Impact of Land-Surface Moisture Variability on Local Shallow Convective Cumulus and Precipitation in Large-Scale Models

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ABSTRACT

Numerical experiments using a state-of-the-art high-resolution mesoscale cloud model showed that land-surface moisture significantly affects the timing of onset of clouds and the intensity and distribution of precipitation. In general, the landscape discontinuity enhances shallow convective precipitation. Two mechanisms that are strongly modulated by land-surface moisture—namely, random turbulent thermal cells and organized sea-breeze-like mesoscale circulations—also determine the horizontal distribution of maximum precipitation. However, interactions between shallow cumulus and land-surface moisture are highly nonlinear and complicated by different factors, such as atmospheric thermodynamic structure and large-scale background wind. This analysis also showed that land-surface moisture discontinuities seem to play a more important role in a relatively dry atmosphere, and that the strongest precipitation is produced by a wavelength of land-surface forcing equivalent to the local Rossby radius of deformation. A general trend between the maximum precipitation and the normalized maximum latent heat flux was identified. In general, large values of mesoscale latent heat flux imply strongly developed mesoscale circulations and intense cloud activity, accompanied by large surface latent heat fluxes that transport more water vapor into the atmosphere.

1. Introduction

It is a consensus that clouds and their associated precipitation play a very important role in the global water and energy cycle. The ability to accurately represent their effects is crucial in atmospheric numerical modeling. However, as indicated by Washington and Parkinson (1986), the incorporation of moisture processes is one of the more difficult aspects of constructing climate models because of two major problems: (i) a lack of fully understanding the cloud and precipitation physics, and (ii) the horizontal resolution of large-scale models—for example, climate models (on the order of hundreds of kilometers)—is much larger than the scales at which clouds are formed. In large-scale models the cloud–precipitation processes are classified as subgrid-scale mechanisms that need to be parameterized.

Deep cumulus convection has been intensively investigated during the past three decades because of the large magnitude of the energy transformations associated with the phase change of water and the strong updrafts and downdrafts extending throughout much of the troposphere. As reviewed by Cotton and Anthes (1989), various parameterization schemes, such as those of Manabe et al. (1965), Kuo (1965), and Arakawa and Schubert (1974) have been proposed to account for the modulation of convection by large-scale forcings and the feedback of cumulus to the environment. These schemes were relatively successful when applied to large-scale models because large-scale variables strongly control the evolution of deep cumulus.

It has been recognized that another category of convection, known as shallow convective cumulus, may affect the atmosphere at both local and global scales. Cotton and Anthes (1989) mentioned that shallow cumulus clouds also considerably affect large-scale motions, usually over longer timescales. On a global scale, shallow cumulus play a major role in determining the planet's radiation budget and hence its climate. Parameterization schemes developed to date, which were originally intended for use in large-scale models, may not be able to represent local shallow convective cumulus, which are affected by local land-surface characteristics. Esbensen (1978) concluded that the timescale for boundary layer cumuli to reach quasi equilibrium with the environment is significantly longer than the adjustment time for deeper clouds. Considering the importance of shallow convection cumulus and the basic difference between shallow and deep cumulus, it is necessary to develop new parameterization schemes of shallow cumulus for large-scale models, independently of deep cumulus.

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Prior to developing a sound parameterization scheme for local shallow cumulus, it is necessary to understand how land-surface characteristics, under different large-scale atmospheric conditions, affect the development of such clouds and associated precipitation. Different types of landscape variability, such as topography, land-surface moisture, albedo, and vegetation, can significantly affect the local flow by redistributing solar energy into turbulent sensible and latent heat fluxes at the ground surface (Collins and Avissar 1994). However, in the present study, we focus on the effects of land-surface moisture heterogeneity on local shallow clouds and precipitation. As reviewed by Anthes (1984), a number of observational and theoretical studies suggest that convective rainfall seems to be associated with increases in vegetation and with variation in surface characteristics on scales ranging from 10 km to large fractions of continents. Recent studies on mesoscale meteorology such as those of Mahfouf et al. (1987), Segal et al. (1988), and Avissar and Pielke (1989) have shown that differential heating between wet and dry land could generate mesoscale sea-breeze-like circulations. Convergence formed near the boundary between wet and dry regions and updrafts formed at the circulation front, which could result in enhanced cloud development.

While it is generally believed that land-surface moisture variability plays an important role in modulating local cloud development, only a few systematic investigations have focused on this problem. Using satellite images in combination with detailed landscape information, Rabin et al. (1990) observed an appreciable effect of spatial variation in land-surface moisture on cumulus cloud formation over relatively flat terrain. The numerical simulations of Wetzel and Argentinii (1990) and Rabin et al. (1990) showed the role of landscape variability in the formation of clouds within a sea-breeze-like circulation forced by differential heating. However, these studies focused on the timing and amount of cloud cover in the presence of weak organized vertical circulations, and the one-dimensional boundary layer shallow-cloud parcel model used in their study was highly simplified. Thus, there is a need to use more realistic dynamical and detailed cloud models to study the effects of landscape variability on shallow clouds and precipitation, the latter playing an important role in local weather prediction and hydrologic applications, which was excluded from previous studies.

The objectives of the present study are (i) to determine whether landscape variability can enhance cloud formation and associated precipitation, and under which atmospheric conditions; (ii) to identify important parameters characterizing the increase in precipitation due to land-surface moisture heterogeneity and; (iii) to assess the effect of horizontal wavelengths of land-surface forcing on precipitation.

To simulate shallow convective cumulus explicitly with reasonable accuracy, high horizontal grid resolution (ideally smaller than 1 km) is necessary. Due to computational limits three-dimensional simulations of shallow convection over a mesoscale domain are not yet readily possible. Still, other investigators have found two-dimensional simulations useful for understanding these mesoscale phenomena. Nicholls and Weissbluth (1988) investigated the differences between two-dimensional and three-dimensional simulations of a tropical squall line and showed that, in spite of some differences, the basic structure of the system appeared reasonably well simulated by the two-dimensional model. Using a high grid resolution two-dimensional version of a nonhydrostatic model, Nicholls et al. (1990) closely simulated the interaction between sea breezes and deep convection over the Florida peninsula. Thus, in this study, two-dimensional simulations were conducted. They will serve as the basis for a more realistic three-dimensional study that we plan to realize in the future, when appropriate computer resources become available.

2. Simulation description

a. Numerical model

The model used for this study was the Colorado State University (CSU) Regional Atmospheric Modeling System (RAMS), which consists of the full set of nonhydrostatic compressible dynamic equations, a thermodynamic equation, and a set of microphysics equations for liquid- and solid-phase clouds and precipitation. The general characteristics of CSU RAMS are summarized in Pielke et al. (1992), and only the major options used in these simulations are given.

Vertical and horizontal turbulent eddy mixing is parameterized using the 2.5-level scheme suggested by Mellor and Yamada (1982), in which the mixing coefficients are determined from a prognostic equation for turbulence kinetic energy (TKE) formulated by Yamada (1983). A 10-level soil model developed by Tremback and Kessler (1985) was used to calculate the diurnal variations of surface temperature and moisture, which were in turn used to evaluate the turbulent sensible heat, latent heat, and momentum fluxes in the surface layer, based on the Louis (1979) scheme. Longwave and shortwave radiation are parameterized according to Chen and Cotton (1983).

Since this study focuses on shallow convective clouds and precipitation generated primarily by atmospheric planetary boundary layer forcing, it is reasonable to consider only warm-cloud precipitation. The precipitation parameterization in RAMS follows the general philosophy outlined by Kessler (1969), who divided total liquid water into two categories: cloud water and rainwater. Cloud water develops from cloud condensation nuclei; the governing equation for vapor depositional growth is derived in standard cloud physics
Table 1. Initial profiles of atmospheric conditions.

<table>
<thead>
<tr>
<th>Pressure (mb)</th>
<th>Potential temperature (K)</th>
<th>ATM1 humidity (g kg⁻¹)</th>
<th>ATM2 humidity (g kg⁻¹)</th>
<th>ATM3 humidity (g kg⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>978</td>
<td>297.8</td>
<td>13.6</td>
<td>11.8</td>
<td>15.4</td>
</tr>
<tr>
<td>960</td>
<td>299.4</td>
<td>13.8</td>
<td>11.9</td>
<td>15.7</td>
</tr>
<tr>
<td>935</td>
<td>301.6</td>
<td>14.2</td>
<td>12.3</td>
<td>16.1</td>
</tr>
<tr>
<td>845</td>
<td>306.6</td>
<td>12.5</td>
<td>10.8</td>
<td>14.2</td>
</tr>
<tr>
<td>750</td>
<td>309.7</td>
<td>8.8</td>
<td>8.3</td>
<td>8.3</td>
</tr>
<tr>
<td>700</td>
<td>313.9</td>
<td>5.2</td>
<td>5.2</td>
<td>5.2</td>
</tr>
<tr>
<td>600</td>
<td>324.9</td>
<td>6.9</td>
<td>6.9</td>
<td>6.9</td>
</tr>
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<td>335.5</td>
<td>0.7</td>
<td>0.7</td>
<td>0.7</td>
</tr>
<tr>
<td>400</td>
<td>352.0</td>
<td>0.4</td>
<td>0.4</td>
<td>0.4</td>
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<tr>
<td>275</td>
<td>376.2</td>
<td>0.2</td>
<td>0.2</td>
<td>0.2</td>
</tr>
</tbody>
</table>


texts (e.g., Pruppacher and Klett 1978). For rainwater, growth by collision and coalescence can occur both with cloud water and within rainwater. The autoconversion of cloud droplets to rain is based on a threshold average diameter in the droplet distribution (Manton and Cotton 1977).

To simulate the development of the daytime convective boundary layer, all simulations were begun at 0600 (local standard time) and run for a 14-h period. The atmosphere was simulated up to a height of 10 km, allowing for the formation of shallow convection. This depth of atmosphere was represented in the model by 40 grid elements with a high resolution near the ground surface and progressively less resolution with height. A rigid lid was used for the upper boundary condition as in the simulation of Nicholls et al. (1991). A weak dissipative layer 2.5 km in depth was included at the top of the simulated atmosphere to reduce the reflection of gravity waves from the upper boundary. The horizontal domain was 250 km wide and was represented in the model by 500 grid elements with a resolution of 500 m. The time step of integration was 2 s. To allow gravity waves to pass through the interior model walls, the Klemp–Wilhelmson (1978) lateral boundary conditions were adopted for the simulations.

b. Initial atmospheric conditions

Similar to Wetzel and Argentinii (1990), the 28 July 1989 case detailed in the First ISLSCP (International Satellite Land Surface Climatology Project) Field Experiment (FIFE) was selected to be the basic atmospheric sounding for our simulations. Hereafter, this sounding is referred to as ATM1. In this case, winds at all levels of the troposphere were light, with virtually no vertical wind shear. The relative humidity in the atmosphere below 2 km was assumed to have a uniform distribution of 75%. To test the influence of the atmospheric moisture profile on clouds and precipitation, two other composite research soundings were used in which the relative humidity in the atmosphere below 2 km was assumed to be (i) 65% (hereafter referred to as ATM2), and (ii) 85% (hereafter referred to as ATM3). Since the emphasis was on analyzing atmospheric moisture profiles, the same vertical profile of temperature was used. This does not, however, dismiss the important role of the vertical profile of temperature in determining instability on cloud–precipitation evolution, as observed and discussed by Blanchard and Lopez (1985). The atmospheric soundings for these three simulations are summarized in Table 1.

c. Land-surface moisture patterns

The soil type selected for our simulations was a sandy loam. Its main characteristics are shown in Table 2. To assess the effect of land-surface forcing resulting from horizontal heterogeneity, numerical experiments were designed to simulate circulations induced by the juxtaposition of irrigated fields to relatively dry land. For this purpose, the 250-km-wide horizontal domain was divided into a western and an eastern part. The western part consisted of a 75-km-wide domain assumed to be relatively wet (irrigated) land. The eastern part consisted of a 175-km-wide domain assumed to be bare and dry. The size of these two parts was chosen to preclude the circulation that moves from the wet to the dry land to reach the eastern lateral boundary. Simulations were initialized with different atmospheric profiles and land-surface moisture patterns. A summary of initial conditions of the various simulations is provided in Table 3.

3. Results and discussions

a. Impact of land-surface moisture discontinuity

Despite the fact that the entire simulated horizontal domain was exposed to the same solar radiation input, the surface temperature over wet soil did not change at the same rate as that over the dry soil. The solar energy absorbed by a wet surface is mostly used to

Table 2. Characteristics of the sandy loam soil type that was adopted for the simulations. The hydraulic properties are soil water content at saturation ($\theta_s$), soil water potential at which water content ($\psi$) starts to be lower than saturation ($\psi_s$), hydraulic conductivity at saturation ($K_s$), and $b$ is an exponent in the function that relates soil water potential and water content. They are provided by the United States Department of Agriculture textural classes (Clapp and Hornberger 1978).

<table>
<thead>
<tr>
<th>Soil property</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density</td>
<td>1600 kg m⁻³</td>
</tr>
<tr>
<td>$\theta_s$</td>
<td>0.435 m⁻³</td>
</tr>
<tr>
<td>$\psi_s$</td>
<td>−0.218 m</td>
</tr>
<tr>
<td>$K_s$</td>
<td>2.99 m day⁻¹</td>
</tr>
<tr>
<td>$b$</td>
<td>4.90</td>
</tr>
<tr>
<td>Roughness</td>
<td>0.10 m</td>
</tr>
<tr>
<td>Albedo</td>
<td>0.20</td>
</tr>
<tr>
<td>Emissivity</td>
<td>0.95</td>
</tr>
</tbody>
</table>
Table 3. Summary of numerical simulation conditions. Here, $\alpha$ is the ratio between actual soil surface water content and the maximum water capacity that the soil can hold. Allowable values for this parameter range from 0.0, representing totally dry soil, to 1.0, which represents saturated soil. Here, $\sigma_1$ is the $\sigma$ in the 75-km-wide western part of the domain, and $\sigma_2$ in the 175-km-wide eastern part of the domain. Equal value of $\sigma_1$ and $\sigma_2$ means that the horizontal domain is homogeneous, and $U$ is the initial uniform background westerly wind.

<table>
<thead>
<tr>
<th>Simulations</th>
<th>Atmospheric profile</th>
<th>$\sigma_1$</th>
<th>$\sigma_2$</th>
<th>$U$ (m s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SM1</td>
<td>ATM1</td>
<td>1.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>SM2</td>
<td>ATM1</td>
<td>1.0</td>
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<td>0.0</td>
</tr>
<tr>
<td>SM3</td>
<td>ATM1</td>
<td>0.5</td>
<td>0.5</td>
<td>0.0</td>
</tr>
<tr>
<td>SM4</td>
<td>ATM1</td>
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<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>SM5</td>
<td>ATM1</td>
<td>1.0</td>
<td>0.0</td>
<td>5.0</td>
</tr>
<tr>
<td>SM6</td>
<td>ATM1</td>
<td>1.0</td>
<td>0.0</td>
<td>10.0</td>
</tr>
<tr>
<td>SM7</td>
<td>ATM2</td>
<td>1.0</td>
<td>0.5</td>
<td>0.0</td>
</tr>
<tr>
<td>SM8</td>
<td>ATM3</td>
<td>1.0</td>
<td>0.5</td>
<td>0.0</td>
</tr>
<tr>
<td>SM9</td>
<td>ATM2</td>
<td>1.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>SM10</td>
<td>ATM2</td>
<td>0.3</td>
<td>0.3</td>
<td>0.0</td>
</tr>
<tr>
<td>SM11</td>
<td>ATM2</td>
<td>0.5</td>
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<td>0.0</td>
</tr>
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<td>ATM3</td>
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<tr>
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<td>0.0</td>
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<tr>
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<tr>
<td>SM16</td>
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<tr>
<td>SM17</td>
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<td>0.3</td>
<td>0.0</td>
</tr>
<tr>
<td>SM18</td>
<td>ATM1</td>
<td>0.5</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>SM19</td>
<td>ATM1</td>
<td>0.25</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>SM20</td>
<td>ATM1</td>
<td>0.1</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>SM21</td>
<td>ATM1</td>
<td>1.0</td>
<td>0.25</td>
<td>0.0</td>
</tr>
</tbody>
</table>

evaporate water, while it heats a dry surface. As a result, the wet surface is characterized by a low Bowen ratio—that is, a large latent heat flux and a small sensible heat flux—and the dry surface is characterized by a high Bowen ratio. Because the entire planetary boundary layer (PBL) is affected by the magnitude of the sensible heat flux at the surface, very different PBLs are obtained over surfaces with different Bowen ratio. Due to this differential heating rate, a horizontal temperature gradient is generated between the wet part and the dry part of the domain, and a sea-breeze-like circulation is induced, as described by Anthes (1984), Mahfouf et al. (1987), and Segal et al. (1988), among others.

Figure 1 illustrates the evolution of the atmospheric fields obtained in SM1, which is initialized with ATM1 and characterized by a strong gradient of soil moisture between the western and the eastern part of the domain (Table 3). The air temperature difference near the ground surface between the dry and the wet land is about 2 K at 1000 LST (Fig. 1a). It increases until early afternoon, reaching 5 K at 1400 (Fig. 1b) and a maximum of 6 K at 1500 (not shown in Fig. 1). At 1000 the maximum west-east component of the mesoscale circulation that develops between the wet and the dry land is 1.5 m s$^{-1}$. Over the dry land, the vertical motion is upward and some thermal cells are developed from Rayleigh–Bénard convective instability. These cells, however, are suppressed by the easterly wind of the mesoscale circulation. At 1200, the horizontal component of the mesoscale circulation is 4 m s$^{-1}$. Thin clouds form at first in the lower atmosphere (at a height of about 500 m) over the wet land. Later, relatively thick clouds appear at the circulation front. Rabin et al. (1990) observed that clouds seem to develop earlier over wet land—for example, over forested zones—where the Bowen ratio is small. The difference in the timing of onset of clouds is due to the fact that water evaporating from the wet surface moistens the atmosphere in the lower, relatively cool PBL, which favors an early appearance of clouds. These thin clouds produced over wet land, however, do not extend to higher elevation (Fig. 1c). This is due to the combined effect of a weak turbulence activity over the wet surface and the subsidence produced by the mesoscale circulation.

Relatively thick clouds (with a depth of about 800 m) appear in the afternoon over the dry land, as a result of the strong vertical turbulent mixing effect in this part of the domain. The diameter of these shallow cumulus ranges from 2 to 4 km, emphasizing the need to use a high horizontal grid resolution for our simulations. The largest clouds (with a depth of approximately 1500 m at 1400 LST) formed at the circulation front, where the vertical motion is strong and transports moist air to relatively high elevation. The maximum mixing ratio of cloud water and rainwater is about 2.1 g kg$^{-1}$ in these cumulus. Rainfall often reaches the ground surface starting at about 1500, after the mixing ratio of cloud and rainwater reaches its maximum. At 1800 the intensity of the mesoscale circulation decreases and clouds dissipate. After 1800, practically no rain reaches the ground surface.

The impact of land-surface moisture discontinuity on the formation of clouds and precipitation is further illustrated with simulations SM2, SM3, and SM4 (Table 3), which are characterized by different combinations of land-surface moisture in the western and eastern parts of the domain. For illustration, 2D sections of mixing ratio of cloud water and rainwater obtained at 1300 and 1600 LST with simulations SM1, SM2, SM3, and SM4 are shown in Fig. 2. Based on the horizontal distribution of cloud, these simulations can be divided into three categories. The first one, represented by simulation SM1, is characterized by the juxtaposition of two extreme land-surface moistures, resulting in a strong mesoscale circulation. As discussed in the previous section, thin clouds develop early in the day over wet land, and relatively thick clouds form above the dry land and at the circulation front. The second category, represented by simulations SM2 and SM3, is characterized by the dominance of relatively wet land in the entire domain. In these cases, the relatively large surface latent heat flux over the entire domain supplies a large amount of water vapor into the atmosphere. As a result, in the early afternoon, thin clouds with
approximately equal diameter appear in the shallow mixed layer over the entire domain. However, in simulation SM2 the most important clouds formed at the mesoscale circulation front, which developed in the afternoon. In SM3, which is characterized by a homogeneous distribution of soil moisture over the entire domain, clouds appeared to be uniformly distributed over the entire domain. The third category, represented
by simulation SM4, is characterized by a dry land in the entire domain. As can be seen in Fig. 2(iv), the formation of clouds in this case is determined primarily by the thermal cells resulting from Rayleigh–Bénard instabilities, and clouds are randomly distributed over the entire domain. These clouds are very thin and, in general, do not produce a significant precipitation, as explained below.

At 1300 LST, the maximum mixing ratio of cloud and rainwater are 0.5, 0.4, 0.3, and 0.3 g kg⁻¹ for SM1, SM2, SM3, and SM4, respectively. Corresponding values at 1600 are 2.0, 2.6, 2.6, and 1.7 g kg⁻¹. By analyzing our other simulations, which have different patterns of land-surface moisture, it appears that the mixing ratio is large in those cases where strong mesoscale circulations develop, in part because these circulations
favor the formation of clouds in the updrafts region at the circulation front. However, in those cases where a large area of land consists of wet soil, the formation of clouds is enhanced by evaporation from the land surface, even when organized sea-breeze-like mesoscale circulations are relatively weak (as in SM2) or absent (as in SM3). By contrast, when the domain consists mostly of dry land (as in SM4), clouds are small in both depth and width.

In most simulations, the onset of rainfall generally occurred after 1500 and the maximum rainfall rate produced by these shallow convective cumulus was about 6 mm h$^{-1}$. Figure 3 shows the accumulated precipitation at 2000 for these four simulations. It is interesting to note
the impact of land-surface moisture discontinuity on the horizontal distribution of precipitation. When a strong sea-breeze-like mesoscale circulation is generated, such as in SM1 and SM2, precipitation falls mostly in the region close to the circulation front. Otherwise, as in the case of SM3 and SM4, the horizontal distribution of precipitation is determined by turbulence and, therefore, appears to be random.

The effect of land-surface moisture on precipitation appears to be complicated by at least two factors: (i) the intensity of mesoscale circulations and (ii) the amount of water vapor available in the atmosphere. In general, the upward motion in mesoscale circulations is stronger than thermal cells induced by turbulence. Thus, we can expect an increase of cloud and precipitation in heterogeneous areas that can produce such circulations. In fact, Fig. 3 indicates that the maximum accumulated precipitation (11.9 mm) is obtained in SM2, where a mesoscale circulation was created by the juxtaposition of a saturated land and a land at only 50% of saturation. However, the intensity of the mesoscale circulation alone cannot explain why the accumulated precipitation in SM3 was larger than in SM1, since no circulation was obtained in SM3. The most likely reason for this enhanced rainfall is that the entire domain in SM3 is much more wet than in SM1, in which a large part of the domain is completely dry. Under the same atmospheric condition, a wetter surface provides more water vapor to the atmosphere than a drier surface and, as a result, supplies more water for the formation of cloud. This increases the amount of rainwater formed through the autoconversion and accretion processes. Hence, it is not surprising that the homogeneous dry land produces the minimum precipitation (Fig. 3).

Thus, from this numerical experiment, it appears that the distribution and intensity of shallow cumulus precipitation is strongly affected by the spatial distribution of land-surface moisture. Mesoscale circulations and water vapor availability, both dependent upon the land-surface moisture pattern, affect significantly the development of shallow convection. The distribution of precipitation appears to be determined by the dominance of either random turbulent motions or organized mesoscale circulations. It is important to emphasize that the precipitation caused by such a land-surface moisture heterogeneity should be parameterized in atmospheric numerical models, which do not have an appropriate grid resolution to resolve these features.

b. Subgrid-scale fluxes in large-scale atmospheric models

Because of limited computational resources, all atmospheric numerical models, from large-eddy simu-
lation (LES) models to general circulation models (GCMs), are formulated on the basic concept that atmospheric variables can be separated into a resolved (mean) part and an unresolved (perturbation) or subgrid-scale part. The divergence of subgrid-scale fluxes resulting from this separation of scales can significantly affect the resolved flow and, therefore, must be parameterized. Such a parameterization involves the use of an averaging operator. As pointed out by Cotton (1986), the averaging operator and the space and timescales used for modeling these processes can have a major influence on the philosophical approach in developing the model. Deardorff (1974) used an LES model with a horizontal resolution of 125 m to demonstrate that, in most parts of the PBL, the resolvable fluxes were much more important than the subgrid-scale fluxes. By contrast, when the grid resolution of the model is so coarse that mesoscale circulations can develop within a single grid element of the model, as in GCMs, subgrid-scale mesoscale and turbulent fluxes can significantly affect the resolvable flow, as was demonstrated by Pielke et al. (1991), and Avissar and Chen (1993, 1994). Thus, parameterization of subgrid-scale fluxes may require different degrees of complexity in different types of numerical models.

In an attempt to improve current parameterizations of subgrid-scale fluxes in large-scale models, Avissar and Chen (1993) proposed a new set of prognostic equations for mean atmospheric variables in large-scale numerical models, which account for both subgrid-scale turbulent (microscale) and mesoscale fluxes. According to their study, if \( D \) is the horizontal domain represented by one grid element in a large-scale model, and \( d_i \) is the horizontal domain represented by one grid element in a mesoscale model used to simulate \( D \) at a higher resolution, then

\[
D = \sum_{i=1}^{n} d_i, \tag{1}
\]

with \( n \) being the number of grid points in the mesoscale model used to represent the large-scale domain \( D \). Within a mesoscale domain \( d_i \), a variable \( \phi(x, y, t) \) can be partitioned into a resolvable mean part \( \phi \) and a subgrid-scale turbulent part \( \phi' \) giving

\[
\phi(x, y, t) = \phi(d_i) + \phi'(x, y, t), \tag{2}
\]

where \( \phi(d_i) \) is the time- and domain-averaged value of \( \phi \) over the mesoscale grid element \( d_i \), which can be defined as

\[
\phi(d_i) = \frac{1}{\tau A(d_i)} \int_{d_i}^{t} \phi(x, y, t) \, dt \, dA(d_i), \tag{3}
\]

with \( A(d_i) \) being the area represented by \( d_i \) in the model and \( \tau \) the time step of integration. Assuming further that a mean variable at the mesoscale resolution \( \phi(d_i) \) consists of a mesoscale perturbation \( \phi'(d_i) \) superimposed on a large-scale mean variable \( \langle \phi \rangle \), we have

\[
\phi(d_i) = \langle \phi \rangle + \phi'(d_i), \tag{4}
\]

where the domain-averaged variable \( \langle \phi \rangle \), which can be considered as a large-scale (or synoptic scale) variable, is defined as

\[
\langle \phi \rangle = \frac{1}{A(D)} \int_{A(D)} \phi(d_i) \, dA(D). \tag{5}
\]

Consequently, we have

\[
\phi(x, y, t) = \langle \phi \rangle + \phi'(d_i) + \phi''(x, y, t). \tag{6}
\]

Using this decomposition, vertical fluxes within \( D \) can be expressed as

\[
w\phi = \langle w \rangle \langle \phi \rangle + w'\phi' + w''\phi'' + \langle w \rangle \phi' + w'\phi', \tag{7}
\]

where, for brevity, \( \phi(d_i) \) and \( \phi'(x, y, t) \) are written as \( \phi' \) and \( \phi'' \), respectively.

When averaged over \( d_i \), the last four terms on the right-hand side of this equation cancel, giving

\[
\overline{w\phi} = \langle w \rangle \langle \phi \rangle + w'\phi' + \overline{w''\phi''} + \langle w \rangle \phi' + w'\phi', \tag{8}
\]

noting that \( \overline{\langle \phi \rangle} = \langle \phi \rangle; \overline{\phi'} = \phi'; \) and \( \overline{\phi''} = 0 \). When further averaged over \( D \), the last two terms on the right-hand side of Eq. (8) cancel as well, giving

\[
\langle w\phi \rangle = \langle w \rangle \langle \phi \rangle + \langle w'\phi' \rangle + \langle w''\phi'' \rangle, \tag{9}
\]

noting that \( \langle \phi \rangle = \langle \phi \rangle; \overline{\langle w\phi \rangle} = \langle w\phi \rangle; \) and \( \overline{\phi'} = 0 \).

In large-scale models, \( \langle w \rangle \langle \phi \rangle \) is the resolved flux, \( \langle w'\phi' \rangle \) is the domain-averaged subgrid-scale mesoscale flux associated with mesoscale circulations, and \( \langle w''\phi'' \rangle \) is the domain-averaged (over \( D \)) subgrid-scale turbulent flux. It is important to emphasize that in the simulations discussed in this paper, which use a horizontal grid spacing of 500 m, mesoscale fluxes are created by resolved eddies, that is, eddies that range from 2 to 250 km. According to Orlanski’s (1975) classification, these mesoscale circulations belong to the meso-\( \gamma \) (2-20 km) and meso-\( \beta \) (20-200 km) scales. Consequently, turbulent eddies, which are assumed to be subgrid scale, have a scale smaller than 2 km.

As an example, Fig. 4 shows a comparison between domain-averaged turbulent heat fluxes and domain-averaged mesoscale heat fluxes at different times for simulations SM1, SM2, SM3, and SM4. Unlike turbulent heat fluxes whose maximum is located near the ground surface, the domain-averaged mesoscale heat flux is zero at the ground surface, where the vertical velocity is zero. It increases with height in the PBL and reaches its maximum at a height ranging from 300 to 1500 m, depending on the development of the PBL. From this height, mesoscale heat fluxes decrease in an
Fig. 4. Vertical profiles of domain-averaged (a) turbulent sensible heat flux, (b) mesoscale sensible heat flux, (c) turbulent latent heat flux, and (d) mesoscale latent heat flux for (i) SM1, (ii) SM2, (iii) SM3, and (iv) SM4.
Fig. 4. (Continued)
FIG. 5. Diurnal variation of global subgrid-scale (a) sensible heat flux and (b) latent heat flux for (i) SM1, (ii) SM2, (iii) SM3, and (iv) SM4. Units are watts per square meter. Solid line is the total flux, dotted line is the turbulent flux, and dashed line is the mesoscale flux.
approximately linear way with height up to the top of the mixed layer, where it becomes negative, indicating a downward transport of heat in the air aloft. It should be noted that when the domain is homogeneously wet, as shown in Fig. 4(iii) and Fig. 4(iv), mesoscale heat fluxes are produced by the release of latent heat due to formation of clouds. In the absence of moist processes, however, mesoscale heat fluxes are zero in homogeneous domains, as discussed in Chen and Avissar (1994).

Clearly, in the afternoon, when mesoscale circulations are well developed, mesoscale heat fluxes are as important as turbulent fluxes and, from a certain elevation in the PBL, they become even larger than turbulent heat fluxes. This is particularly true for the latent heat. This interesting feature of moisture fluxes agrees in general with the early observations of Nicholls and LeMone (1980), who collected data by aircraft equipped with turbulence measuring instrumentation during the GARP (Global Atmospheric Research Program) Tropical Atlantic Experiment (GATE). They found that the largest contribution to the total moisture variance came from mesoscale eddies having wavelengths on the order of 10 km. Sommeria and Lemone

![Diagrams](https://example.com/diagrams.png)

**Fig. 6.** Diurnal variation of (i) global latent heat flux (solid line is the total flux, dotted line is the turbulent flux, and dashed line is the mesoscale flux); (ii) surface turbulent fluxes (solid line is the total flux, dotted line is the sensible heat flux, and dashed line is the latent heat flux); and (iii) accumulated precipitation (mm) at 2000 LST as affected by a west–east horizontal background wind of (a) 10 m s$^{-1}$ and (b) 5 m s$^{-1}$.
Table 4. Accumulated precipitation $R$ at 2000 LST for the various simulations described in Table 3.

<table>
<thead>
<tr>
<th>Simulations</th>
<th>$R$ (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SM1</td>
<td>5.0</td>
</tr>
<tr>
<td>SM2</td>
<td>8.1</td>
</tr>
<tr>
<td>SM3</td>
<td>8.0</td>
</tr>
<tr>
<td>SM4</td>
<td>0.2</td>
</tr>
<tr>
<td>SM5</td>
<td>1.6</td>
</tr>
<tr>
<td>SM6</td>
<td>0.6</td>
</tr>
<tr>
<td>SM7</td>
<td>5.7</td>
</tr>
<tr>
<td>SM8</td>
<td>16.7</td>
</tr>
<tr>
<td>SM9</td>
<td>0.0</td>
</tr>
<tr>
<td>SM10</td>
<td>0.0</td>
</tr>
<tr>
<td>SM11</td>
<td>0.3</td>
</tr>
<tr>
<td>SM12</td>
<td>4.0</td>
</tr>
<tr>
<td>SM13</td>
<td>12.6</td>
</tr>
<tr>
<td>SM14</td>
<td>14.4</td>
</tr>
<tr>
<td>SM15</td>
<td>10.2</td>
</tr>
<tr>
<td>SM16</td>
<td>15.3</td>
</tr>
<tr>
<td>SM17</td>
<td>5.0</td>
</tr>
<tr>
<td>SM18</td>
<td>0.8</td>
</tr>
<tr>
<td>SM19</td>
<td>0.2</td>
</tr>
<tr>
<td>SM20</td>
<td>16.0</td>
</tr>
<tr>
<td>SM21</td>
<td>34.0</td>
</tr>
<tr>
<td>SM22</td>
<td>40.0</td>
</tr>
<tr>
<td>SM23</td>
<td>5.6</td>
</tr>
</tbody>
</table>

(1978), using an LES model, simulated explicitly the turbulent structure of the trade-wind boundary layer (including the cloud layer) and compared the simulated and observed values of fluxes and variances. They found that the model underestimated the moisture variance when they considered only the turbulence. This implies that mesoscale circulations are more efficient than turbulent eddies in transporting water vapor to higher elevation. This, of course, has a very strong impact on cloud formation.

It is useful to examine the variation of the global subgrid-scale flux, which is defined here as the flux averaged over the entire domain (i.e., horizontally and vertically), since it provides a measure of the total amount of heat and moisture transported into the atmosphere by subgrid-scale motions. The global total subgrid-scale flux is the sum of the global mesoscale and the global turbulent fluxes, which are defined, respectively, as

$$ \langle \langle w' \phi' \rangle \rangle_V = \frac{1}{Z} \sum_{k=1}^{N} \langle w' \phi' \rangle_k \Delta Z_k $$

(10)

and

$$ \langle \langle \overline{w^2 \phi^2} \rangle \rangle_V = \frac{1}{Z} \sum_{k=1}^{N} \langle \overline{w^2 \phi^2} \rangle_k \Delta Z_k, $$

(11)

where the $\langle \langle \rangle \rangle_V$ indicates a global (i.e., volume) averaging, $\langle \langle \rangle \rangle_k$ is the horizontal average at the vertical level $k$, $N$ is the number of vertical layers in the model, and $\Delta Z_k$ is the depth of the layer $k$. In this study, $Z = 5$ km, since both mesoscale and turbulent fluxes are negligible above this height.

The diurnal variation of the global subgrid-scale heat fluxes obtained in simulations SM1, SM2, SM3, and SM4 is illustrated in Fig. 5. From 1400 LST, the mesoscale sensible heat flux becomes larger than the turbulent sensible heat flux. It is interesting to note that the transport of heat by turbulent eddies is very weak during the morning in SM2 and SM3, due to a low turbulence activity. This transport is downward in the afternoon, due to the release of latent heat in clouds, while the mesoscale transport of sensible heat remains upward. Except in SM2, the maximum mesoscale latent heat flux is larger than the corresponding maximum turbulent flux.

c. Impact of large-scale atmospheric conditions

1) Background wind

Interactions between large-scale background wind, turbulence, and mesoscale circulations are highly non-linear and are complicated by various factors, including atmospheric instability. A case study of the effect of atmospheric wind profiles on deep convection has been conducted by Nicholls et al. (1991). A systematic analysis of this effect is beyond the scope of this paper, yet two cases have been chosen to provide some insights into the potential effect of large-scale background wind on rainfall. In one case, the atmosphere was initialized with a vertically uniform west–east wind of 5 m s$^{-1}$ (simulation SM5) and in the other case, with a uniform wind of 10 m s$^{-1}$ (simulation SM6). Mesoscale perturbations of vertical velocities were slightly reduced by the moderate background wind (5 m s$^{-1}$), and perturbations of temperature remain practically unchanged. As a result, the mesoscale sensible heat flux was only slightly affected by the background wind. However, strong background winds larger than 10 m s$^{-1}$ in the direction of the mesoscale flow tend to eliminate the temperature gradient generated by the wet-dry land contrast, and, as a result, the associated mesoscale circulation. Consequently, the mesoscale latent heat flux is reduced by strong background wind, as shown in Fig. 6a.

A strong background wind (i.e., 10 m s$^{-1}$) increases the turbulent sensible heat flux due to increased mixing (Pielke et al. 1991; Chen and Avissar 1994). Note in Fig. 6a(ii) that the maximum surface turbulent sensible and latent heat fluxes in the strong wind case are relatively large. However, this increase in turbulent latent heat flux appeared to have no apparent effect on precipitation. A comparison of the accumulated precipitation obtained in SM5, SM6 [Fig. 6(iii)], and SM1 (Fig. 3a) shows that the maximum accumulated precipitation (4.3 mm in SM1 and only 0.6 mm in SM6) is reduced by the intensity of a background wind. Not surprisingly, a strong background wind causes a shift...
of the maximum precipitation in the direction of the wind.

2) MOISTURE PROFILE

In a drier atmosphere, the surface turbulent latent heat flux increases because more water vapor can evaporate from the ground surface into the atmosphere if, of course, water is available at the soil surface for evaporation. This local supply of water vapor is very important for the formation of clouds and precipitation in a dry atmosphere. As indicated in Table 4, in simulations SM9 and SM10, in which the averaged soil moisture over the entire domain is relatively low, only a small amount of thin clouds appear in the early afternoon and no rainfall is predicted. In simulation SM11 initialized with the same atmospheric profile—namely ATM2—a weak rainfall of 0.4 mm was obtained in spite of the fact that the turbulence activity was much reduced over this relatively wet land, which, on the other hand, releases more water in the dry atmosphere. A maximum rainfall of 5.7 mm was ob-
of maximum accumulated precipitation between SM1 (relatively dry atmosphere) and SM12 (moist atmosphere) is small, and both SM1 and SM12 produce precipitation larger than that obtained in simulation SM9. This implies that if the entire land surface is relatively dry, the presence of a moist atmosphere does not necessarily increase rainfall. In a moist atmosphere, the maximum rainfall is obtained in the case where the land surface is a wet land, as illustrated in Fig. 7(iii). The accumulated precipitation obtained in SM8 is larger than that obtained in SM7 by 11 mm.

**d. Relation between mesoscale latent heat flux and maximum precipitation rate**

As discussed in the previous sections, the distribution and intensity of precipitation is a complex function of land-surface moisture variability, atmospheric instability, atmospheric moisture, and large-scale wind patterns. Thus, to develop a sound parameterization of shallow convective clouds and precipitation accounting for these factors, a considerable combined effort in field experiments, theory, and numerical modeling is required. As a first step toward the development of such a parameterization, we investigated numerically the relation between the maximum precipitation and mesoscale latent heat flux. To further simplify this analysis, we concentrated on some cases, namely, light background wind.

In general, precipitation rate is primarily determined, for a given atmospheric condition, by two factors: (i) the intensity of upward velocity and (ii) the local supply of water vapor. The vertical drafts can be either initiated by the random turbulent eddies (Rayleigh–Bénard thermal cells) or created in the convergence region of organized mesoscale circulations at the circulation front. Therefore, the mesoscale latent heat flux, which is the covariance between the perturbation of vertical velocity and the perturbation of moisture, contains information on both mesoscale circulation and the water in the atmosphere. On the other hand, the surface turbulent latent heat flux is a good indicator of the quantity of water vapor evaporated from the ground surface to the atmosphere. As discussed previously, water vapor can be more efficiently transported to higher elevation by the mesoscale circulation than by turbulent eddies. Thus, the sum of maximum mesoscale latent heat flux and maximum surface turbulent latent heat flux, hereafter referred to as the maximum latent heat flux (MLHF), is a relevant parameter to characterize the variation of maximum accumulated precipitation.

The MLHF obtained in SM5 and SM6 are 522 and 476 W m\(^{-2}\), respectively (see Fig. 6). It is interesting to note that in these two simulations, the maximum accumulated precipitation increases with the MLHF. The difference of MLHF between SM5 and SM6 is due primarily to the difference of mesoscale latent heat flux, as can be seen in Fig. 6b(i). In SM7 and SM8, the
MLHF is 614 and 518 W m\(^{-2}\), respectively (Fig. 7). In this case, the maximum accumulated precipitation does not increase with the MLHF, indicating that this flux alone may not be good enough to accurately represent the variation of maximum accumulated precipitation. Probably, the main reason is that, even with the same MLHF, the development of the precipitation process may be quite different due to the difference of energy input into the atmosphere. This input of energy is well characterized by the total surface turbulent heat fluxes (i.e., the sum of sensible and latent heat flux). Thus, one possibility of resolving this problem may be to normalize the MLHF by the maximum total surface heat flux (hereafter referred to as MSHF).

Results of the 21 simulations presented in Table 4 are summarized in Fig. 8, which illustrates the relation between the MLHF normalized by the MSHF and the maximum accumulated precipitation at 2000 LST. Clearly, these results show that the accumulated precipitation increases with normalized MLHF. The physical explanation for this relation is straightforward; namely, large values of mesoscale latent heat fluxes are typically associated with a strong mesoscale circulation and cloud activity. An increase in surface turbulent
latent heat flux provides more water vapor into the atmosphere and favors the formation of clouds and the quantity of condensed water necessary to initiate rainfall. Furthermore, the upward transport of water vapor is enhanced by mesoscale circulations.

Note, however, that a very weak rainfall (0.3 mm) was obtained with a strong normalized MLHF (0.91) in simulation SM14. In this simulation, the evaporation of water from the ground surface was considerably enhanced due to a combination of a wet land surface and a dry atmosphere. As a result, the maximum surface latent heat flux was as high as 550 W m\(^{-2}\), while the maximum mesoscale latent heat flux was only 12 W m\(^{-2}\), indicating a weak vertical motion and no cloud activity.

A relatively large range of accumulated precipitation is depicted in Fig. 8 when the normalized MLHF approaches 1. But such a dispersion is not surprising, since the 21 simulations include various atmospheric moisture profiles and background winds. When both mesoscale and turbulent transport of moisture are very strong (i.e., normalized MLHF $\geq 1.0$), the response of cloud and rainfall to various atmospheric and land-surface moisture condition is highly nonlinear. This emphasizes the need to account for additional atmospheric conditions in such a relationship.

e. Impact of wavelength of land-surface forcing

The numerical experiments described in the previous sections provided a general description of the impact of land-surface moisture and atmospheric conditions on the intensity and distribution of precipitation. However, Dalu et al. (1991) and Chen and Avissar (1994) showed that the horizontal length scale of the land-surface discontinuity is another important parameter in determining the intensity of mesoscale circulations. In these earlier studies, however, moist processes were ignored. To understand what is the smallest wavelength of land-surface discontinuity that can still significantly affect the intensity of precipitation, three additional simulations were produced. In these simulations, a completely dry land located in the middle of a 250-km-wide domain was flanked by saturated land. The size of the dry patch was 25, 100, and 180 km in SM22, SM23, and SM24, respectively. The initial atmospheric profile adopted for these simulations was ATM1.

The evolution of the global latent heat flux, surface turbulent heat flux, and accumulated precipitation at 2000 LST are depicted in Fig. 9. The maximum surface latent heat flux is considerably reduced by increasing the size of the dry land, from 476 W m\(^{-2}\) in SM22 to 160 W m\(^{-2}\) in SM24. This is because the solar radiation received on the dry land was primarily converted into sensible heat. It is interesting to note that the maximum global mesoscale heat flux and precipitation first increase with the wavelength of dry land, reaching the largest values in the 100-km-wide dry land case (SM23), and decrease with a further increase of the size of the dry patch. The local Rossby radius of deformation varies between 80 and 140 km (depending on the time of the day and the elevation in the PBL) in our simulations. Thus, this result confirms the linear analyses of Dalu et al. (1991), Dalu and Pielke (1993), and the numerical experiments of Chen and Avissar (1994), which indicated that the largest mesoscale circulation should be obtained for a wavelength of forcing corresponding to the local Rossby radius of deformation. Furthermore, Anthes (1984) suggested that circulations forced by a length scale of 100 km appeared...
to be capable of initiating and enhancing moist convection under appropriate atmospheric conditions.

To explain this process, it is helpful to examine the 2D section of horizontal velocity at 1300 and at 1600 LST obtained in SM23 presented in Fig. 10. Two separate mesoscale circulations developed over the two dry–wet land discontinuities in the early afternoon and tended to converge over the central dry area, as can be seen in Fig. 10a. Later, the collision of two mesoscale circulations produced a strong convergent area (Fig. 10b), which extended to higher elevation and produced strong rainfall. The maximum accumulated precipitation was 41 mm. However, as explained by Chen and Avissar (1994), in cases where the dry patch is larger than 180 km, no collision of the two mesoscale circulations is obtained before 2000 and, as a result, precipitation is not enhanced. This also explains why two precipitation centers with approximately equal rainfall can be seen in Fig. 9c(iii), whereas in the other two cases the maximum rainfall was concentrated over the center of the dry area where the collision of the two circulations occurred.

The maximum mesoscale latent heat flux in SM22 is almost identical to that in SM24, but a difference of 28.4 mm in the maximum accumulated precipitation is found between them. This difference can be explained by the large difference in the surface latent heat fluxes between the two simulations. Indeed, in SM22, most of the simulated domain consists of saturated land, which provides a large amount of water vapor, which is essential for the maintenance of clouds and production of heavy rain.

4. Summary and conclusions

The simulations described in this study indicate that a land-surface moisture discontinuity, which generates mesoscale circulations, can significantly affect the timing and onset of clouds and the intensity and distribution of precipitation. The horizontal distribution of precipitation appears to be dominated by one of two processes: turbulent motions or mesoscale circulations. When turbulence only is involved in the heating and transport of moisture in the PBL, the precipitation is randomly distributed with approximately equal intensity. However, intense rainfall is produced at the front of organized mesoscale circulations. The dominance of these two regimes is strongly modulated by the land surface. Thus, it is necessary to incorporate landscape information in large-scale and regional numerical models to accurately predict temperature and cloud fields, which have a strong influence on local weather, climate, and the hydrology cycle.

The response of the atmosphere to land-surface moisture is highly nonlinear and is complicated by various factors, including atmospheric thermodynamic structure and large-scale background wind. Our results indicate that both cloud and precipitation fields are strongly affected by the horizontal distribution of land-surface wetness, in particular in a drier atmosphere. Furthermore, large-scale background wind in the direction of mesoscale flow decreases the intensity of rainfall, and the length scale of the horizontal landscape discontinuity plays also an important role in determining the precipitation. The results suggest that the local Rossby radius of deformation is the optimal wavelength of land-surface forcing to produce precipitation. However, even when the size of the landscape discontinuity is on the order of 20 km, mesoscale circulations can still produce strong rainfall. Note that in absence of moist processes, Chen and Avissar (1994) mentioned that with a wavelength of land-surface forcing smaller than 30 km, mesoscale effects could be ignored as compared to turbulent effects.

In an attempt to provide some insights into the parameterization of shallow convective cumulus and precipitation in large-scale atmospheric models, we discussed the relation between the maximum latent heat flux (MLHF) normalized by the maximum surface turbulent heat flux (MSHF) and the maximum accumulated precipitation. In general, we found that MLHF/MSHF increases with precipitation due to the
clear relation found between the mesoscale latent heat flux and the mesoscale circulations. However, the dispersion of the results indicated that other parameters, such as atmospheric moisture profile and background wind, affect also this relation and should be considered in a quantitative manner for the purpose of parameterization of large-scale models.

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