Local cloud enhancement associated with urban morphology: evidence from observations and idealized large-eddy simulations

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Previous studies have noted that cities enhance cloud cover, but the mechanisms of urban morphological types on cloud formation remain elusive. Observations of cloud climatology from 44 major U.S. cities show that cloud enhancement increases with the street-canyon aspect ratio and decreases with building density. Here, to explain these observations, we conducted numerical experiments using urban morphology-resolving large-eddy simulations. In these simulations, urban and rural surfaces retain their respective heat-flux differences, while the moisture sources and background atmospheric water vapor are prescribed to be identical, allowing us to isolate the morphological controls on moist convection. Results show that urban morphology influences cloud formation through two mechanisms: taller buildings intensify urban-breeze circulations at the urban-rural interface, while denser buildings, acting as momentum sinks, reduce vertical tur-

bulent transport at the urban core. These vertical motions modify the transport of moisture in the urban atmospheric boundary layer, causing different cloud amounts across different urban morphology. This study highlights the mechanistic link between urban form, vertical motions, and cloud enhancement, thus providing a basis for city-specific boundary-layer convective parameterizations in large-scale weather and climate models.

Introduction

By 2050, it is projected that nearly 70% of the global population will reside in urban areas, resulting in a multitude of challenges for urban sustainability (*1*). Urbanization is marked by the transformation of natural landscapes into built environments, which affects the local hydroclimates (*2, 3*). Unlike the extensively studied urban heat island (UHI) effect and its mechanistic controlling factors (*4, 5*), urban effects on clouds remain understudied (*6*). It is well-known that land-atmosphere coupling affects the atmospheric boundary-layer (ABL) cloud formation (*7–9*). Meanwhile, recent studies (*10*) show that it is a widely occurring phenomenon that cities enhance the formation of local clouds. Yet, the effects of urban morphology, such as a city's average building height and density of the built-up areas, on the boundary-layer cloud formation through modifying the urban surface-atmosphere exchanges still remains elusive.

Compared to rural surroundings, an urban area with its stronger momentum and thermal surface heterogeneities (*6, 11*) can potentially affect the formation of clouds. The sizes of urban surface heterogeneities range from momentum heterogeneity represented by urban morphology at the micro-scale (~ 100 m) to thermal heterogeneity related to urban-rural contrast of surface heating at the convective scale (~ 10 km). At the convective scale that is commensurate with the city scale, larger and more prolonged sensible heat flux compared to rural surroundings (*12*) leads to a deeper and more unstable boundary-layer, transporting moisture to a higher atmospheric level and increasing cloud amount above the urban surface (*6, 13*). Such thermal heterogeneity also leads to the urban breeze circulations (UBCs), (*14–16*), which transport moisture from rural surfaces to the urban atmosphere (*17*). Studies pointed out that the structure of UBC depends on the difference in heating rates between urban and rural surfaces, the background stratification, urban morphology such as the dimension of the city, and urban canopy parameters such as building height and building density (*15, 18*). However, the role of urban morphology on UBCs and the subsequent impact on local cloud patterns is still not well understood.

Compared to the influence of surface heating on moisture transport and boundary-layer clouds in cities (*6, 10*), the impact of micro-scale momentum surface heterogeneity, i.e., roughness elements such as buildings and trees, on local clouds is even less studied (*19, 20*). However, at the scale of the roughness elements, their geometric structures and spatial arrangement can significantly affect the momentum and scalar transport in urban environments (*21–27*). When background flow exists in addition to urban surface heating, the enhanced moisture convergence due to roughness at the urban center (*28–31*) could increase vertical mixing and boundary-layer height (*32*), facilitating the formation of clouds. On the other hand, lower scalar transfer efficiency compared to momentum over urban canopy has been noted by previous studies (*25*). The ventilation efficiency, represented by the air exchange rate of a city decreases with increasing city size (*33*) and increasing building height (*34*). As a result, the presence of urban canopy may also suppress vertical mixing, decrease ABL height, and impede the vertical transport of moisture and formation of clouds over urban surface. Therefore, the contrasting effects of urban roughness on scalar transport could lead to distinct modifications of the local cloud formation process, especially considering a wide range of geometric characteristics of urban roughness. Yet, the underlying mechanisms remain unclear.

A recent study using observational data over 400 cities in the US revealed the urban signatures on local clouds (*10*), but the mechanisms of city-scale surface heating and micro-scale urban morphology affecting the local cloud formation cannot be disentangled from the observations. As a city's micro-scale surface heterogeneities can be characterized by different morphological parameters, it could be speculated that if micro-scale urban surface heterogeneities do affect cloud patterns, observational data will show the "signatures" of cities corresponding to different morphological types. Using the concept of local climate zones (LCZs) (*22*) might provide one potential way to characterize urban morphological types within the urban morphology parameter space (Figure 1 (a)). At the city-scale, a simple geometric mean of the urban morphology parameters (Figure 1 (b)) characterizes the entire city as one type of morphology. Note that despite the known caveats of LCZs for land-cover, here we only apply it to systematically differentiate morphological types measured by some key geometric parameters, such that each individual city can be described by a "basic unit" of morphological type. Thus, this consideration motivates the current study to set up numerical simulations that explicitly resolve such "basic unit" of urban morphological types and micro-scale turbulence using large-eddy simulations (LES) (*35–37*). An idealized setup of free convective condition, i.e., in absence of a background horizontal wind, is considered in LES. The idealized setup provides the first necessary step to understand the mechanisms of micro-scale surface heterogeneity impacting the formation of shallow cumuli, which will help to interpret satellite observations of cloud climatology (*10*) across cities characterized by different morphological types. Two research questions will be examined in this study:

1. Under free-convective conditions, how do different aspects of urban morphology modify vertical velocity, and how do these changes influence the formation of boundary-layer clouds?

2. Can the mechanistic relationships identified in the LES be detected in satellite-based cloud climatology, and how can they provide a basis for improving the parameterization of cloud formation in larger-scale weather and climate models?

Results

Simulated results of flows and formation of clouds over different LCZs

Numerical simulations using LES (FastEddy®) were performed (*35–37*) for seven morphological types of roughness based on the concept of local climate zone, which is a widely adopted way to represent different urban land cover land use types in numerical modeling (*38*). Applying the concept of LCZ in numerical simulation setup also ensures the consistency between numerical modeling and observational analysis shown in later section. The computational domain is shown in Figure 2. The summary of study cases and corresponding parameters are presented in Table 1. Details of the LES model and simulation setup can be found in Materials and Methods. Due to the idealized condition, we expect the findings remain valid for the urban-rural flow regimes that can be categorized as 'bubble' in (*29*). Specifically, if the ratio between background wind and the convective velocity scale at rural area, i.e., the background w_*, is smaller than 0.7, the results are expected to be applicable. The background w_* is defined as:

$$w_* = (\frac{g}{\theta_0} H_r z_i)^{1/3}, \tag{1}$$

where $\theta_0 = 300$ K is a referential temperature; H_r is the surface flux at rural area; and z_i is the initial boundary-layer height.

Two cases (LCZ1 and LCZ8) with the most contrasting spatial patterns of cloud formation are first analyzed in Figure 3. LCZ1 refers to the case with compact high-rise urban canopy and LCZ8 refers to sparse low-rise urban canopy. The distinct cloud patterns in LCZ1 and LCZ8 (Figure 3 (a) and (b)) are quantified by the liquid water path (LWP), defined as

$$LWP = \int_0^{L_z} \rho_d q_l dz, \tag{2}$$

where ρ_d is dry air density; q_l is liquid water mixing ratio; and L_z is the height of the computational domain. The cloud patterns over rural area in both cases are similar, which is expected because of the same surface heat and moisture fluxes being imposed. However, cloud patterns differ over the urban area. For compact high-rise urban canopy (LCZ1), most of the clouds form close to the urban-rural interface; whereas in sparse low-rise urban canopy (LCZ8), clouds form over the entire urban area, especially over the street intersections. Other cases of different morphological types are shown in supplementary information (Figure S7-11). Noticeably, the most compact high-rise case (LCZ1_dense) shows enhanced cloud formation at the corners of the urban region. Cloud formation over urban core is enhanced for cases with wider canyon width (d), i.e., cases LCZ5 and LCZ6. These results imply that spatial patterns by the contrast in roughness height at the rural-urban interface, as well as the geometry of building arrangement within the urban core, which will be further explored in the next section by considering two velocity scales.

To explain the differences in the spatial distribution of LWP, we examine the x and z components of the velocity averaged in y direction (between the white dotted lines in Figure 2) at the vertical plane (Figure 4 (a)-(d)). For both cases, two distinctive regimes of flow patterns can be identified over the urban area: the urban-rural interface, where UBC is present and the urban core (See schematic in Figure 5 (a)). For compact high-rise urban canopy, the distinctive structure of UBC appears in the urban-rural interface with large magnitude of updraft and downdraft (Figure 4 (a) and (c)). Within the urban core, w in compact high-rise urban canopy has a smaller magnitude than that in the interface, implying reduced turbulence mixing. In contrast, for sparse low-rise urban canopy (Figure 4 (b) and (d)), magnitudes of w in the rural-urban interface and urban core are similar. The coherent updrafts and downdrafts, especially below z = 250 m, are imposed by the length scales of the surface morphology (D and L). Throughout the urban core, updrafts of length scale $L-D$, i.e., the street width, can even extend beyond the boundary layer height, in contrast to the magnitude and spatial organization of w in compact high-rise urban canopy.

The difference in velocity fields between sparse low-rise urban canopy and compact high-rise urban canopy implies that the compact high-rise urban morphology (LCZ1) intensifies UBC while suppresses the vertical turbulent transport of momentum inside the urban core. The transport of heat in the two cases also differ, as a deeper boundary layer is seen in sparse low-rise urban canopy over the urban area (black solid lines in Figure 4 (a)-(d)). Given identically imposed surface sensible heat flux and initial profile of potential temperature in all simulation cases, the coherent thermals over the urban core in sparse low-rise urban canopy more efficiently transports heat than the turbulent eddies in compact high-rise urban canopy, resulting in a deeper boundary layer.

The turbulent water vapor flux and liquid water fluxes averaged in y direction for compact high-rise urban canopy and sparse low-rise urban canopy are shown in Figure 4 (e)-(h). The vertical turbulent water vapor flux $\langle w'q_v'\rangle_y$ is positive near the surface (q_v is water vapor mixing ratio). The vertical turbulent liquid water flux $\langle w'q_l'\rangle_y$ (q_l is liquid water mixing ratio) is zero below cloud base (L_{CB} = 614 m), where L_{CB} is diagnosed from the simulated fields as the lowest level at which the horizontally averaged q_l exceeds 10^{-4} g kg^{-1}. The regions above cloud base with a large positive $\langle w'q_l'\rangle_y$ corresponds to a large positive $\langle w'q_v'\rangle_y$, which also corresponds to the updrafts extending from the cloud base to heights beyond the boundary layer, i.e., the penetrating thermals. Therefore, the cloud depth, which can be approximately determined by the height difference between the cloud base and boundary layer height (*9*), is determined by the magnitude of updrafts above the cloud base. These observations imply that in addition to previously recognized effect of city-scale surface heating that leads to persistent low clouds (*6*), different urban surface morphological types modulate the spatial patterns of shallow cumuli by modifying the spatial coherence and magnitude of updrafts under free convective condition. Therefore, in compact high-rise urban canopy, UBC dominantly contributes to the upward transport of moisture, leading to most of the shallow cumuli formation close to the urban-rural interface. While in sparse low-rise urban canopy, the transport of moisture is due to strong coherent updrafts over the urban core, unlike the UBC-dominated mechanism in compact high-rise urban canopy. Clouds form over the entire urban area, especially over the street intersections, where the vertical transport of water vapor by updraft is not impeded by buildings.

The effect of urban morphology on vertical velocity

The flows and turbulent moisture fluxes for compact high-rise urban canopy and sparse low-rise urban canopy highlight the close connection between urban morphology and the vertical motions. All of the simulated LCZ cases are further explored in this section to understand the importance of energetic thermals that transport surface moisture beyond the cloud base causing cloud formation. We consider the correlation between LWP and the mean kinetic energy of turbulence contributed by the vertical motions ($TKE_{w,urban}$) per unit area, which is given by half of the variance of vertical velocity fluctuation for the air column over the urban area from $z = L_{CB}$ to L_z. Figure 3 (c) shows that under free convective condition, these two quantities are linearly correlated for all LCZ cases. Therefore, we focus on the characteristics of the near-surface vertical motions, which are strongly modulated by surface morphological types. In addition, the bulk roughness length for each simulated roughness type is used as a diagnostic metric to examine its correlation with LWP (see Figure S12). The value of LWP shows a general decreasing trend with increasing bulk roughness length, consistent with the idea that higher surface roughness may suppress vertical moisture transport due to increased drag. However, since both roughness-induced convergence and drag suppression are simultaneously at play, some intermediate roughness height and density cases show deviations from this trend. This also implies that we need to define new velocity scales to reflect the two mechanisms, rather than relying on the bulk roughness length.

Here we define two velocity scales, w_{UBC} and $\sigma_{w,core}$, which characterize the vertical motions in the urban-rural interface and the urban core, respectively. The magnitude of vertical component of UBC, w_{UBC}, is defined as the average of vertical velocity within the length scale of UBC, L_{UBC}. L_{UBC} represents the distance between the centers of updraft and downdraft of the UBC and is obtained by taking the lag distance that the auto-correlation function of $\langle w \rangle_y$ reaches the minimum value. $(\langle w'w' \rangle_{core})^{1/2}$ varies with z and the maximum of $(\langle w'w' \rangle_{core})^{1/2}$ above the building height is defined as $\sigma_{w,core}$. Analytical scaling analyses of w_{UBC} and $\sigma_{w,core}$ are considered to understand the effect of urban morphology on w_{UBC} and $\sigma_{w,core}$. To further validate the physical relevance of these two velocity scales, we compare their values normalized by the convective velocity scale w_*

(see Figure S13), which confirms that both w_{UBC}/w_* and $\sigma_{w,core}/w_*$ fall within typical convective boundary-layer ranges.

Close to the surface, the direction of UBC is from rural to urban, i.e., from areas with lower sensible heat to high sensible heat as shown in Figure 5 (a). The increased momentum roughness from rural to urban area enhances convergence at the urban rural interface and increases the magnitude of w_{UBC}. This can be shown by considering the mass conservation equation (*31*). Figure 5 (c) indicates that cases with higher building height h tend to have higher values of w_{UBC}, except cases LCZ6 and LCZ1_dense for reasons to be explained later. This means that more intense updrafts are found at the urban-rural interface for cases with taller buildings. In addition, in the urban core, the rate of work done against the drag force imposed by urban roughness reduces the production of vertical velocity variance. More densely packed buildings with a higher λ_p will lead to lower vertical velocity variance. Such a trend is well reflected by $\sigma_{w,core}$ in Figure 5 (d). In fact, in some cases, the magnitude of vertical velocity variance also affects the strength of UBC, as surface heating generated upward motions near the urban-rural interface are less obstructed by buildings with wider spacing. As a result, LCZ6 with lower λ_p has a higher w_{UBC} than LCZ8; LCZ1_dense with a higher λ_p has a lower w_{UBC} than LCZ1.

These observations can be used to derive a parameterization for $\sigma_{w,core}$. Considering a horizontal plane at some height z below the building top shown in Figure 5 (b), $\sigma_{w,b}$ is the standard deviation of vertical velocity fluctuation averaged over regions bounded by the dashed-lines and the lateral surfaces of the buildings. This bounded region has a dimension of δ, as indicated in Figure 5 (b). Beyond δ, the standard deviation of vertical velocity fluctuation $\sigma_{w,0}$ is dominated by surface heating, as this region lies within the urban core but outside the building-influenced zone, where neither urban–rural interface circulations nor building-induced roughness effects are present. For regions within a certain distance, δ, to the lateral surfaces of the buildings, the standard deviation of vertical velocity fluctuation $\sigma_{w,b}$ is smaller than $\sigma_{w,0}$ due to the flows doing work against the drag force. An area weighted average of the standard deviation of vertical velocity over this horizontal plane is

$$\sigma_{w,core} = \frac{(D+2\delta)^2 - D^2}{L^2 - D^2}\sigma_{w,b} + \frac{L^2 - (D+2\delta)^2}{L^2 - D^2}\sigma_{w,0}, \tag{3}$$

in which D is building width and L is distance between two building centers (L is also the repeating unit size as shown in Figure 2).

The analysis of σ_w with respect to the distance to the lateral surfaces of the buildings confirms our assumption that for a certain height, σ_w further away from a certain distance to the buildings ($\delta = 15$ m in this case) approaches a constant value at $z = 15$ m, which is half of the building height for the case with the lowest h (Figure S1 (a)). As shown by the LES results, σ_w increases logarithmically with distance from the building walls (Figure S1 (a)). Increasing σ_w with distance away from the building walls is also found in experimental results in natural convective flows over a vertical plane in wind tunnel and field studies around a high-rise building (*39–41*). Thus, without loss of generality, near the wall:

$$\sigma_{w,b}(x) = \frac{\sigma_{w,0}}{\ln \frac{\delta}{z_{0u}}} \ln \frac{x}{z_{0u}}, \tag{4}$$

in which x is the distance to the wall; $z_{0u} = 0.1$ m is the roughness length of urban surface (including all walls of buildings); and $\delta \ll L$, the expression of $\sigma_{w,core}$ can be written as

$$\sigma_{w,core} = \sigma_{w,0}(1 - \frac{\delta}{L \ln(\frac{\delta}{z_{0u}})} \frac{4\sqrt{\lambda_p}}{1 - \lambda_p}), \tag{5}$$

in which the relation $\lambda_p = D^2/L^2$ is used to replace D. In Equation 5, $\sigma_{w,core}$ decreases with λ_p, which is consistent with our simulated results (Figure 5 (d)). To further validate our model, we compare the results obtained with the proposed parameterization (modeled results) against those from the LES (simulated results). From analysis of the LES results, we took $\delta = 15$ m and $\sigma_{w,0}$ as the vertical turbulence averaged from δ to half of the canyon width $d/2$, the comparison between analytical and simulated results shows good agreements (Figure S1 (b)). Therefore, Equation 5 quantifies the modulating effect of urban morphology on $\sigma_{w,0}$, which can be seen as the turbulence vertical motion above the urban core due to surface heating only. In practice, $\sigma_{w,0}$ could be approximated from ground-based remote sensing (e.g., Doppler lidar) measurements of vertical velocity profiles in urban areas (*6*) or diagnosed from planetary boundary layer schemes (*42*). The geometry-related parameters such as building width D and size of repeating unit L can be extracted from urban morphology dataset. It might be tempting to extend the parameterization to predict LWP over urban regions. However, it should be approached with caution as cloud amount is influenced not only by morphology but also by many other factors, such as city size, the magnitude of surface heat flux and its contrast with rural surroundings, boundary-layer height, background temperature and humidity profiles, and background winds. Nevertheless, current study reveals that the vertically integrated TKE contributed by vertical motions over the urban core (i.e., Figure 3) correlates with the LWP. Future work could consider factors modifying the profiles of σ_w to develop more realistic parameterization for cloud enhancement.

In summary, by examining across distinct urban morphological types, this section has demonstrated that the amount of shallow cumuli is closely connected to the vertical velocity, showing a linear relationship between the liquid water path and the depth integrated turbulence kinetic energy above the cloud base. We also show that magnitude of the vertical velocity characterizing the UBC at the urban-rural interface increases with the building height h due to roughness-enhanced convergence, while its magnitude at the urban core decreases with the building density λ_p. A simple parameterization is proposed to quantify the effect of building block dimension and building density on σ_w.

Observational evidence of urban cloud cover varying with urban morphology parameters

The findings in the previous section based on idealized LES results are useful to understand the importance of vertical motions under free convective conditions for shallow cumuli formation, especially highlighting the distinct roles of the coherent vertical updrafts associated with the UBC at the urban-rural interface and the turbulent mixing above the urban core. It can also be hypothesized from the numerical simulation results that comparing across cities with distinct morphological types, cities with taller and less densely packed buildings tend to show increased amount of shallow cumuli. With this hypothesis in mind, we analyzed the relative cloud enhancement Δ_{Cloud} derived from MODIS cloud masks (MYD35_L2 C6.1) from 2002 to 2020 (See details in Materials and Methods) in 44 major US cities. The analysis focuses on nighttime observations during the warm season months (April–September).

Rather than using simple citywide averages of building height and plan area fraction, which may not accurately represent the morphological structures responsible for turbulent generation and cloud modulation, we define an effective morphology for each city. In this study, the simulations are designed as idealized experiments to theoretically understand the underlying physical processes by which urban morphology influences vertical motions and cloud formation. Given the complex and highly heterogeneous morphological distributions in real cities, using the most representative morphological characteristics is essential. Specifically, we utilize LCZ classifications to identify the dominant urban types within each city, assigning representative values of building height (h) and plan area fraction (λ_p) to each LCZ. These LCZ-weighted values reflect the spatially coherent and thermodynamically meaningful morphology, enabling more realistic and physically interpretable cross-city comparisons. The magnitude of Δ_{Cloud} is analyzed with respect to the street-canyon aspect ratio h/d, building density λ_p, and city size. Given that there are many compounding factors other than urban form parameters, such as the effect of urban aerosols across different cities (*43, 44*) and different nocturnal local climatology (*10, 45*), which jointly impact the variation of relative cloud enhancement, the variation of morphological parameters cannot be regarded as the *only* factor impacting urban cloud formation. Instead, mechanisms revealed by idealized LES will be used to help us interpret the observations.

The Pearson correlation coefficient matrix of analyzed parameters is shown in Figure 6 (a). Δ_{Cloud} is positively (negatively) correlated with h/d (λ_p) with p-values smaller than 0.001 as indicated in Figure 6 (b). Although h/d and λ_p have statistically significant (p-value < 0.05) positive correlations with city size, city size is insignificantly correlated with Δ_{Cloud}. Other statistical tests, such as Spearman correlation coefficient and partial correlation coefficient computed between h/d , λ_p, and Δ_{Cloud}, also show similar results (Table S1). Overall, the mean, median, and standard deviation of Δ_{Cloud} for the quartiles of h/d and λ_p suggest that Δ_{Cloud} increases with h/d and decreases with λ_p (Figure 6 (c) and (d)). Linear regressions of Δ_{Cloud} against h/d and λ_p are performed (Figure S2). Data show that Δ_{Cloud} is always positive (in the range of 0 to 15%). The differences between inland, coastal, and mountain cities do not account for the scatter. The values of h/d and λ_p in majority of the large US cities cluster around 0.5 and 0.3, perhaps due to similarity in urban planning. This could also be an inherent limitation of the LCZ map dataset, where only categorical values of h/d and λ_p are available. High-resolution data products (*46*) of building heights and street dimensions could be applied in future studies to improve the accuracy of h/d and λ_p. To address this limitation, we performed bootstrap sampling of half the cities without replacement, repeated ten thousand times. Distributions of p-values and R-square values are shown in Figure S2, indicating over 75% and 80% of the linear regression results for h/d and λ_p show statistically significant positive and negative correlations, thus confirming the robustness of the results.

Relating the previously discussed mechanisms to the observed trends here: it can be concluded that the increasing urban cloud enhancement for cities with higher h/d might be related to roughness-enhanced convergence; whereas the decreasing trend with λ_p could be a manifestation of the reduced vertical turbulent transport over a denser urban canopy. We do not directly compare the simulated cloud amount quantified by LWP with the observed results because a city consists of different LCZ types (Figure 1 (a)) and many confounding factors could influence Δ_{Cloud}. The LES results over highly idealized LCZ types under simplified boundary conditions are useful for isolating the effect of urban morphology and reveal mechanistic processes, which highlight the importance of the vertical motions generating coherent updrafts favorable for the formation of shallow cumuli.

Discussion

In this study, the effect of different urban morphology on boundary-layer cloud formation is investigated by means of a combination of numerical simulation using LES modeling (FastEddy®) and observations. In the numerical simulations, urban and rural surfaces retain their respective differences in surface heat fluxes, while the moisture sources and background atmospheric water vapor are prescribed to be identical, so that the influence of morphological dynamics on moist convection can be isolated from moisture heterogeneity. Despite the idealized set-up in the numerical simulation and the simplified assumptions in the parameterization of σ_w, useful insights on the effect of urban morphology on local cloud formation are obtained. The key findings are summarized below.

1. Both urban surface heating (*6, 10*) and roughness contribute to the local cloud enhancement. LES results show that under conditions of no background wind, surface heating and roughness-induced convergence modulate the vertical motions in the atmospheric boundary layer. Heat and moisture transport by organized updrafts play a dominant role in the cloud amount and their spatial patterns. For example, higher variance of the vertical velocity fluctuation is positively correlated with the cloud amount as shown in Figure 5 (a). Organized updrafts either at the urban-rural interface ((Figure 4 (g)) or the urban core (Figure 4 (h)) lead to substantial cloud formation.

2. Results from building-resolving LES in this study provide insight into the plausible mech-

anisms by which urban morphological types influence the boundary-layer cloud formation. These results also corroborate cloud climatology observations based on satellite data. Given no background flow and the boundary-layer development solely via sensible heating, taller urban canopy could intensify UBC by convergence. When the building density, λ_p of an urban area increases, the variance of vertical velocity fluctuation over the urban core is reduced because of increased drag. Turbulent vertical transport of heat and moisture in such case is reduced compared to the transport by organized updrafts.

3. Two velocity scales w_{UBC} and $\sigma_{w,core}$ associated with UBC at the urban-rural interface and the strength of upward motions at the urban core can be defined and calculated based on LES results. w_{UBC} increases with h and $\sigma_{w,core}$ decreases with λ_p. These two velocity scales provide a mechanistic basis for parameterizing the effect of urban morphological types on shallow cumulus convection. Capturing such morphology-driven modulation of vertical velocity fluctuations is essential for representing the dynamical effects of different cities under convective conditions in coarser-scale weather and climate models. Due to the limitations of free convective conditions considered in this study, future studies could expand the simulations to mixed convective conditions. Despite limitations in this work, the insights into these two velocity scales could guide future improvements to σ_w parameterizations in models that do not explicitly resolve buildings, ultimately leading to improved PBL schemes and more realistic cloud formation simulations.

The study is restricted in the following aspects. Firstly, due to the assumption of quasi-steady state forcing, the numerical model neglected the diurnal cycle of cloud formation, when the effect of time-varying surface flux is important, potentially leading to intermittent updrafts and more complex interactions with urban morphological effects. Secondly, idealistic boundary conditions, such as a square-shape city, buildings with the same height and width, and urban surface with uniform surface heating are imposed. The moisture source heterogeneities are ignored, which are known to be prevalent in an urban setting (*6, 47*). Thirdly, the urban morphology data are limited by simplifying assumptions in the parameters for LCZ categories. Higher fidelity data might be helpful to adequately distinguish cities of different urban forms. Last but not least, all simulations are conducted under free-convective conditions without background wind. In real urban environments, background wind is often present and can play a critical role in modulating the relative importance of buoyancy- and shear-driven turbulence, thereby influencing urban breeze circulations and local cloud formation. Future studies should investigate how varying background wind speeds affect the transition from free to mixed convection, and how this transition interacts with urban morphological parameters.

Methods

Description of the numerical model

The FastEddy® model, which is a graphics processing unit (GPU)-resident LES model, is used to conduct numerical experiments. The model solves the fully compressible Navier-Stokes equations cast in flux-conservative form on a non-uniform mesh. The full set of non-hydrostatic governing equations are given by:

$$\frac{\partial \rho_d}{\partial t} = -\frac{\partial \rho_d \tilde{u}_i}{\partial x_i} + F_\rho, \tag{6}$$

$$\frac{\partial \rho_d \tilde{u}_i}{\partial t} + \frac{\partial \rho_d \tilde{u}_i \tilde{u}_j}{\partial x_j} = -\frac{\rho_d}{\rho_m}\left(\frac{\partial \tilde{p}'}{\partial x_i} + g\rho_m' \delta_{i3}\right) - \frac{\partial \tau_{ij}}{\partial x_j} + F_{u_i}, \tag{7}$$

$$\frac{\partial \rho_d \tilde{\theta_d}}{\partial t} + \frac{\partial \rho_d \tilde{u_j} \tilde{\theta_d}}{\partial x_j} = -\frac{\partial \tau_{\theta j}}{\partial x_j} + \frac{\tilde{\theta_d} L_v}{T_d c_p} f_{cond} + F_\theta, \tag{8}$$

$$\frac{\partial \rho_d \tilde{q_v}}{\partial t} + \frac{\partial \rho_d \tilde{u_j} \tilde{q_v}}{\partial x_j} = -\frac{\partial \tau_{q_v j}}{\partial x_j} - f_{cond} + F_{q_v}, \tag{9}$$

$$\frac{\partial \rho_d \tilde{q_l}}{\partial t} + \frac{\partial \rho_d \tilde{u_j} \tilde{q_l}}{\partial x_j} = -\frac{\partial \tau_{q_l j}}{\partial x_j} + f_{cond} + F_{q_l}, \tag{10}$$

where $\sim$ indicates the filtered quantity (omitted in main text); the subscripts "d" and "m" denote dry and moist, respectively; u_i (m s^{-1}) is the velocity vector in Cartesian coordinate (i=1, 2, or 3); t (s) is time; ρ_d (kg m^{-3}) is dry air density; T_d (K) is air temperature; q_v and q_l (kg kg^{-1}) are water vapor and liquid water content as a ratio to the mass of dry air; $g = 9.8$ m s^{-2} is gravitational acceleration; $L_v = 2.5 \times 10^6$ J kg^{-1} is the latent heat of vaporization; $c_p = 1005$ J K^{-1} kg^{-1} is the specific heat of dry air at constant pressure; τ_{ij} (m^2 s^{-2}) is the sub-grid scale (SGS) stress tensor; $\tau_{\theta j}$ (K m s^{-1}), $\tau_{q_v j}$ (kg kg^{-1} m s^{-1}) and $\tau_{q_l j}$ (kg kg^{-1} m s^{-1}) are the SGS temperature, water vapor and liquid water fluxes, respectively. ρ_m (kg m^{-3}) is the moist air density given by

$$\rho_m = \rho_d(1 + q_v + q_l). \tag{11}$$

The moist perturbation of density ρ_m' (kg m^{-3}) is calculated as the departure from the dry hydrostatic component ρ_{dh} (Pa) as

$$\rho_m' = \rho_m - \rho_{dh}. \tag{12}$$

θ_d (K) is dry potential temperature, and is related to temperature and pressure by

$$\theta_d = T_d(p_0/p)^{R_d/c_p}, \tag{13}$$

where p (Pa) is air pressure; $p_0 = 10^5$ Pa is a reference pressure; $R_d = 287.053$ J K^{-1} kg^{-1} is dry air ideal gas constant. p' (Pa) is the pressure perturbation given by

$$p' = [\frac{R_d \rho_d \theta_m}{p_0^{R_d/c_p}}]^{\gamma} - p_{dh}, \tag{14}$$

where $\gamma = c_p/c_v = 1.4$ is the ratio of specific heat of dry air at constant pressure to constant volume; and θ_m (K) is the modified potential temperature given by

$$\theta_m = \theta_d[1 + (R_v/R_d)q_v], \tag{15}$$

where $R_v = 461.5$ J K^{-1} kg^{-1} is water vapor ideal gas constant.

The f_{cond} (kg m^{-3} s^{-1}) term in Equation 8, 9, and 10 represents the transfer between water vapor and liquid water (cloud) from condensation ($f_{cond} > 0$) and evaporation ($f_{cond} < 0$) processes. A saturation adjustment assumption is adopted assuming a homogeneous distribution of thermodynamical properties within a grid volume, and condensation only occurs when the grid solution becomes supersaturated with respect to liquid. The condensation forcing term is taken from Rutledge and Hobbs (1983) (*48*), and has the following form:

$$f^*_{cond} = \frac{\rho_v - \rho_{vs}}{1 + \frac{L_v^2 q_{vs}}{c_p R_v T_d^2}}, \tag{16}$$

where $\rho_v = \rho_d q_v$, $\rho_{vs} = p_{vs}/(R_d T_d) = \rho_d q_{vs}$. q_{vs} is the saturation water vapor mixing ratio. p_{vs} is the saturation vapor pressure computed using the 8th-order polynomial fitting from Flatau et al. (1992) (*49*). The amount of evaporated liquid is bounded by the available cloud water:

$$f_{lim} = \rho_d q_l, \tag{17}$$

and the final forcing becomes

$$f_{cond} = \frac{max[f^*_{cond}, -f_{lim}]}{\tau_{cond}}, \tag{18}$$

where τ_{cond} is the relaxation time scale which is an adjustable parameter that provides a time scale for the condensation forcing that is consistent with the resolvable scales. Here $\tau_{cond} = 1.0$ s is adopted and the sensitivity of the resulting liquid water distribution and turbulence quantities to the choice of τ_{cond} was tested by Muñoz-Esparza et al. (2022) (*37*). Despite idealized numerical setup, the equations and parameters were tested in Muñoz-Esparza et al. (2022) (*37*), such that local cloud formation under free convective conditions can be achieved.

F_{u_i} (kg m^{-2} s^{-2}) represents the force imposed by the buildings. F_{u_i} is modeled using the immersed body force method (IBFM) from Chan and Leach (2007) (*50*):

$$F_{u_i} = -C_d \rho_d |\tilde{u}_i| \tilde{u}_i, \tag{19}$$

where C_d is the drag coefficient modeled as

$$C_d = \alpha_m \beta_r \Delta^{-1}, \tag{20}$$

where $\Delta = \sqrt[3]{\Delta x \Delta y \Delta z}$ is the nominal grid size; $\alpha_m = 1000$ is a constant coefficient; and β_r represents the ratio of volume in a cell occupied by the immersed body. To rectify the small error due to the leakage of air flow into the immersed body, which tends to homogenizes the scalar fields across the building interface, similar forcing terms (F_ρ (kg m^{-3} s^{-1}), F_θ (kg K m^{-3} s^{-1}), F_{q_v} (kg m^{-3} s^{-1}), and F_{q_l} (kg m^{-3} s^{-1})) are added to the mass and energy conservation equations:

$$F_\rho = -C_t |U_s| (\rho_d - \rho_{ref}), \tag{21}$$

$$F_\theta = -C_t |U_s| (\rho_d \tilde{\theta}_d - (\rho_d \theta_d)_{ref}), \tag{22}$$

$$F_{q_v} = -C_t |U_s| (\rho_d \tilde{q}_v - (\rho_d q_v)_{ref}), \tag{23}$$

$$F_{q_l} = -C_t |U_s| (\rho_d \tilde{q}_l - (\rho_d q_l)_{ref}), \tag{24}$$

where $U_s = 1$ m s^{-1} is the velocity scale, $C_t = \alpha_t \beta_r \Delta^{-1}$ for $\alpha_t = 10$, and the subscript 'ref' refers to the reference conditions. Details and validation of the IBFM are provided in Muñoz-Esparza et al. (2020) (*35*).

The SGS stress tensor and scalar fluxes are modeled using a prognostic turbulence kinetic energy (TKE) equation (*51, 52*). All prognostic quantities are discretized at the grid cell center. The fifth-order weighted essentially non-oscillatory (WENO) scheme (*53*) is implemented as advection scheme. The third-order Runge-Kutta scheme (*54*) is used for time advancement. Heat and moisture sources are imposed at the ground surface (not including the walls of the buildings), detailed set-up is given in Domain configuration and simulation set-up.

Domain configuration and simulation set-up

A total of seven simulations are performed. The details of the simulation setup are presented in Table 1. The configuration of rural and urban areas are depicted in Figure 2. The length, width, and height of the computational domain are $L_x = 11.28$ km, $L_y = 7.455$ km, and $L_z = 3$ km, respectively. The number of grids in horizontal x and y, and vertical z directions are $nx = 1504$, $ny = 994$, and $nz = 122$, respectively. The grid size in the horizontal direction is $dx = dy = 7.5$ m; and the grid sizes in the vertical direction stretch from $dz = 7.5$ m close to the ground to $dz = 58$ m at the model top. Time step, dt, for all simulations is 0.02 s. The initial boundary-layer height, z_i, for all simulations is 520 m. The urban area has a horizontal width of $L_c = 3750$ m and is located at the center of the domain. The rows of buildings align with each other. Free convective condition is assumed for all simulations, i.e., no mean background horizontal wind.

All four lateral boundaries are periodic. Given the domain dimensions defined above, the urban footprint and its induced circulations remain well separated from the lateral boundaries. As a result, the influence of the urban area on the boundaries is negligible. A zero gradient top boundary condition is imposed for momentum and all scalars. Monin-Obukhov similarity theory is adopted to impose the momentum wall model for the bottom surface. In all simulations, the momentum roughness lengths of the rural and urban surfaces (excluding all walls of the buildings) are set as $z_{0r} = 0.02$ m and $z_{0u} = 0.1$ m, respectively (*35*). The effect of the buildings on the flows are modeled using the IBFM, as described in the previous section. The scalar roughness length z_{0s} is set to be 1/10 as large as the momentum roughness lengths in corresponding regions. A heat flux bottom boundary condition is applied in the simulation. Specifically, the spatially-averaged surface heat fluxes for rural and urban areas are $H_r = 0.05$ Km s^{-1} and $H_u = 0.056$ Km s^{-1}, respectively, following the report of (*55, 56*). For urban area, the heat flux for all the walls of the buildings is zero, and the heat flux for the ground surface varies with building density, λ_p, in order to keep the same total heating rate for all cases. The sources of water vapor are introduced into the simulation through a constant surface water vapor mixing ratio (in both space and time) of $q_{v,0} = 0.017$ kg kg^{-1} for both rural and urban surfaces in all simulated cases, in order to control variables and isolate the effects of urban morphology on cloud formation. Zero flux liquid water boundary conditions are applied at the rural, urban, and all walls of the buildings.

The flow is initialized with random noise added to a zero velocity profile. A four-layer atmospheric boundary layer (ABL) structure for initial temperature and water vapor profiles is imposed, as shown in Figure S3. Below the initial boundary layer height $z_i = 520$ m, a constant initial temperature of $\theta_i = 298.7$ K is imposed; the water vapor concentration at $z = z_i$ is 0.0163 kg kg^{-1}. This ABL structure is identical to that in (*37*), which was directly compared with the LES results of (*57*), and those LES results were validated against field measurements from the BOMEX campaign (*58*). The initial liquid water concentration is set to zero across the entire domain. After the simulations reach quasi-steady state, instantaneous outputs of velocity, temperature, water vapor, and liquid water are analyzed.

The simulated results are the instantaneous outputs at $t = 1950$ s for all simulated cases. The pseudo-color plots in Figure 4 (a)-(h) are averaged in y direction within the two dashed lines shown in Figure 2. The width between the two dashed lines is $1/5L_C$. The horizontal averaged quantities are denoted with $\langle\ \rangle$. Subscript x (y) indicates averaging in x (y) direction; subscript $urban$ ($core$) indicates horizontal average over urban area (urban core). The turbulent fluctuation of a variable is calculated by subtracting its horizontal mean from that variable (e.g., $w' = w - \langle w \rangle_{x,y}$). Additional analyses at earlier times (e.g., 1650 s and 1800 s) show lower cloud amounts and a shallower boundary layer, but no qualitative differences in the results or conclusions.

Observational dataset

According to findings in (*10*), the urban-rural land surface temperature difference and urban size are the most relevant controlling factors for the nighttime cloud enhancement during warm seasons. Therefore, we expect that the effect of urban morphological types of each city manifest the most in this subset of observational data. Nighttime cloud cover during warm seasons (April to September) over 44 cities with sizes greater than 1000 km^2 are analyzed. This sampling also aligns our observations with the free-convective LES setup, because nocturnal winds are typically weaker than daytime winds, making calm-wind conditions more likely (*59*). The selected cities are listed in Table S2, which contain 23 inland, 13 coastal cities, and 8 mountainous cities. Coastal cities, which are potentially subjected to land-sea breeze influences, are defined as those located within 70 km from the sea (*60, 61*). Mountainous cities are identified with a greater than 1,000-m elevation difference with the highest point within 70-km buffered surrounding areas. Global Multi resolution Terrain Elevation Data 2010 (GMTED2010) is used to identify the elevation difference. The rest of the cities are identified as inland cities. The cloud cover is obtained from MODIS cloud masks (MYD35_L2 C6.1) from 2002 to 2020. The cloud enhancement, Δ_{Cloud}, is defined as the relative difference between cloud cover over an urban area , $cloud_{urban}$, and that over the surrounding rural area, $cloud_{rural}$, i.e., $\Delta_{Cloud} = (cloud_{urban} - cloud_{rural})/cloud_{rural}$. According to (*10*), the urban areas are defined using US Census urban areas product, and the background is each urban domain's surrounding region but excludes a transition buffered area to avoid urban-influenced clouds in upwind and downwind directions in the atmosphere. The transition and background areas are buffered proportional to the urban size. Waterbodies are removed across the urban-background domains to avoid nonurban influences. A t test is conducted to identify the significance of Δ_{Cloud} for individual cities in each given month over 18 years at $p = 0.05$ level (*10*). The uncertainties and potential bias of the cloud cover from MODIS are reported by (*62–64*).

Two urban form parameters are selected for analysis: (1) street-canyon aspect ratio h/d, and (2) building density λ_p. Here h and d are building height and canyon width, respectively. Other urban form parameters such as sky view factor and impervious area fraction are correlated with the selected parameters, and are therefore not selected for analysis. The selected urban form parameters are calculated based on the LCZ map of the continental United States at a 100 m spatial resolution, derived from multiple Earth Observation datasets and expert and crowd-sourced LCZ class labels (*65, 66*). The LCZ map for each city is derived from the continental United States LCZ map, as shown in Figure S4. Since the urban morphology near the center of the city (referred to as urban center hereafter) has the most variations among different cities, the spatially averaged h/d and λ_p over urban center are calculated to represent the urban morphology. The urban center of each city is defined as a circular area of 400 km^2, centered at the city's geographic center. A sensitivity test is conducted to show that the results are not sensitive to the selected urban center size (Figure S5 and S6). The spatially averaged h/d and λ_p are derived from the distribution of h/d and λ_p over the urban center of each city, converted from the city scale LCZ map according to the typical values of each LCZ type (Table 3 in (*22*)).

Data Availability

The Python data post-processing scripts used in this study are available on Zenodo at https://doi.org/10.5281/zenodo The full original LES output data are not publicly available due to their large size. The cloud cover is obtained from MODIS cloud masks (MYD35_L2 C6.1) from 2002 to 2020. The LCZ map is derived from multiple Earth Observation datasets.

Code Availability

The FastEddy® model is publicly available via github: https://github.com/NCAR/FastEddy-model, with the urban capabilities utilized in this study incorporated starting with version 4.0.

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Author Contributions Statement

Y.C. and Q.L. designed and conceptualized this research. D.M. and J.S. developed the LES code, designed the LES setup, and run the simulation. L.H. provided the observational dataset. Y.C. analyzed the numerical and observational dataset and developed the analytical models. Y.C. produced the visualizations, and wrote the manuscript draft. Y.C., S.C., L.X., D.M., J.S., L.H., J.A., and Q.L. revised the manuscript.

Competing Interests Statement

All authors declare that they have no conflicts of interest.

Table 1: Summary of study cases and corresponding parameters. LCZs represents local climate zones. h: building height; D: building width; d: canyon width; λ_p: building density.

Case number	Description	h (m)	D (m)	d (m)	λ_p	h/D	h/d
LCZ1	compact high-rise	200	120	60	0.444	1.667	3.333
LCZ1_dense	compact high-rise	200	120	45	0.524	1.667	4.444
LCZ3	compact low-rise	45	120	60	0.444	0.375	0.75
LCZ4	open high-rise	100	60	75	0.196	1.667	1.333
LCZ5	open mid-rise	75	90	120	0.18	0.833	0.625
LCZ6	open low-rise	30	60	75	0.196	0.5	0.4
LCZ8	sparse low-rise	30	300	262.5	0.284	0.1	0.114

Main Text Tables

Figure 1: The local climate zone (LCZ) map of an example city. (a) Color scale indicates the LCZ type with corresponding urban morphology parameters such as building height (h), street width (d), and building density (λ_p), defined as ratio of plan area to the total urban surface area. (b) The average of the parameters across all LCZ types represents the city-scale urban morphology type that differs across different cities. LCZ 1: compact highrise; LCZ 2: compact midrise; LCZ 3: compact lowrise; LCZ 4: open highrise; LCZ 5: open midrise; LCZ 6: open lowrise; LCZ 7: lightweight low-rise; LCZ 8: large lowrise; LCZ 9: sparsely built; LCZ 10: heavy industry; LCZ 11: dense trees; LCZ 12: scattered trees; LCZ 13: Bush, scrub; LCZ 14: low plants; LCZ 15: bare rock or paved; LCZ 16: bare soil or sand; LCZ 17: water.

Figure 2: Schematics of large-eddy simulation (LES) computational domain. L_x and L_y are length and width of the computational domain, respectively; L_C is the size of the urban area; L is the size of a repeating unit; D is width of the building. The width between the two dashed lines is $L_C/5$. The 9 buildings are only shown for illustration purpose and each case contains different numbers of aligned buildings.

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Figure 3: Simulated results of cloud pattern and cloud amount. (a)-(b): Distribution of liquid water path (LWP) for LCZ1 and LCZ8. LCZs represents local climate zones. Black dashed lines represent urban-rural interface. (c) Horizontally averaged LWP over urban area with respect to the integrated turbulent kinetic energy (TKE) contributed by vertical turbulent velocity from cloud base to domain top. $TKE_{w,urban} = \langle \frac{1}{2}\rho_d w'w' \rangle_{urban}$. Black solid line is the linear fit of the dataset, $y = 0.418x - 0.998$, $R^2 = 0.94$, $p = 2.6 \times 10^{-4}$. L_{CB} is cloud base height, defined as the minimum height that $q_l > 0$. $L_{CB} = 614$ m for all simulated cases.

Figure 4: Simulated results of velocity field, water vapor and liquid water fluxes. (a), (b) show the vertical profiles of x velocity u; (c), (d) show the z velocity w; (e), (f) show the vertical turbulent water vapor flux $w'q_v'$ averaged in y direction; and (g), (h) show the vertical turbulent liquid water flux $w'q_l'$ averaged in y direction. Two extreme cases in terms of urban form are displayed: (a), (c), (e), (g) correspond to compact high-rise urban canopy (LCZ1); (b), (d), (f), (h) correspond to sparse low-rise urban canopy (LCZ8). LCZs represents local climate zones. Plots are averaged in y direction between the two white dashed lines in Figure 2. Black dashed lines represent urban-rural interface. Black solid lines represent the top of boundary-layer over the urban area. The height of boundary-layer is defined as the height at which the variance of the vertical velocity fluctuation over urban area $\sigma_{w,urban}^2$ decreases to 10% of its near-surface maximum (*67*).

Figure 5: Schematic of urban morphology influences vertical motions and simulated results of two vertical velocity scales. (a) Schematic of urban breeze circulation (UBC) model with urban canopy. u_{UBC} and w_{UBC} are the streamwise and vertical velocity scales of UBC and L_{UBC} is the length scale of UBC. (b) Schematic of the model for vertical turbulence σ_w. The annular areas between the building wall and the dashed box indicate regions that vertical turbulence is influenced by the buildings with width δ. $\sigma_{w,0}$ and $\sigma_{w,b}$ are the standard deviations of the vertical velocity fluctuations without urban canopy and influenced by urban canopy, respectively; D is building width and L is the distance between two building centers (size of the repeating unit). (c) Vertical velocity scale of UBC w_{UBC} for all simulated cases. w_{UBC} is defined as the average vertical velocity within the range of the horizontal length scale of L_{UBC}. L_{UBC} is defined as the lag that the auto-correlation function of $\langle w \rangle_y$ reaches the minimum value. (d) Vertical turbulence above urban core for all simulated cases. $\sigma_{w,core}$ is defined as the maximum standard deviation of w over urban core. Urban core is urban area exclude a region with width L_{UBC} surrounding urban-rural interface.

Figure 6: Observational evidance of urban cloud cover varying with urban morphology parameters. (a) Heatmap of the Pearson correlation coefficient matrix; (b) Heatmap of p-values associated with each correlation coefficient; (c) Relative cloud enhancement Δ_{Cloud} averaged over quartiles of street-canyon aspect ratio h/d, where standard deviation and median for each quartile are also indicated; (d) Δ_{Cloud} averaged over quartiles of building plan area fraction λ_p, where standard deviation and median for each quartile are also indicated. $Q1$ to $Q4$ represents 0-25%, 25-50%, 50-75%, and 75-100% percentiles of h/d and λ_p in (c) and (d), respectively. Error bars represent the range between the minimum and maximum values.