The Convective Instability Pathway to Warm Season Drought in Texas. Part II: Free-Tropospheric Modulation of Convective Inhibition

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ABSTRACT

This study is concerned with the modulation by convective instability of summertime precipitation in Texas as a mechanism for maintaining or enhancing drought. The important role of convective inhibition (CIN), its dependence on the temperature at 700 hPa and the surface dewpoint, and the mechanism by which soil moisture modulates precipitation through CIN were described in Part I of this two-part series study. This study, Part II, examines the dynamic and physical processes controlling the temperature at 700 hPa and elucidates the large-scale influences on convective instability and precipitation integrating the principal processes found in both Parts I and II.

Back-trajectory analysis indicates that a significant contributor to warming at 700 hPa is the inversion caused by warm air transport from the Rocky Mountains and the Mexican Plateau where the surface potential temperature is greater than 307.5 K, rather than by subsidence. It was found that downward motion and warm air transport are enhanced in Texas when an upper-level anticyclonic circulation develops in the southern United States.

Upper-level anticyclonic circulations in the southern United States, one of the distinctive features of central U.S. droughts, strongly affect Texas summertime precipitation by modulating the thermodynamic structure of the atmosphere and thus convective instability. Stationary anticyclonic anomalies increase CIN not only by enhancing warm air transport from the high terrain but also by suppressing the occurrence of traveling disturbances. The resulting reduced precipitation and dry soil significantly modulate surface conditions, which elevates CIN and decreases precipitation. The aforementioned chain reaction of upper-level anticyclone influences that is expected to play an important role in initiating and maintaining Texas summer droughts can be understood within the context of CIN.

1. Introduction

The central United States has experienced a number of severe droughts during the last century. The major droughts of the 1930s and the 1950s were the most severe and long lasting and affected the southwestern United States and the Great Plains. The dust storms associated with the catastrophic decadal drought in the 1930s motivated the term ''Dust Bowl.'' The Great Plains drought of 1988 was associated with record high temperatures and widespread forest fires that burned nearly 4.1 million acres of forest, a Mississippi River discharge that was 40% of normal (a record low), and roughly $40 billion in economic losses (Sud et al. 2003). The 1980 drought was responsible for nearly 1300 deaths due to the intense heat (Hao and Bosart 1987). The frequent occurrence of multiyear droughts in the Great Plains implies that multiyear droughts are a regular characteristic of the Great Plains climate (Woodhouse and Overpeck 1998) and are expected to recur as normal climate fluctuations in the future as well.

One of the distinctive features of central U.S. droughts is a positive geopotential height anomaly over regions affected by drought (Namias 1982; Chang and Wallace 1987; Trenberth et al. 1988; Mo et al. 1991; Trenberth and Branstator 1992; Lyon and Dole 1995; Chen and Newman 1998; Schubert et al. 2004). The anticyclonic circulation associated with this geopotential height anomaly serves as a block to cyclonic activity from the Pacific by displacing storm tracks north of the U.S.–Canadian border,

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which results in a precipitation deficit and triggers droughts (Namias 1982). This positive geopotential height anomaly is often accompanied by other anomalies over the Pacific and/or over the Atlantic. This characteristic has been inferred to originate from anomalous sea surface temperatures (SSTs) in the tropical and extratropical Pacific through the propagation of stationary planetary waves.

While the droughts listed above were initiated by remote forcing (i.e., SST anomalies), they were maintained by local forcing, that is, the feedback of soil moisture after the remote forcing disappeared (Namias 1982; Lyon and Dole 1995; Trenberth and Cuillemot 1996; Hong and Kalnay 2002; Sud et al. 2003; Schubert et al. 2004). Traditional budget analyses (Hao and Bosart 1987; Lyon and Dole 1995) found that decreased evaporation associated with reduced soil moisture was responsible for the precipitation deficit over the drought regions but the divergence of the moisture flux was not significant. A significant role of soil moisture in maintaining drought conditions was also shown in several GCM simulation studies (Oglesby and Erickson 1989; Oglesby 1991; Hong and Kalnay 2002; Sud et al. 2003). In simulations of the 1988 (Atlas et al. 1993) and 1998 (Hong and Kalnay 2002) U.S. droughts, reduced soil moisture resulted in decreased rainfall as well as increased surface air temperature.

However, although the observational and modeling studies described above found a critical influence of anomalous SST over oceans and soil moisture over land on precipitation, those studies have not made clear the physical and dynamical contributions of the remote and local forcings to the precipitation deficit. For example, from the fact that precipitation is reduced as subsidence is enhanced (Hao and Bosart 1987; Chang and Smith 2001), we cannot immediately determine whether convection is suppressed by the subsidence inversion or by decreased instability induced by the subsidence. The response of precipitation to the remote and local forcings is not linear because of the complicated precipitation processes including planetary boundary layer (PBL) processes, moisture transports, temperature inversions above the PBL, and so on (Betts et al. 1996; Schär et al. 1999).

Texas droughts may differ from those experienced in the Great Plains due to the different latitudes (Lyon and Dole 1995) and proximity to moisture sources (Higgins et al. 1997b), implying different levels of relative importance for tropical and midlatitude dynamics. In particular, since rainfall in the Great Plains is often associated with baroclinicity and storm track dynamics, rainfall anomalies over the Great Plains are greatly affected by changes in the large-scale circulations in the upper troposphere. However, in Texas, cold fronts and similar disturbances are rare or nonexistent in summer so the amounts of rainfall in Texas should be directly influenced by the modulation of factors controlling convective precipitation (e.g., convective instability). Therefore, to better understand and model the summer drought in Texas, it is necessary to comprehend how land–atmosphere feedbacks and large-scale circulations directly affect the warm season convective precipitation.

On a monthly time scale, thermodynamic characteristics were found to be crucial in controlling the interannual variability of convective precipitation not only in the tropics/subtropics but also in the summer continents (Myoung and Nielsen-Gammon 2010a). In a companion paper (Myoung and Nielsen-Gammon 2010b, hereafter Part I), the modulation of summer precipitation by convective parameters and associated key surface processes were examined in Texas. In both studies, the significant role of convective inhibition (CIN) over convective available potential energy (CAPE) and precipitable water (PW) was evident in Texas (Part I) and other extratropical land areas throughout the globe (Myoung and Nielsen-Gammon 2010a) using not only National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data but also observed precipitation datasets. The importance of CIN over CAPE and PW in Texas is due to the fact that convection tends to be inhibited when CIN, a measure of the amount of energy needed to initiate convection, is large even with large CAPE and moisture availability. Part I further found that surface dryness and warming at 700 hPa independently contribute to large CIN totals and that the surface dewpoint temperature (Td) is greatly affected by soil moisture.

Although these studies of convective precipitation examined the overall variability of monthly mean precipitation, they provide significant implications for Texas drought. Thermodynamic characteristics resulting in precipitation deficits and the principal processes controlling those characteristics are expected to play important roles in initiating and maintaining summer droughts in Texas. The reduction in soil moisture that decreases precipitation by reducing Td and then increasing CIN in Part I is the primary factor maintaining drought conditions in the south-central United States as well as in the Great Plains. These results confirm that the most direct physical influences on summertime drought in Texas can be represented by convective parameters, especially CIN. Therefore, it is expected that upper-level anticyclonic circulations that often trigger warm season droughts in the United States may influence precipitation deficits due to their influence on CIN.

Part II has two main objectives. One is to examine free-tropospheric atmospheric processes governing the temperature at 700 hPa (Tlt) since, as described in Part I, Tlt significantly covaries with CIN and monthly precipitation in Texas. The other is to collect the information from the analysis of land surface processes in Part I and free-tropospheric processes analyzed here to show how land–atmosphere feedbacks and large-scale circulations that were inferred to play crucial roles in the previous U.S. drought studies directly and indirectly affect the convective parameters and precipitation.

This work is organized as follows. The data and methods used in this study are outlined in section 2. Section 3 examines the most significant processes affecting the interannual variability of temperature at 700 hPa. Sections 4 and 5 elucidate the large-scale influence on convective instability and precipitation through flow patterns and trajectories. Section 6 integrates the principal processes controlling the interannual variability of summer precipitation in Texas, including the cascade of influences and pathways between large-scale circulations and local precipitation. The major results and conclusions of this study are summarized in section 7.

2. Data and methods

Most of the datasets and analysis methods used in this study are the same as those in Part I. Readers are directed to Part I for more detailed descriptions of the datasets and analysis methods.

The monthly precipitation dataset used in this study is composed of U.S. climate division precipitation data obtained from the National Climatic Data Center for the period 1948–2003. From this dataset, an area-weighted, log-transformed, and state-wide averaged precipitation in Texas for each month (PRCP, hereafter) was computed.

The surface and tropospheric variables that may play a significant role in modulating the convective instability and associated precipitation are listed in Table 1. They are obtained from NCEP–NCAR reanalysis dataset (Kalnay et al. 1996) and are gridded at 2.5° × 2.5° resolution. This study uses monthly mean values averaged at the 11 grid points within Texas. However, to compute the temperature advection (TA700), 6-hourly temperatures and winds are used and the advection values are then averaged monthly. CIN is computed from the reanalysis based on monthly mean vertical profiles of temperature and dewpoint averaged across Texas using the General Meteorological Package (GEMPAK).

As in Part I, linear correlation analysis and linear regression analysis are employed to reveal statistical relationships among the variables. For the correlation analysis, ± 0.34 is roughly the 99% significance limit for a nonzero correlation with $N - 2 = 54$ degrees of freedom, assuming independent, normally distributed data. Such statistically significant relationships will be indicated by a star in the graphics that follow. The significance of the differences between the dependent correlations is measured using Williams's (1959) T statistic (Steiger 1980; Part I). In addition to these statistical analyses, potential physical mechanisms of interaction between two variables are also considered as a possible means of assigning causality.

TABLE 1. Frequently used acronyms and symbols.

Acronym or symbols	Meaning
SM	Soil moisture
SH	Sensible heat flux
LH	Latent heat flux
Td	Surface dewpoint
Ts	Surface temperature
DP	Dewpoint depression
OMG	500–850-hPa mean vertical motion
TA700	Temperature advection at 700 hPa

To investigate the influences of upstream air on temperature at 700 hPa, we used back-trajectory analysis, employing Green's functions of the mass continuity equation for a conserved tracer substance. The three-dimensional trajectory model was provided by Dr. K. Bowman of Texas A&M University. Detailed information and discussions are found in Bowman and Carrie (2002) and Bowman and Erukhimova (2004). The 6-hourly winds from the NCEP–NCAR reanalysis are interpolated to the particle locations linearly in both space and time. Trajectories for a rectangular region containing the state of Texas (28° ~ 35°N and 104° ~ 95°W) are computed. Particles are initialized on a regular latitude–longitude grid with 0.25° × 0.25° spacing at 700 mb, which gives 1073 particles in total. Trajectories are initialized each day at 1800 UTC at the grid locations and run 4, 8, and 12 days backward for every day in July from 1948 to 2003. A set of 1736 (31 days × 56 yr) trajectories is provided for each grid point.

3. Causes of low-tropospheric warming

Figure 1 shows the relationships of the temperature at 700 hPa in July in Texas (Tlt) with surface variables (SM, SH, LH, and Ts), vertical motion (OMG), and lower-tropospheric temperature advection (TA700). As shown in Part I, Tlt is not strongly correlated with any surface variables except Ts ($r = 0.71$, $p < 0.0001$). The linkage with Ts was found to be due to the entrainment of free-tropospheric air into the PBL (Part I). On the other hand, Tlt has a statistically significant positive correlation with OMG ($r = 0.45$, $p = 0.0005$). Since air warmed diabatically would tend to ascend due to buoyancy or geostrophic adjustment, the positive relationship between Tlt and OMG implies that downward motion

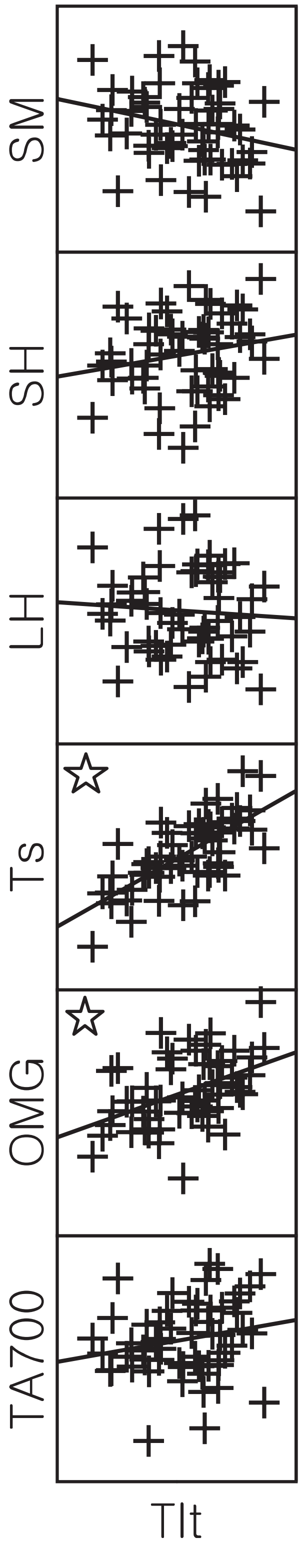


FIG. 1. Scatterplots showing the relationship between Tlt and various tropospheric variables. Stars indicate that the two variables are correlated at the 99% significance level.

causes warming at 700 hPa. Adiabatic heating by subsidence is often observed in the clear-sky regions of the tropics where the local downward motion and radiative cooling are in equilibrium without advection (Gray 1973; Gray and Jacobson 1977). Subsidence also results in a net drying of the lower troposphere. Since water vapor leads to longwave radiative cooling in the troposphere (Manabe and Strickler 1964), a drier troposphere (lower specific humidity) would induce less longwave radiative cooling and contribute to warming. However, despite the statistically significant correlation, OMG accounts for only 20% of the interannual variability of Tlt and the specific humidity at 700 hPa is nearly independent of Tlt in this study (not shown). Therefore, vertical motion does not seem to be primarily responsible for the temperature variation at 700 hPa.

Advection of warm air into a region can cause warming at 700 hPa as well. However, a notably poor correlation between temperature advection and Tlt in Fig. 1 ($r = 0.20$, $p = 0.14$) rejects the idea of a direct role for advection.

Since temperature advection occurs only when there is a temperature difference between two regions, there may not be strong warm advection even when there is steady flow from a pseudo-permanent heat source or sink that may exist on the periphery of Texas. Strong temperature advection might occur only when the local flow is transitioning from one state to another, not when it is locked into a particular state. For example, if there is steady airflow over an elevated heat source and the air arrives over Texas at the same altitude, the air would be expected to be anomalously warm, but the temperature advection would be minimal and would depend only on the rate of radiative cooling after the air left the heat source. The long-term mean potential temperature at 700 hPa for July 1968–1998 illustrated in Fig. 2a features a warm region over the high terrain to the west of Texas that is affected by surface diabatic heating. Although, climatologically, cool and moist southeasterlies from the Gulf of Mexico or Caribbean are dominant at 700 hPa in July in Texas (Fig. 2b), anomalous changes in wind direction, that is, from southeasterlies to southwesterlies in Texas, may have a major impact on Tlt. This speculation motivates us to examine the influences of the upstream thermodynamic characteristics on Tlt.

To investigate the influences of upstream air on Tlt, a back-trajectory analysis was conducted for the particles at 700 hPa as described in section 2. Figure 3 shows the origins of the back trajectories that terminate in Texas 8 days later. Most particles originate from oceanic regions to the east including the Caribbean Sea and the subtropical Atlantic (eastern origins), but other particles originate over the northeastern Pacific, the southwestern United States, and the Gulf of California to the west

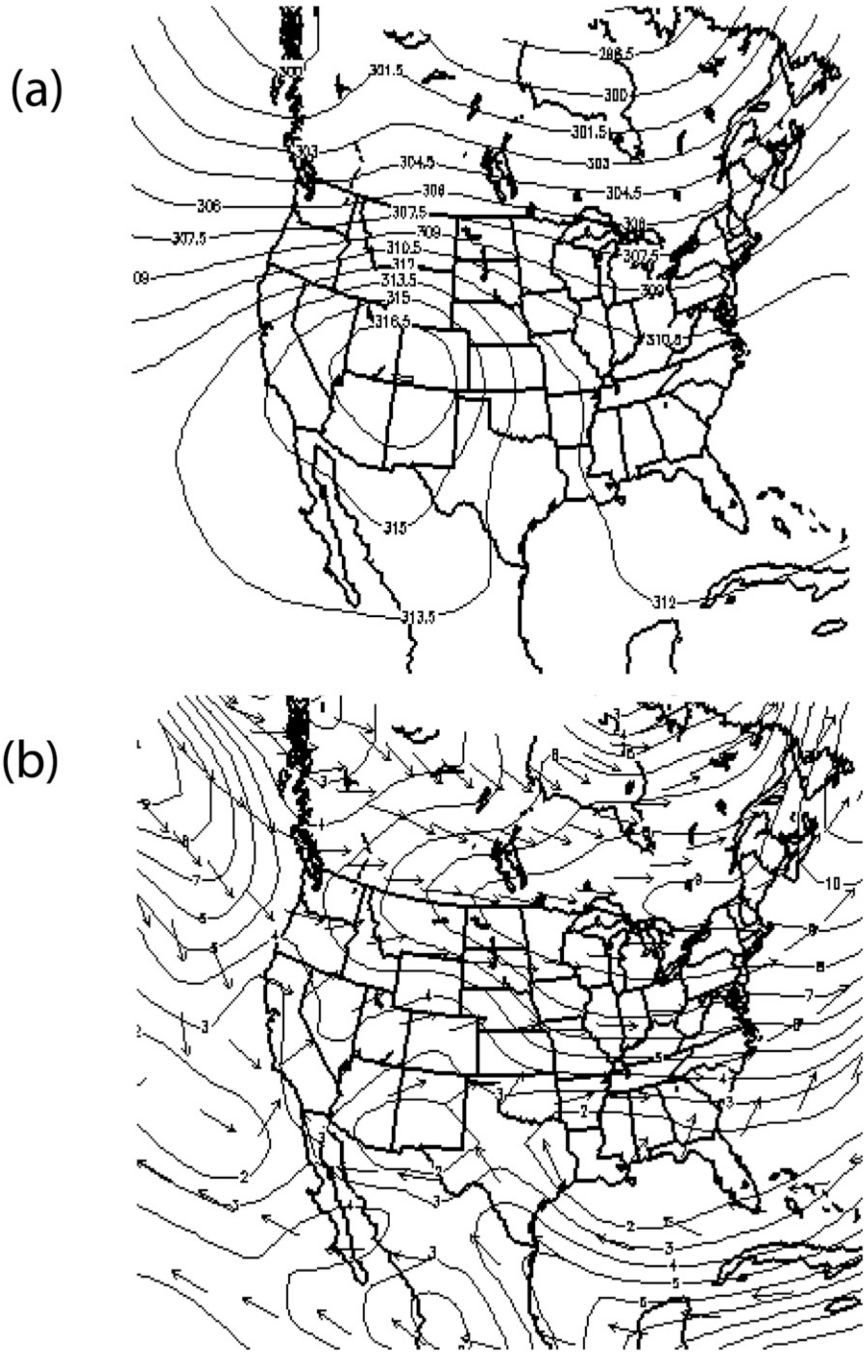


FIG. 2. Climatological 700-hPa (a) potential temperature (K) and (b) wind direction (arrows) and speed (m s^{-1}) for 1968–98. The contour intervals are 1.5 (K) and 1 (m s^{-1}).

(western origins). Figure 4 illustrates the origins for particles within the highest and lowest 20th percentiles of all 700-hPa temperatures in Texas, considering all 0.25° × 0.25° grid points over the 56 Julys. We call the particles within the highest 20th percentile warm particles and those within the lowest 20th percentile cold particles. The warm particles (Fig. 4a) are associated with mostly western origins although some of them come from the eastern origin areas. In contrast, most of the cold particles come from the eastern origins (Fig. 4b). Note that Fig. 4 represents the cumulative pattern over 56 yr.

Although the origins of the warm and cold particles vary in location and intensity interannually, the result in Fig. 4 suggests that the more particles originating from

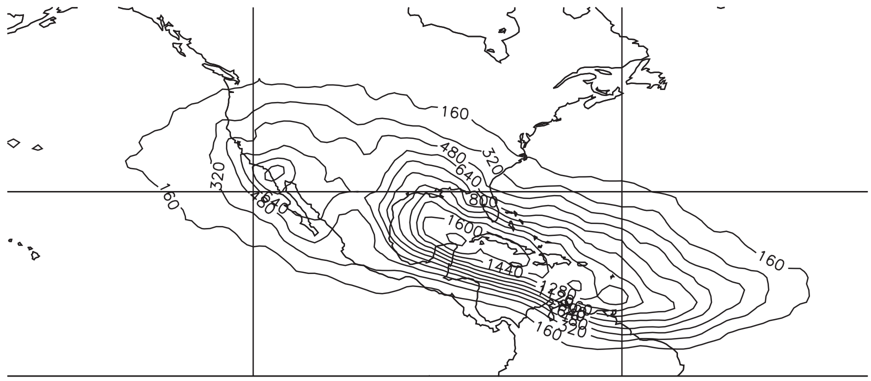
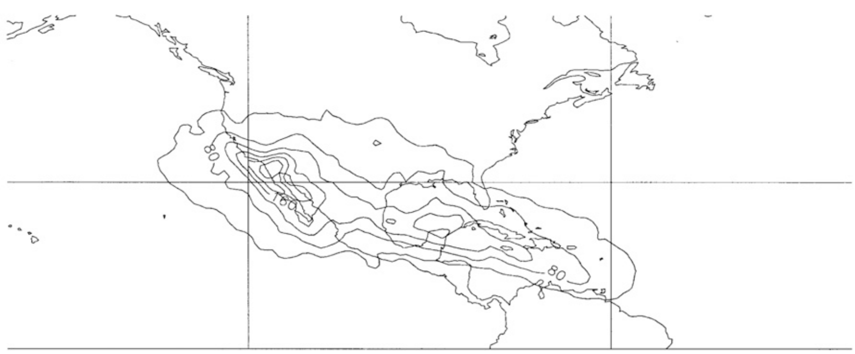
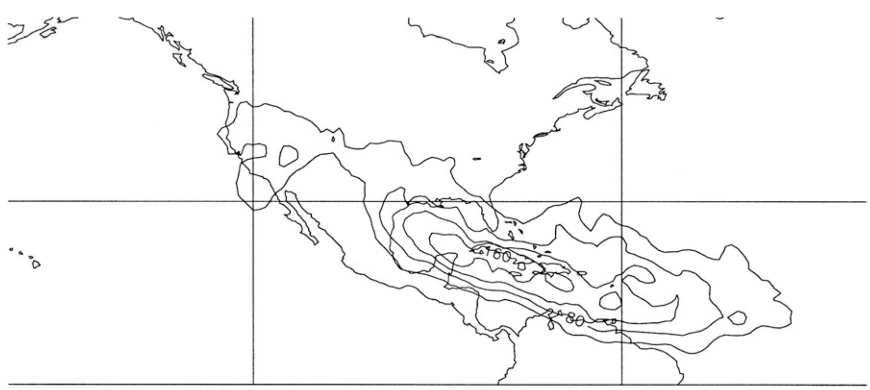


FIG. 3. Origins of 8-day back trajectories. The contour interval is 160 origins per unit area ($1° \times 1°$ grid box) for 1736 (31 days $\times$ 56 yr) sets of 1073 particles.

the western origins, the higher is Tlt. This leads us to hypothesize that the variability of Tlt is controlled by the number of particles from specific origins that preferentially produce warm destination temperatures (DT, which refers to the 700-hPa temperature of a particle in Texas). To test this hypothesis, we examined locations that preferentially were the source of warm particles. Warm particle origins were relatively common over the southwestern United States, northern Mexico, and the eastern subtropical North Pacific (not shown), with over half of the particles originating there becoming warm particles over Texas. The number of particles originating from such areas in a given year is positively correlated with Tlt ($r = 0.58$, $p < 0.0001$), but the correlation



(a)



(b)

FIG. 4. Origins of the (a) warm and (b) the cold particles. The contour interval is 40 per unit area ($1° \times 1°$ grid box).

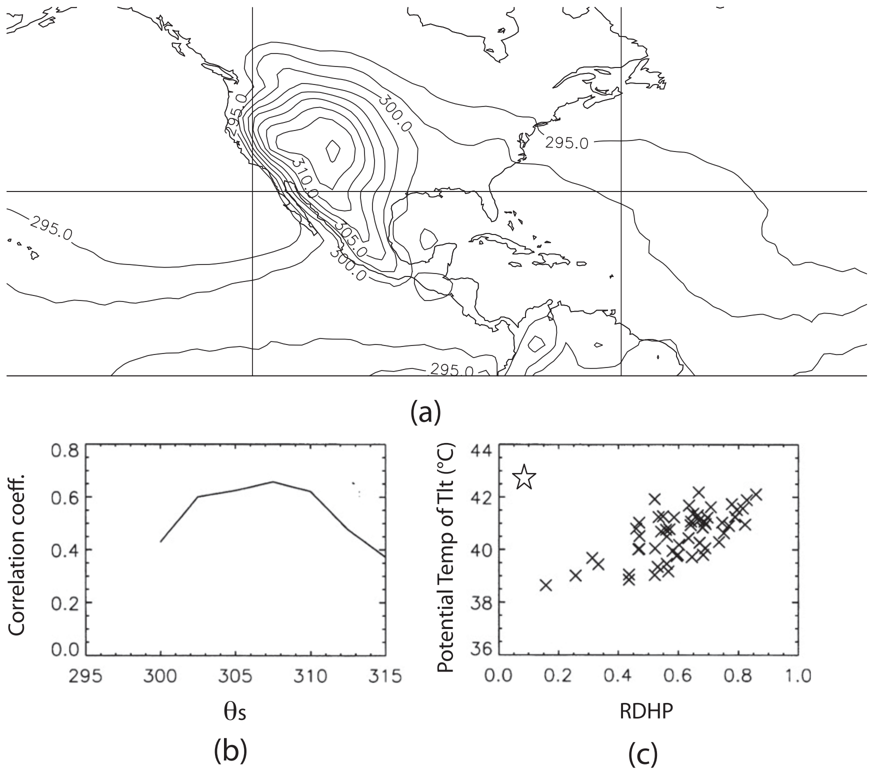


FIG. 5. (a) Climatological average of θs (K) for 1948–2003. The contour interval is 2.5 K. (b) Correlation coefficient of Tlt and RDHP according to θs, and (c) scatterplot of RDHP (x axis) and potential temperature (°C) computed from Tlt (y axis) with 307.5 K θs.

is not sensitive to the definition of "relatively common," nor are such locations cleanly separated from areas that tend to produce cooler particles. Furthermore, there is considerable interannual variability in the preferred origin locations of warm particles.

Nevertheless, the fact that a high correlation coefficient was found suggests that particle trajectories are related to Tlt somehow. If the diabatically heated area to the west of Texas is important, the paths rather than the origins of particles may be more relevant for affecting Tlt. According to Lanicci and Warner (1991), the elevated mixed layer (EML) that develops over the high terrain of the Mexican Plateau and the Rocky Mountains tends to induce a capping inversion and suppresses convection over the south-central United States most frequently in spring and early summer. Deep and strong capping inversions tend to suppress convection directly. In addition, on a longer time scale, they may cause the elevation of CIN not only by elevating low-tropospheric temperature but also by influencing the thermodynamics of the boundary layer. Motivated by this, we hypothesize that the proportion of particles experiencing diabatic heating over the Rocky Mountains and the Mexican Plateau regulates Tlt.

To test the hypothesis, the correlation coefficient was investigated between Tlt and the ratio of the particles passing over a diabatically heated area during their 12-day back trajectories. Rather than use reanalysis-inferred diabatic heating rates, which may not be strongly constrained by the observations, the diabatically heated area is defined as a region where the monthly mean of the surface potential temperature (θs) exceeds a certain threshold in a given month. The spatial extent of such a region varies interannually, and Fig. 5a shows the climatological average of θs. The highest θs is found over the central Rocky Mountain to the west of Colorado, and θs over the Mexican Plateau is also relatively high. Figure 5b represents the correlation coefficients between Tlt and the ratio of the diabatically heated particles to the total number of particles arriving at 700 hPa in each July in Texas (hereafter RDHP). The highest correlation ($r = 0.66, p < 0.0001$) occurs at 307.5 K θs, whose scatterplot is illustrated in Fig. 5c. This result clearly indicates that Tlt increases with more air from the diabatically heated area above 307.5 K.

The diabatically heated area corresponds to regions of high altitude. The possibility exists that the high correlation is due to air particles experiencing compression

TABLE 2. Adjusted R^2 of the linear regression models of Tlt on RDHP and OMG.

	R_a^2
RDHP	0.422
OMG	0.189
RDHP and OMG	0.420

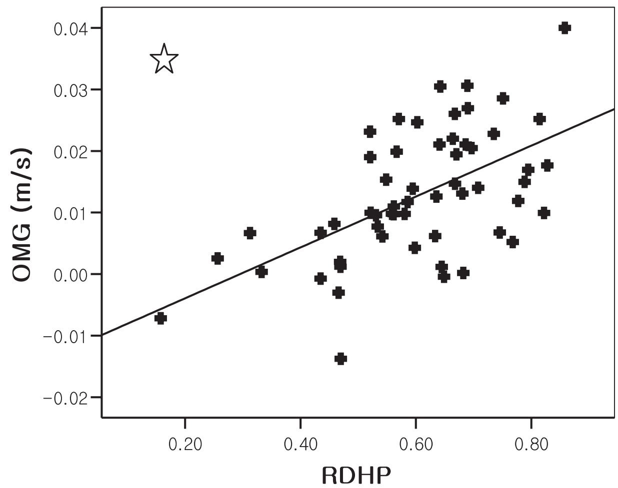


FIG. 6. Scatterplot of RDHP and OMG.

warming as they descend to lower altitudes in the lee of the high topography. However, the highest correlation coefficient between the ratio of high-terrain particles and Tlt ($r = 0.37$, $p = 0.005$) was much less than that for RDHP. The correlation difference suggests that high altitude does not correspond precisely to greater θs. This may be due to the existence of a difference in sensible heating according to the latitudinal variation of solar radiation and precipitation. In addition, there is interannual spatial variation in θs according to the surface thermodynamic conditions. When the diabatically heated region is fixed as covering all areas with a long-term mean θs of 307.5 K, the correlation coefficient ($r = 0.38$, $p = 0.004$) is much lower than if the actual θs for individual years is used. This is consistent with a possible role for soil moisture over the high-altitude areas of the southwestern United States and northwestern Mexico influencing Tlt, CIN, and PRCP, similar to what was found by Beljaars et al. (1996) for the 1993 central United States flood event.

As previously demonstrated, on the other hand, warming at 700 hPa is weakly linked with descending motion. Since both processes can increase Tlt, it is possible that warm air transport (high RDHP) and downward motion (large OMG) contribute to Tlt independently. However, the results of our regression analysis on Tlt using RDHP and OMG described in Table 2 shows that 1) RDHP and OMG explain 42% and 19% of the total variance of Tlt, respectively, with the difference being significant at the 95% level ($p = 0.04$), and 2) adding OMG to RDHP as an independent variable does not improve the predictability of Tlt. The latter discards the hypothesis that RDHP and OMG independently control Tlt. This is partially due to the fact that RDHP is not statistically independent from OMG, as shown in Fig. 6. The positive relationship between RDHP and OMG ($r = 0.57$, $p < 0.0001$) is an indication that subsidence occurs when an influx of diabatically heated air prevails in Texas.

Some of the results in this section merit further comment. First, subsidence is enhanced in Texas when the influx of diabatically heated particles is dominant, but the elevation of Tlt is caused by the latter rather than the former. This process has been ignored in the previous studies of the 1980 and 1998 droughts that strongly affected Texas. Anticyclonic circulations in the upper troposphere have been assumed to trigger drought by serving as a block to cyclonic activity from the Pacific by displacing storm tracks north of the United States–Canadian border and/or by reducing low-level moisture by interfering with the influx of moist air from the Gulf of Mexico into the Great Plains (Namias 1982). This study finds that anomalous circulations associated with the upper-tropospheric anticyclone induce increased transport of warm air from the elevated terrain into Texas, having a significant impact on initiating and maintaining warm season droughts in Texas by increasing CIN. Certain related aspects of the tropospheric thermodynamic environment over the southern plains have occasionally been noted, such as the increased level of free convection (LFC) reported across the southern plains in the 1988 drought by Hao and Bosart (1987). Second, although it is generally expected that an air parcel undergoing diabatic heating would ascend due to buoyancy or geostrophic adjustment, it is found in dry months in Texas that diabatically heated particles tend to descend. The next section will examine what induces the influx of diabatically heated air and subsidence during dry months in Texas.

4. Influences of upper-tropospheric circulation on the ratio of diabatically heated particles and omega

The combination of strong subsidence with enhanced warm air transport from the west does not seem to be merely the consequence of mechanical descent downstream as the air leaves the mountain area. The downward motion is not confined to the lower troposphere but extends to the upper troposphere above 500 hPa (not shown). Moreover, descent is more pronounced in the upper troposphere than in the lower troposphere. These features

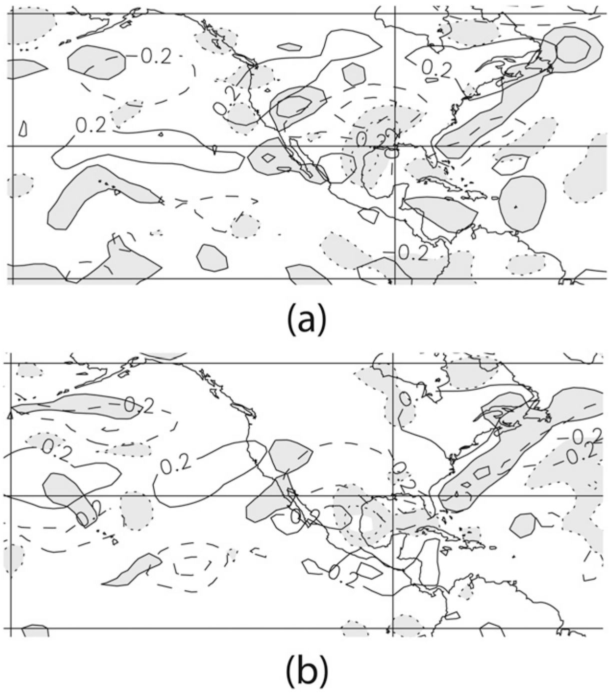


FIG. 7. Spatial map of the temporal correlation coefficient of (a) OMG and (b) RDHP with vorticity (contours) and divergence (shaded contours) at 200 hPa in July. The contour interval is 0.2, corresponding to $p = 0.14$ for the 0.2 contour and $p = 0.002$ for the 0.4 contour.

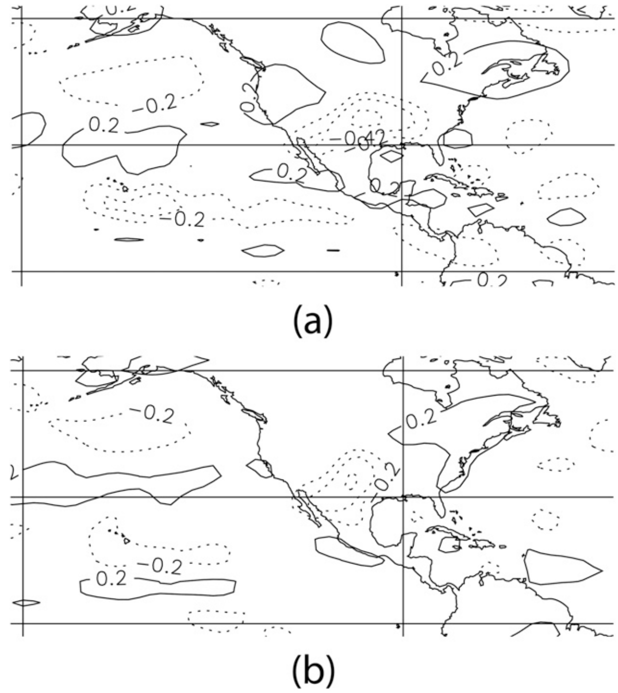


FIG. 8. Spatial map of the temporal correlation coefficient of the 500-hPa vorticity with (a) OMG and (b) RDHP in July. The contour interval is 0.2.

suggest that some large-scale feature that prevails in the troposphere may be involved in the linkage of RDHP with OMG.

The July mean 200-hPa vorticity and divergence fields were computed for each year in the period 1948–2003, and their correlation maps with OMG and RDHP are presented in Figs. 7a and 7b, respectively. The larger correlation values occur mainly in North America (Fig. 7a). More specifically, a statistically significant negative relationship with vorticity is found in the southwestern and the south-central United States (contours). This indicates that subsidence at 850–500 hPa in Texas is strongly linked with a possible anticyclonic circulation. For divergence fields (shaded contours), it is confirmed that convergence at the top of the troposphere is associated with descent in the lower to midtroposphere from the negative relationship between divergence and OMG found in Texas and its vicinity. RDHP is also linked with the upper-tropospheric vorticity and divergence at 200 hPa (Fig. 7b), although the location is slightly different from Fig. 7a. The linkage of RDHP with those upper-tropospheric variables is somewhat weak at 200 hPa, but much stronger and statistically significant at the 99% level at 500 hPa, which will be described below. These features confirm that upper-tropospheric circulations in the southern United States that are linked with the variations of OMG are also correlated with RDHP.

The correlation of OMG and RDHP with vorticity is stronger at 500 hPa than at 200 hPa. Figure 8 depicts the spatial patterns of their correlations with vorticity at 500 hPa. To further investigate the nature of this relationship, we defined a 500-hPa anomalous circulation region as the combined area adjacent to Texas where the correlation coefficient is greater than 0.3 in magnitude in either Figs. 8a or 8b and computed the interannual mean vorticity over the region (VOR_{500}). Figure 9 demonstrates the relationship of VOR_{500} with RDHP and OMG. The stronger the anticyclonic circulation, the larger are RDHP and OMG in Texas. Statistically, the correlation with OMG ($r = -0.79, p < 0.0001$) is significantly higher ($T = 2.7$, $p = 0.01$) than the correlation with RDHP ($r = -0.57, p < 0.0001$).

Which dynamical and physical mechanisms would explain the linkage of VOR_{500} with OMG and with RDHP? One of the direct influences of the upper-level anticyclonic circulation is to change the flow patterns in the troposphere. Due to the existence of a diabatic heating source over the Rocky Mountains and the Mexican Plateau to the west of Texas, Tlt is sensitive to the wind direction. Thus, enhancement of warm air transport from the west induced by anticyclonic circulation can increase Tlt significantly.

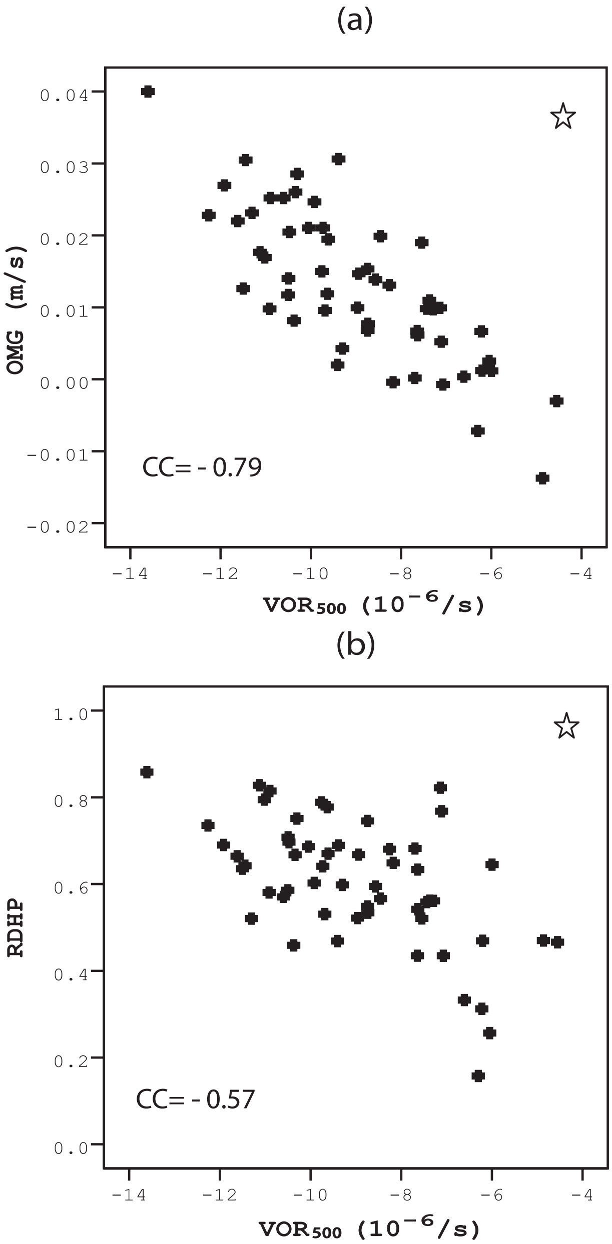


FIG. 9. Scatterplots between VOR_{500} and (a) OMG and (b) RDHP.

One example is shown in Fig. 10. We can see the 700-hPa geopotential heights (Fig. 10b) and 1000–500-hPa thickness (Fig. 10d) in July 2000 when VOR_{500} and RDHP were extremely high, while the mean July conditions are depicted in Figs. 10a and 10c. The anomalous northwesterlies in western Texas are expected to be warm due to the diabatic heating over the Rocky Mountains. Northerly and northwesterly winds in eastern Texas and the Gulf of Mexico indicate the weakening of climatological southeasterlies (Fig. 2b) over those regions. These anomalous wind patterns are associated with an upper-level anticyclonic anomaly illustrated by the positive anomaly in the 1000–500-hPa thickness in New Mexico and northwestern Texas, extending to the western Great Plains (Fig. 10d). When an anticyclone develops over the south-central United States, it is likely that climatological southeasterlies become less frequent and, instead, an influx of diabatically heated air from the west is enhanced at 700 hPa.

On the other hand, the physical and dynamical reasons why downward motion is enhanced in Texas when an anticyclonic circulation prevails in the south-central United States are not straightforward. In the literature concerning the 1988 drought in the Great Plains (Namias 1982; Trenberth and Branstator 1992; Lyon and Dole 1995; Trenberth and Cuillemot 1996), strong subsidence in drought regions and alternating anticyclones and cyclones like a wave train in their periphery had been inferred to be associated with a quasi-stationary Rossby wave, although it is not examined explicitly.

According to the quasigeostrophic omega equation, downward (upward) motion occurs downshear of an upper-tropospheric vorticity minimum (maximum) due to the negative (positive) differential vorticity advection in the upper troposphere over the surface high (Holton 1992). Therefore, for dynamically forced subsidence, thermal winds have to substantially advect negative vorticity in the southern United States into the subsidence region in the vicinity of Texas. The climatological mean speed and direction of the vertical wind shear (i.e., thermal wind) between 200 and 700 hPa in July in Fig. 11 indicate that a relatively weak northerly shear occurs in the regions where upper-level convergence is predominant in the vicinity of Texas in Fig. 7a. The northerly shear is present in almost all individual Julys (not shown). Since the center of the negative correlation of vorticity with OMG is north of Texas, the pattern is consistent with the subsidence being at least in part dynamically driven. Further diagnosis would be needed to determine the relative importance of this effect.

The arrow of causation is unlikely to move in the other direction. Downward motion directly leads to the production of cyclonic vorticity aloft. The fact that downward motion is instead correlated with anticyclonic vorticity implies that OMG does not cause VOR_{500}.

5. Physical connection between upper-level vorticity and precipitation

The aforementioned strong relationships among VOR_{500}, OMG, and RDHP imply that they influence the tropospheric variables, CIN, and precipitation as well. In this section, we will discuss and summarize how the

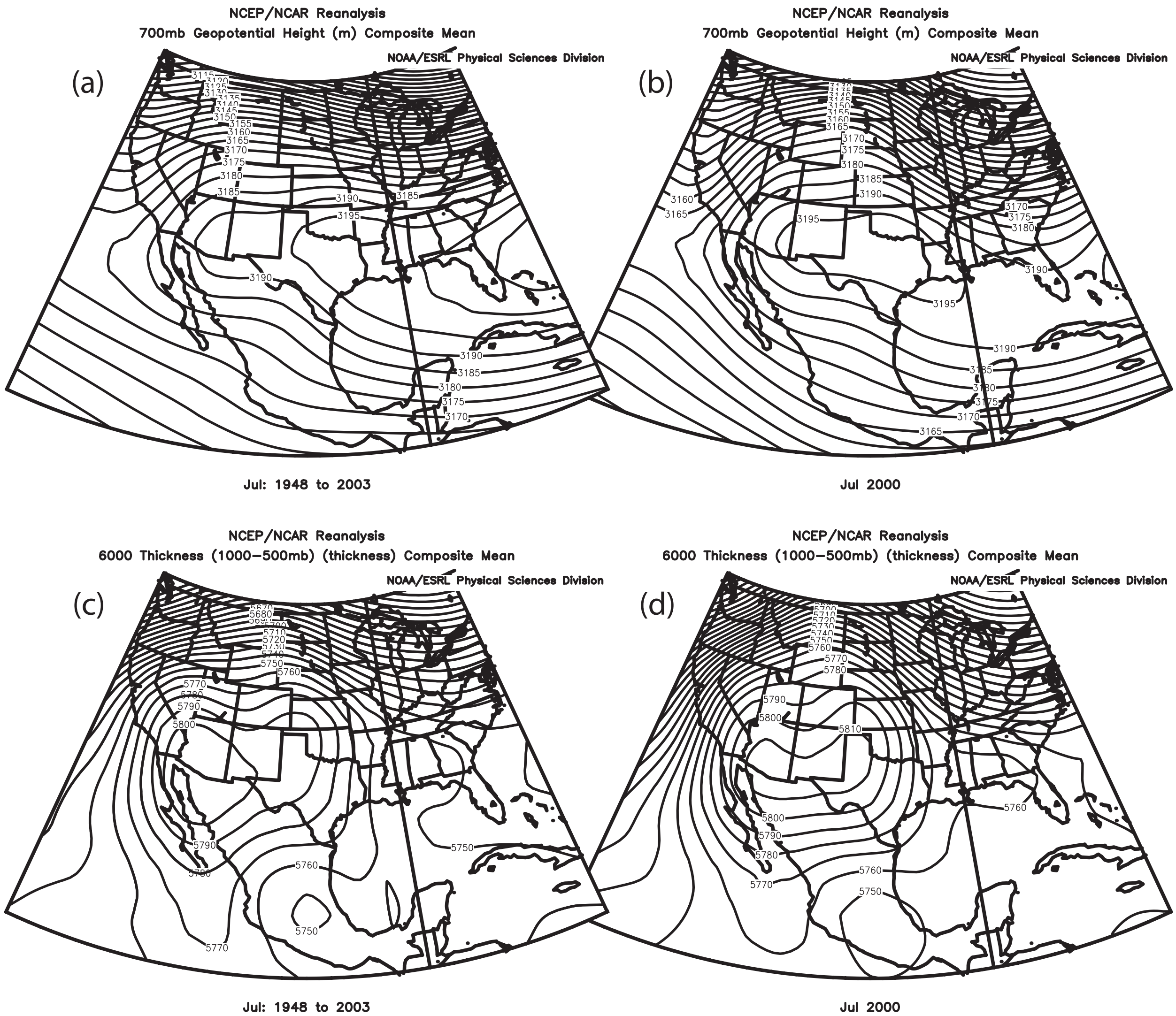


FIG. 10. Comparisona of (a),(b) 700-hPa geopotential heights (contour interval is 5 m) and (c),(d) 1000–500-hPa thickness (contour interval is 10 m) for (a),(c) the study-period mean July conditions and (b),(d) the year 2000, a representative July drought year.

500-hPa vorticity, vertical motion, and warm air transport are linked with precipitation by examining the cascade of pathways between VOR_{500} and Texas precipitation.

The correlations of OMG, RDHP, and VOR_{500} with surface variables (SM, SH, LH, Td, Ts, Tlt, and DP), CIN, and PRCP are given in Table 3. All the correlations except between RDHP and Td are statistically significant at the 99% level, indicating that atmospheric processes responsible for precipitation deficits during the warm season in Texas are interrelated on a monthly time scale. However, comparing the magnitudes of correlations and understanding the physical mechanisms between two variables allows for the identification of important physical pathways causing precipitation deficits and drought in Texas.

Sharing several similarities and differences, correlations of VOR_{500} and RDHP in Table 3 provides insights into the cascade of influences and pathways between large-scale circulations and local precipitation. First, they are most tightly correlated with CIN, next with PRCP, and least with SM, indicating VOR_{500}, OMG, and RDHP preferentially affect CIN and precipitation and only indirectly affect soil moisture. (The p values for the difference in correlation versus CIN and versus SM are $p = 0.008$ for VOR_{500}, $p = 0.05$ for OMG, and $p = 0.002$ for RDHP.) This feature is due to the direct effects of RDHP on CIN and precipitation by modulating Tlt, which is affected by VOR_{500} through the alteration of trajectories. The influence of warm air at 700 hPa on the surface is negligible on a short time scale; however, on a longer

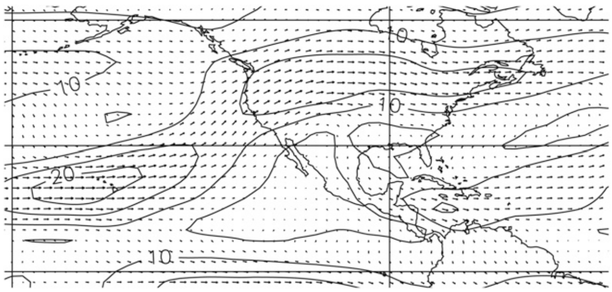


FIG. 11. Climatological mean speed (m s^{-1}) and direction of vertical wind shear between 200 and 700 hPa in July for 1948–2003. The contour interval is 5 m s^{-1}.

TABLE 3. Correlation coefficients of RDHP, VOR_{500}, and OMG with RRCP, CIN, and the tropospheric variables. Boldface entries are statistically significant at the 99% level.

	RDHP	VOR_{500}	OMG
PRCP	**−0.61**	**0.73**	**−0.73**
CIN	**0.73**	**−0.74**	**0.76**
SM	**−0.56**	**0.61**	**−0.66**
SH	**0.52**	**−0.62**	**0.65**
LH	**−0.48**	**0.49**	**−0.52**
Td	−0.33	**0.44**	**−0.55**
Ts	**0.78**	**−0.69**	**0.69**
Tlt	**0.66**	**−0.61**	**0.45**
DP	**0.65**	**−0.67**	**0.73**

time-scale, it affects the thermodynamic properties in the boundary layer. It increases the surface temperature (Ts) by entrainment of warm air into the PBL (Stensrud 1993; Part I) and decreases soil moisture by reducing the convection and the likelihood of precipitation. This inference is consistent with the stronger correlations of RDHP and VOR_{500} with Tlt and Ts than with SH, LH, and Td that are primarily controlled by SM (Table 3), although the only differences significant at the 99% level are with RDHP and Ts compared to RDHP and SH, LH, and Td. Not only is the difference in correlations not significant with VOR_{500}, but VOR_{500} is tightly correlated with both CIN ($r = -0.74, p < 0.0001$) and PRCP ($r = 0.73$, $p < 0.0001$), with an indistinguishable difference between the two correlations ($T = 0.29$, $p = 0.78$), while RDHP is more highly correlated ($T = 1.8, p = 0.07$) with CIN ($r = 0.73$, $p < 0.0001$) than with PRCP ($r = -0.61, p < 0.0001$). This is likely to be due to the upper-level anticyclonic circulations affecting the precipitation not only by increasing RDHP/Tlt and CIN but also by suppressing the occurrence of traveling tropical and extratropical disturbances. Since RHDP tends to have a direct impact on Tlt and Ts, it tends to be more tightly correlated with Tlt and Ts and less tightly with other surface variables than is VOR_{500}.

Although OMG shows strong linkages to CIN, PRCP, and other variables, we have not identified the critical impacts of subsidence on the precipitation deficit. While both OMG and RDHP can directly contribute to Tlt by adiabatic warming and warm air transport, respectively, this study found that increments in Tlt in dry months are dominated by increases in RDHP rather than by OMG. These results suggest that vertical motion does not efficiently modify the thermodynamic structure and convective instability, especially CIN, in summer in Texas. The lack of influence of OMG on the convective instability is consistent with studies of the thermodynamic structures in the tropics (Ye et al. 1998; Gray 1973; Gray and Jacobson 1977). According to Ye et al. (1998), vertical motion rarely changes the thermodynamic characteristics not only in the boundary layer but also in the free troposphere because of the canceling effects of adiabatic cooling and subsidence warming in the rising and sinking branches of the Walker and Hadley circulations. Instead, the surface flux and convection are likely to modulate the thermodynamic conditions by changing the surface temperature and moisture substantially in the tropics.

Other feedbacks have been proposed. Charney (1975) asserted that dry soil increases the albedo of the surface in the Sahel, causing downward motion to be enhanced to compensate for radiative cooling in order to reach thermodynamical equilibrium. However, this theory may not apply to more heavily vegetated regions in the subtropics (Charney et al. 1977; Sud and Fennessy 1984). On the other hand, vertical motions and upper-level circulations in Texas can be modified by convective heating in the periphery of Texas such as the southwestern United States and northern Mexico, the eastern tropical Pacific, and the Caribbean Sea, where convection is enhanced in the summertime (Johnson et al. 1987; Schaack and Johnson 1994; Higgins et al. 1997a). Further investigation of whether remote tropical heating is responsible for the strong subsidence over Texas in dry months is beyond the scope of this study.

6. Discussion

Figure 12 demonstrates the principal processes and pathways examined in Parts I and II that lead to July precipitation deficits in Texas. Processes verified in this study to be significant are indicated by solid lines and unverified but possibly significant processes are given by dashed lines in Fig. 12. The arrows of causality are inferred based upon physical principles. The following discussion begins in the lower-left portion of Fig. 12.

Part I showed that of the three convective parameters tested (CAPE, CIN, and PW) only CIN was significantly

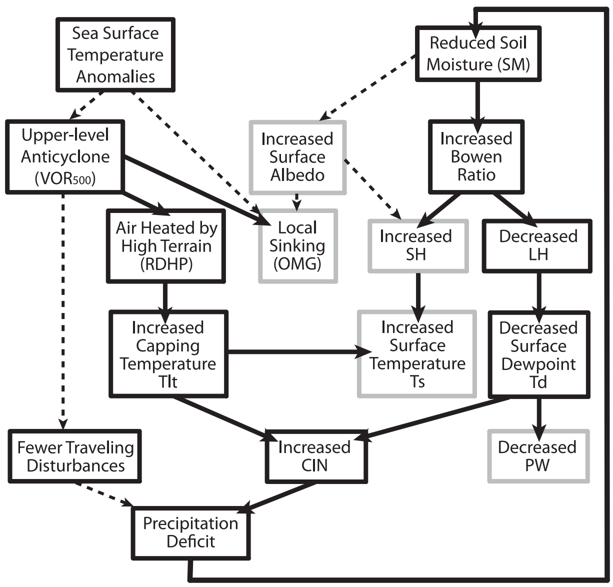


FIG. 12. Schematic diagram showing the primary processes affecting drought that have been verified by this study (solid) and unverified but possibly significant processes (dashed). Factors not part of a primary causal chain leading to precipitation deficit are depicted in gray boxes.

related to the resulting monthly precipitation. Other processes directly affecting precipitation were not tested; the most plausible impact might be from the disturbance frequency, since much summertime precipitation originates from tropical disturbances. One might imagine a situation in which the environment does not support frequent convection, but occasional disturbances with reduced CIN govern the overall precipitation amount. However, monthly mean CIN alone explains 56% of the variance of the monthly precipitation (Part I), so it is clear that CIN is the dominant environmental characteristic.

CIN, in turn, is almost entirely determined by the surface dewpoint Td and 700-hPa temperature Tlt (Part I). The surface temperature Ts does not independently affect CIN, but it is closely related to Td through variations in the Bowen ratio (SH–LH), which in turn is governed on an interannual basis by soil moisture (SM; see Part I). While the nocturnal low-level jet (LLJ) and associated moisture flux convergence tend to enhance the deep convection and precipitation in the Great Plains during warm seasons, they do not seem to modulate the surface humidity and precipitation on monthly time scales in Texas (section 4b in Part I). Part I showed for the first time that, at least in Texas, the mechanism by which the soil moisture feedback on precipitation operates is through modulation of CIN.

The surface temperature Ts is also independently affected by the entrainment of warm air from aloft, as diagnosed by variations in Tlt (Part I). The largest direct impact on Tlt was found in Part II to be the recent history of air parcel trajectories arriving over Texas. The greater the fraction of trajectories passing over the elevated heat source of the Rocky Mountains and the Altiplanicie Mexicano (RDHP), the greater the temperature Tlt and the greater the CIN. The upper-level wind patterns associated with a relatively large RDHP correspond to anticyclonic flow over Texas, as manifested by a statistically significant correlation between RDHP and the mean 500-hPa vorticity (VOR_{500}) (Table 3).

Ultimately, many North American droughts have been associated with spatial and temporal variations of oceanic sea surface temperature anomalies. Tropical and midlatitude sea surface temperature anomalies affect the atmosphere most directly through tropical diabatic heating anomalies, which are associated with vertical circulations as well as Rossby wave trains. Although not tested here, the Rossby wave generation mechanism has greater potential as a modulator of Texas droughts, since horizontal wind patterns exert a larger influence on CIN than does vertical motion. The key role of horizontal trajectories rather than vertical motion may be evidence against an important role for surface albedo variations. The formation of an upper-level anticyclone over Texas encourages the transport of diabatically heated air from high-terrain areas to the vicinity of Texas, producing a capping inversion whose associated CIN inhibits precipitation and leads to precipitation deficits. Future research on the prediction of summertime drought should focus on processes controlling the formation and persistence of this upper-tropospheric anticyclone.

7. Summary and conclusions

This study investigated the modulation of convective instability on summertime precipitation in Texas and the significant processes controlling convective instability. In Part I, the monthly mean precipitation during the summertime in Texas was found to be primarily modified by CIN rather than by CAPE or PW and the temperature at 700 hPa and the surface dewpoint account for most of the variability in CIN and the precipitation. While surface variables including surface dewpoint were greatly affected by soil moisture, this study examines the dynamic and physical processes controlling the temperature at 700 hPa and elucidates the large-scale influence on the convective instability and precipitation, integrating the principal processes revealed in both Parts I and II.

Back-trajectory analysis indicates that a significant contributor to warming at 700 hPa is the inversion caused by warm air transport from the Rocky Mountains and Mexican Plateau where the surface potential temperature is greater than 307.5 K. This also increases the surface temperature by the entrainment of warm air into the PBL. These effects of warm air transport induce an increase in CIN and a reduction in the precipitation despite the climatological presence of substantial low-level moisture and CAPE.

The transport of diabatically heated warm air from the high terrain is associated with enhanced descent and an upper-level anticyclonic circulation in the south-central and southwestern United States. It seems that, at 700 hPa, an anticyclonic flow impedes the climatological southeasterly flow, and then air transport from the west is enhanced, which results in the negative relationship between VOR_{500} and RDHP. While the physical reasons why downward motion is enhanced in Texas when anticyclonic circulation prevails in the south-central United States are unclear, the pattern is locally consistent with a contribution due to the dynamical forcing of the vertical motion.

Upper-level anticyclonic circulations in the southern United States strongly affect Texas's summertime precipitation by modulating the principal processes as follows. Enhanced warm air transport from the high terrain reduces the rainfall by increasing CIN. The existence of an upper-level anticyclone tends to inhibit convection by suppressing the occurrence of disturbances. The resulting reduced precipitation decreases the soil moisture. Reduced soil moisture significantly modulates the surface temperature and dewpoint and then elevates CIN. The aforementioned chain reactions of upper-level anticyclone influences are well understood within the context of CIN.

The statistically significant and tight linkages among OMG, VOR_{500}, SM, and PRCP include the key characteristics of the 1980, 1988, and 1998 droughts in surrounding states: strong upper-tropospheric anticyclone, enhanced subsidence, and dry soil. A monthly time scale was found to be useful for representing important processes such as the influence of precipitation on soil moisture, the feedback of soil moisture on precipitation, upper-tropospheric large-scale influences, and land–atmosphere interactions that cannot be characterized on either seasonal or weekly time scales. This study succeeded in verifying the detailed processes, resulting in precipitation deficits and drought quantitatively within the context of the convective instability.

While no critically direct impacts of subsidence on the precipitation deficit were found, this study emphasizes the significant role of warm air transport from the Mexican Plateau and the Rocky Mountains associated with an upper-level anticyclone in modulating the convective instability and convective precipitation in Texas. Although this study focused on the convective precipitation processes in Texas, the impacts of EMLs on convection emphasized in this study are not limited to Texas and can be applied to the south-central United States and the southern Great Plains (Lanicci and Warner 1991). In NOAA (2001), it was reported that Texas and the adjacent areas have had a higher frequency of droughts than other states in the western United States, northern plains, and southeastern United States during the 15-yr period from 1985 to 2000. The more frequent occurrence of drought in Texas may be attributed to the influence of the warm air over the high terrain. Therefore, attempts to model or predict summer precipitation and drought in the south-central United States must adequately represent the transport of diabatically heated air from the high terrain.

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