Investigating the initiation and propagation processes of convection in heavy precipitation over the western Sichuan Basin

LI Qin\textsuperscript{a,b}, YANG Shuai\textsuperscript{a,c}, CUI Xiao-Peng\textsuperscript{a,b} and GAO Shou-Ting\textsuperscript{a}

\textsuperscript{a}Laboratory of Cloud-Precipitation Physics and Severe Storms (LACS), Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China; \textsuperscript{b}College of Earth Science, University of Chinese Academy of Sciences, Beijing, China; \textsuperscript{c}Plateau Atmosphere and Environment Key Laboratory of Sichuan Province (PAEKL), Chengdu University of Information Technology, Chengdu, China

\textbf{ABSTRACT}
Heavy precipitation events occur often over the western Sichuan Basin in summer, near the transition zone between the Sichuan Basin and the steep terrain of the Tibetan Plateau. One such event — a heavy precipitation process that occurred on 18–20 August 2010, with clear nocturnal peaks — is chosen as a case to tentatively explore how the convection associated with convective-scale precipitation is initiated and propagated. By utilizing the vertical momentum equation from the viewpoint of separating perturbation pressure into dynamic and thermal parts, it is demonstrated that the vertical momentum is induced by the imbalance of several forces, including the dynamic/buoyant part of the perturbation pressure gradient force and the buoyancy force, with the latter dominating during the nocturnal-peak period. Although a negative value of the dynamic perturbation pressure gradient force partly offsets the positive buoyant forcing inside the strong updraft, the pattern of vertical motion tendency is largely attributable to its buoyancy because of its larger magnitude. Relative to the buoyancy component, the dynamic part of the vertical perturbation pressure gradient is also examined, revealing a smaller order of magnitude. Thus, it is the thermal effect that should be responsible for the initiation and propagation of convection. As for the convective-scale precipitation, it always presents a trailing morphology relative to the strong leading-side updraft. Furthermore, overlapping strong signals of vertical motion and its tendency point towards strong precipitation in the future.

1. Introduction
Heavy precipitation events frequently occur over the western Sichuan Basin, China, near the transition zone between the Sichuan Basin and the steep terrain of the Tibetan Plateau (TP), especially in the warm season from June to September. Owing to the extremely complex underlying surface conditions, with distinctly steep topography, these events have represented a considerable challenge for weather forecasters. To comprehensively understand the characteristics and mechanisms of the precipitation in this region, many researchers have investigated its spatial distributions and temporal variations (Yu, Zhou, and Xiong 2007; Zhou et al. 2008; Bai, Liu, and Liu 2011; Hu, Lu, and Su 2014), as well as the effects of the orography (Ge, Zhong, and Du 2008; Lu, Yu, and Zhou 2009; Shen and Zhang 2011) and the characteristics of the associated large-scale circulation and weather systems (Yu 1984, 1986; Jiang et al. 2008) in southwestern China. One interesting phenomenon uncovered by these studies is that the frequency and intensity of heavy rainfall over the western Sichuan basin, especially west of 106°E, is higher than in the east (Zhu and Yu 2003). Furthermore, these heavy rainfall events usually involve distinct eastward propagation under the thermal and dynamic coaction of the TP (Lu, Yu, and Zhou 2009; Xue, Bai, and Li 2012; Qian et al. 2015).
Given the frequent triggering and eastward propagation of convective-scale precipitation around this region, and the fact that precipitation is closely related to vertical velocity, how the convection associated with precipitation is initiated and propagates should be more closely investigated. Also, since the movement of a rainbelt is connected to both the thermal and dynamic effects of the TP (Zhang, Zhang, and Sun 2014), which one plays the dominant role?

To address these issues, a heavy precipitation process that occurred in western Sichuan on 18–20 August 2010 is chosen as a case for analysis. The heavy precipitation brought major flooding, landslides, debris flows, and other geological disasters, resulting in a large number of human casualties and considerable economic loss. Because of the relatively paucity of observational data over steep terrain, high-resolution numerical modeling is a better tool for providing the convective-scale results that we need to analyze this case. Indeed, in our previous work on this topic (Li, Cui, and Cao 2014; Yang, Gao, and Lu 2014, 2015; Huang and Cui (2015), Huang, Cui, and Li (2016) explored the dominant cloud microphysical processes involved. In the present study, based on the same simulation, the mechanism responsible for the continuous initiation and propagation of convective cells and the movement of the accompanying convective-scale precipitation is investigated from the viewpoint of the vertical motion equation and the perturbation pressure equation.

2. Data, model, and methodology

2.1. Data and model

To achieve a high level of detail and identify the local characteristics of the convective-scale precipitation in the target region, a high-resolution numerical simulation is necessary. As such, the Weather Research and Forecasting model, V3.7.1, is configured with three-nested domains of resolution 27, 9, and 3 km, separately. Global Forecast System reanalysis data, with a resolution of 0.5° and 6-h intervals, are used as the initial and boundary fields. The simulation is integrated from 1800 UTC 17 August 2010 and lasts for 55 h, with the first 6 h as the spin-up time. A more detailed description of the simulation and microphysical schemes employed can be found in our previous studies (Li, Cui, and Cao 2014; Yang, Gao, and Lu 2014, 2015; Huang and Cui 2015; Li et al. 2016). The model outputs from the inner domain are utilized for analysis. Hourly precipitation data, at a 0.1° horizontal resolution, observed by automatic weather stations across China, merged with satellite data using the Climate Prediction Center Morphing technique (Shen, Xiong, and Wang 2010; Pan et al. 2012), are used to examine the observed precipitation pattern during the event.

2.2. Methodology

To explore the mechanism responsible for the continuous initiation and propagation of convection and the movement of associated convective-scale precipitation, the vertical momentum equation and perturbation pressure equation (Rotunno and Klemp 1982; Weisman and Klemp 1986; Parker 2010; Wang et al. 2015) are revisited.

Starting from the full vector equation of motion,

$$\frac{d\mathbf{u}}{dt} = -\frac{1}{\rho} \nabla p - f\mathbf{k} \times \mathbf{u} - \mathbf{g}k + \mathbf{F},$$  \hspace{1cm} (1)

the vertical acceleration can be approximated as (Rotunno and Klemp 1982; Parker 2010; Wang et al. 2015)

$$\frac{dw}{dt} = -\frac{1}{\rho} \frac{d\rho'}{d\rho} \frac{d\rho'}{dz} - \frac{\rho'}{\rho}g + \mathbf{F}_z \approx -\frac{1}{\rho_0} \frac{d\rho}{dz} \frac{\rho'}{\rho_0}g + \mathbf{F}_z,$$  \hspace{1cm} (2)

where \(\mathbf{u}(u, v, w)\) is the three-dimensional velocity vector, and \(\rho\) and \(\rho_0\) are the air density and pressure. The superscript \(\rho'\) denotes perturbation of a particular meteorological variable. The symbols \(\rho_{p}, p'_{o}, p'_{a}\) are the background value of the air density, the dynamical and buoyancy pressure perturbations, respectively. The terms \(f, g, \mathbf{F}(\mathbf{F}_x, \mathbf{F}_y, \mathbf{F}_z)\) and \(\mathbf{k}\) are the Coriolis parameter, gravitational acceleration, turbulent mixing, and unit vector in the \(z\) direction, respectively, and the following equation relations exist:

$$\mathbf{B} = -\frac{\rho'}{\rho_0}g = \frac{\theta'}{\theta_0}g - g \sum q;$$  \hspace{1cm} (3)

$$p' = p'_{o} + p'_{a}. \hspace{1cm} (4)$$

The buoyancy acceleration \(\mathbf{B}\) is composed of two effects—one of gaseous phase \((\frac{\theta'}{\theta_0}g)\) and the other of condensates \((-g\sum q, \text{ with } q = q_c + q_r, \text{ with each } q_c, q_r, q_s, q_g)\), where \(\theta'\) is the virtual potential temperature perturbation \((\theta_v = \theta_0 + \theta'_v)\) and \(q_c, q_r, q_s, q_g\) are the mixing ratios of cloud water, cloud ice, rain, snow, and graupel, respectively. The term \(-g\sum q\) shows the drag by cloud particles and precipitation. The perturbation pressure \(p'\) includes a dynamic part \(p'_{o}\) and a buoyancy part \(p'_{a}\) which are derived from the perturbation pressure equations (Equations (4) and (6)) (Rotunno and Klemp 1982; Klemp 1987):

$$\nabla^2 p' = \nabla^2 p'_{o} + \nabla^2 p'_{a} + \nabla^2 p'_{d}, \hspace{1cm} (5)
$$

with

$$\nabla^2 p'_{o} = -\rho_0 \nabla \cdot (f \mathbf{k} \times \mathbf{u}) = \rho_0 f \zeta \hspace{1cm} (6.1)$$

and

$$\nabla^2 p'_{a} = \frac{d}{dz}(\rho_0 B) \hspace{1cm} (6.2)$$

$$\nabla^2 p'_{d} = -\nabla \cdot [\rho_0 (\mathbf{u} \cdot \nabla) \mathbf{u}] \hspace{1cm} (6.3)$$
Here, the three components, $\nabla^2 p'_{\alpha}$, $\nabla^2 p'_{\beta}$, and $\nabla^2 p'_{\gamma}$, are the geostrophic part, buoyancy part, and dynamic part, respectively. Certainly, if we assume incompressible atmosphere ($\rho_0 = \text{constant}$) on a smaller scale, Equations (6.2) and (6.3) can be simplified to $\nabla^2 p'_{\beta} = \rho_0 \nabla \cdot (u \cdot \nabla u)$ and $\nabla^2 p'_{\gamma} = -\rho_0 \nabla \cdot (\mathbf{u} \cdot \nabla) \mathbf{u}$. Note that the geostrophic part, Equation (6.1), is not considered for the convective scale in this study. Therefore, $p' = p'_{\alpha} + p'_{\beta}$ in Equation (4) is used herein.

Thus, the vertical acceleration is driven by an imbalance among the dynamic $(-1/\rho_0)(\partial p'_{\alpha}/\partial z)$ and buoyancy $(-1/\rho_0)(\partial p'_{\beta}/\partial z)$ components of the vertical perturbation pressure gradient, the buoyancy force ($B$), and the turbulent mixing ($F_z$), according to Equation (2). However, which of these factors plays the more dominant role is unclear, as is the term responsible for the convection growth and precipitation amplification ($dw/dt > 0$ indicates the case is favorable for the strengthening of mature convective cells and the initiation of new convection). Furthermore, by separating the dynamic and buoyancy components of perturbation pressure, the dynamic $(-1/\rho_0)(\partial p'_{\alpha}/\partial z)$ and thermodynamic $-((1/\rho_0)(\partial p'_{\beta}/\partial z) - B)$ roles are quantitatively estimated. Based on the above equations, $dw/dt$ will be derived by utilizing the relaxation method to obtain $p'_{\beta}$ and $p'_{\gamma}$ according to Equation (6). For more detail, please refer to Parker (2010) and Wang et al. (2015).

3. Results

We begin by providing a general description of the chosen heavy precipitation event over western Sichuan on 18–20 August 2010. Owing to the complex local geography and the special geological structure over these mountainous areas, mudslides accompanied the rainstorm. The catastrophic conditions that ensued led to a complete interruption for local transportation and communications, with more than 2200 people trapped in the mountainous areas. The fact that the mudslides induced by this rainstorm event broke out in many separate valleys of the mountainous area is indicative of the highly localized nature of the weather event. Therefore, high-resolution numerical simulation is necessary to resolve the convective-scale processes.

The circulation background and synoptic situation have been analyzed previously by Li, Cui, and Cao (2014), Li et al. (2016). In short, the precipitation occurred over the western Sichuan basin to the eastern edge of the TP (Figure 1(a)). The simulated precipitation shows a similar pattern and order of magnitude as observed (Figure 1(b)), especially for the left-hand branch of the rainbelts.

Based on the simulation and preliminary analyses of the heavy precipitation process (Li et al. 2016), the convection initiation and spread, and related precipitation growth, are analyzed by utilizing the vertical momentum equation from the viewpoint of separating the perturbation pressure into its dynamic and thermal parts.

Figure 1. The (a) observed and (b) simulated 48-h total rainfall amount (mm) over the western Sichuan Basin from 0000 UTC 18 August 2010 to 0000 UTC 20 August 2010. The terrain height above 3 km is outlined. (c, d) The evolution of area-averaged (c) hourly precipitation intensity (mm h$^{-1}$) and (d) vertical motion intensity at the 5.57 km level (m s$^{-1}$) from 0000 UTC 18 August 2010 to 0000 UTC 20 August 2010.
Since similar diurnal cycles for the evolution of the intensity of precipitation and vertical motion emerge on 18 and 19 August 2010, the first nocturnal precipitation peak is used as an example for analysis. According to Equation (2), the vertical momentum is induced by the imbalance of several forces, including the dynamic/buoyancy component of the perturbation pressure gradient force and the buoyancy force after neglecting the turbulent mixing. Figure 3(a)–(i) show their vertical distributions along the strong convection centers shown in Figure 2(a)–(c). Figure 4 exhibits the total perturbation pressure ($p'$) from the background separation and the dynamic/buoyancy component ($p'_d/p'_b$) by solving the perturbation pressure equation (Equation (6)), and the major contributing terms of the buoyancy forces ($-\frac{1}{\rho_0} \frac{\partial \rho_0}{\partial z}, B$) along the same vertical cross sections as Figure 3.

Figures 3(a)–(c) examine the role of the combined forcing (contours) in the vertical momentum equation (Equation (2)) on the vertical motion itself (shaded). At 1600 UTC, a pair of strong convection cells (> 5 m s$^{-1}$) are stretched downwards, with their centers at (7 km, 102.9°E) and (5 km, 103.02°E), respectively. Weak sinking motion is located at both flanks of the strong updrafts. The precipitation occurs behind the leading-side (LS, i.e. the right-hand side) branch of the convection, between two updrafts. Positive $dw/dt$ values dominate east of 102.9°E, with the center superposing the LS strong convection center, meaning new cells might be initiated or existing mature cells might strengthen or move towards the direction where $dw/dt > 0$. From the evolution of the convection (Figure 3(b) and (c)), it is indeed the case that the eastward direction is the leading side of the propagating convection. At 1700 UTC (Figure 4(b)), the left-hand-side convection cell is weakened because of the negative value of the $dw/dt$ at 1600 UTC (Figure 3(a)), while the LS convection maintains and moves eastwards with positive $dw/dt$ overlaying strong convection at 103.15°E, and then extending to 103.3°E and stretching upwards to middle and higher levels. Correspondingly, at 1800 UTC (Figure 3(c)), the left-hand-side convection at 1700 UTC (Figure 3(b)) disappears and is replaced by a wide region of sinking motion (Figure 3(c)), and the LS convection propagates to 103.3°E, where its intensity reaches 7 m s$^{-1}$. This modal distribution might be related to the west–negative/east–positive pattern of the $dw/dt$ forcing. East of 103.3°E, the fact that the $dw/dt < 0$ forcing is transformed into a negative value suggests the suppression of the updraft at the leading side after the first peak at 1800 UTC. Furthermore, the downdraft, with an intensity of −3 m s$^{-1}$, at 103.3°E and the 2 km level (below the strong convection and cloud), is induced by the drag effect of hydrometeors due to precipitation. During this period (Figure 3(a)–(c)), precipitation always presents a trailing morphology relative to the convection.

![Figure 2](image-url)  
**Figure 2.** Horizontal distribution of vertical velocity (shaded; m s$^{-1}$) and hourly precipitation (isolines with 10 mm intervals; mm) at (a) 1600 UTC, (b) 1700 UTC, and (c) 1800 UTC 19 August 2010.
to the strong LS updraft. The positive $dw/dt$ signifies the development and propagation of a strong convection cell. Furthermore, the overlapping strong signals of vertical motion and $dw/dt > 0$ point to strong precipitation in the future. For example, the strong precipitation instances at 103.02°E/103.15°E at 1700 UTC/1800 UTC are attributable to the strong convection and positive $dw/dt$ forcing at the 5 km level at 1600 UTC/1700 UTC.

Since both the initiation and propagation of strong convection and the movement of precipitation bear a close relationship with the $dw/dt$ modality, which factor — thermal or dynamic — could be responsible for the variation of the $dw/dt$ distribution? According to Equation (2), the buoyancy forces and the thermal/dynamic perturbation pressure gradient force are analyzed. From Figure 3(d)–(f), broad positive buoyancy ($B > 0$) regions dominate the middle level, with larger values within updrafts, while negative buoyancy ($B < 0$) covers above and below strong convection cells (Figure 3(d) and (e)). The sandwich structure further produces lower-to-mid-level increasing $\rho_0 \frac{\partial p}{\partial z}$ (i.e. $\nabla^2 \rho_0$ according to Equation (6.2)) and mid-to-higher-level decreasing $\rho_0 \frac{\partial p}{\partial z}$ then leads to a nearly out-of-phase relation between the buoyancy $\left(-\frac{1}{\rho_0} \frac{\partial \rho_0}{\partial z}\right)$ components of the vertical perturbation pressure gradient (Figure 3(g)–(i)) and $B$, especially inside strong updrafts. Therefore, the buoyancy $B$ should not be considered alone; a negative value of $\left(-\frac{1}{\rho_0} \frac{\partial \rho_0}{\partial z}\right)$ partly offsets positive buoyancy forcing. Thus, the pattern of $dw/dt$ is largely attributable to its buoyancy $B$, because of its larger magnitude than $\left(-\frac{1}{\rho_0} \frac{\partial \rho_0}{\partial z}\right)$. The dynamic $\left(-\frac{1}{\rho_0} \frac{\partial \rho_0}{\partial z}\right)$ component of the vertical perturbation pressure gradient is also examined (Figure 4(f)). This presents a similar pattern within convection cells (negative value) to $\left(-\frac{1}{\rho_0} \frac{\partial \rho_0}{\partial z}\right)$ (as shown in Figure 3(h)), except for a smaller order of magnitude. Therefore, only one figure and not a time series are shown herein. By solving the perturbation pressure equation (Equation (6)), the dynamic and buoyancy components are separated from one another (Figure 4(a) and (b)). In Figure 4(a), $\nabla^2 \rho_0$ is positive at the lower level (below 4 km) and negative at the middle level (shaded), corresponding to $\rho_0 < 0$ below and $\rho_0 > 0$ above (isolines), respectively. Furthermore, $\nabla^2 \rho_0$ is an order of magnitude larger than $\nabla^2 \rho_d$ (shaded in Figure 4(b)). Correspondingly, $\rho_0$ is also larger than $\rho_d$ by solving Equation (6). Therefore, the distribution of perturbation pressure (Figure 4(e)) is determined by $\rho_0$. Note that the upstream westerly (downslope jet) and the easterly at both flanks produce convergence and upward motion, which lead to mass accumulation and positive dynamic perturbation pressure (Figure 4(b)), therefore suppressing further development of convection by negative $\left(-\frac{1}{\rho_0} \frac{\partial \rho_0}{\partial z}\right)$ force.

Figure 3. Zonal–vertical cross sections of vertical velocity (shaded; m s$^{-1}$), the total forcing term on the right-hand side of Equation (2) on $dw/dt$ (isolines; 10$^{-3}$ m s$^{-2}$), and the hourly precipitation (histogram; mm), along (a) 29.82°N at 1600 UTC, (b) 29.7°N at 1700 UTC, and (c) 29.8°N at 1800 UTC (compare the lines shown in Figure 2(a)–(c)). (d–f) As in (a–c), except for the vertical velocity (shaded; m s$^{-1}$) and the buoyancy forcing term (Equation (2)) (isolines; 10$^{-3}$ m s$^{-2}$). (g–i) As in (a–c), except for the vertical velocity (shaded; m s$^{-1}$) and the buoyancy part of the perturbation pressure gradient $-(1/\rho_0)\frac{\partial \rho_0}{\partial z}$ (Equation (2)) (isolines; 10$^{-3}$ m s$^{-2}$).
and propagation of convection cells and the movement of accompanying convective-scale precipitation are explored in this study by utilizing the vertical motion equation and perturbation pressure equation. The main findings can be summarized as follows:

1. The evolution of the area-averaged hourly rainfall intensity shows two clear nocturnal peaks, at 1800 UTC 18 August 2010 and 1600 UTC 19 August 2010, accompanied by two large increases in amplitude of vertical motion. Furthermore, throughout the precipitation process, the variation in rainfall intensity is basically consistent with the swing in vertical velocity intensity. The rain clusters shown in the horizontal charts also connect with convection cells nearby. This demonstrates a close correlation between them in both magnitude and pattern. Therefore, the initiation and movement of convection is key for identifying the development of convective-scale precipitation.

4. Summary

Based on the numerical simulation for a mountainous heavy precipitation process in western Sichuan, the mechanisms responsible for the continuous initiation and propagation of convection cells and the movement of accompanying convective-scale precipitation are explored in this study by utilizing the vertical motion equation and perturbation pressure equation. The main findings can be summarized as follows:

1. The evolution of the area-averaged hourly rainfall intensity shows two clear nocturnal peaks, at 1800 UTC 18 August 2010 and 1600 UTC 19 August 2010, accompanied by two large increases in amplitude of vertical motion. Furthermore, throughout the precipitation process, the variation in rainfall intensity is basically consistent with the swing in vertical velocity intensity. The rain clusters shown in the horizontal charts also connect with convection cells nearby. This demonstrates a close correlation between them in both magnitude and pattern. Therefore, the initiation and movement of convection is key for identifying the development of convective-scale precipitation.
Since precipitation is closely related to vertical velocity, the reason behind the frequent triggering of convective-scale precipitation in the study region is also investigated, via the vertical momentum equation. According to Equation (2), the vertical momentum is induced by the imbalance of several forces, including the dynamic/buoyancy part of the perturbation pressure gradient force and the buoyancy force. Similar diurnal cycles for the evolution of the intensity of precipitation and vertical motion emerge on 18 and 19 August 2010; therefore, the first nocturnal precipitation peak is used as an example to analyze. During the first peak period, precipitation always presents a trailing morphology relative to the strong LS updraft. The positive \( \frac{dw}{dt} \) signifies the development and propagation of a strong convection cell. Furthermore, the overlapping strong signals of vertical motion and \( \frac{dw}{dt} > 0 \) point to strong precipitation in the future.

Owing to the fact that both the initiation and propagation of strong convection and the movement of precipitation bear a close relationship with the \( \frac{dw}{dt} \) modality, which factor — thermal or dynamic — could be responsible for the variation of the \( \frac{dw}{dt} \) distribution is compared by separating the thermal/dynamic perturbation pressure equation. Broadly positive buoyancy \((B > 0)\) dominates inside updrafts, with negative values above and below. This sandwich structure further produces lower-to-mid-level increasing \( \frac{p_b}{\omega_0} (\partial B/\partial z) \) and mid-to-high-level decreasing \( \frac{p_b}{\omega_0} (\partial B/\partial z) \), thus further causing a nearly out-of-phase relation between the buoyancy \( -\left(1/\rho_0\right) (\partial p_b/\partial z) \) component of the vertical perturbation pressure gradient and the \( B \) inside strong updrafts. Therefore, the buoyancy \( B \) should not be considered alone; a negative value of \( -\left(1/\rho_0\right) (\partial p_b/\partial z) \) partly offsets the positive buoyancy forcing. Thus, the pattern of \( \frac{dw}{dt} \) is largely attributable to its buoyancy \( B \), because of its larger magnitude than \( -\left(1/\rho_0\right) (\partial p_b/\partial z) \). The dynamic \( -\left(1/\rho_0\right) (\partial p_b/\partial z) \) component of the vertical perturbation pressure gradient is also examined, and we find that it presents a similar pattern within convection cells (negative value) to \( -\left(1/\rho_0\right) (\partial p_b/\partial z) \), except for a smaller order of magnitude.

By solving the perturbation pressure equation (Equation (6)), \( \nabla^2 p_b^0 \) is an order of magnitude larger than \( \nabla^2 p_a^0 \). Consequently, \( p_b^0 \) is also larger than \( p_a^0 \). Therefore, the distribution of the perturbation pressure is determined by \( p_b^0 \). As for the dominant term affecting the variation of vertical motion (Equation (2)), the buoyancy \( B \) and its contributing terms (Equation (3)) are also shown. The positive buoyancy \( B \) depends on the combined effects of the positive virtual potential temperature perturbation inside the cloud from the release of latent heat, and the role of downward drag by hydrometeors inside cumulonimbus, with the former dominating. Note that the conclusions presented in this paper are also valid for the second peak (not shown for brevity).

**Disclosure statement**

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