fig. 2c). The second period (27 July–29 August) clearly is wetter than the first period, but the brief wet spell in early July moistened the soils enough to compel us to include the period 5–8 July.
FIG. 3. (a) Time series of daily (1200–1200 UTC) precipitation from CPC [(b),(c) spatially averaged over the domain] and from the 10 ISFF stations. In the six color panels below (a), the DSP is shown on the left, the WSP on the right. These panels are geolocated by latitude and longitude, and terrain is contoured in blue at 0.5-km intervals. (b),(c) CPC, ISFF, and Bigelow mean daily precipitation in a small domain [(d),(e) shown as a box]. (d),(e) CPC mean daily precipitation in a larger domain. (f),(g) AMSR-E/Aqua volumetric soil moisture in the larger domain.
Table 1. Comparison of the dry and wet soil periods. All parameters are averages for all ISFF stations, unless stated otherwise. Values are averaged for a 4-h period centered on local solar noon (1930 UTC ± 2 h), except daily precipitation, which is a 24-h total. Some parameters involve other data sources, as listed [Bigelow (BGL)].

<table>
<thead>
<tr>
<th>Property</th>
<th>DSP</th>
<th>WSP</th>
<th>Difference (WSP − DSP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soil moisture (vol %)</td>
<td>6.6</td>
<td>15.6</td>
<td>9.0</td>
</tr>
<tr>
<td>Air temperature (°C)</td>
<td>33.7</td>
<td>27.4</td>
<td>−6.3</td>
</tr>
<tr>
<td>Soil temperature (°C)</td>
<td>40.7</td>
<td>31.8</td>
<td>−8.9</td>
</tr>
<tr>
<td>( \theta_e ) (K)</td>
<td>346.5</td>
<td>351.1</td>
<td>4.6</td>
</tr>
<tr>
<td>( \theta ) (K)</td>
<td>317.7</td>
<td>311.2</td>
<td>−6.5</td>
</tr>
<tr>
<td>( \Delta \theta ) (K) (BGL−ISFF)</td>
<td>0.9</td>
<td>2.2</td>
<td>1.3</td>
</tr>
<tr>
<td>Mountain-scale convergence (10(^{-4}) s(^{-1}))</td>
<td>1.1</td>
<td>0.6</td>
<td>−0.5</td>
</tr>
<tr>
<td>Vapor mixing ratio (g kg(^{-1}))</td>
<td>9.2</td>
<td>13.3</td>
<td>4.1</td>
</tr>
<tr>
<td>RH (%)</td>
<td>26</td>
<td>53</td>
<td>27</td>
</tr>
<tr>
<td>LCL (m MSL)</td>
<td>4045</td>
<td>2568</td>
<td>−1498</td>
</tr>
<tr>
<td>Daily precipitation (mm)</td>
<td>0.9</td>
<td>5.3</td>
<td>4.4</td>
</tr>
<tr>
<td>CPC daily precipitation (mm)</td>
<td>1.1</td>
<td>5.5</td>
<td>4.4</td>
</tr>
<tr>
<td>Surface albedo (%)</td>
<td>20</td>
<td>17</td>
<td>−3</td>
</tr>
<tr>
<td>Net SW radiation (W m(^{-2}))</td>
<td>694</td>
<td>568</td>
<td>−126</td>
</tr>
<tr>
<td>Net LW radiation (W m(^{-2}))</td>
<td>−201</td>
<td>−100</td>
<td>101</td>
</tr>
<tr>
<td>Net radiation (W m(^{-2}))</td>
<td>481</td>
<td>453</td>
<td>−28</td>
</tr>
<tr>
<td>Soil heat flux (W m(^{-2})) (GH)</td>
<td>−75</td>
<td>−47</td>
<td>28</td>
</tr>
<tr>
<td>(&lt;0 → into soil)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SH flux (W m(^{-2}))</td>
<td>205</td>
<td>101</td>
<td>−104</td>
</tr>
<tr>
<td>LH flux (W m(^{-2}))</td>
<td>58</td>
<td>168</td>
<td>110</td>
</tr>
<tr>
<td>SH + LH − GH (W m(^{-2}))</td>
<td>338</td>
<td>316</td>
<td>−22</td>
</tr>
</tbody>
</table>

in the WSP. That leaves a relatively short DSP, from 1 to 4 July and from 9 to 26 July. We emphasize that this study uses soil moisture to distinguish the wet period, instead of atmospheric parameters such as precipitation or dewpoint. Thus, our wet period is distinct from the Arizona “monsoon period,” which the National Weather Service bases on the dewpoint data. In 2006 the WSP started the day after significant rains, and included several days without rainfall.

The contrast between the dry and wet periods is obvious also in a regional context (Figs. 3d,e): the daily precipitation across southern Arizona during the DSP is only 38% of that during the WSP, which experiences heavy rain mainly in southeast Arizona, yielding a strong west–east gradient with the SCR at the western edge of the wet region. Around the SCR the daily precipitation is about fivefold higher during the WSP (Table 1). In the soils around the four ISFF stations, a maximum soil moisture content of about 22.6% is recorded at a depth of 5 cm, after the heavy rains of late July. During the WSP (DSP) the soil moisture averages 69% (29%) of this peak value. Many Sonoran desert plants depend on moisture deep below the surface (e.g., Shreve and Wiggins 1964), but the vegetation is sparse and transpires little, thus the LH flux is largely controlled by surface and near-surface water (Kurc and Small 2004). Seasonal greening does occur in response to significant rainfall (Méndez-Barroso et al. 2009), such as the late July 2006 wet spell. This vegetation response may serve to extend the memory of precipitation pulses (Watts et al. 2007).

b. Soil moisture, surface energy balance, and airmass characteristics

The surface and lower atmosphere differ substantially between the two periods, especially around noon (Fig. 2 and Table 1). The higher soil moisture (Fig. 2c) implies a higher LH flux (Fig. 2f)\(^1\) and a lower land surface temperature (Fig. 2a) during the WSP. The 6.3-K lower air temperature during the WSP (Table 1) is at least partly due to the reduced SH, especially in late July and early August as heavy precipitation was recorded across south and central Arizona. The Bowen ratio (SH:LH flux) averages just 0.7 during the WSP, compared to 6.2 during the DSP.

The SH and LH fluxes plus the soil heat flux may be underestimated: their sum is about 140 W m\(^{-2}\) below the measured net radiation (Table 1). Several other studies have documented a significant imbalance between net radiation and the three flux terms, most likely due to eddy correlation technique uncertainties (e.g., Panin et al. 1998; Twine et al. 2000; LeMone et al. 2002; Finnigan et al. 2003; Oncley et al. 2007). The lack of closure of the surface energy balance is more likely over the complex terrain where the flow is neither stationary nor horizontally homogenous (Panin et al. 1998). The flux deficit is about the same in the DSP and the WSP, thus the relative magnitude of the heat fluxes probably is about right.

The higher water vapor mixing ratio \( r \) during the WSP (Table 1) is in part due to higher regional-scale evapotranspiration from the surface (water vapor recycling), although synoptic-scale airmass advection likely explains a dominant part of the difference between the two periods, as it does in typical Arizona “monsoon” periods (e.g., Adams and Comrie 1997). That is because climatologically, in summer, southern Arizona is close proximity to a rather deep high-moisture air mass to the south.

As a result of the lower temperature and higher mixing ratio, the relative humidity in the WSP is about

\(^1\) Note that the LH and SH traces in Fig. 2f are discontinuous because the average is shown only if all four stations report. The all-station requirement is not imposed on averages based on all 10 ISFF stations in Fig. 2, since 9 reporting stations (and on 1 day just 8) still give a robust average.
twice that in the DSP, and the lifting condensation level (LCL), measured AGL, is about half as high in the WSP. The LCL (Fig. 2d) is calculated from surface temperature and dewpoint, assuming well-mixed conditions to cloud base. This is a fair assumption as cumuli are present almost every day around LSN, even during the DSP, and as the MGAUS soundings reveal a deep CBL around noon.

On many days in the WSP, cumulus clouds obscure the top of the SCR. On these days Mt. Bigelow experiences near-100% relative humidity, and the LCL estimated from the ISFF stations is substantially below that from the Bigelow temperature and dewpoint values (Fig. 2d). The potential temperature excess at Bigelow over the ISFF stations (Δθ) is also much higher on these days (Fig. 2e and Table 1), either because the CBL top over the ISFF stations is below Mt. Bigelow, or else because θ is increased by condensation of water vapor (i.e., the equivalent potential temperature θ_e is more closely conserved than θ in a cloud-topped CBL). Only on days with an LCL and a CBL top well above the elevation of Mt. Bigelow, is Δθ a good measure of solenoidal forcing strength.

The near-surface air has a higher θ_e during the WSP (Table 1). As will be shown later, the θ_e excess of 4.6 K during the WSP is not driven by diurnal surface heating, as it is persistent throughout the night. Thus, it is an air-mass characteristic. A θ_e excess (~3 K) during the WSP is found also in the 0000 (Table 2) and 1200 UTC KTUS soundings in the boundary layer. One can partition Δθ_e into temperature and moisture terms by linearizing the differential expression for θ_e:

\[ Δθ_e \approx Δθ + \frac{L}{C_p}Δr, \]

where \( L \) is the latent heat of condensation and \( C_p \) is the specific heat under constant pressure. The θ_e excess results from a persistently higher BL humidity during the

The net surface radiation [the sum of net surface shortwave (SW) radiation and net surface long-wave (LW) radiation] is lower during the WSP than the DSP, although the difference is relatively small (Table 1). These findings disagree with the theory of Eltahir (1998) and Pal and Eltahir (2001), which expects not only a higher θ_e, but also a higher net radiation and when the soil is moister, irrespective of cloudiness. This suggests that the WSP θ_e excess primarily is due to advection, and not primarily driven by the surface. High θ_e moisture surges from the south are the primary drivers of North American monsoon variability (Higgins and Shi 2000).

The surface albedo is slightly lower during the WSP (Table 1), as moist soils tend to look darker (Idso et al. 1975). But the net SW radiation is lower during the WSP, and so is the net LW radiation loss. This suggests higher low-level cloud coverage around LSN during the WSP. This is qualitatively confirmed by webcam data and visible satellite imagery.

c. Soil moisture and thunderstorm potential

We now explore the relation between soil moisture and deep-convective parameters. We use the 0000 UTC KTUS soundings since typically the CBL is still well developed at that time, although it may modified by orographic deep convection. The composite soundings for the WSP and the DSP in Fig. 4 are computed by interpolating data from individual soundings to fixed levels and then averaging. All available soundings (59 in total) are used in Fig. 4 and Table 2. Significant lightning activity and radar-detected near-surface precipitation was present over the SCR between LSN and 0000 UTC on 16 of these 59 days, most of them during the WSP (see below). Afternoon thunderstorms over the SCR may have spread cold pools across the Tucson valley, increasing CIN and decreasing CAPE. None of the 0000 UTC soundings reveal a shallow cold pool, and KTUS radar data do not reveal any outflow boundaries propagating over the sounding site between 2200-0000 UTC. The exclusion of these 16 soundings decreases the average CAPE a little in both periods, on

### Table 2. Average stability parameters for all 59 individual 0000 UTC KTUS soundings. The KTUS ground level is at 751 m MSL.

<table>
<thead>
<tr>
<th>Parameters (No. of soundings)</th>
<th>DSP (21)</th>
<th>WSP (38)</th>
<th>Difference (WSP – DSP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CBL height (m MSL)</td>
<td>4431</td>
<td>2818</td>
<td>-1613</td>
</tr>
<tr>
<td>LCL height (m MSL)</td>
<td>4241</td>
<td>2872</td>
<td>-1369</td>
</tr>
<tr>
<td>LFC height (m MSL)</td>
<td>4904</td>
<td>3587</td>
<td>-1317</td>
</tr>
<tr>
<td>LNB height (m MSL)</td>
<td>10 988</td>
<td>12 689</td>
<td>1701</td>
</tr>
<tr>
<td>CAPE (J kg⁻¹)</td>
<td>420</td>
<td>707</td>
<td>287</td>
</tr>
<tr>
<td>CIN (J kg⁻¹)</td>
<td>69</td>
<td>73</td>
<td>4</td>
</tr>
<tr>
<td>Precipitable water (mm)</td>
<td>29.0</td>
<td>34.5</td>
<td>5.5</td>
</tr>
<tr>
<td>θ_e @ 900 mb (K)</td>
<td>344.4</td>
<td>347.6</td>
<td>3.2</td>
</tr>
</tbody>
</table>
FIG. 4. (a) Composite 0000 UTC KTUS soundings in the WSP and DSP. (b) Potential temperature and (c) mixing ratio gradient of this composite sounding in the lower troposphere.
account of the higher preexisting CAPE on days with afternoon thunderstorms.

The WSP profile is more similar to a typical monsoon sounding in the Sierra Madre Occidental farther south in Mexico (Higgins and Gochis 2007). It suggests a history of modification by deep convection. Compared to the DSP profile, it has a slightly higher tropopause, a mid-tropospheric temperature lapse rate close to the moist adiabatic rate, and higher mixing ratio in the mid- and upper troposphere. But the largest differences between the two periods are found near the surface (Fig. 4).

Both the \( \theta_e \) and \( r \) profiles indicate that the CBL is generally deeper in DSP, consistent with the higher SH flux in that period. The smaller slope of the \( \theta_e \) profile during the DSP (Fig. 4b) indicates a weaker static stability. The lapse rate is superadiabatic [\( \partial \theta_e / \partial z < 0 \)] during this period, not just near the ground (which can be an instrument bias, related to the lack of ventilation just before the sonde is released), but also well above the surface, up to 1.0 km AGL. This and the more uniform \( r \) (near-zero slope) suggest more vigorous mixing through the CBL during the DSP, consistent with a higher surface SH flux (Table 1). The high \( r \) lapse rate during the WSP suggests a high LH flux (Fig. 4c); this lapse rate is almost as large near the surface as that in the entrainment zone near the CBL top. The CBL top in the composite sounding (Fig. 4) is not sharply defined, as a peak in the \( \theta_e \) rise rate and/or the \( r \) lapse rate. It is defined practically as the first level with a significant increase of \( \theta \) [0.1 K (100 m)\(^{-1}\)] and a significant decrease of \( r \) [0.2 g kg\(^{-1}\) (100 m)\(^{-1}\)]. The CBL top is generally much more obvious in individual KTUS soundings. The CBL top also is generally better defined at KTUS at 0000 UTC, than over the SCR closer to local noon. This may be because of subsidence over the Tucson valley, and ascent over the SCR. The soundings over the SCR are MGAUS units, released from various points around the mountain between 1500–2100 UTC, on just 16 days in the two-month period [i.e., the CuPIDO intensive operations periods (IOPs) (more on this below)]. Six different balloon launch sites were used, all within 10 km from the mountain top (Mt. Lemmon; Fig. 1 in Damiani et al. 2008). They were selected in order to capture the upwind side of the mountain. KTUS is located farther from the SCR (\( \sim 40 \) km) in a broad valley (Fig. 1).

The CBL top and other sounding parameters in Table 2 are computed from the individual KTUS soundings and then averaged for the DSP and the WSP. The CBL is a remarkable 1.6 km deeper during the DSP, and extends well above the SCR top. The day-to-day variability of the CBL depth during the DSP is considerable (Fig. 5c). In some cases the CBL top may be underestimated due to a rather shallow stable layer within a possibly much deeper CBL (e.g., on 15 July). In such cases we ascertained that this stable layer is not the top of a convective cold pool.

The precipitable water is higher and the LCL lower during the WSP (Table 2), consistent with surface observations (Table 1). On many WSP days, the SCR top is obscured by cumulus clouds. The CBL water vapor generally suffices for some CAPE to be present throughout the two-month period: surface-based CAPE is present on all but 3 days in DSP and all but 2 days in the WSP (Fig. 5b). The CAPE and near-surface (900 mb) \( \theta_e \) are higher during the WSP at 0000 UTC, suggesting that more fuel remained for evening convection at that time. Thus, the WSP has a higher potential of deep convection and heavy precipitation.

The KTUS soundings in both soil periods suggest that the initiation of deep convection over the valley must be rare. The CIN is rather substantial (70 J kg\(^{-1}\)) over KTUS at 0000 UTC, about thrice the average CIN in the MGAUS sounding released closer to the mountain closer to local noon. This is unlikely because of earlier convection (as cold pools were rarely observed over KTUS at 0000 UTC), but rather because of a tendency for subsidence to occur over the Tucson valley. This also explains the better definition of the CBL top at KTUS. Also, the distance from the LCL to the LFC is quite high, about 700 m on average, at KTUS in both periods (Table 2). This makes it unlikely for a cumulus cloud to reach the LFC, lacking buoyancy and suffering from entrainment. In fact the average CBL top remains well below the LFC.

d. Soil moisture and diurnal evolution of the boundary layer

We now examine the difference between the DSP and WSP diurnal cycles. The surface energy parameters vary as expected, with a peak around LSN and very small values at night (Figs. 6e–j). The net shortwave radiation (and also the net radiation) peak slightly before LSN, because cloudiness rapidly increases over the SCR during the 4 h centered at LSN, according to visible satellite imagery. The diurnal temperature range (Fig. 6a) is larger during the DSP, on account of the large daytime SH flux and the lower PW and cloudiness at night, leading to a higher longwave radiation loss. The same applies to the soil temperature range (Fig. 6d), which is driven by a larger soil heat flux (Fig. 6h). The diurnal water vapor range is larger as well during the DSP (Fig. 6b), because the daytime CBL grows deeper, and thus more dry free-tropospheric air is entrained into the BL. However, the diurnal range of \( \theta_e \) is slightly smaller during the DSP (Fig. 6c), as the temperature and...
humidity vary in opposition, and thus two relatively large terms in Eq. (1) oppose each other during the DSP. The heat flux terms at Mt. Bigelow are shown also in Fig. 6. The main reason is to demonstrate the importance of heating over elevated terrain to drive orographic cumulus growth. To quantify the evolution of the solenoidal forcing, we could examine the potential temperature excess at Mt. Bigelow over the foothill stations (Δθ, Table 1). But this is a suitable measure of solenoidal forcing only when the CBL is well mixed and deeper than Mt. Bigelow, and when the cloud base is above Mt. Bigelow. This is the case on all days in the DSP (Figs. 5a,c), at least during part of the day, but not on many WSP days. On these days the CBL top does not exceed the elevation of Mt. Bigelow and/or the cloud base is below Mt. Bigelow. Thus, we can estimate the peak solenoidal forcing (i.e., the isobaric temperature difference at Mt. Bigelow pressure level, between mountain and adjacent valley) to be about 1.0 K during the DSP (Table 1). Since this estimate is not available during the WSP, and the difference is only meaningful during the brief period when the CBL is well developed (as evident from Fig. 6a), the diurnal cycle of SH flux at Mt. Bigelow (Fig. 6j) is a more useful measure, as it drives elevated heating that drives this solenoidal forcing.

The Mt. Bigelow heat flux terms appear to be quite noisy, and are available over only 2/3 of the two-month period. The large perturbations in SH and LH fluxes are due to rather low-frequency terrain-driven coherent eddies, according to the vertical velocity power spectra (not shown): the Bigelow flux tower is located on a narrow, steep ridge, which explains the small diurnal temperature range (Fig. 6a). These eddies are more intense during the day as thermals grow in size toward the CBL top, but they may occur at night as well, in stratified flow impinging on the mountain. So the Bigelow flux data carry considerable uncertainty, but basic patterns are consistent with the ISFF record (Figs. 6i,j).
One remarkable feature is the peaking of the SH flux well before solar noon at Mt. Bigelow in both periods (Fig. 6j), even earlier than the SH flux peak at the ISFF stations. The Bigelow ridge is east–west (or east-southeast–west-northwest) oriented (Fig. 1), and the prevailing low-level flow was from the south-southeast, so the asymmetry of SH around solar noon is unlikely to be due to terrain facing the morning sun. Instead, it is due to a drop in SW radiation and thus net radiation around noon, associated with orographic Cu development (Figs. 6e,g). The near-surface SH flux in the morning hours over the mountain drives a mountain-scale solenoidal circulation (Demko and Geerts 2010a) and destabilizes the atmosphere, leading to orographic Cu.

The surface SH flux drives the development of the CBL. The CBL clearly deepens around solar noon (Figs. 7a,b). In Fig. 7 the CBL top is expressed AGL, because MGAUS launch sites vary in elevation. CBL depth is affected not just by time of day, but also by local terrain and prevailing wind. Unfortunately no time-matched mountain–valley soundings are available to examine the CBL “topography,” but numerical simulations indicate significant doming of the CBL top over the SCR, albeit less than the terrain doming (De Wekker 2008; Demko and Geerts 2010a). Thus, terrain height matters in examining CBL height (AGL). The higher launch points yield a lower CBL height (Fig. 7), but the launch sites did not change during IOPs.

In short, there is a clear CBL deepening trend at all sites during the CuPIDO IOPs. This trend continued to 0000 UTC, especially during the DSP, given the large increase in CBL depth from the MGAUS soundings (Figs. 7a,b) to the KTUS soundings (Table 2), which were released ~5 h later in the afternoon, on average. This deepening averages 2.0 km on the 4 IOP days in the DSP, and 1.2 km on the 10 WSP IOP days. Both the CBL depth and the cloud base (LCL) are much greater in the DSP than in the WSP, but in both periods the air tends to reach saturation at or very near the CBL top, not only according to KTUS (Table 2) soundings, but also in the MGAUS soundings. Camera animations and visible satellite imagery reveal that on most DSP days towering Cu pop up over the high terrain, while on many
The larger diurnal range of BL depth, $\theta$ (Fig. 6) and CBL depth during the DSP may imply that the diurnal variation of CAPE is larger as well, and, in turn, that convective precipitation is more strongly diurnally modulated during the DSP. Indeed, there is a clear trend in CAPE around noon during the DSP, and no such trend in the WSP (Figs. 7c,d). We believe this difference is significant (although the MGAUS-measured change in CAPE is limited to the 14 CuPIDO IOPs and a narrow part of the diurnal window) as it is consistent with the difference in the timing of convective precipitation, as discussed below.

### e. Diurnal variation of thermally forced mountain circulation

Elevated surface heating leads to a thermally forced orographic circulation with low-level anabatic wind converging over the mountain, and upper-level divergence near the CBL top (e.g., Fig. 1 of Geerts et al. 2008). One reason for the location of the ISFF stations in the foothills surrounding the SCR during CuPIDO was to estimate the presence and strength of this thermally direct circulation, or at least its surface component. Not surprisingly, the kinetic energy of this circulation is overwhelmed by turbulent kinetic energy in the CBL. So, following Geerts et al. (2008) and Demko et al. (2009), we examine the integrated convergence along the polygon defined by the 10 ISFF stations around the mountain. The “mountain scale” convergence ($s^{-1}$) is computed as

\[
- \nabla \cdot \mathbf{v} = \frac{1}{A} \oint_{\Gamma} \mathbf{v}_n ds \approx \frac{1}{A} \sum \mathbf{v}_n \cdot \delta s.
\]  

Here $\mathbf{v}$ is the horizontal wind vector; $\mathbf{v}_n(\mathbf{v}_n)$ is the velocity normal to the closed loop (polygon), whose

![Diagram](https://via.placeholder.com/150)
The flow even becomes slightly divergent in the early afternoon (Fig. 8). This reflects a tendency of orographic thunderstorms to be maturing and producing divergent cold pools in the afternoon. Indeed, some relation between the afternoon divergence spell (Fig. 9b) and precipitation (Fig. 9c) is evident in both periods: the brief recovery of convergent flow in the late afternoon coincides with a lull in precipitation and lightning frequency (Fig. 9b). This recovery of both convergence and temperature becomes more evident in a composite of thunderstorms where time is expressed relative to the storm’s first lightning strike over the SCR (Fig. 10).

Nocturnal divergent flow around the mountain establishes before sunset. Significant precipitation occurs in the evening in both periods, and throughout the night during the WSP, on account of a few long-lived mesoscale convective systems (MCSs) in late July. In general the early afternoon thunderstorms tend to be small, while nighttime systems tend to be larger and less confined to the mountain footprint, so the impact of thunderstorms on mountain-scale convergence is evident only during the afternoon.

f. Diurnal variation of orographic Cu growth, lightning, and precipitation

A stereo-pair of cameras was operational from rooftops on the southwest side of the SCR during CuPIDO (Fig. 1), in order to geolocate orographic Cu clouds (Quicktime animations can be found at the CuPIDO archive at http://www.eol.ucar.edu/projects/cupido/). On some days low-level or widespread cloudiness prevented the identification of orographic clouds. The daily evolution of cumulus cloud growth could be tracked on most CuPIDO days. For those days, the height of the tallest Cu tower over the SCR was estimated manually, at 10-min intervals. As soon as a spreading cumulonimbus anvil forms, the cloud-top height recording is terminated, but may be resumed if the anvil vanishes and new towers grow. The cloud-top determination is discontinued also when orographic convection becomes too complex or becomes obscured by interspersed clouds.
Fig. 9. (a) Stereo-grammetrically estimated cumulus cloud top over the Santa Catalina Mountains, as seen from the southwest side (Fig. 1) on those days on which the view was not obscured by low-level clouds, plus LCL-estimated cloud base and NLDN lightning strike frequency on matching days, for the WSP and the DSP. (b) Diurnal cycle of NLDN lightning strikes on all days in a box confined by stations north, east-northeast, west, and south (i.e., ~20 km of Mt. Lemmon). (c) Rain rate at the ISFF sites. (d) Day-to-day diurnal cycle of the lightning strikes.
We also track occurrences of lightning strikes over the SCR, using the NLDN database.

Cumulus cloud tops generally grow rapidly before LSN in both periods, but the cumulus deepening process and growth in lightning frequency tend to occur 1–3 h earlier during the WSP than during the DSP (Fig. 9). This is consistent with the higher CAPE (Table 2) and the higher humidity above the cloud base (Fig. 4) during the WSP. The latter implies lower cumulus erosion by entrainment of dry air in the midtroposphere (e.g., Wang et al. 2009). The cloud depth growth is offset somewhat by the increase in cloud-base height during the morning hours, on account of surface heating (Fig. 9a). Note that the cloud-top height is an average only, during the prelightning growth phase. During the DSP, convective cloud tops grow, at first with almost no lightning, until ~2.5 h after LSN, at which time lightning is remarkably frequent, not just for the 44 days for which both lightning and cloud-top data are available (Fig. 9a), but also in the 60-day record (Fig. 9b). Around this time (2100–2200 UTC), the mountain-scale convergence and the Bigelow SH flux become suppressed, not just in the DSP, but also in the WSP. The late-afternoon lull in precipitation (around 0000 UTC) is consistent with a reduced lightning frequency, a lower cloud-top height, and a recovery in mountain-scale convergence.

Precipitation correlates rather well with the lightning strikes in the afternoon and early evening. It is strongly diurnally modulated during the DSP (Fig. 9c), with virtually no precipitation between midnight and LSN, a clear afternoon maximum (consisting of two separate peaks), and a weak maximum a few hours after sunset. An afternoon summer precipitation maximum in the southern Arizona is well established (e.g., Wallace 1975; Watson et al. 1994b; Carbone et al. 2002; an animation of the diurnal variation of radar-based precipitation in summer can be viewed at http://locust.mmm.ucar.edu/episodes/Hovmoller). Precipitation is diurnally more evenly distributed during the WSP. The nocturnal precipitation maximum during the WSP is surprising. This maximum derives almost entirely from a series of nocturnal MCSs between 27 and 31 July 2006. Such episodes are unusual but not unprecedented (e.g., McCollum et al. 1995).

This nocturnal precipitation maximum is not mirrored by an equally pronounced peak in lightning frequency during WSP nights (Fig. 9). The storms on 27–31 July 2006, unlike storms on other days in the 60-day period, were almost all nocturnal (Fig. 9c). They had a large, continuous anvil on satellite infrared imagery, with widespread precipitation according to the KTUS radar. The relatively lack of lightning strikes (Fig. 9b) under copious rainfall (Fig. 9c) suggests that these nocturnal MCSs were relatively devoid of lighting, possibly because of a lower aerosol concentration in the BL (Williams et al. 2002), or because of weaker updrafts and/or shallow cloud tops (Nesbitt et al. 2008; Rowe et al. 2012). The overall number of daily lightning strikes over the SCR is about the same during both periods (12% difference), yet the daily precipitation rate is 5 times higher during the WSP than the DSP. We do not attempt to estimate precipitation efficiency, but the composite soundings suggest that the DSP convection is less efficient, given the deep, dry BL (Fig. 4). Convection shortly after sunset (~0300–0600 UTC) generates copious lightning during the DSP (on 26 July, Fig. 9d), but appears to be inefficient in terms of precipitation. The WSP also shows a precipitation peak with much lightning around this time: this is due to an event on 9 August, whose sounding is similarly dry at low levels as the corresponding one in the DSP (i.e., 26 July).

4. Discussion

Some of the above results suggest a negative relation between soil moisture and the parameters that affect convective initiation: the WSP corresponds with lower net radiation, weaker sensible heat flux, lower CBL temperature, more stable profile within the lowest 1000 m AGL, and shallower CBL. Yet in effect, deep convection tends to grow earlier in the day over the mountain and produce more surface precipitation during the WSP, on account of a lower cloud base, and higher precipitable water, equivalent potential temperature, and CAPE.

FIG. 10. Composite evolution of (a) average temperature at the 10 ISFF stations and at Mt. Bigelow (expressed as a departure from the 4-h mean) and (b) mountain-scale convergence, for select thunderstorm events over the SCR. Time is relative to the first lightning strike. Storms are selected to occur between 1800–0000 UTC, and are relatively isolated (no lightning <1 h before the first lightning and <1 h after its end).
The relationship between soil moisture and convection over complex terrain clearly is not straightforward. Airmass characteristics such as precipitable water, $\theta_v$, and cloud-active aerosol concentration are not primarily driven by soil moisture, but rather by advection, especially in view of the more tropical airmass in close proximity to the south. But the variation of other characteristics, such as SH flux, diurnal temperature range, solenoidal forcing, and mountain-scale convergence are strongly impacted by soil moisture. Variations in cumulus development and orographic precipitation are affected by both large-scale advection and regional soil moisture variations.

Orographic thunderstorms are common during the Arizona monsoon during both dry and wet soil periods, but the nature of the convection is different. This difference is at least partly due to soil moisture, as revealed in this study. Orographic deep convection can be triggered fairly easily in dry soil conditions, on account of a higher surface SH flux, more elevated heating, and stronger solenoidally forced circulation. It occurs mostly in the afternoon and early evening, and produces much lightning, but little precipitation. Under moist soils orographic deep convection tends to initiate rather early in the day, not because convergent, anabatic flow develops earlier, but rather because CAPE is larger, the low- to midtroposphere more humid (reducing cumulus erosion by dry-air entrainment), and thus growth from shallow to deep orographic convection more rapid.

Several nights with organized convection occurred during the WSP in 2006. Clearly the moist soils were not the cause of the organization or longevity of the convection, rather, the MCSs caused moist soils. But in turn, the moist soils alter the surface energy balance and daytime orographic cumulus growth, which affects the lightning frequency and precipitation efficiency of thunderstorms.

We repeat that the observed differences between the WSP and the DSP cannot exclusively be attributed to soil moisture. In Part II of this study, the two-month period will be numerically simulated and, following validation against the observations presented in Part I, a sensitivity experiment will be discussed in which the cause of the organization or longevity of the convection is at least partly due to soil moisture, as revealed in this study. Orographic deep convection can be triggered fairly easily in dry soil conditions, on account of a higher surface SH flux, more elevated heating, and stronger solenoidally forced circulation. It occurs mostly in the afternoon and early evening, and produces much lightning, but little precipitation. Under moist soils orographic deep convection tends to initiate rather early in the day, not because convergent, anabatic flow develops earlier, but rather because CAPE is larger, the low- to midtroposphere more humid (reducing cumulus erosion by dry-air entrainment), and thus growth from shallow to deep orographic convection more rapid.

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5. Conclusions
A two-month series of surface station and sounding data, collected around the Santa Catalina Mountains in Arizona during the summer monsoon, was used to study the effect of soil moisture on boundary layer evolution, thermally driven circulations, and cumulus development over complex terrain. The 60-day period was divided in two periods, depending on local soil moisture: the DSP and the WSP. The main findings are as follows:

- Wet soil conditions correspond with a dominance of latent over sensible heat flux at the surface and a slightly lower net radiation during the daytime because of the larger cloud amount. This yields a smaller diurnal temperature range and a lower daytime surface temperature. Also, compared to the DSP, the CBL is more stable and shallower around noon, and the solenoidal forcing weaker, resulting in weaker anabatic flow.

- The WSP air mass has more water vapor, and thus a higher equivalent potential temperature, a lower cumulus cloud base, and more CAPE. Orographic Cu towers appears rather early during the day, and in a high-humidity environment they grow rapidly to the cumulonimbus stage with divergent cold pools, on average 2–3 h earlier than during the DSP. The higher humidity during the WSP makes nocturnal organized convection more likely, when a suitable tropospheric flow pattern arises.

- Orographic precipitation occurs in the afternoon and early evening during the DSP, from thunderstorms that tend to be prolific lightning producers. Yet, with a high cloud base, the storms’ precipitation efficiency is relatively low.

- Observations alone do not allow discrimination between soil moisture and advected airmass characteristics in explaining these differences; hence, the need for a numerical sensitivity experiment.

In Part II of this study, high-resolution numerical simulations will be validated in terms of observed precipitation, soil moisture, surface, and upper-air characteristics, and then analyzed further to explore the effect of soil moisture on BL development, thermally forced circulations, and orographic cumulus convection.

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