

Influences of Desert Afforestation on Boundary Layer Convergence Lines and Related Convection and Convective Precipitation over a Desert–Oasis Border

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ABSTRACT: Land-use and land-cover changes significantly influence weather and climate by altering surface albedo, roughness, heat flux, and boundary layer dynamics. These changes can modify convection and precipitation through shifts in near-surface temperature, moisture, and atmospheric instability. Vegetation heterogeneity across a desert–oasis divide has been observed to generate boundary layer convergence lines (BLCLs) and initiate convection. Afforestation may affect these behaviors through land–air interaction and boundary layer processes. Since 1986, the Hetao area of the Yellow River in northern China, where an irrigated oasis is surrounded by extensive deserts, has undergone afforestation for over 30 years. This study examines how this desert afforestation may affect BLCLs and their associated convection and precipitation through ensemble simulation by perturbing a composite of 34 cases in the synoptic pattern that has the most frequent formation of BLCLs and their associated convection initiation (CI). The results indicate that the afforestation weakens the desert–oasis contrast in 2-m air temperature and moisture, causing BLCLs to form later, attain maximum length later, and weaken. Although delaying and weakening convective triggering, the afforestation increases the near-surface water vapor content and vertical transport of water vapor, and thus the resultant larger instability, which accelerates CI, intensifies and spatially broadens convection, and causes more frequent and heavier convective precipitation in the Hetao area. This study suggests that afforestation can enhance convection and precipitation at the desert–oasis boundary even though it seems to do the opposite by delaying and weakening convection triggers. These findings have important implications for weather forecasting, land management, and climate adaptation in arid and semiarid regions.

SIGNIFICANCE STATEMENT: Afforestation in the Hetao region of the Yellow River has significantly altered local atmospheric processes by reducing the desert–oasis contrast in surface air temperature and moisture. This study reveals a dual effect: Afforestation delays and weakens boundary layer convergence lines (BLCLs), which typically act as triggers for convection and whose delay is expected to have a negative impact on convection initiation (CI). However, afforestation paradoxically accelerates CI by enhancing near-surface moisture content and vertical transport. This improved understanding of the complex impacts of afforestation on CI offers valuable insights for accurate predictions of convective weather in heterogeneous vegetated regions, supporting regional climate adaptation strategies and sustainable land management in arid and semiarid environments.

KEYWORDS: Convection; Vegetation-atmosphere interactions; Precipitation; Secondary circulation; Moisture/moisture budget; Desert meteorology

1. Introduction

Land-use and land-cover changes have important impacts on weather and climate from local to global scales through changing surface albedo, surface roughness, the partition of energy flux, and planetary boundary layer features (Raddatz 2007). The associated complex land–atmosphere interaction may alter convection and precipitation through changing near-surface temperature, moisture, and instability (Chen et al. 2020). For example, based on large-eddy simulation (LES), Avissar and Schmidt (1998) found that regions with strong thermal heterogeneity in the underlying surface exhibit significant cloud formation potential.

Over heterogeneous land cover with weak environmental wind, strong gradients in soil temperature and moisture may

cause the formation of thermal circulations with the updraft branch over the dry side and the downdraft branch over the wet side (Seth and Giorgi 1996; Taylor et al. 2011). The updraft branch may initiate deep convection when moisture and instability conditions are favorable (Zhang and Klein 2010; Kang and Bryan 2011; Taylor et al. 2011; Kang and Ryu 2016). For example, Taylor and Ellis (2006) found that in the Sahara Desert, cold cloud activity is significantly suppressed over moist soil during the afternoon and evening compared to adjacent dry areas. Emanuel (2023) proposed that strong convection may not form on excessively humid soils, as increased soil moisture can enhance convective inhibition (CIN), thereby offsetting the increase in convective available potential energy (CAPE) and suppressing the formation of strong convection. Previous studies show that the behavior of the thermal circulation and its associated convection are closely related to the magnitude of the gradient in soil and air temperature (Chen et al. 2005) and moisture (Taylor et al. 2007) across the border of

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different land covers, the size of inhomogeneity (Courault et al. 2007), and environmental wind (Lee et al. 2019).

Regarding the impact of land-use and land-cover changes on weather and climate, many efforts have been focused on deforestation in the Amazon (Shukla et al. 1990) and overgrazing in the Sahel (Charney 1975; Xue and Shukla 1993, 1996), among other hotspot regions. It is widely believed that reduced vegetation cover decreases regional precipitation by increasing surface albedo and low-level static stability, as well as inhibiting evapotranspiration and near-surface relative humidity. For example, through numerical simulation, Xue (1996) showed that deforestation in Inner Mongolia, China, can reduce precipitation and increase temperature, which is consistent with observations. Collow et al. (2014) conducted land surface sensitivity experiments over the Great Plains of the United States using the Weather Research and Forecasting (WRF) Model, revealing that complete deforestation substantially reduces convective precipitation. However, Yoon and Hohenegger (2025) simulated a scenario of complete deforestation of the Amazon rainforest over a 3-yr period. They found no significant reduction in the average annual precipitation in the Amazon rainforest, as changes in the 700-hPa circulation and associated moisture convergence compensated for the reduction in evapotranspiration.

Compared to the impact of deforestation, studies on the impact of afforestation on weather and climate are relatively limited. Since afforestation could be an important option for reducing the negative effects of anthropogenic climate change (Wulfmeyer et al. 2014), its impacts on weather and climate are drawing increasing attention. Increased precipitation by afforestation has been observed in uniform arid areas such as Israel (Alpert and Mandel 1986; Otterman et al. 1990) and the Sahel (Notaro et al. 2010). Similar features were also noticed in northern China (Zhao and Zeng 2002) and the Great Plains in the United States (Collow et al. 2014). However, the impacts of afforestation on thermal circulation and their associated convection over heterogeneous vegetation regions have received much less attention.

Afforestation may alter the temperature and moisture gradient in heterogeneous vegetation regions and subsequently affect the characteristics of the associated thermal circulation and convection initiation (CI). However, conflicting results have been found regarding whether denser vegetation promotes or inhibits convection. On the one hand, Klein et al. (2017) found that areas with more vegetation coverage impede daytime deep convection due to the diffusion of moist air and lower sensible heat flux in West Africa. On the other hand, Anthes (1984) found that planting bands of vegetation with widths of about 50–100 km in semiarid areas could lead to more convective precipitation under favorable large-scale atmospheric conditions. Considering substantial revegetation efforts have been made around the world in restoring degraded desert ecosystems, exploring how such efforts may potentially change local moisture, convection, and precipitation is important for understanding vegetation–precipitation feedback and essential for environmental management and sustainable development in arid and semiarid regions. This study aims to explore how a 30-yr-long real-world desert afforestation in

northern China may modify the thermal circulation and the associated convection.

Located in the arid and semiarid region of northern China, where the response of vegetation to precipitation is most pronounced (Fensholt et al. 2012; Yang et al. 2015; Chen et al. 2020; Zhang et al. 2022), the Hetao area in Inner Mongolia features a vegetation heterogeneity where the largest one-head irrigation oasis in Asia is adjacent to widespread deserts (Fig. 1a).

Along the desert–oasis border in the Hetao area, boundary layer convergence lines (BLCLs) have been observed as a radar fine line in warm seasons. BLCLs represent a narrow zone of low-level wind convergence associated with the updraft branch of the thermal circulation. It is defined on radar as a combination of the two lowest elevation angles (0.5° and 1.5°) with a linear echo reflectivity ranging from -5 to 10 dBZ, where the width is less than 10 km and the length is greater than 10 km (Wilson and Schreiber 1986; Koch and Ray 1997; Huang et al. 2019). Huang et al. (2019) showed that approximately 60 BLCLs form in the Hetao area from June to August each year with 44% providing sufficient uplift for CI. Indeed, most of the lightning around the desert–oasis border in the Hetao area is associated with the BLCLs (Meng et al. 2024). BLCLs generally form after the early morning peaking at 1000–1100 LST, attaining their maximum length, and beginning to initiate convection about 2 h later in the early afternoon (Huang et al. 2019). Notably, convective precipitation associated with BLCLs in this area primarily concentrates in the Kubuqi Desert–Ordos Plateau transition zone, gradually expanding northwestward with the expansion of the grassland area into the desert area from June to August (Huang et al. 2019), which provides a more widespread water vapor supply. This expansion underscores the significant role of vegetation in convective precipitation.

Since 1986, as part of the Yellow River basin ecological protection strategy, the Hetao area has undergone over 30 years of afforestation. After the treatment of the desert area, there have been significant changes in vegetation distribution with large areas of afforestation over the Kubuqi and Ulan Buh deserts (Fig. 1b). According to Gai (2022), the Kubuqi Desert, with a total area of approximately $13\,900\text{ km}^2$, has seen its vegetation coverage increase from 16.2% in 2002 to 53% in 2021. This study aims to explore the impacts of the desert afforestation in the Hetao area on the behaviors of BLCLs and their associated convection and precipitation.

2. Data and methods

a. Data

The data utilized in this study encompass precipitation from satellite and automatic weather station (AWS), reanalysis data, topography, and land-use data. The satellite data are the Integrated Multi-satellitE Retrievals for Global Precipitation Measurement (GPM) (IMERG) (Huffman et al. 2023), covering the time period from June to August from 1998 to 2022. The spatial resolution of GPM-IMERG data is $0.1^\circ \times 0.1^\circ$, with the time resolution of half an hour. The hourly

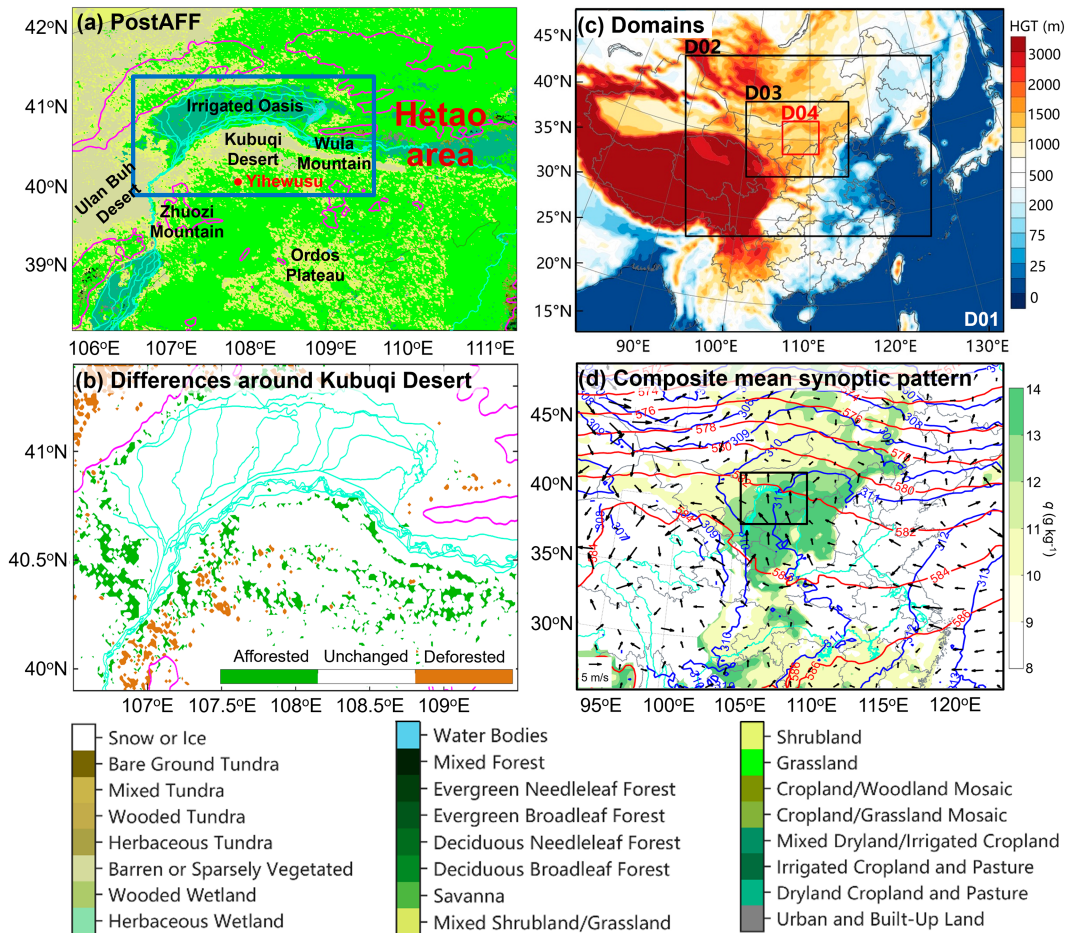


FIG. 1. (a) Land use after afforestation in 2019 over D04 in (c). The blue rectangle in (a) represents the main region affected by convection in the Hetao area. (b) Land-use changes from 1985 to 2019 (shaded) around the Kubuqi Desert over the domain denoted by the blue rectangle in (a). (c) Locations of WRF simulation domains with terrain height above sea level (m; shaded). (d) Composite mean synoptic pattern including 700-hPa specific humidity (shaded; g kg^{-1}), 700-hPa geopotential height (blue lines; dagpm), 500-hPa geopotential height (red lines; dagpm), and surface wind (black vectors) over D02 in (c). The black rectangle in (d) represents the range of D04. The pink lines in (a) and (b) represent topography at 1500 m above sea level, and the cyan lines in (a), (b), and (d) represent the Yellow River and irrigation channels.

precipitation observations at Yihewusu AWS (location given by the red dot in Fig. 1a) in the Kubuqi Desert are provided by the China Meteorological Administration, covering 1973–77, 2007–09, and 2011–22. The hourly precipitation data from 1978 to 2006 and 2010 were unfortunately missing. The period of hourly precipitation denotes the rainfall accumulation in the hour starting from the given time.

ERA5 reanalysis data (Hersbach et al. 2021a,b) from the European Centre for Medium-Range Weather Forecasts (ECMWF), with the spatial resolution of $0.25^\circ \times 0.25^\circ$ and the temporal resolution of 1 h, were used to analyze the synoptic environment and drive the WRF Model. The ETOP01 topography data (1-arc-min grid) from the National Geophysical Data Center (NGDC) (NGDC/NESDIS/NOAA/U.S. Department of Commerce 2011), and the 30-m, yearly land-use dataset in China (Fig. 1a) from Yang and Huang (2021), were employed in WRF

simulations. The land-use maps were produced on the basis of the topography, spectral properties, and seasonal variations of vegetation and hydrological characteristics.

b. Methods

Numerical simulations were performed using WRF (ARW) V3.7.1 (Skamarock 2008). WRF Model configuration involves four one-way nested domains with spatial resolutions of 27, 9, 3, and 1 km from the outermost to the innermost (Fig. 1c), each comprising 40 vertical layers with a model top at 50 hPa and the model time step of 45 s. The height of the bottom model layer is approximately 20 m above the ground. The physical parameterization schemes include Morrison double-moment microphysics scheme (Morrison et al. 2009), CAM longwave and shortwave schemes (Collins 2004), Pleim–Xiu

surface layer scheme (Pleim 2006) and land surface scheme (Xiu and Pleim 2001), Asymmetric Convective Model, version 2 (ACM2), planetary boundary layer scheme (Pleim 2007) for all domains, and Grell–Freitas (GF) cumulus parameterization scheme (Grell and Freitas 2014) for the outermost domains (D01 and D02). No cumulus parameterization schemes are used for the innermost domains (D03 and D04).

The experiments include two sets of ensemble simulations with an ensemble size of 30. Experiment PostAFF was performed using the land-use data in 2019 after afforestation (Fig. 1a), while experiment PreAFF was performed using exactly the same initial and boundary perturbations as the PostAFF except replacing the greened vegetation in the area covered by D04 with the preafforestation state (i.e., the vegetation in 1985) for all four domains. The vegetation differences between 2019 and 1985 were defined as follows: Considering that the afforestation changes correspond to decreased skin temperature, we determined the afforestation grid points by combining changes in both the land-use types and simulated skin temperature. Based on the deterministic simulations initiated from the unperturbed composite fields with the vegetation in 2019 and 1985, we mapped the changes in land-use type to changes in skin temperature at 1400 LST. We marked a grid point as “Afforested” if there is a change in land use at the grid point, and the changed land use corresponds to a lower skin temperature in 2019 relative to that in 1985. Nearly 99% of the Afforested grids consist of 1) “barren or sparsely vegetated” transformed to “shrubland,” “grassland,” or “dryland cropland and pasture”; 2) shrubland transformed to grassland or dryland cropland and pasture; and 3) grassland transformed to dryland cropland and pasture. “Deforested” was determined in the opposite way.

Since this study aims to examine the impact of real-world afforestation on BLCLs and their associated CI, the ensemble simulations were designed in a quasi-idealized scenario. First, we generated a composite of BLCL-related convection cases under the synoptic pattern (T1) that has the most frequent BLCLs and their associated CIs in the Hetao area, based on Huang et al. (2022). To understand the environment of BLCLs in the Hetao area, Huang et al. (2022) classified the synoptic patterns into five categories, labeled from T1 to T5, based on the spatial distribution of mean sea level pressure (MSLP) at 0800 LST (after which BLCLs begin to form substantially, LST = UTC + 8 h) during the 2012–16 summers [Fig. 1 in Huang et al. (2022)]. T1 is characterized by a high pressure ridge at the surface, higher near-surface temperature and thus higher CAPE and temperature gradients across desert and oasis, and low-level southerly winds over the Kubuqi Desert with rich moisture supply, enhancing the near-surface convergence with northerly oasis winds. Among all the BLCL-related CI cases during the 2012–16 summers, 34 of them occurred under the T1 synoptic pattern (Fig. 1d).

The composite of the 34 cases was produced by averaging hourly ERA5 analyses from 2000 LST the day before each case to 2000 LST the day of the case. The integration time for D01 is from 2000 LST the previous day to 2000 LST the day of interest, while for D02, D03, and D04, it runs from 0200 to 2000 LST the day of interest. Then, we used the composite of

the 34 BLCL-CI cases to represent the most common synoptic pattern, perturbed it by a magnitude that guaranteed all 30 members captured the BLCL-related CIs. The initial and boundary conditions of the two ensembles were generated by perturbing the composite mean atmospheric conditions at the initial time in D01. These perturbations represent random sampling of the climatologically based background error (BE) covariance using the default control variables [CV option 3 (CV3)] from the WRF data assimilation (WRFDA) package (Barker et al. 2004). CV3 is a method for generating BE based on the statistical differences of forecasts with different lead times. Ensembles are generated by perturbing control variables streamfunction ψ , unbalanced velocity potential χ_u , unbalanced temperature T_u , specific humidity q , and unbalanced surface pressure $P_{s,u}$. These perturbations are then transformed into physical variables u , v , t , q , and p . Averaged over all grids, the standard deviations of zonal wind, meridional wind, temperature, specific humidity, and pressure are 0.6 m s^{-1} , 0.6 m s^{-1} , 0.3 K , 0.2 g kg^{-1} , and 23 hPa , respectively. Under these perturbations, all the 30 members captured BLCLs and their associated convection, guaranteeing not only the representativeness of the ensemble for the targeted convective BLCL processes but also the robustness of the results with certain uncertainties in the behaviors of BLCLs and the initiated convections as well as the convection-related parameters and water vapor transport processes. A larger perturbation would either make BLCL or CI fail to occur.

3. Results

a. Observations of convective precipitation, BLCL, and convection before and after afforestation

The areal-mean hourly convective precipitation was calculated using GPM-IMERG data from 1998 to 2022 over the region denoted by the blue rectangle in Fig. 1a, which is the main area affected by convection in the Hetao area. Increasing trends are observed in both magnitude and frequency at the 70% confidence level (Fig. 2a). Here, convective precipitation is identified when hourly accumulated precipitation exceeds 6 mm, according to the research on precipitation in Inner Mongolia, China, by Chang (2015). Consistent increase is also indicated in the observations from Yihewusu AWS in the Kubuqi Desert from 1973 to 2022, in the frequency (Figs. 2b,c, counted in the hour starting from the given time) and magnitude of hourly convective precipitation (Figs. 2b,d), although the trends are not statistically significant. Since the GPM-IMERG data used in this study represent spatiotemporal averages, they are affected by nonprecipitating periods and grids. As a result, the convective precipitation magnitude and frequency from the GPM-IMERG data (Fig. 2a) are nearly two orders of magnitude lower than those from the AWS (Fig. 2b). Note that climate and vegetation changes in the Hetao area may both contribute to the increased precipitation. Global warming has been demonstrated by previous studies to make northern China warmer and wetter (Wu et al. 2020; Aihaiti et al. 2021), leading to increased precipitation (Sun et al. 2018; Zhou et al. 2019). The following sections will examine the potential role

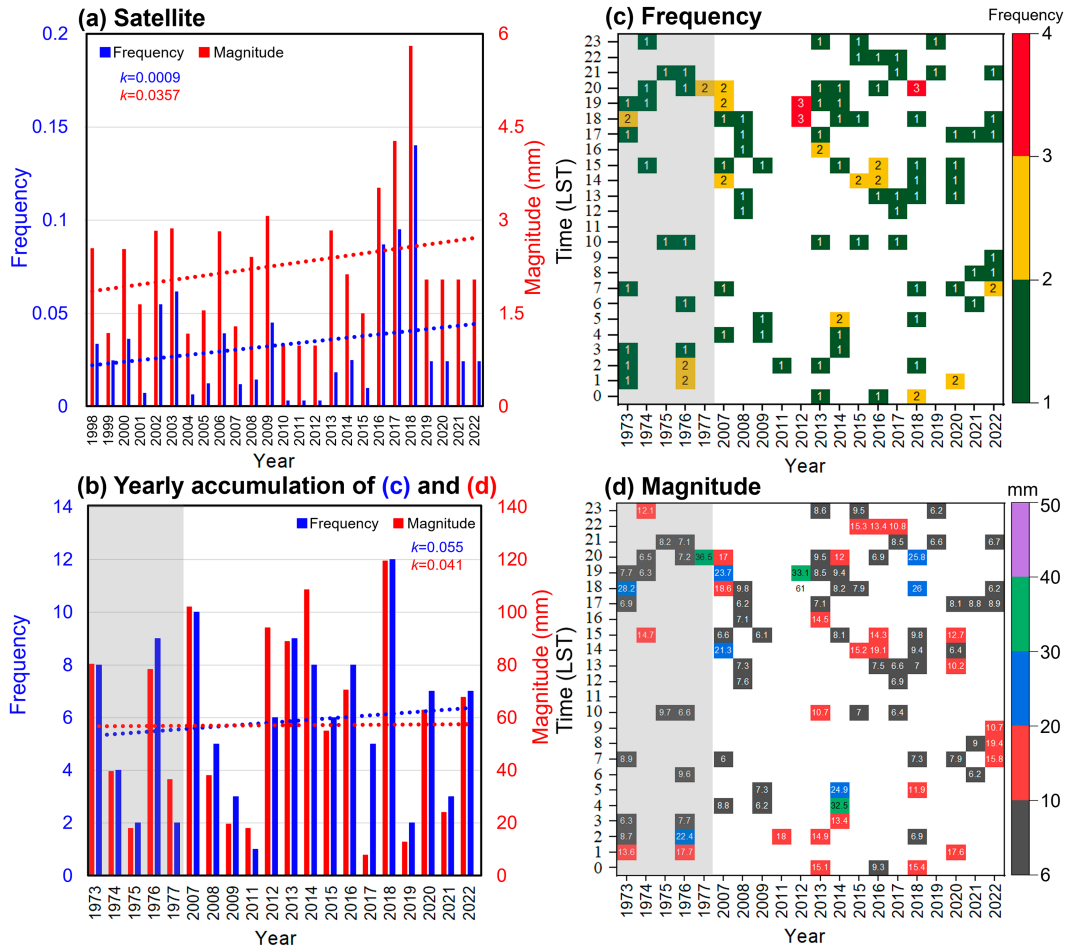


FIG. 2. (a) Satellite data of yearly accumulations of the frequency and magnitude of hourly convective precipitation averaged over the area denoted by the blue rectangle in Fig. 1a. Also given are the statistics of observed hourly convective precipitation at Yihewusu AWS (location given in Fig. 1a) in the Kubuqi Desert including (b) the yearly accumulations of the frequency and magnitude and the time–year variations of the (c) frequency and (d) magnitude. The area shaded in gray (i.e., before 1985) represents data before afforestation. For (a), the whole period of time is after 1985. The term k in (a) and (b) represents the slope of the fitted line (dashed lines).

that afforestation may have played in these variations of convective precipitation and the physical mechanism.

In the Hetao area, convective precipitation is primarily driven by BLCL-associated convection. During June–August from 2012 to 2016, radar observations identified 34 BLCL-associated convective events. Statistical analysis shows that BLCL formation in this region generally occurs between 1030 and 1100 LST, with the BLCL reaching its maximum length mainly between 1330 and 1400 LST (Fig. 3a). In this work, CI is defined as the time when the composite radar reflectivity first reaches 30 dBZ. Convection dissipation is defined as the time when the 30-dBZ composite radar reflectivity disappears. Convection duration is defined as the time span from the first appearance of the 30-dBZ composite radar reflectivity to its disappearance. Results show that CI most frequently takes place between 1330 and 1400 LST, while peak convective intensity is most common between 1430–1500 and 1700–1730 LST. Convection typically dissipates between 1700 and 1730 LST (Fig. 3b). The

lifetimes of convective events ranged from 0.5–1 to 4–4.5 h (Fig. 3c).

b. Simulated BLCL and CI behaviors with postafforestation vegetation

PostAFF well captured the behaviors of BLCLs and their associated convection. In our simulation, BLCL is defined as a linear convergence zone with the horizontal wind divergence at the bottom of the model less than $-0.4 \times 10^{-3} \text{ s}^{-1}$, with a width less than 10 km and a length greater than 10 km. Relative to the timing of the 34 observed cases (Fig. 3), PostAFF shows an approximately 2-h later peak in BLCL formation, the attainment of maximum length, and the timing of CI and the point of strongest intensity (Fig. 4). PostAFF results show that BLCLs mainly form at 1300 and 1310 LST and attain their maximum length at 1410 LST (Fig. 4a). Convection mainly initiates at 1520 LST, attains maximum intensity at 1830–1850 LST, and dissipates after 2000 LST

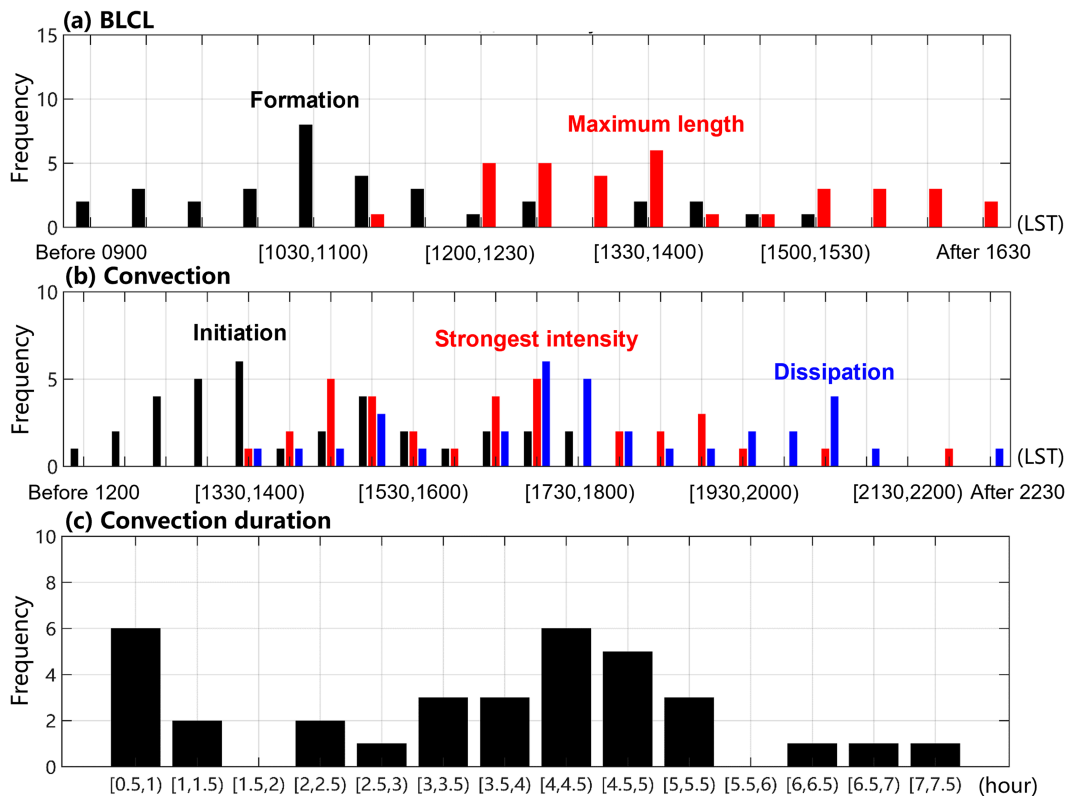


FIG. 3. Radar statistics of (a) frequency–time variations of the 34 observed BLCL formation (black) and the attainment of maximum length (red) from June to August 2012 to 2016, under T1 pattern. (b) As in (a), but for CI (black), the point of maximum intensity (red), and dissipation (blue). (c) Convection duration (black) of the 34 cases.

(Fig. 4b). The duration of convection ranges 3–5 h with the highest frequency occurring for durations longer than 4.5 h (Fig. 4c). The timing differences between the simulated and observed BLCLs and convection characteristics may be because the composite simulation tends to smooth variations among individual events. Similar features were also observed in Grimit and Mass (2002) and Lojko et al. (2025), where composite fields by averaging over multiple cases exhibit weaker gradients and later peaks.

We used the deterministic simulation initiated from the unperturbed composite field of PostAFF to show the detailed structural features of BLCL and convection together with the ensemble spread of BLCL and convection (Fig. 5). The results show BLCLs (Fig. 5a) and CI (pink points at the southwest corner of Fig. 5b) first occur north of the Zhuozi Mountain. Later on, BLCLs form and develop north of the Kubeqi Desert (shown by the orange rectangle in Fig. 5b) around 1350 LST (Fig. 5b), which initiates convection at around 1530 LST (shown by the orange rectangle in Fig. 5c). However, the convection north of the Kubeqi Desert becomes much stronger than the convection initiated near Zhuozi Mountain (Fig. 5d). In addition to the above two convective systems, a third convective system initiates before 1530 LST and develops in the southeast of the Kubeqi Desert over the BLCL along the southeast desert–oasis border (Figs. 5c,d).

Considering this work is focused on the effects of the thermal contrast between desert and oasis on BLCLs and their associated convection, the BLCL and convection generated north of the Kubeqi Desert, where the contrast in vegetation is most pronounced, will be the target of analysis in the following sections. The convection on the western side is mainly influenced by Zhuozi Mountain, while the convection on the eastern side is too far from the oasis, and thus, they are not quite relevant.

c. Effects of afforestation on BLCL

A comparison between the simulations of PostAFF and PreAFF shows that afforestation delays the time when the BLCL forms and attains the maximum length (Fig. 4a). The peak time of BLCL formation is at 1250 LST for PreAFF, which is 10–20 min earlier than PostAFF. Similarly, the peak time of the BLCLs attaining their maximum length is at 1350 LST in the PreAFF, which is approximately 20 min earlier than that in the PostAFF. According to Cohen (1988), a timing shift that is comparable to the standard deviation of BLCL initiation and maximum length timing) can be considered meaningful.

The afforestation not only delays the formation time of the BLCL but also influences its intensity. BLCL strength is defined as horizontal wind divergence averaged at the bottom

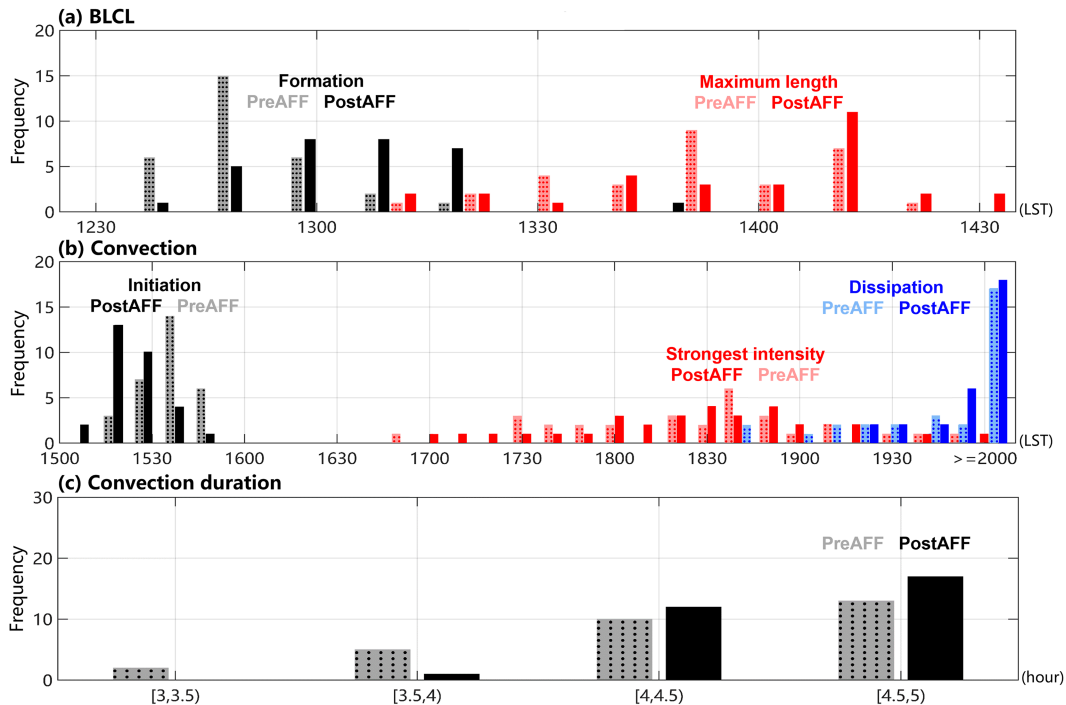


FIG. 4. (a) Simulated frequency–time variations of BLCL formation (black and stippled gray) and the attainment of maximum length (red and stippled light red). (b) As in (a), but for CI (black and stippled gray), the point of maximum intensity (red and stippled light red), and dissipation (blue and stippled light blue). (c) Convection duration (black and stippled gray). PostAFF is denoted in dark color, while PreAFF is denoted in stippled light color.

layer of the model over the BLCL region before CI. The ensemble mean and spread of BLCL strength show that the BLCLs in PreAFF are stronger than those in PostAFF by about $0.5 \times 10^{-4} \text{ s}^{-1}$ (around 10%) from 1240 to 1500 LST (Fig. 6a). This difference passes the two-sample t test at the 90% confidence level.

The impact of afforestation on the formation and mature times and intensity of BLCL is possibly due to the decreased gradient in 2-m air temperature. BLCL tends to extend along the area with the highest gradient in 2-m air temperature and moisture (Liu et al. 2023). Afforestation causes a consistently smaller desert–oasis difference in ensemble-mean 2-m air temperature and moisture averaged in the blue ellipses in Fig. 5b, from 0700 LST to around noon (Fig. 6b). During this period of time, the differences in temperature and moisture are both between 5% and 10%, and the differences pass the two-sample t test at the 90% confidence level.

The negative impact of this real-world afforestation on BLCL is consistent with results from making more extreme land-use changes. We performed two additional sets of ensemble experiments using a larger area of afforestation in the desert (ReduceDesert) and replacing the whole oasis with the desert (RemoveOasis). These larger changes in land use both result in less sharp land-use contrast and substantially change the BLCL behaviors. In ReduceDesert, BLCLs do not form in more than half of the ensemble members across the desert–oasis border, and no BLCLs form across the desert–oasis border in the RemoveOasis. For the sake of brevity, the results of these additional tests are not shown.

d. Effects of afforestation on convection

In PostAFF, although BLCLs form later, as a trigger for convection, they initiate convection at earlier times (Fig. 4b). The peak of CI in PostAFF appears at 1520 LST, about 20 min earlier than in PreAFF. This time difference is meaningful considering that it has the same magnitude as its standard deviation in CI time (Cohen 1988). Although CI occurs earlier after afforestation, there is no apparent difference between the two experiments in terms of timing when convection attains its maximum intensity and subsequently dissipates (Fig. 4b). The convection attains its peak at 1840 LST in PreAFF and at 1830–1850 LST in PostAFF, while most convection dissipates after 2000 LST in both experiments. As a result, the afforestation extends the duration of the convection (Fig. 4c).

The earlier development of convection after afforestation is likely due to higher mixed-layer CAPE (MLCAPE) caused by more abundant water vapor (Fig. 6b). MLCAPE and mixed-layer CIN (MLCIN) were calculated by lifting air parcels from a mixed layer between the ground and 100 hPa above the ground at the CI position. The results show that before CI from 1020 to 1500 LST, the ensemble-mean MLCAPE (Fig. 7a) and column water content (CWC) in PostAFF (Fig. 7b) are consistently higher than in PreAFF, whereas the ensemble-mean MLCIN exhibits the opposite behavior. This favorable water vapor condition persists until the CI time. During this period of time, the increases in MLCAPE and CWC pass the two-sample t test at the 95%

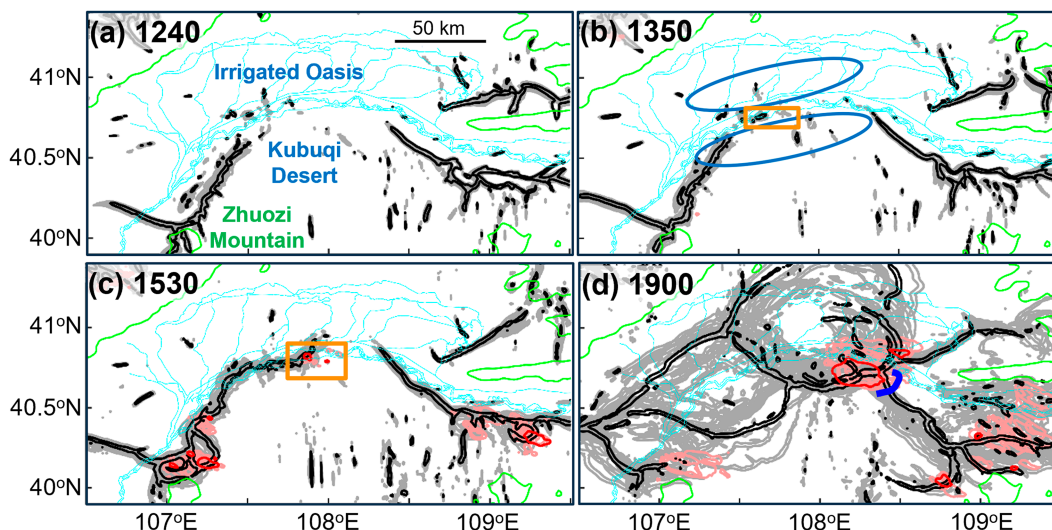


FIG. 5. Evolution of simulated BLCL and convection at (a) 1240, (b) 1350, (c) 1530, and (d) 1900 LST over the domain denoted by the blue rectangle in Fig. 1a. The green lines represent topography at 1500 m MSL. The cyan lines represent the Yellow River and irrigation channels. The gray horizontal wind divergence at the bottom of the model ($-4 \times 10^{-4} \text{ s}^{-1}$) and pink composite radar reflectivity lines (30 dBZ) represent the BLCLs and convection of each individual ensemble member, respectively. The black horizontal wind divergence at the bottom of the model ($-4 \times 10^{-4} \text{ s}^{-1}$) and red composite radar reflectivity (30 dBZ) lines represent the deterministic simulations of BLCL and convection initiated from the unperturbed composite field using the land use of PostAFF. The blue ellipses in (b) indicate the area over which the differences in 2-m air temperature and specific humidity between desert and oasis are calculated for Fig. 6b. The orange rectangles in (b) and (c) denote the location of the target BLCL and convection of this study. The thick blue curve in (d) represents the points used to calculate MLCAPE, MLCIN, CWC, MLLCL, and MLLFC in Fig. 9.

confidence level, while the difference in MLCIN is not statistically significant.

The averaged water vapor flux at the bottom layer of the model over the BLCL region reveals that although PreAFF

has stronger BLCL strength (Fig. 6a), which means stronger uplift, PostAFF has an earlier CI because of a stronger vertical transport of water vapor with higher water vapor flux after 1100 LST until CI (Fig. 7c). In PostAFF, water vapor flux can

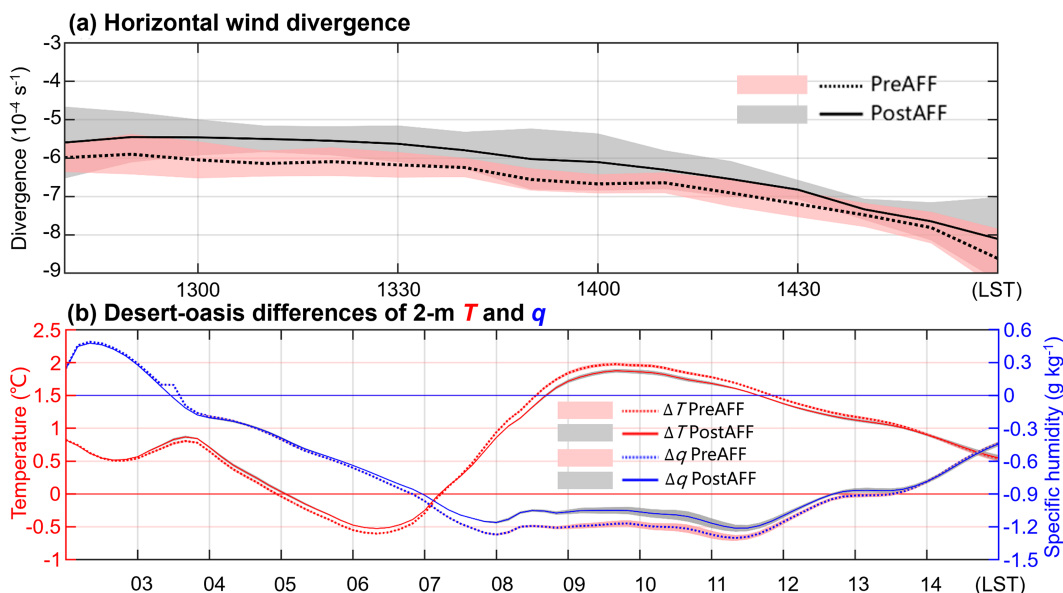


FIG. 6. Ensemble-mean (a) BLCL strength (horizontal wind divergence before CI at the bottom layer of the model averaged over the target BLCL region where the wind divergence is less than $-4 \times 10^{-4} \text{ s}^{-1}$ within the orange rectangle in Fig. 5b) and (b) desert-oasis differences in 2-m air temperature (red lines) and specific humidity (blue lines) for PreAFF (dashed lines) and PostAFF (solid lines), as well as the uncertainty in terms of the ensemble mean plus and minus one standard deviation for PreAFF (light red shading) and PostAFF (gray shading).

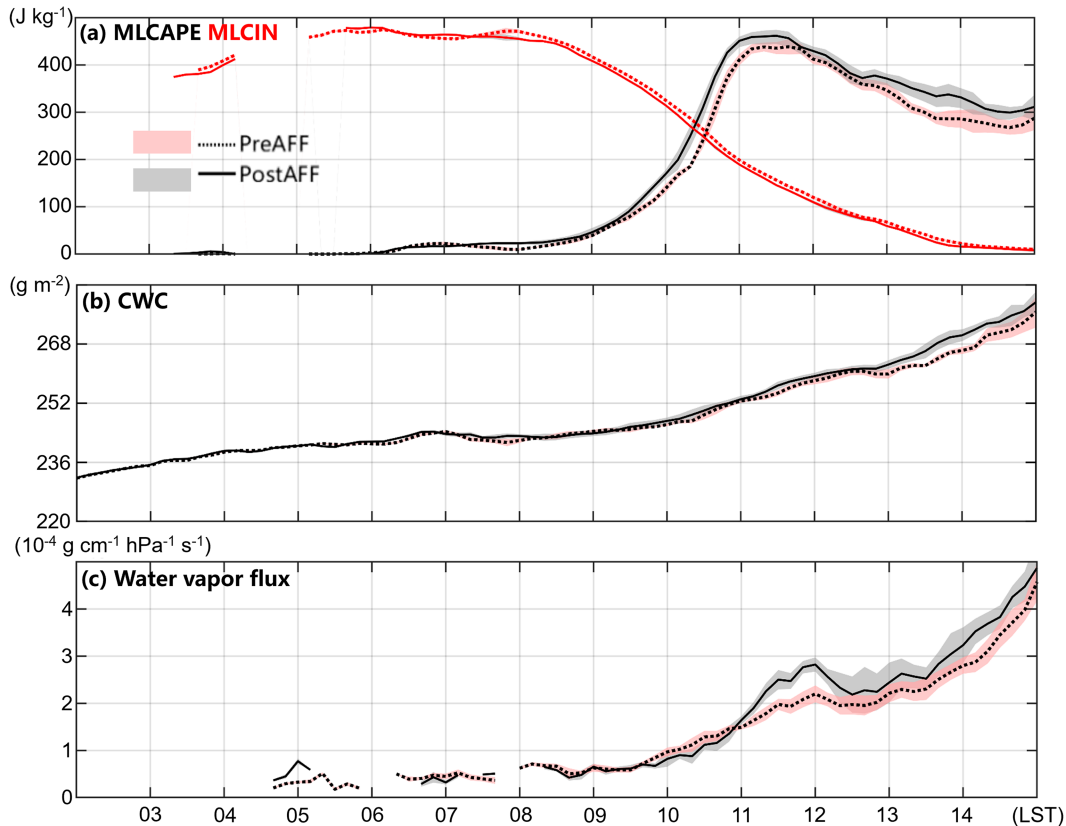


FIG. 7. Evolution of simulated (a) MLCAPE (J kg^{-1} ; black lines) and MLCIN (J kg^{-1} ; red lines) and (b) CWC (g m^{-2}) at CI location, and (c) water vapor flux ($10^{-4} \text{ g cm}^{-1} \text{ hPa}^{-1} \text{ s}^{-1}$) at the bottom layer of the model averaged over the target BLCL region where the wind divergence is less than $-4 \times 10^{-4} \text{ s}^{-1}$ within the orange rectangle in Fig. 5b. The lines denote the ensemble mean in PreAFF (dashed lines) and PostAFF (solid lines), while the shaded area denotes the uncertainty in terms of the ensemble mean plus and minus one standard deviation of the respective ensemble with light red shading for PreAFF and gray shading for PostAFF.

be $0.6 \times 10^{-4} \text{ g cm}^{-1} \text{ hPa}^{-1} \text{ s}^{-1}$ higher than PreAFF from 1100 to 1200 LST, with the increase of nearly 50%. This increase passes the two-sample t test at the 90% confidence level.

Consistent with the earlier convection development, afforestation also intensifies and expands the convection. In PostAFF, when convection attains maximum intensity, the area where composite reflectivities are greater than 30 dBZ is significantly larger (Fig. 8a) and the maximum convection intensity is slightly higher (Fig. 8b), as evidenced by the ensemble mean. The stronger convection after afforestation is possibly due to the greater MLCAPE (Fig. 9a) averaged over an arc (denoted in Fig. 5d) approximately 2 km from the leading edge of the storm (in terms of the 30-dBZ contour of the convection, and the arc in PreAFF is in a similar relative position to the leading edge of convection). Figure 9a shows that ensemble-mean MLCAPE in PostAFF is consistently higher than in PreAFF, with a maximum difference of more than 80 J kg^{-1} (approximately 25%) during the whole life period of the convection, while ensemble-mean MLCIN is not significantly different in the two experiments (Fig. 9b). Besides,

Figs. 9c and 9d show that the ensemble means of mixed-layer lifted condensation level (MLLCL) and mixed-layer level of free convection (MLLFC) averaged over the same arc as MLCAPE and MLCIN are consistently lower in PostAFF by about 2 and 5 hPa, respectively, which is also conducive to the development of convection under a later BLCL formation and weaker lifting condition. Except for MLCIN, the differences in other convective parameters in Fig. 9 all pass the two-sample t test at the 90% confidence level.

The larger instability in PostAFF relative to PreAFF may have resulted from increased water vapor transport, in particular, the vertical water vapor transport due to enhanced transpiration from additional vegetation following afforestation. We conducted a water vapor budget analysis using the method following Qi et al. (2021) over the rectangular area in Fig. 1a from 700 to 200 hPa, considering that the horizontal transport of water vapor occurs mainly at 700 hPa, and there is almost no water vapor above 200 hPa (Fig. 10a). The total QT_t , horizontal QT_h , and vertical QT_v water vapor transport were calculated.

The ensemble means show that QT_t is larger in PostAFF than in PreAFF in the early development stage of the

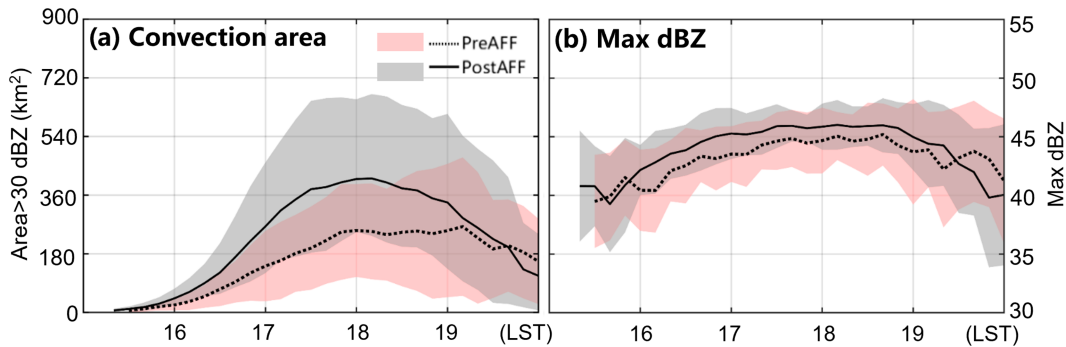


FIG. 8. Evolution of ensemble-mean (a) area of >30 dBZ in terms of composite reflectivities (dBZ) and (b) maximum composite reflectivities (dBZ) of the target convection in PreAFF (dashed line) and PostAFF (solid line). The shaded areas denote the uncertainty in terms of the ensemble mean plus and minus one standard deviation of the respective ensemble with gray shading for PostAFF and light red shading for PreAFF.

convection from 1630 to 1930 LST, with a maximum difference of about $1 \times 10^7 \text{ g s}^{-1}$ around 1700 LST (Fig. 9e). More importantly, by dividing the total water vapor budget into horizontal and vertical components, we found that the trend and magnitude of QT_v are the closest to those of QT_h , suggesting that QT_v is the main contributor. In particular, QT_h does not differ much between PostAFF and PreAFF before 1700 LST (Fig. 9e), while QT_v is higher in PostAFF than in PreAFF from 1630 to 1730 LST and remains consistently positive, reaching a maximum difference close to $1 \times 10^7 \text{ g s}^{-1}$ (Fig. 9f). This result indicates that during the development stage of the convection, the increasing water vapor that contributes to the stronger convection in PostAFF primarily comes from the upward transport. This larger upward transport is possibly the result of increasing transpiration from additional vegetation following afforestation. This is confirmed by the consistently higher ensemble-mean latent heat (LH) flux in PostAFF than in PreAFF throughout the whole simulation period from 0200 to 2000 LST (Fig. 10b), which is averaged over the same region as for water vapor budget analyses.

e. Effects of afforestation on convective precipitation

Possibly due to the stronger and larger convection, the afforestation increases the total precipitation and especially the convective precipitation. Figure 11a shows the spatial distribution of significant differences in convective precipitation between the PostAFF and PreAFF during the entire integration period, with no-rain hours in both experiments excluded. The regions with significant changes (shaded areas, at the 95% confidence level) are generally consistent with the spatial distribution of convection shown in Fig. 5.

The areal-mean hourly precipitation in the blue rectangle in Fig. 11a (exactly the same as the blue rectangle in Fig. 1a) increases by about 0.03 mm (20%) (Fig. 11b). Following Douglas et al. (2009) and Collow et al. (2014), the areal-mean calculation includes nonprecipitating regions. This increase passes the two-sample t test at the 95% confidence level. This result is different from Liu and Liu (2018) where no significant change in precipitation due to vegetation greening and

increased forest area is observed in northern China but consistent with Collow et al. (2014) where areal-mean hourly precipitation increases by 0.05–0.1 mm (about 10%–20%) after all vegetation types on the Great Plains of the United States were converted to forest, as well as Pielke (2001) and Santanello et al. (2018), which showed that in arid and semiarid regions, the vertical motion of water vapor caused by vegetation evapotranspiration indeed leads to enhanced convection and increased rainfall.

To compare with the observed trends in convective rainfall, we analyzed the magnitude and frequency of hourly convective precipitation (i.e., precipitation rate $> 6 \text{ mm h}^{-1}$, defined in the same way as precipitation observations). The frequency of the hourly convective precipitation is defined by dividing the accumulated times of convective precipitation in the area of the blue rectangle by the total number of grid points within the blue rectangle in Fig. 11a. Results show that after afforestation, the areal-mean hourly convective precipitation increases by about 0.01 mm (Fig. 11c), and the frequency of convective precipitation increases by about 0.002 (Fig. 11d), and both increase by nearly 100%. The differences pass the two-sample t test at the 95% confidence level. The simulated increasing trend aligns with that of the observations.

4. Conclusions and discussion

This study investigates the impacts of more than 30 years of afforestation on BLCLs and the associated convection and precipitation in deserts adjacent to an irrigated oasis in semiarid Inner Mongolia, China. The analysis is based on WRF ensemble simulations conducted under a synoptic pattern that most frequently favors the formation of BLCLs and their associated CIs.

The results show that afforestation weakens the 2-m air temperature and moisture gradient between oasis and desert by about 0.1°C (5%) and 0.25 g kg^{-1} (10%), respectively, which delays BLCL formation by 10 min and maturation by 20 min as well as weakens their intensity by about $-0.5 \times 10^{-3} \text{ s}^{-1}$ (around 10%). Although the weakened BLCL provides weaker

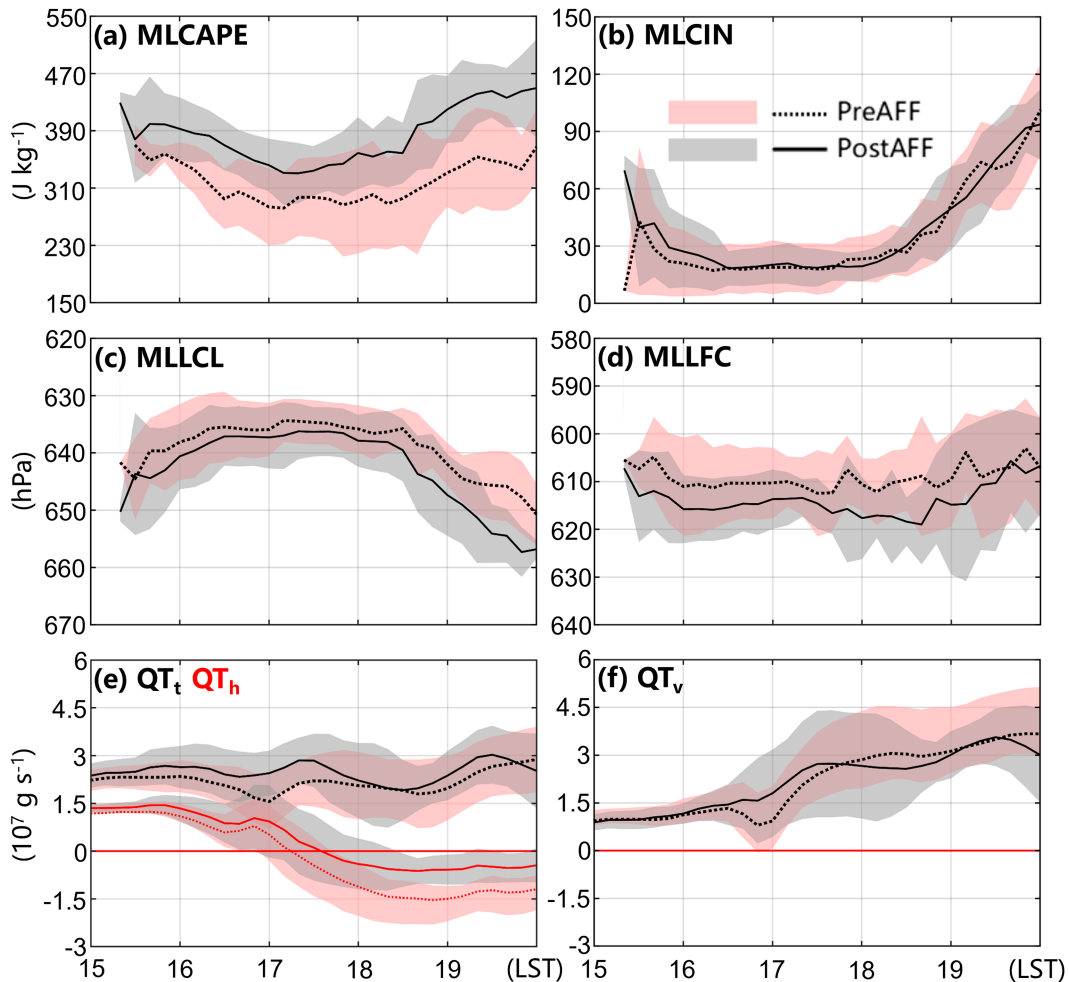


FIG. 9. (a) MLCAPE (J kg^{-1}), (b) MLCIN (J kg^{-1}), (c) MLLCL (hPa), (d) MLLFC (hPa) averaged over an arc approximately 2 km from the leading edge of the target convection, and (e) total QT_t (10^7 g s^{-1} ; black lines) and horizontal QT_h (10^7 g s^{-1} ; red lines) water vapor transport over the rectangular area in Fig. 1a from 700 to 200 hPa. (f) As in (e), but for vertical water vapor transport QT_v (10^7 g s^{-1}). The lines denote the ensemble mean in PreAFF (dashed lines) and PostAFF (solid lines), while the shaded area denotes the uncertainty in terms of the ensemble mean plus and minus one standard deviation of the respective ensemble with light red shading for PreAFF and gray shading for PostAFF.

uplift, the denser vegetation enhances evapotranspiration, which facilitates greater vertical transport of water vapor from lower levels. The increased water vapor flux over the target BLCL and CWC at the CI location then makes the convection initiate 20 min earlier. The resultant higher MLCAPE with the increase of around 80 J kg^{-1} (25%) and lower MLLCL and MLLFC by around 2 and 5 hPa, respectively, in the near-storm environment foster the development of stronger convection and increase the magnitude and frequency of hourly convective precipitation significantly by nearly 100%. Experiments with more extreme land-use modifications further confirm that extremely low desert-oasis contrast further weakens BLCLs or even prevents BLCL formation.

Relative to most idealized modeling studies in the literature, this work made the first effort in exploring the anthropogenic impact of large-scale real-world afforestation over the desert

area on the behavior of BLCLs, CIs, and convective precipitation. This study suggests that afforestation can enhance convection and precipitation at the desert-oasis boundary even though it seems to delay BLCL formation. These findings are meaningful and important not only in helping to understand the anthropogenic impact on complex interactions among land surface, boundary layer processes, and severe weather but also in offering valuable insights into climate regulation, agriculture, water resource management, ecological restoration, and sustainable development in arid and semiarid regions.

Finally, it is important to note that the results were derived from a composite of 34 BLCL-related convection cases under the synoptic pattern that has the most frequent BLCLs and their associated CIs. Results may differ under other synoptic patterns, which warrant further investigation.

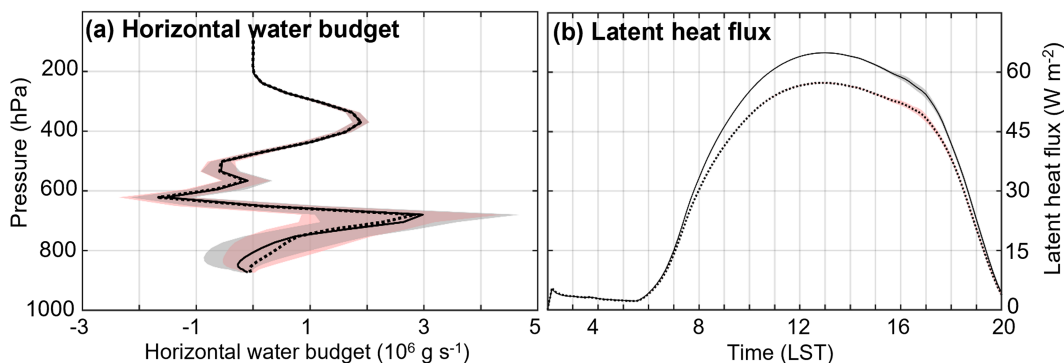


FIG. 10. (a) Vertical profile of ensemble-mean horizontal water vapor budget (10^6 g s^{-1}) over the rectangle area in Fig. 1a at 1400 LST of PreAFF (dashed lines) and PostAFF (solid lines), and (b) evolution of ensemble-mean LH flux (W m^{-2}) averaged over the rectangle area in Fig. 1a, in PreAFF (dashed line) and PostAFF (solid line). The shaded areas denote the uncertainty in terms of the ensemble mean plus and minus one standard deviation of the respective ensemble with gray shading for PostAFF and light red shading for PreAFF.

On the other hand, the increase in precipitation in the Hetao area over the past 30 years may have been influenced not only by changes in vegetation but also by global warming. It is important to disentangle the individual role that afforestation and global warming play in the precipitation

changes from both observational and modeling perspectives. The impact of climate change on vegetation inhomogeneity, BLCLs, and their associated convection and precipitation in the Hetao area also needs to be explored in future research.

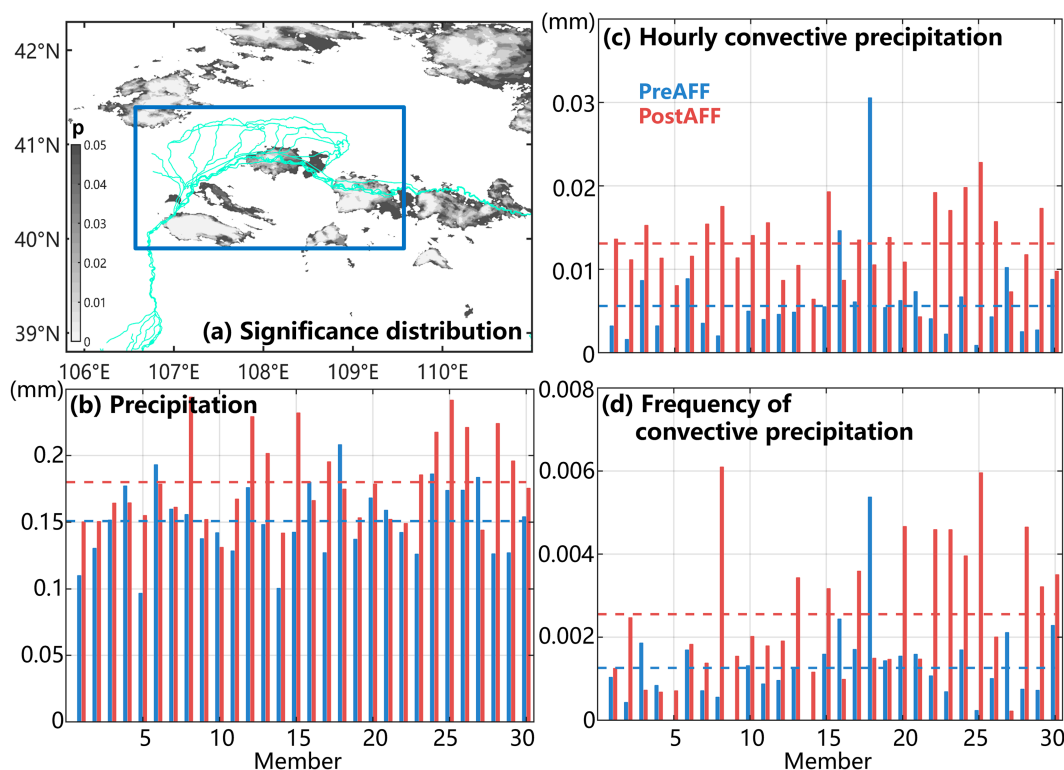


FIG. 11. (a) Spatial distribution of significance of simulated hourly convective precipitation differences in the whole integration period with no-rain hours in both experiments removed. Areal means of (b) hourly precipitation, (c) hourly convective precipitation, and (d) frequency of hourly convective precipitation, over the region in the blue rectangle in (a), which is exactly the same as the blue rectangle in Fig. 1a, based on accumulations of hourly model output for each ensemble member of PreAFF (blue) and PostAFF (red). The dashed lines in (b)–(d) represent the ensemble mean.

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Data availability statement. The GPM-IMERG satellite data (Huffman et al. 2023) were downloaded from <https://doi.org/10.5067/GPM/IMERG/3B-HH/07>. The ERA5 hourly data (Hersbach et al. 2021a,b) were also downloaded from NCAR’s official website <https://rda.ucar.edu/datasets/d633000/dataaccess/#>. The ETOPO1 topography data (NGDC/NESDIS/NOAA/U.S. Department of Commerce 2011) were downloaded from NCAR’s official website <https://rda.ucar.edu/datasets/d759004/dataaccess/#>. The 30-m, yearly land-use dataset in China (Yang and Huang 2021) was downloaded from <https://zenodo.org/records/4417810>. Additional data related to this paper can be requested from the authors.

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