A Simple Lagrangian Parcel Model for the Initiation of Summertime Mesoscale Convective Systems over the Central United States

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ABSTRACT: Mesoscale convective systems (MCSs) account for more than 50% of summertime precipitation over the central United States and have a significant impact on local weather and hydrologic cycle. It is hypothesized that the inadequate treatment of MCSs is responsible for the long-standing warm and dry bias over the central United States in coarse-resolution general circulation model (GCM) simulations. In particular, a better understanding of MCS initiation is still lacking. Here a single-column Lagrangian parcel model is first developed to simulate the basic features of a rising parcel. This simple model demonstrates the collective effects of boundary layer moistening and dynamical lifting in triggering convective initiation and reproduces successfully its early afternoon peak with surface equivalent potential temperature as a controlling factor. It also predicts that convection is harder to trigger in the future climate under global warming, consistent with the results from convection-permitting regional climate simulations. Then, a multicolumn model that includes an array of single-column models aligned in the east–west direction and incorporates idealized cold pool interaction mechanisms is developed. The multicolumn model captures readily the cold pool–induced upscale growth feature in MCS genesis from initially scattered convection that is organized into a mesoscale cluster in a few hours. It also highlights the crucial role of lifting effects due to cold pool collision and spreading, subsidence effect, and gust front propagation speed in controlling the final size of mesoscale clusters and cold pool regions. This simple model should be useful for understanding fundamental mechanisms of MCS initiation and providing guidance for improving MCS simulations in GCMs.

KEYWORDS: North America; Deep convection; Mesoscale systems; Summer/warm season; Moisture/moisture budget; Single column models

1. Introduction

Mesoscale convective system (MCS) is a collection of thunderstorms that are typically organized over a few hundred kilometers and produces large amounts of rainfall, inducing tremendous impact on the local weather and hydrologic cycle in both the tropics and midlatitudes (Houze 2004, 2018). By using 13 years of high-resolution ground-based radar and satellite observations, Feng et al. (2019) found that long-lived intense MCSs account for over 50% of warm-season precipitation in the Great Plains and over 40% of cold-season precipitation in the southeast United States. Due to the high intensity and larger area coverage, MCS rainfall plays a more important role in generating surface runoff than non-MCS rainfall (Hu et al. 2020), and MCSs account for majority of slow-rising and hybrid floods during the warm season (Hu et al. 2021). As the dominant organized convection, MCSs show significant interactions with the large-scale circulation through diabatic heating and convective momentum transport (Moncrieff et al. 2017). Recognizing the self-similarity of tropical convection, i.e., a self-similar progression from shallow to deep convection to stratiform anvils across multiple spatiotemporal scales, an intriguing question is whether MCSs are building blocks or a prototype of large-scale

waves (Mapes et al. 2006). Accurately understanding and simulating the MCS life cycle, particularly its initiation, has important practical applications.

However, present-day general circulation models (GCMs) still suffer significant biases in realistically simulating MCSs and the associated large-scale environments (Houze 2018; Feng et al. 2021). In general, coarse-resolution GCMs using cumulus parameterizations typically have difficulty in simulating MCSs that are in the convective gray zone (Jeworrek et al. 2019; Feng et al. 2021). Furthermore, GCMs poorly represent the convective/stratiform proportions and cold pool dynamics (Arakawa 2004; Cao and Zhang 2017; Moncrieff et al. 2017), both of which play essential roles in the life cycle and propagation mechanisms of MCSs (Feng et al. 2015). A long-standing warm and dry bias in the summertime central United States presents a great challenge for GCMs (Liu et al. 2017). It is hypothesized that the bias is associated with the failure of GCMs in simulating MCSs and the associated precipitation (Dai et al. 1999; Klein et al. 2006; Lin et al. 2017; Morcrette et al. 2018; Van Weverberg et al. 2018). In contrast, convection-permitting regional climate models that explicitly resolve fine-scale clouds and mesoscale processes present much better skills in simulating realistic MCS properties, including its precipitation amount, diurnal cycle, and intensity distribution (Prein et al. 2017; Feng et al. 2018). For global simulations, the superparameterized GCMs that replace traditional cumulus parameterization by a cloud-resolving model (CRM) in each model grid also show promising results in capturing realistic features of MCSs

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(Khairoutdinov and Randall 2001; Randall et al. 2003; Grabowski 2004; Randall et al. 2016; Lin et al. 2019; Hannah et al. 2020).

A better understanding of the MCS from a theoretical perspective should be useful for not only illustrating the underlying mechanisms of its initiation and development but also providing essential guidelines to improve GCM simulations and projecting future changes in MCSs with warming. Early theoretical studies about MCSs date back to Raymond and Jiang (1990), where the maintenance of certain long-lived MCSs is attributed to the interaction between quasi-balanced vertical motions and the diabatic effects of moist convection in a sheared environment. Mapes (1993) demonstrated theoretically that the heating source from MCSs causes upward displacement at low levels through inviscid gravity waves and thus predicts that the associated cloud clusters are gregarious. Recently, Agard and Emanuel (2017) developed an idealized one-dimensional model to demonstrate the thermodynamic constraint on convective available potential energy (CAPE) in continental environments, providing intriguing implications for MCS initiation. Nevertheless, we still lack a simple model to examine the MCSs and their initiation mechanisms, although some progress has already been made along this direction. For example, Singh and O’Gorman (2013) developed an entraining plume model in the limit of zero cloud buoyancy to reproduce the increase of CAPE with warming, and highlighted the crucial role of entrainment in determining the thermal stratification of the tropical atmosphere. By considering a spectrum of entraining plumes instead, Zhou and Xie (2019) reformulated this conceptual model and successfully captured the C-shape equivalent potential temperature profile as well as its upward stretch with warming. Tian et al. (2021) developed a plume model with thermodynamic and dynamical constraints to explore the relative importance of various convection-controlling factors for the diurnal cycle of clouds and precipitation. It is still unclear whether these theoretical models can capture the basic features of MCS initiation.

Here we are particularly focusing on the summertime MCS initiation over the central United States, including convective initiation and MCS genesis in the early stages of MCS life cycle (Feng et al. 2019). This focus is partly motivated by the larger challenge in modeling summertime MCS than springtime MCS, even in convection-permitting regional models (Prein et al. 2017) and differences in the role of large-scale environments on MCS initiation in summer versus spring (Song et al. 2019). A better understanding of summertime MCS initiation should provide useful guidelines to further improve weather forecast for precipitation and climate projection for atmospheric conditions. In general, convective initiation means the first hour when a cold cloud system is detected, and MCS genesis means the first hour after the convective feature major-axis length reaches 100 km (Feng et al. 2019). For convective initiation, previous observational studies have highlighted the important role of environmental vertical shear and thermodynamic conditions in determining the structure and evolution of MCSs (LeMone et al. 1998). Recent studies also suggested that convection over the mid-latitude continents could be possibly triggered by dynamical lifting effects from eastward-moving large-scale environments and subsynoptic perturbations (Li and Smith 2010;

Wang et al. 2011a,b; Pokharel et al. 2019; Tuttle and Davis 2013) and boundary layer moistening effects from low-level jets (Rasmussen and Houze 2016) and soil moisture (Klein and Taylor 2020). It should be interesting to investigate the collective effects of boundary layer moistening and dynamical lifting in triggering convective initiation. In contrast, Song et al. (2019, 2021) suggested that a subset of summertime MCSs initiate under unfavorable large-scale environments with weak baroclinic forcing, and convective self-aggregation may play more important roles compared to those more favorable large-scale conditions. For MCS genesis, it is still challenging to illustrate the underlying mechanisms for the upscale growth of convection into mesoscale clusters. The self-aggregation phenomenon, in which moist convection spontaneously organizes into one or several isolated mesoscale clusters, occurs in idealized radiative–convective equilibrium simulations and highlights the interactions between clouds, moisture, radiation, surface fluxes, and circulation (Bretherton et al. 2005; Muller and Held 2012; Bretherton and Khairoutdinov 2015; Wing et al. 2017). Another body of literature attributes the organization of convection to the cold pool dynamics (Torri et al. 2015; Haerter 2019; Haerter et al. 2019; Feng et al. 2015), in which new convection is triggered in the surrounding area through either mechanical lifting by the cold pool gust fronts or thermodynamic forcing by the accumulation of moisture around the edge of cold pools.

The overall goals of this study is to develop a simple model of MCS initiation and use it to understand what salient mechanisms are needed to reproduce some key features of MCSs, such as their upscale growth, and explore how MCSs may be influenced by global warming. This goal is achieved through the following five aspects: 1) develop a simple single-column model to capture the basic features of summertime convective initiation over the central United States, 2) investigate the collective effects of boundary layer moistening and dynamical lifting on convective initiation, 3) study the diurnal cycle of convective initiation and identify the differences between cases with and without MCS initiation, and 4) explore the impact of global warming on convective initiation and theoretically predict the convection population under climate change. Last, we also develop a multicolumn model with spatial dependence to reproduce the upscale growth and identify key mechanisms in determining final states of MCSs.

The simple model we developed here is based on the Lagrangian parcel model by Romps and Kuang (2010). To assess the role of fusion for diluted parcels to reach the tropopause, Romps and Kuang (2010) constructed a model of an entraining Lagrangian parcel that shows some skills in simulating a parcel traveling through the troposphere and overshooting the tropopause. Here we further improved and simplified this model in several ways. This new single-column model describes the ascent trajectory of a small-scale parcel within an updraft in a Lagrangian view and provides a useful tool to evaluate the impact of ambient environmental conditions and microphysical processes on the convective initiation. Then we expanded the model into a multicolumn configuration where 1000 such single-column models are aligned in the east–west direction. To allow the interactions between different columns, we

FIG. 1. Schematic diagram for the simple Lagrangian parcel model in the (a) single-column and (b),(c) multicolumn versions. In (a) for the single-column version, a parcel is first initiated at the surface and lifted up by a certain distance above the cloud-base height. Due to the imbalance between buoyancy, gravity, and momentum drag, this parcel then starts to move upward under the impact of entrainment, condensation, and precipitation processes. The multicolumn version in (b) considers multiple such single-column models that are aligned in an east–west direction. For each column, two gust fronts are added and set to propagate in both directions as soon as the parcel goes beyond the cloud-base height. The version in (c) demonstrates the three cold pool interaction mechanisms that are introduced in the multicolumn version, including 1) the strong lifting effect due to cold pool collision (large upward red arrow), 2) weak lifting effect due to gust fronts from isolated cold pools (small upward red arrow), and 3) weak subsidence effect (small downward blue arrow).

further considered cold pool interaction mechanisms, including the lifting effect due to cold pool collision and gust front spreading over active convection regions, and subsidence effect over suppressed-convection regions. This multicolumn model introduces spatial dependence in the solution and provides a useful tool to assess the spatial variance of MCS genesis.

The rest of the paper is organized as follows. Sections 2a and 2b summarizes the details of the model governing equations, physical parameters, cloud microphysical closure, and numerical setup in the single-column and multicolumn configurations, while section 2c describes the data used in this study. Section 3 uses the single-column model to simulate the summertime convective initiation over the central United States and discusses the impact of boundary layer moistening and dynamical lifting on convective initiation, its diurnal cycle, and future changes under global warming. Section 4 uses the multicolumn model to study the upscale growth of the MCS genesis and investigates the key mechanisms and controlling factors in determining the final size of mesoscale clusters. The paper concludes with a discussion in section 5. The derivation of some cloud microphysical closure is included in the appendix.

2. Model and data

In this section, we first summarize all the details about the simple Lagrangian parcel model in the single-column and multicolumn configurations. The single-column model is based

on the Lagrangian parcel model developed by Romps and Kuang (2010) but modified in a few aspects for the sake of simplicity. Then we consider multiple single-column models that are aligned in the east–west direction and incorporate the cold pool interaction mechanisms in the multicolumn model. Last, we document all the information about the reanalysis data for the current climate, GCM projection for future climate, and MCS initiation timing from observations.

a. The single-column Lagrangian parcel model for convective initiation

As shown by the schematic diagram in Fig. 1a, the single-column Lagrangian parcel model describes the whole ascent trajectory of a buoyancy-driven parcel from the surface to the upper troposphere. The ascent motion of the parcel involves several key physical processes, including the imbalance between buoyancy, gravity, and momentum drag, the phase change between water vapor and liquid water, and the environmental mass exchange through entrainment. All the equations for governing the Lagrangian parcel dynamics are shown in Table 1. The model is written in the form of conservation laws and solves seven physical variables related to the parcel, including volume V , density ρ , water vapor mass fraction q_v , liquid water mass fraction q_l , vertical velocity w , vertical displacement z , and temperature T .

In detail, the equation in the first row of Table 1 describes the conservation of mass, where the tendency of mass $(d/dt)(V\rho)$ is forced by liquid water fallout $-Vf_l$ and entrainment $V\epsilon$

TABLE 1. Governing equations of the simple Lagrangian parcel model. This model solves seven physical variables related to the parcel, including volume V (unit: m^3), density ρ (unit: kg m^{-3}), water vapor mass fraction q_v (unit: kg kg^{-1}), liquid water mass fraction q_l (unit: kg kg^{-1}), vertical velocity w (unit: m s^{-1}), vertical displacement z (unit: m), and temperature T (unit: K). All variables with subscript e stand for the environmental profiles with their only dependence on height z , including density $\rho_e(z)$, water vapor $q_{ve}(z)$, temperature $T_e(z)$, and pressure $p_e(z)$. The variables $E^{\text{tot}} = c_v T + gz + Lq_v + (1/2)w^2$, $E_l^{\text{tot}} = c_v T + gz + (1/2)w^2$, $E_e^{\text{tot}} = c_v T_e(z) + gz + Lq_{ve}(z)$ stand for the specific energies of the parcel, liquid water, and environmental air. The notation $|x|$ means the absolute value of x .

Physical law	Symbol	Equation
Total mass	$V\rho$	$\frac{d}{dt}(V\rho) = -Vf_l + V\epsilon$
Total water vapor mass	$V\rho q_v$	$\frac{d}{dt}(V\rho q_v) = Ve + V\epsilon q_{ve}(z)$
Total liquid water mass	$V\rho q_l$	$\frac{d}{dt}(V\rho q_l) = -Ve - Vf_l$
Total vertical momentum	$V\rho w$	$\frac{d}{dt}(V\rho w) = -V\rho g - V\frac{d}{dz}p_e(z) - Vf_l w - \frac{1}{4}V\left(\frac{9\pi}{2V}\right)^{1/3} c_d \rho_e(z) w w $
Vertical displacement	z	$\frac{d}{dt}z = w$
Total energy	$V\rho E^{\text{tot}}$	$\frac{d}{dt}(V\rho E^{\text{tot}}) = -Vw\frac{d}{dz}p_e(z) - Vp_e(z)\psi - Vf_l E_l^{\text{tot}} + V\epsilon E_e^{\text{tot}} - \frac{1}{4}V\left(\frac{9\pi}{2V}\right)^{1/3} c_d \rho_e(z) w^2 w $
Ideal gas law	—	$p_e(z) = \rho RT$

(see [appendix](#)). The equation in the second row describes the conservation of water vapor with its tendency $(d/dt)(V\rho q_v)$ forced by evaporation/condensation Ve and entrainment of water vapor from the environment $V\epsilon q_{ve}(z)$, while the one in the third row is for the conservation of liquid water with its tendency $(d/dt)(V\rho q_l)$ forced by evaporation/condensation $-Ve$ and liquid water fallout $-Vf_l$. The equation in the fourth row governs the vertical momentum budget where the tendency of momentum $(d/dt)(V\rho w)$ is driven by the imbalance between gravity $-V\rho g$ and buoyancy $-V(d/dz)p_e(z)$, the momentum loss due to liquid water fallout $-Vf_l w$, and momentum drag $-(1/4)V[9\pi/(2V)]^{1/3}c_d\rho_e(z)w|w|$. The equation in the fifth row simply states that the vertical velocity of the parcel w is defined as its vertical displacement $(d/dt)(z)$. The one in the sixth row shows the total energy budget where its tendency $(d/dt)(V\rho E^{\text{tot}})$ is forced by energy gain due to buoyancy force $-Vw(d/dz)p_e(z)$ and entrainment $V\epsilon E_e^{\text{tot}}$, and energy loss due to volume expansion $-Vp_e(z)\psi$ (see [appendix](#)), liquid water fallout $-Vf_l E_l^{\text{tot}}$, and momentum drag $-(1/4)V[9\pi/(2V)]^{1/3}c_d\rho_e(z)w^2|w|$. The equation in the seventh row includes the ideal gas law. It is worth mentioning that the pressure of the parcel is assumed to be the same as its environmental pressure $p_e(z)$. All variables from environmental profiles with the subscript e are fixed in time with only dependence on height z . Similar to [Romps and Kuang \(2010\)](#), this model ignores the detrainment process, which assumes the large-scale thermodynamical environments are not modified by convection. All physical constants, model parameters and cloud microphysical closure are summarized in [Table 2](#).

When compared with the starting model by [Romps and Kuang \(2010\)](#), here we modified a few aspects to further improve and simplify this single-column model. First, instead of using a prognostic equation for volume, we choose to guarantee the ideal gas law so that the volume of the parcel can be inferred from total mass $m = V\rho$ and density ρ . Second, except for the momentum drag, we ignored the other momentum

forcing term for the force felt by the ascent parcel as it accelerates the ambient environmental air. Third, we derived some of the cloud microphysical closure for entrainment ϵ , volume expansion ψ , evaporation e and kept the liquid water fallout f_l the same. Fourth, the ice phase change is not included in the model for simplicity, although [Romps and Kuang \(2010\)](#) concluded that fusion provides the kinetic energy required for diluted parcels to reach tropopause. Based on CRM simulations, [Fan et al. \(2017\)](#) found that ice-related processes, including riming, deposition, and freezing, also contribute to heating at different altitudes, and their effects would be neglected in this study. Furthermore, we made a few additional assumptions to simplify the model, including (i) using constant heat capacity coefficients c_v and c_p , (ii) assuming the hydrostatic balance for the environmental profiles with no liquid water $q_{le} = 0$ and vertical motion $w_e = 0$, (iii) considering a fixed mass-specific gas constant with only dependence on height $R(z)$. Theoretically, the mass-specific gas constant R for a moist parcel should take water vapor q_v into account. Here we simplify this relation by assuming its value is the same as that inferred from the ambient environmental air, that is, $R = R_e(z)$. Since the parcel may differ from the environmental profile in specific humidity by a few gram per kilogram, this simplification assumption may lead to some biases ([Yang and Seidel 2020](#)). For the sake of simplicity, we followed [Romps and Kuang \(2010\)](#) to ignore the detrainment process that usually occurs at the cloud top. The inclusion of the detrainment in this model should only have impacts on the final height and lifetime of the parcel dynamics, but not its convective initiation, which is the focus of this study.

The complete trajectory of an ascent parcel is divided into two stages, including (i) the early stage with dynamical lifting and (ii) the later stage with free convection. During the dynamical lifting stage, a parcel is first initiated at the surface (geopotential height $\sim 110\text{ m}$), the lowest level defined in the

TABLE 2. Physical constants, model parameters, and cloud microphysical closure in the simple Lagrangian parcel model.

Category	Symbol	Value	Description
Physical constants	c_v	$718.0 \text{ J kg}^{-1} \text{ K}^{-1}$	Heat capacity at constant volume
	c_p	$1.0 \times 10^3 \text{ J kg}^{-1} \text{ K}^{-1}$	Heat capacity at constant pressure
	g	9.8 m s^{-2}	Gravitational acceleration constant
	L	$2.26 \times 10^6 \text{ J kg}^{-1}$	Latent heat of evaporation
	R	Determined by environmental profiles ($\approx 288 \text{ J kg}^{-1} \text{ K}^{-1}$)	Mass-specific gas constant
Model parameters	τ_{en}	300 s	Entrainment time scale (Hannah 2017)
	V^*	$1.0 \times 10^9 \text{ m}^3$	Reference volume
	Γ	0.5	Liquid water fallout rate (Hernandez-Duenas et al. 2013)
	τ_e	135 s	Evaporation time scale (Hernandez-Duenas et al. 2013)
	q^*	0.01 kg kg^{-1}	Reference specific humidity (Hernandez-Duenas et al. 2013)
	c_d	0.2	Momentum drag coefficient (Blevins 1984)
	s_{gf}	5 m s^{-1}	Propagation speed of gust fronts
	σ_{iso}	$\frac{10}{3} \text{ m s}^{-1}$	Lifting effect by a gust front from isolated cool pools
	σ_{col}	$\frac{20}{3} \text{ m s}^{-1}$	Lifting effect by cool pool collision
	σ_{sub}	$-\frac{20}{3} \text{ m s}^{-1}$	Subsidence effect coefficient
	ϵ	$\epsilon = \frac{1}{\tau_{\text{en}}} \rho \left(\frac{V^*}{V} \right)^{1/3}$	Mass entrainment rate (see appendix)
	q_{vs}	$q_{\text{vs}} = 0.622 \frac{e_s}{p_e(z)}$, $e_s = 6.1094e^{17.625T/(T+243.04)}$	Saturation specific humidity (note e_s is in hPa and T is Celsius) based on August–Roche–Magnus formula (Alduchov and Eskridge 1996)
	E	$e = \begin{cases} -\frac{1}{\tau_e}(q_v - q_{\text{vs}}), & \text{if } q_v \geq q_{\text{vs}} \\ \frac{1}{\tau_e}(q_{\text{vs}} - q_v) \frac{q_l}{q^*}, & \text{if } q_v < q_{\text{vs}} \end{cases}$	Evaporation rate (Hernandez-Duenas et al. 2013)
Cloud microphysical closure	f_l	$f_l = \gamma \max(0, -e)$	Liquid water fallout rate (Romps and Kuang 2010)
	Ψ	$\psi = \frac{Le}{c_p T \rho} - \frac{c_v w}{c_p p_e(z)} \frac{d}{dz} p_e(z)$	Volume expansion rate (see appendix)
	S	$S = \sigma_{\text{sub}} \frac{\text{No. of active columns}}{\text{No. of suppressed columns}}$	Subsidence effect

model. The initial volume of the parcel is set at 10^9 m^3 , similar to the typical size of a thermal (Stull 1988). The initial water vapor mass fraction of the parcel is set to be larger than that in the environmental air by δq_v , while its liquid water mass fraction is zero. Besides, its initial vertical velocity is 2.5 m s^{-1} , a typical updraft speed at the cloud base (Zheng et al. 2015). This vertical velocity is fixed during the dynamical lifting stage. The initial temperature of the parcel is assumed to be the same as the environment. Here the environmental profile includes the surface ($\sim 110 \text{ m}$) as well, and it is fixed throughout the whole simulation. Then the parcel gets lifted by a certain distance, δz . Similar to the typical parcel argument with the skew T diagram, the parcel temperature first follows the dry adiabatic lapse rate in the conservation of dry static energy $\text{DSE} = c_v T + gz$. After it reaches the moisture saturation at the lifting condensation level (LCL), its temperature then follows the moist adiabatic lapse rate in the conservation of moist static energy $\text{MSE} = c_v T + gz + Lq_v$. That said, several studies in the literature

found that the conservation of DSE/MSE for an adiabatically lifted parcel is only an approximate (Romps 2015; Peters and Chavas 2021). Also, it is assumed that all liquid water due to condensation falls out from the parcel during the dynamical lifting stage. During the free convection stage, the parcel is released freely in the troposphere and its dynamics is completely governed by the model as shown in Table 1. It is worth mentioning that the two parameters δz and δq_v represent the dynamical lifting and boundary layer moistening effects, respectively, both of which should have a significant impact on the parcel dynamics.

The model is numerically solved by the fourth-order Runge–Kutta scheme with a time step of 1 s. The numerical scheme solves the equations in Table 1 directly without rewriting them, but infers all the physical variables ($V, \rho, q_v, q_l, w, z, T$) at each time step to assign the value of cloud microphysical closure. Since the environmental profiles have only values at discrete vertical levels that are unevenly spaced by either 25 or 50 hPa below 100 hPa, their actual value at the height of the

parcel is obtained through spline interpolation. Thanks to the small time steps, the parcel's height only changes by at most several meters after each time step, making it possible for the model to accurately resolve the detailed dynamics of the parcel. All the simulations are run for 1 h, which is long enough for the solution to reach an equilibrium state.

b. The multicolumn Lagrangian parcel model for MCS genesis

To investigate the upscale growth after convection initiation to the MCS genesis, we further developed a multicolumn version of the Lagrangian parcel model. This additional spatial dimension of the solution allows us to discuss the spatial variability of multiple convective clusters, particularly their collective behaviors. As shown by Fig. 1b, the multicolumn model considers 1000 single-column Lagrangian parcel models that are aligned in the east–west direction. Inside each column, we only consider one single parcel that moves vertically within that column, as governed by the above single-column model. The horizontal length of each column is set as 1 km, the same as the initial size of the parcel. It is worth mentioning that without any additional mechanisms, the dynamics of all parcels in this multicolumn model would be independent of each other. However, as discussed below, the multiple columns will be connected by the cold pool dynamics, as Feng et al. (2015) found that cold pools play an important role in triggering new convection and promoting convective cloud organization via aggregating convective clouds near cold pool boundaries.

Here we introduce three interaction mechanisms associated with the cold pool dynamics, including (i) the lifting effect by gust front spreading, (ii) the lifting effect by cold pool collision, and (iii) the clear-sky subsidence effect produced by active convection. As shown by the light red plane in Fig. 1c, we first set the level at 2 km above the surface as the cloud-base height (CBH) (Kalb et al. 2004), which is high enough so that the parcel obtains enough buoyancy after getting lifted to this level and would rise to deeper levels afterward. A shallower choice of CBH in the model may only trigger shallow convection without much cold pool dynamics. Any column with a parcel above (at or below) the CBH is recognized as an active (suppressed) convection column. For each parcel with active convection, we also attached two more variables to record the positions of its two gust fronts in both directions, respectively. These two gust fronts are initially located at the middle of the column and set to propagate in both eastward and westward directions as soon as the parcel moves above the CBH to mimic precipitation-driven cold pools that spread laterally. During the passage of the gust front through the neighboring suppressed convection columns, the model also applies a weak lifting effect σ_{iso} to the parcels within those columns, mimicking the lifting effect by gust front spreading from isolated cold pools. When two gust fronts from different columns collide, their positions are frozen to mimic the stationary intersecting cold pools. Meanwhile, they apply a stronger lifting effect σ_{col} to the parcel within that column, mimicking the lifting effect by cold pool collision. In the simulations, the

values of strong lifting effect due to cold pool collision σ_{col} and weak lifting effect due to gust front spreading σ_{iso} are set to be comparable with the updraft velocity of gust fronts from the large-eddy simulation (Fournier and Haerter 2019; Henneberg et al. 2020), and a factor of 2 is chosen to reflect their difference in magnitude. After the parcel reaches the equilibrium state with negligible liquid water mass fraction ($q_l < 0.1 \text{ g kg}^{-1}$), its life cycle is set to terminate and its position is returned to the CBH with both gust fronts be relocated to the middle of the column. For suppressed convection column, the parcel is influenced by an additional weak descending effect σ_{sub} , mimicking the subsidence effect due to radiative cooling and compensating descending circulation over the ambient clear-sky regions. The magnitude of the subsidence effect $S = \sigma_{\text{sub}}(\text{number of active columns/number of suppressed columns})$ is proportional to the number ratio between active and suppressed convection columns so that more active convection corresponds to stronger subsidence effect, and vice versa. The default propagation speed of gust fronts (see Fig. 9) are set at 5 m s^{-1} similar to that in the large-eddy simulation (Fournier and Haerter 2019; Henneberg et al. 2020), but varied for sensitivity experiments (see Fig. 11). The constant propagation speed of gust fronts is in line with the results from the large-eddy simulation model (Fournier and Haerter 2019). Nevertheless, it should still be interesting to investigate whether a time-varying gust front propagation speed would help to capture more realistic features of cloud clustering. The values of all physical parameters are summarized in Table 2. It is worth mentioning that a better way to determine the strength of cold pool is to obtain a cold pool closure based on the quasi-equilibrium theory (Emanuel et al. 1994).

The model is numerically solved with a time step of 1 s. The domain has periodic boundary conditions in the zonal direction. At each time step, we first compute the evolution of all individual columns in parallel, including the parcels and their gust fronts. Then we calculate the total vertical displacement of parcels in the suppressed convection columns due to lifting effects by gust front spreading and cold pool collision as well as the clear-sky subsidence effect. This vertical displacement will be applied during the evolution of the next time step. The initial vertical positions of all parcels are randomly picked from the uniform distribution, which ranges from the surface to the level 2.5 km above that. The environmental profiles of all physical variables are obtained from the reanalysis data and fixed at 1500 local time. All simulations are run for 12 h, which is long enough for the solution to reach an equilibrium state.

c. The data for current and future environmental profiles as well as MCS initiation timing

Three types of data are used in this study. To align with the previous study by Song et al. (2021) on the environments of MCS initiation in the central United States, here we only focus on the 14-yr (2004–17) summertime (June–August) data over the central United States (35° – 45°N , 100° – 90°W). The first type of data is from the fifth-generation ECMWF reanalysis (ERA5) hourly data product (Hersbach et al. 2018)

and downloaded from the Climate Data Store (CDS) website (<https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-pressure-levels?tab=form>). We average all the data of specific humidity and temperature at each specific hour of the diurnal cycle. The resulting climatological mean diurnal profiles are regarded as the summertime environmental profiles over the central United States and integrated into the Lagrangian parcel model in section 3. It has been shown that ERA5 successfully reproduces severe local storm environments with strong spatiotemporal correlations and low biases, especially over the Great Plains (Li et al. 2020). Although we choose to use the pressure-level ERA5 data here, it is worth noting that ERA5 also provides higher-resolution model-level data that may better capture environmental profiles of severe thunderstorms, especially boundary layer features such as capping inversions (Taszarek et al. 2021). This higher-resolution model-level data should potentially benefit simulations of the diurnal cycle when compared with the observed MCS initiation. The second type of data is the temperature and specific humidity anomalies between future climate (2056–99; RCP8.5) and current climate (1962–2005; historical) based on 37 CMIP5 models (Song et al. 2018). This dataset is used in section 3d to assess how changes of the environmental thermodynamic profiles under global warming may influence convection initiation. The third type of data is the diurnal timing of the 14-yr summertime MCS initiation based on the MCS tracking data (Feng 2019), which is used to validate the simple model simulations. Here we count the total number of MCS initiations at each hour of the diurnal cycle and compare that with the model results in section 3c.

3. Convective initiation in the single-column Lagrangian parcel model

The goals of this section are to first investigate the capability of the single-column Lagrangian model in simulating the rising parcel and then discuss the impact of various atmospheric conditions, diurnal cycle, and climate change on convective initiation. The model setup is the same as the description in section 2a. Unless stated otherwise, the initial moisture anomaly δq_v of the parcel is set at 6.7 g kg^{-1} to mimic convective instability due to low-level moisture convergence, while its dynamical lifting distance δz is set at 2 km to be comparable with typical CBH. The environmental profiles of temperature and relative humidity are fixed at 1500 local time averaged over the 14-yr summertime ERA5 data over the central United States (see the description in section 2c). Note that a parcel with final height higher than 8 km will be regarded as deep convection.

a. Characteristic features of convective initiation and its dependence on atmospheric conditions

Figure 2 shows the time series of all physical variables related to the rising parcel from the single-column simulation. In Fig. 2a, after releasing from the level 2 km above the surface, the parcel rises gradually to the higher levels, overshoots the height of 10.5 km at time $t = 23 \text{ min}$, and oscillates about the equilibrium

level (EL) with a decaying amplitude. Correspondingly, the vertical velocity of the parcel in Fig. 2b remains positive before it reaches the EL, reaching its peak magnitude of 8 m s^{-1} at time $t = 8 \text{ min}$ and oscillates about the state of rest afterward. Due to the mass gained through entrainment, the mass of the parcel in Fig. 2c shows an exponential growth such that its final mass at time $t = 40 \text{ min}$ is 26 times larger than its initial mass. Such a fast increase in the parcel's mass is related to the rapid volume expansion, while its density in Fig. 2d drops gradually and is slightly smaller than the environmental air density as it rises. The similarity between the parcel density and environmental air density in Fig. 2d reflects the fact that the parcel dynamics is efficiently driven by its buoyancy so that the less dense parcel gets lifted to higher levels. Note that we assume the parcel pressure is the same as the environmental air pressure at the same level. As implicated by the ideal gas law, the parcel temperature in Fig. 2e is slightly higher than the environmental temperature and only shows a significant warming anomaly in the first few minutes due to the initial dynamic lifting. As shown by Fig. 2f, the parcel stays oversaturated of water vapor before it reaches the EL, gaining more buoyancy due to the latent heat release from condensation. After reaching the EL, the specific humidity of the parcel becomes less than its saturation value due to dry air intrusion through entrainment. Meanwhile, the water vapor condensation results in some amount of liquid water, part of which is stored in the parcel and the rest is precipitated out in the form of rain. The liquid water mass fraction of the parcel in Fig. 2g reaches its peak magnitude at time $t = 15 \text{ min}$ and decays gradually afterward. The precipitation rate reaches its peak magnitude a few minutes earlier than the liquid water mass fraction and vanishes completely after the parcel reaches the EL. The simple Lagrangian parcel model can capture the basic properties of a rising parcel such as its vertical trajectory, water vapor condensation, and overshooting, resembling those from the cloud-resolving simulation (Roms and Kuang 2010) and observation (Zheng et al. 2015).

The energy budget of the parcel is used to gain a better understanding of its behaviors and identifying the key processes that determine the rising trajectory of the air parcel. Figure 3a shows the time series of all four specific energy components associated with the parcel. As indicated by the upper boundary in red, the total specific energy first decreases to its minimum value at time $t = 5 \text{ min}$, then recovers gradually before the parcel reaches the EL and remains constant afterward. The internal energy $c_v T$ dominates the total specific energy of the parcel. Its value decreases gradually as the temperature of the rising parcel declines. In contrast, the potential energy gz is the second-largest component and keeps increasing as the parcel moves to higher levels. Meanwhile, the latent heat Lq_v has a similar magnitude as the potential energy initially but decreases quickly afterward due to water vapor condensation. Last, the kinetic energy $(1/2)(w^2)$ is in a much weaker magnitude than the other three specific energy components. The significant variation of these specific energy components results from not only the energy transition among them but also the external forcing. Thus, it is worthwhile to investigate which physical processes are responsible for the change of total specific energy. By rewriting the energy

FIG. 2. Time series of various physical variables related to the parcel (blue curves) in the single-column model simulation, including (a) height z , (b) vertical velocity w , (c) mass $V\rho$, (d) density ρ , (e) temperature T , (f) water vapor mass fraction q_v , (g) liquid water mass fraction q_l , and (h) precipitation rate Vf_l . In addition, three red thin curves are overlaid in (d)–(f) and correspond to (d) environmental air density, (e) environmental air temperature, and (f) saturation specific humidity of the parcel.

equation in the sixth row of Table 1, we can obtain the specific energy budget for the parcels,

$$\begin{aligned} \frac{d}{dt}E^{\text{tot}} = & -\frac{1}{\rho}w\frac{d}{dz}P_e - \frac{1}{\rho}P_e\psi - \frac{1}{\rho}f_l(E_l^{\text{tot}} - E^{\text{tot}}) \\ & + \frac{1}{\rho}\epsilon(E_e^{\text{tot}} - E^{\text{tot}}) - \frac{1}{4\rho}\left(\frac{9\pi}{2V}\right)^{1/3}c_d\rho_e w^2|w|, \end{aligned} \quad (1)$$

where the term on the left-hand side is the time tendency of total specific energy, and the terms on the right-hand side correspond to energy gain due to buoyancy force, energy loss due to volume expansion, liquid water fallout, entrainment, and momentum drag, respectively. Thanks to the simplicity of the model, all budget terms in Eq. (1) are calculated accurately so that the term on the left-hand side is exactly balanced by the sum of those on the right-hand side without a residual. Figure 3b shows the time series of all these budget terms related to the rising parcel. First, the energy tendency (black

curve) switches its signs from negative to positive at $t = 5$ min, indicating the increase of the parcel specific energy afterward. Before the parcel reaches the EL, the energy gain due to buoyancy force (red curve) is mostly balanced by the energy loss due to volume expansion (blue curve) and entrainment (green curve). Specifically, from an energetic view, the less dense parcel obtains buoyancy from the environmental air and increases its potential energy as it rises. Meanwhile, due to the drop in environmental air pressure, the parcel expands its volume and loses its energy by doing work to the environmental air. Also, the cold and dry environmental air intrusion through entrainment lowers the parcel's temperature and water vapor mass fraction, resulting in smaller internal energy and latent heat. After the parcel reaches the EL, the energy budget is mostly balanced between energy gain due to buoyancy force and energy loss due to volume expansion in alternate signs.

Besides the default simulation as discussed above, it is also interesting to study the impact of various atmospheric

FIG. 3. Time series of (a) specific energy components of the parcel and (b) forcing terms in its specific energy budget in the single-column model simulation. In (a), the height of color layers correspond to the magnitude of all specific energy components, including internal energy $c_v T$ (red), potential energy gz (green), latent heat Lq_v (blue), and kinetic energy $(1/2)(w^2)$ (black). Note that the black layer is much smaller than other layers and thus close to the x axis. The total height of all layers corresponds to the total specific energy of the parcel. In (b), the curves in color correspond to different forcing terms in the specific energy budget, including energy tendency $(d/dt)(E^{\text{tot}})$ (black), energy gain due to buoyancy force $-(1/\rho)w(d/dz)p_e$ (red), energy loss due to volume expansion $-(1/\rho)P_e\psi$ (blue), energy loss due to liquid water fallout $-(1/\rho)f_l(E_l^{\text{tot}} - E^{\text{tot}})$ (cyan), energy loss due to entrainment $(1/\rho)\epsilon(E_e^{\text{tot}} - E^{\text{tot}})$ (green), and energy loss due to momentum drag $-(1/4\rho)[9\pi/(2V)]^{1/3}c_d\rho_e w^2|w|$ (magenta). The dashed vertical line indicates the time when the parcel reaches an equilibrium state.

conditions and initial parcel conditions on convective initiation by conducting sensitivity experiments. Figure 4 shows the final height and total precipitation amount in the simulations with various magnitudes of entrainment time scale, initial parcel size, initial parcel vertical velocity, and initial parcel temperature. As the entrainment gets weaker (longer entrainment time scale) in Fig. 4a, the convective initiation switches from shallow to deep, and the final height of the parcel increases gradually. In the strong entrainment regime (entrainment time scale less than 5 min), shallow convection releases negligible precipitation, resulting from the dilution effects of entrainment with dry environmental air intrusion. In contrast, the moderate and weak entrainment regime (entrainment time scale more than 5 min) sees more than 50% of initial water vapor being precipitated out. Interestingly, the total amount of precipitation decreases as the entrainment further gets weaker, presumably due to less environmental water vapor supply from the lower troposphere. Similar features are also seen in Fig. 4b for the initial parcel size, which is related to the fact that larger parcel size corresponds to weaker entrainment in the cloud microphysical closure (see Table 2). In general, larger thermals have smaller fractional lateral entrainment because larger areas have a relatively smaller perimeter per unit area. As for initial parcel vertical velocity in Fig. 4c, it becomes easier for the faster parcel to overcome gravity and momentum drag and thus reach higher levels eventually. Besides, the total precipitation increases dramatically as the initial vertical velocity increases to 2.5 m s^{-1} , mainly related to the drier final state at higher levels. However, the total precipitation decreases as the initial vertical velocity further increases, resulting from the quicker passage of the parcel in the lower troposphere with

less environmental water vapor supply. Similar features are also seen in Fig. 4d, presumably because warmer initial parcel temperature corresponds to larger buoyancy so the parcel could reach faster updraft velocity in the early stage. To sum up, all four atmospheric and initial conditions promote the parcel to reach higher levels, while the total precipitation depends on not only the final height but also the entrainment of environmental water vapor in the lower troposphere.

b. Collective effects of boundary layer moistening and dynamical lifting on convective initiation

The convective initiation over the midlatitude land could be possibly influenced by many environmental factors (Song et al. 2019, 2021). For example, by using 15-yr TRMM Precipitation Radar dataset and a high-resolution Weather Research and Forecasting (WRF) simulation, Rasmussen and Houze (2016) found that the occurrence of intense storms over subtropical South America is generally associated with a strengthening of the South American low-level jet (SALLJ). Based on a numerical water-tagging technique, Hu et al. (2020) identified the distinct impacts of MCS and non-MCS rainfall on the surface water balance and the crucial role of soil moisture–precipitation feedback. Molina and Allen (2019) also highlighted the importance of land moisture in contributing to the U.S. thunderstorms. From theoretical perspectives, Agard and Emanuel (2017) obtained a thermodynamic constraint on CAPE for continental severe convection by considering a dry adiabatic column with a moist boundary layer, which was later validated by using reanalysis data over the United States (Li and Chavas 2021). Based on that, Chavas and Dawson (2021) developed a theoretical model for steady thermodynamic and kinematic profiles for severe convective storm environments. On the other hand, several

FIG. 4. Sensitivity experiments by varying four configuration parameters in the single-column model simulations, including (a) entrainment time scale τ_{en} , (b) initial parcel size V_{ini} , (c) initial parcel vertical velocity w_{ini} , and (d) initial parcel temperature T_{ini} . In each panel, the height of blue bars (left y axis) indicates the final parcel height in an equilibrium state, while that of red bars (right y axis) is for the ratio between total precipitation released by the parcel and its initial water vapor mass.

previous studies suggest that the MCS initiation could also be triggered by large-scale environments and subsynoptic perturbations through dynamical lifting, including eastward-propagating waves (Li and Smith 2010), short-wave trough (Tuttle and Davis 2013), and midtropospheric perturbations (Wang et al. 2011a,b; Pokharel et al. 2019). It should be interesting to figure out the collective effects of boundary layer moistening and dynamical lifting on convective initiation. Here we conduct totally $101 \times 101 \sim 10^4$ experiments by varying δq_v every 0.08 g kg^{-1} from 0 to 8 g kg^{-1} and varying δz every 30 m from 0 to 3 km.

Figure 5 shows the phase diagrams for the final height and total precipitation from simulations with different combinations of boundary layer moistening anomaly δq_v and dynamical lifting distance δz . According to Fig. 5a, both boundary layer moistening and dynamical lifting effects promote the parcel to reach higher levels, and a clear transition from shallow to deep can be found in the intermediate regimes. The deepest final height occurs when both effects reach their maximum strength. Also, within the range of boundary layer moistening (less than 8 g kg^{-1}), the parcel should get lifted by at least 1 km so that

deep convection can be triggered. Similarly, within the range of dynamical lifting distance (less than 3 km), the parcel should obtain at least 5 g kg^{-1} moisture anomaly relative to the environmental air so that deep convection can be triggered. The white dashed curve stands for the level of free convection (LFC) that are theoretically calculated based on the conservation of DSE and MSE. With the fixed boundary layer moistening anomaly, the parcel needs to be lifted to a much higher level than the theoretical LFC so that deep convection will be triggered, presumably due to the dilution effect of entrainment. Similar features in the shallow convection regime are also seen in Fig. 5b for total precipitation. However, with boundary layer moistening anomaly fixed, the total precipitation amount in the deep convection regime decreases as the dynamical lifting distance increases. This can be simply explained by the model assumption that in the early stage with dynamical lifting, the parcel reaches saturation as it gets lifted above the LCL. Then a large amount of water vapor in the parcel precipitates out from the parcel before the later stage with free convection. This amount of precipitation during the early stage with dynamical lifting is not counted in the

FIG. 5. Sensitivity experiments about boundary layer moistening anomaly δq_v (x axis, g kg^{-1}) and dynamical lifting distance δz (y axis, km) in the single-column model simulation. (a) Each pixel in the phase diagram corresponds to a simulation with specific values of δq_v and δz , and the color stands for the parcel height (km) in an equilibrium state. The dashed curve indicates the level of free convection with respect to a specific value of δq_v in the x axis. (b) As in (a), but for the total precipitation ($\times 10^7$ kg).

precipitation amount in Fig. 5b. Overall, this result in Fig. 5 emphasizes the importance of both effects in triggering convective initiation, potentially providing useful guidance for improving the simulation of convective initiation in the GCMs.

c. Diurnal cycle of convective initiation

The diurnal cycle is the dominant variability of convective initiation during the warm season in the central United States (Feng et al. 2019), providing a challenging benchmark for validating the skills of present-day GCMs (Feng et al. 2018, 2021). Figure 6 shows the 17-yr summertime mean vertical profiles of density, temperature, and specific humidity over the central United States at each specific hour. For density in Fig. 6a, its mean profile decreases gradually with height. The maximum diurnal cycle of density occurs at the pressure level 950 hPa in Fig. 6b with its maximum value reached at early morning (0600 LST) and the minimum value at early afternoon (1400 LST). Correspondingly, the maximum diurnal cycle of temperature occurs at the pressure level 970 hPa in Fig. 6d with its maximum value reached at early afternoon (1400 LST) and the minimum value at early morning (0600 LST). The mean profile of temperature in Fig. 6c decreases gradually with height and reverses to increase near the tropopause at the pressure level 100 hPa. As for specific humidity in Fig. 6e, its mean profile decreases exponentially with height with most of water vapor contained in the lower troposphere. The maximum diurnal cycle of specific humidity occurs at the pressure level 850 hPa in Fig. 6f with opposite phases between the surface and lower troposphere. For example, the specific humidity at early afternoon (1400 LST) features a strong moist anomaly at the pressure level 850 hPa and a weak dry anomaly at the surface, resulting from the upward transport of water vapor by convective activities.

Here we use the climatological mean profiles at each specific hour as the fixed environmental profile in the model and compare the results with the observed diurnal cycle. Note that the initial water vapor mass fraction of the parcel is fixed at a constant value throughout all specific hours by assuming the diurnal cycle of moisture at the surface is negligible. Figure 7a shows the final height of the parcel from the simulation and the total number of the observed MCS initiation (i.e., convective initiation associated with MCSs) at each hour. For the model solution (green bar), deep convection is only triggered during the period between 1100 and 1800 LST, while shallow convection prevails in the remaining time. The peak of final height is reached at early afternoon (1500 LST), consistent with the previous observation study on the diurnal cycle of MCS convective initiation (Feng et al. 2019). This peak in the early afternoon is in phase with the maximum environmental temperature at the surface in Fig. 6d, which is a consequence of the fact that warmer initial temperature provides the parcel with larger CAPE and buoyancy for it to rise to higher levels. On the other hand, the observed MCS initiation also reaches its maximum counts in the early afternoon (1500 LST), reflecting the favorable conditions for convection to be triggered during this period. The other smaller peak of the observed MCS initiation during midnight should be attributed to the mechanisms of nocturnal MCS (Grasmick et al. 2018), which the simple model here is unable to capture.

Then we divide all the hourly environmental profiles into two categories with/without observed MCS initiation and rerun the model with the mean environmental profiles at each specific hour from each category. As shown by Fig. 7b, the convection in the cases driven by environmental profiles without MCS initiation remains shallow and no

FIG. 6. Vertical profiles of (a),(b) density, (c),(d) temperature, and (e),(f) specific humidity based on 14-yr (2004–17) June–August ERA5 hourly data averaged over the central U.S. region (35° – 45° S, 100° – 90° W). In (a), the curves in color correspond to the vertical profiles of climatological mean density at different local solar time (LST) hours. (b) The deviation of these vertical profiles from its climatology in the lower troposphere (1000–800 hPa). (c),(d) and (e),(f) As in (a) and (b), but for temperature and specific humidity, respectively.

significant diurnal cycle in parcel height can be found. In contrast, deep convection is triggered in the cases driven by environmental profiles with MCS initiation during the afternoon between 1100 and 1900 LST, reflecting the key role of environmental profiles in triggering deep convection. The prevailing deep convection in the cases with MCS initiation should be attributed to surface-level warm and moist atmospheric conditions as shown in Fig. 7c. The relatively shallow convection at 1600 LST in Fig. 7b results from the warm environmental temperature in the lower troposphere (see Fig. 6d) so that deep convection is suppressed due to large convective inhibition (CIN). The relatively deep convection at 0800 LST can be explained by a similar argument with cold environmental temperature in the lower troposphere and small CIN. Overall, the

results here recognize the environmental thermodynamic conditions, particularly temperature and moisture, as the controlling factors in determining convective initiation and its diurnal cycle.

d. Impact of climate change on convective initiation

Another interesting research topic is to investigate how the favorability of convection would change in the future climate. Here we introduce two additional parameters δT_e and γ to mimic the impact of global warming on environmental profiles. Specifically, the environmental warming anomaly δT_e describes a uniform shift of environmental temperature profiles throughout the whole troposphere (add a constant warm anomaly to the current temperature profile at all levels), while the environmental moistening ratio γ quantifies an enhancement of specific humidity between future

FIG. 7. Diurnal cycle of convective initiation based on the single-column model simulations and the observation. (a) The green bars (left y axis) shows the final height of the simulated parcel at each hour using the ERA5 environmental profiles (see Fig. 6), while the purple bars (right y axis) shows the total number of the observed MCS initiation based on 14-yr (2004–17) June–August MCS tracking data over the central U.S. region (35° – 45° S, 100° – 90° W). (b) The blue (red) bars show the final height of the simulated parcel at each hour using the ERA5 environmental profiles that are only averaged over hours without (with) MCS initiation. (c) The blue (red) bars show the ERA5 surface equivalent potential temperature anomaly from its climatology in the cases without (with) MCS initiation.

and current climates (multiply the current specific humidity profile at all levels by the ratio). Over land with limited moisture supply from the surface, humidity enhancement may vary independently of the warming anomaly. Here we conduct totally $41 \times 41 \sim 1.6 \times 10^3$ experiments by varying δT_e every 0.125 K from 0 to 5 K and varying γ every 0.005 from 1 to 1.2. It is also worth noting that the parcel's initial water vapor mass q_{par} is assumed to exceed the environmental surface moisture q_{env} by a moisture anomaly δq_v , that is, $q_{\text{par}} = q_{\text{env}} + \delta q_v$.

Figure 8 shows the phase diagram for the minimum dynamical lifting distance δz to trigger deep convection with different combinations of environmental warming anomaly and environmental moistening ratio. The regime in red (blue) color corresponds to shorter (longer) dynamical lifting distance to trigger deep convection and reflects more (less) favorable conditions for convective initiation. When the environmental moistening ratio is fixed, increasing environmental warming anomaly provides more unfavorable conditions for triggering deep convection, presumably due to the larger CIN as a barrier for the parcel to overcome. In contrast, when the environmental warming anomaly is fixed, increasing environmental moistening ratio corresponds to more

favorable conditions for triggering deep convection, which can be simply explained by the enhanced CAPE as the initial water vapor mass fraction in the parcel increases. Based on 37 CMIP5 model projections, the future climate is projected to become warmer and moister with $\delta T_e = 4.52$ K and $\gamma = 1.18$ and it becomes harder to trigger deep convection compared to the present day (represented by the lifting distance at the origin). Figure 8b shows similar results as Fig. 8a but for the minimum boundary layer moistening anomaly to trigger deep convection. Overall, This suppressing of convective initiation in the projected future climate is consistent with the previous study by Rasmussen et al. (2020) concluding that in a warming climate, an increase in CIN suppresses weak to moderate convection and provides an environment where CAPE can build to extreme levels.

4. Cold pool–driven MCS genesis in the multicolumn Lagrangian parcel model

As the second stage of MCS life cycle, the MCS genesis stage refers to the first hour after the convective feature major-axis length reaches 100 km (Feng et al. 2019). It is still a challenging

FIG. 8. The impact of climate change on convective initiation over the central U.S. region (35° – 45° N, 100° – 90° W) in the single-column model simulations. Here we consider two possible environmental thermodynamic profile changes under climate change, including environmental warming anomaly δT_e (increase the temperature profile uniformly by δT_e) and environmental moistening ratio γ (amplify the specific humidity profile by γ). (a) Each pixel corresponds to a simulation using modified environmental temperature and specific humidity profiles due to climate change, and the color shows the minimum dynamical lifting distance δz for deep convection (the parcel reaches a level above 8 km in the equilibrium state). (b) As in (a), but for the minimum boundary layer moistening anomaly δq_e . The star symbol in both panels corresponds to the projected temperature and specific humidity profile changes between future climate (2056–99, RCP8.5) and current climate (1962–2005, historical), based on 37 CMIP5 models.

problem to understand the upscale growth phenomenon when individual convective clouds spontaneously organize into large-scale cloud clusters. To allow spatial dependence, we use the multicolumn model that considers an array of single-column models aligned in the east–west direction. In this study, we focus on studying the roles of cold pool in MCS genesis. The derivation details and model setup are summarized in section 2b. The goal of this section is to reproduce the upscale growth feature, identify key underlying mechanisms, and investigate the impact of various physical processes on MCS genesis.

a. Upscale growth of convection due to cold pool interaction mechanisms

Figure 9 shows the Hovmöller diagrams for the final height of parcels from the multicolumn simulations with/without all the three cold pool interactions mechanisms including (i) the lifting effect by cold pool collision, (ii) the lifting effect by gust front spreading, and (iii) subsidence effect. When all three mechanisms are switched on ($\sigma_{\text{col}} \neq 0$, $\sigma_{\text{iso}} \neq 0$, $\sigma_{\text{sub}} \neq 0$) in Fig. 9a, the solution is characterized by an aggregation behavior that randomly distributed individual convection simultaneously occurring at the initial time organizes later into two large-scale clusters, resembling the upscale growth of MCS genesis. The cluster to the west spans as large as more than 400 km, while that to the east is of a size less than 100 km. Meanwhile, the suppressed convection regimes emerge at time $t = 2$ h after the initiation and keep expanding afterward. It should be interesting to figure out the relative importance of all three mechanisms represented in the model in reproducing this aggregation behavior. Figure 9b shows the solution under the

same setup as Fig. 9a except that the lifting effect by cold pool collision is switched off ($\sigma_{\text{col}} = 0$). The absence of cold pool collision dramatically changes the model output where only several small-scale clusters survive after 12 h and suppressed convection regions dominate the whole domain. This result highlights the key role of cold pool collision in triggering new convection in active convection regions and promoting large-scale organized convection. This is consistent with the previous study by Feng et al. (2015) concluding that intersecting cold pools trigger more convection than isolated ones. The important role of cold pool collision on the upscale growth results from not only its strong lifting strength but also its tendency to trigger new convection between active convection regions, the latter of which promotes the clustering of convection eventually. In Fig. 9c, the lifting effect by gust front spreading is switched off ($\sigma_{\text{iso}} = 0$) instead. The presence of the aggregation behavior indicates that the lifting effects from individual cold pools are less important than the intersecting ones, although its absence does reduce the final size of mesoscale clusters at time $t = 12$ h. This is consistent with the fact that the weak lifting effect from individual cold pools tends to trigger new convection on both sides and help expand the active convection regions (Torri et al. 2015). Figure 9d shows the case when the subsidence effect is switched off ($\sigma_{\text{sub}} = 0$). It is not surprising to find that active convection regions prevail in the whole domain, presumably due to the absence of the subsidence effect to suppress convection in less convective regions. To sum up, this result confirms the dominant role of cold pool collision in promoting large-scale organized convection and that of subsidence effect in maintaining the balance between active and

FIG. 9. Hovmöller diagrams for the parcel height in the multicolumn model simulations (a) with all cold pool lifting effects and subsidence effects, (b) without lifting effects due to cold pool collision, (c) without lifting effects due to gust fronts from isolated cool pools, (d) without subsidence effects. The environmental profiles are fixed at 1500 LST (see Fig. 6).

suppressed convection regions, while the lifting effect by gust front spreading helps expand active convection regions.

Figure 10 shows a detailed diagnostic analysis of the solution in Fig. 9a. As indicated by the blue curve in Fig. 10a, the solution from the multicolumn simulation experiences two stages in the convective column fraction. During the stage in the first hour, the total number of convective columns increases sharply because new convection is triggered continuously while old convection has not terminated yet. After that, this number declines gradually and reaches an even lower level of 17% than its initial value 20% under the impact of the subsidence effect. The second peak of convective column number appears at time $t = 2.5$ h, presumably due to the triggering of the second-generation convection. On the other hand, the geometric mean distance as indicated by the red curve keeps decreasing since the initiation, reflecting the aggregation behavior that individual convection tends to stay increasingly closer to each other to form a large-scale cluster. Figure 10b shows the relative strength of the first 10 spatial modes at every 2 h using Fourier mode decomposition. During the first few hours, the location of all convective columns is more or less randomly distributed; thus, no significant peak in large-scale modes is found. At time $t = 6$ h, the spatial modes with characteristic length scale 500 and 167 km emerge gradually and dominate in the spatial spectrum, corresponding to the two large-scale

clusters in Fig. 9a. This 6-h delay for the appearance of mesoscale clusters since the initiation matches well with the observation that the peak of mesoscale genesis is generally a few hours later than that of convective initiation Feng et al. (2019). After that, the first few spatial modes maintain their dominant strength in the power spectrum, while the remaining modes keep decreasing in strength.

b. Dependence of MCS genesis on subsidence strength and gust front propagation speed

In general, midlatitude MCSs are manifested by a variety of organizational modes in their size and structure (Parker and Johnson 2000). Thus, it should be interesting to identify the key controlling factors in determining the final state of mesoscale clusters from this multicolumn simulation. Here we focus particularly on two parameters associated with the cold pool interaction mechanisms, which are the subsidence strength and gust front propagation speed. Except for these two parameters, all the simulations are under the same setup and compared with each other in terms of three diagnostic metrics. For the sake of robustness, we run four ensemble simulations for each case with only differences in their random initial conditions. In particular, we determine the size of mesoscale clusters based on an algorithm that determines spatially continuous convective columns. There are two tunable parameters in the algorithm,

FIG. 10. Diagnostic analysis about the clustering of convective columns in the multicolumn model simulation as shown in Fig. 9. (a) The blue curve (left y axis) shows the time series of total convective column fraction, while the red curve (right y axis) is for the geometric mean distance of all convective columns ($\sqrt[n]{\prod_{i=1}^n d_i}$, where n is the number of convective column pairs). (b) The spectrum of the first 10 spatial modes at every 2 h using Fourier mode decomposition. Note that the length scale in the x axis corresponds to the half wavelength in each spatial mode.

the parameter *gap* means the minimum number of suppressed convection columns allowed within a mesoscale cluster and the parameter *minimum cluster threshold* exclude small-scale ones to be counted as mesoscale clusters. Physically, these two parameters mimic the MCS convective regions (≥ 50 km) that contain multiple convective updraft cores with spacings of downdrafts in between.

Figure 11 shows the bar diagrams for the total area of cold pools, the total number of convective columns, and the averaged size of mesoscale clusters at time $t = 12$ h from the simulations with various magnitudes of subsidence strength and gust front propagation speed, respectively. As the subsidence strength is enhanced in Fig. 11a, both the total area of cold pools and the total number of convective columns decrease substantially, reflecting the key role of the subsidence effect in maintaining the balance between active and suppressed convection regions. Similarly, the average size of mesoscale clusters also decreases dramatically with the subsidence strength when the subsidence strength is less than 1.0, but at a slower rate when its strength is larger than 1.0. Such a sublinear decrease should be related to the fact that the total subsidence effect depends on not only the constant σ_{sub} but also the ratio between active and suppressed convection columns (see Table 2). Figure 11b shows the cases with fixed subsidence strength but varied gust front propagation speed. As gust fronts propagate at a faster speed, both the total area of cold pools and mesoscale cluster size increase, but the total number of convective columns remains more or less the same. This larger cold pool region and mesoscale cluster size should be related to the further propagation distance of gust fronts. However, due to the fast speed, gust fronts spend less time at each interior column as they pass by, thus inducing fewer lifting effects to those parcels and resulting in a larger gap of suppressed convection within mesoscale clusters. Overall, this result

highlights the important role of subsidence strength and gust front propagation speed in determining the final state of MCSs, providing useful guidance for improving the MCS genesis simulations in the GCMs.

5. Concluding discussion

The state-of-the-art GCMs suffer a long-standing warm and dry bias in simulating the summertime atmospheric conditions over the central United States. It is hypothesized that this bias is mainly due to the poor simulation of MCSs and their initiation in the coarse-resolution GCMs. Here we studied the initiation of summertime MCSs over the central United States based on a simple Lagrangian parcel model.

We first developed a single-column Lagrangian parcel model based on the study by Romps and Kuang (2010). Despite its simple form, this model successfully captures many basic features of an entraining parcel that arises from the surface to the upper level, including mass exchange with the environment through entrainment, water vapor condensation, and precipitation, as well as the imbalance between buoyancy force, gravity and momentum drag. Diagnostic analysis of the specific energy budget highlights the dominant balance between energy gain due to buoyancy force and energy loss due to entrainment and volume expansion. The sensitivity experiments regarding entrainment time scale, initial parcel volume, initial parcel vertical velocity, initial parcel temperature indicate their significant role in promoting deep convection with large amounts of rainfall. This simple model may be a useful tool for predicting convection initiation by adding environmental temperature and specific humidity profiles as input fields. In contrast to the conventional parcel analysis based on the skew T diagram, the simple model also considers the cloud microphysical processes such as entrainment, condensation, and momentum drag and outputs the detailed dynamics of the

FIG. 11. Sensitivity experiments varying (a) subsidence strength and (b) gust front propagation speed in the multicolumn model simulations. In each panel, the blue bars show the total area of cold pool regions (the cold pool region for each parcel is surrounded by its two gust fronts on both directions), the yellow bars show the total number of convective columns (equivalent to total kilometers as each column is assumed to be 1-km width), and the red bars are for the averaged size of mesoscale clusters (maximum nonconvective gap = 20 km and minimum cluster size threshold = 50 km) in four ensemble simulations with random initial conditions.

rising parcel, providing a computationally cheap way to capture the prominent features of convective initiation.

This simple model is then used to study the collective effects of boundary layer moistening and dynamical lifting on convective initiation. The result confirms that both effects could trigger deep convection and result in large amounts of rainfall, although a minimum strength requirement exists for each of them to trigger deep convection in the physically reasonable range. The origins of boundary layer moistening could come from the Great Plains low-level jet (GPLLJ) and soil moisture, while that of dynamical lifting could arise from diurnally varying boundary layer turbulence, convectively generated gravity waves and/or cold pools, and the eastward-propagating large-scale environments and subsynoptic perturbations (Song et al. 2021) such as eastward-propagating waves (Li and Smith 2010), short-wave trough (Tuttle and Davis 2013), and mid-tropospheric perturbations (Wang et al. 2011a,b; Pokharel et al. 2019). It has been found that these large-scale environments and subsynoptic perturbations as propagating environments are correlated with summertime MCS initiation over the U.S. Great Plains (Song et al. 2021). Besides, this simple model is also used to study the diurnal cycle of convective initiation leading up to MCSs and reproduces its early afternoon peak akin to the observation. The direct comparison between cases corresponding to environmental profiles with/without observed MCS initiation indicates surface-level equivalent potential temperature as the controlling factor in determining convective initiation. By replacing the current climate environmental profiles with those from the projected future climate, this simple model predicts that global warming will create environments that are harder to trigger convection, potentially suppressing weak to moderate convection but promoting more frequent severe convection as suggested by the convection-permitting regional climate simulations (Rasmussen et al. 2020).

This model should provide useful insights for improving the dynamic CAPE triggers in convective parameterization for GCMs (Xie and Zhang 2000; Zhang 2002; Song and Zhang 2017, 2018; Xie et al. 2019).

To study the upscale growth of convection to MCS genesis, we further developed a multicolumn model that includes an array of single-column models aligned in the east–west direction and incorporates several cold pool interaction mechanisms. The results emphasize the crucial role of the lifting effect due to cold collision in inducing the clustering of convection and that of subsidence effects in maintaining the balance between active and suppressed convection regions. Besides, the lifting effect due to gust front spreading helps expand the boundary of cold pool regions and increase the size of mesoscale clusters. It takes about 6 h for mesoscale clusters to emerge from initially scattered convection, matching the typical delay between the observed convective initiation and MCS genesis. The sensitivity experiments illustrate that weaker (stronger) subsidence strength and faster (slower) gust front propagation speed promote larger (smaller) cold pool regions and mesoscale clusters. This multicolumn model shares similar features as the self-aggregation phenomenon (Wing et al. 2017). In contrast to the self-aggregation theory that emphasizes the key role of longwave radiation and typically requires tens of days to reach the equilibrium state, the results from the simple model suggest that the cold pool interaction mechanisms provide a more efficient way to promote the upscale growth of MCS genesis. The simple model also suggests a possible way to improve the MCS simulations in GCMs by explicitly resolving the cold pool dynamics or including it in the cumulus parameterization.

The idea that treats convective self-aggregation as a cold pool-driven phenomenon is not brand new. For example, Haerter (2019) proposed a simple model for aggregation by cold pool interaction and concluded with a phase transition

between a continuum and an aggregated state. Haerter et al. (2019) identified the key role of cold pool interactions in triggering new convection and conceptualized cold pool dynamics into a parameter-free mathematical model. Our multicolumn model provides a simpler but more realistic way to simulate the aggregation behavior by resolving the updraft dynamics of parcels. That said, there still seems to be no consensus on whether cold pool dynamics is essential for favoring convective aggregation. For example, Jeevanjee and Romps (2013) found that cold pools are responsible for preventing self-aggregation in small domains. Muller and Bony (2015) showed that convective aggregation is favored when downdrafts below clouds are weak, which is sufficient to trigger aggregation in the absence of longwave radiative feedbacks. Yang (2018) showed that switching off cold pools in the CRM still gives rise to the self-aggregation even without radiative feedbacks, and concluded that boundary layer diabatic processes are essential to convective self-aggregation. On the other hand, although the cold pool effects on organized convection is poorly represented in the cumulus parameterization, present-day coarse-resolution GCMs are still able to simulate large-scale organized convection such as the Madden–Julian oscillation (MJO) (Jiang et al. 2015).

Besides, this multicolumn model neglects a few dynamical factors that may have impacts on organized convection, such as gravity waves and the Coriolis force. In particular, at a scale of the domain size (1000 km), atmospheric gravity waves are likely the most important dynamics that communicates across columns to produce organized convection. In fact, the important role of gravity waves on promoting organized convection has been emphasized in previous studies. For example, Mapes (1993) found that a heat source similar to that of observed MCSs causes, through inviscid gravity wave dynamics, upward displacement surrounding the heating, theoretically predicting that cloud clusters should be gregarious. Stechmann and Majda (2009) showed that in the presence of wind shear or barotropic wind, the gravity waves can create a more favorable environment on one side of preexisting convection than the other side, determining the preferred propagation direction of organized convection. Yang and Ingersoll (2013) suggested that the MJO is an interference pattern of westward and eastward inertia-gravity waves that interact directly with the convection. Thus, it should be interesting to parameterize gravity waves in this multicolumn model to reproduce more realistic features of organized convection in the observation.

The single- and multicolumn models in the current form can be elaborated and extended in several ways. For example, instead of using constant c_v and c_p , we should use moisture-dependent heat capacity coefficients to capture more realistic features of convective initiation. Also, it is better to include the detrainment process to simulate the whole life cycle of parcel dynamics. It is worth noting that the model is driven by ERA5 environmental profiles, and thus, any bias in ERA5 may contribute to the deficiency of the model results when compared to the observed MCSs. To evaluate the uncertainty of model results, one may conduct a set of ensemble experiments based on the hourly ERA5 environmental profiles from each day instead of the climatological mean. Another interesting research direction is to study the

impact of some key synoptic features on convective initiation, such as drylines (Ziegler et al. 1997; Ziegler and Rasmussen 1998), GPLLJs (Gebauer et al. 2018), and the elevated mixed layer (Lakhtakia and Warner 1987; Banacos and Ekster 2010). For the multicolumn simulations, it should be interesting to allow the two-way feedback between convection and cold pools for mimicking the complex interactions in the large-scale organization of convection. One particularly interesting research topic is to study the impact of climate change on the initiation of summertime MCSs over the central United States.

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Data availability statement. The first ERA5 data can be downloaded from the Climate Data Store (CDS) website (<https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-pressure-levels?tab=form>). The second data based on 37 CMIP5 models and the third one for MCS initiation timing as well as all model output in this study are available on application to the corresponding author.

APPENDIX

Derivation of the Cloud Microphysical Closure

Here we write down all the details about deriving the cloud microphysical closure that are used in this study. The goal is to find simple forms for the cloud microphysical closure that capture the essential physical mechanisms. The final forms of cloud microphysical closure as well as the related references are summarized in Table 2.

Based on laboratory water tank experiments of thermal plumes, a precise quantitative description of entrainment ϵ' for a steady updraft structure reads as follows (De Rooy et al. 2013),

$$\frac{1}{M} \frac{\partial M}{\partial z} = \epsilon' \simeq \frac{0.2}{R}, \quad (\text{A1})$$

where ϵ' denotes the fractional entrainment, R is the radius of the rising plume, and $M = \rho w_c a_c$ denotes the upward mass flux, including the density ρ , updraft velocity w_c and fractional area a_c . Here we assume both w_c and a_c vary much slower than ρ with height so that Eq. (A1) can be simplified as

$$\frac{1}{\rho} \frac{\delta \rho}{\delta z} = \epsilon' \simeq \frac{0.2}{R}. \quad (\text{A2})$$

By using Eq. (A2), we can obtain a simple form for the entrainment rate ϵ (note that its unit is $\text{kg m}^{-3} \text{s}^{-1}$) in this model,

$$\epsilon \equiv \frac{\delta \rho}{\delta t} = \frac{\delta z}{\delta t} \frac{1}{\rho} \frac{\delta \rho}{\delta z} = w_c \rho \frac{0.2}{R} = \frac{1}{\tau_{\text{en}}} \rho \left(\frac{V^*}{V} \right)^{1/3}, \quad (\text{A3})$$

where the entrainment time scale $\tau_{\text{en}} = 5R^*/w_c$, $V^* = (4/3)(\pi R^3)$ is the volume of a reference sphere, and w_c is the updraft velocity. Equation (A3) indicates that the entrainment rate is higher when the parcel's density is larger or volume is smaller. In other words, the entrainment rate is mostly significant in the lower troposphere.

The derivation of the parcel's fractional expansion rate starts from the ideal gas law, $pV = mR_{\text{specific}}T$, where p , V , m , and T are the parcel's pressure, volume, mass, and temperature, respectively. R_{specific} is the specific gas constant. When combining with the definition of potential temperature $\theta = T(p_0/p)^{R/c_p}$ and taking time derivative, we can obtain the following equation,

$$\frac{c_v}{c_p} \frac{dp/dt}{p} + \frac{dV/dt}{V} = \frac{d\theta/dt}{\theta}, \quad (\text{A4})$$

where c_v and c_p are the heat capacity coefficient at constant volume and pressure, respectively. Finally, by using the heat budget with latent heat release $c_p T[(d\theta/dt)/\theta] = Le/\rho$ and time tendency of pressure $dp/dt = w(d/dz)p_e(z)$ as well as Eq. (A4), we can obtain a simple form for the fractional expansion rate ψ (note that its unit is s^{-1}),

$$\psi \equiv \frac{dV/dt}{V} = \frac{Le}{c_p T \rho} - \frac{c_v w}{c_p p_e(z)} \frac{d}{dz} p_e(z), \quad (\text{A5})$$

where L is the latent heat of evaporation, e is the evaporation rate, w is the parcel's vertical velocity, and $p_e(z)$ is the environmental pressure profile. Equation (A5) indicates that the expansion of the parcel is driven by both the latent heat release and environmental pressure drop as it gets lifted up to higher levels.

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