Control Parameters for Orographic Precipitation Associated with a Conditionally Unstable Flow over a Two-Dimensional Mesoscale Mountain

S.-H. Chen[1] and Y.-L. Lin[2]
[1]Land, Air, and Water Resources, University of California, Davis, CA 95616
[2] Marine, Earth, and Atmospheric Sciences, North Carolina State University, Raleigh, NC

1. Introduction

Generation and propagation of orographic precipitation systems may be controlled by different factors, such as the basic wind speed, moist Brunt-Vaisala frequency, mountain height, vertical wind shear, amount of moisture, etc. Based on idealized simulations for a two-dimensional conditionally unstable airstream over a mesoscale mountain, Chu and Lin (2000; denoted as CL hereafter) identified three moist flow regimes: (I) upstream propagating convective system (low F_w), (II) quasi-stationary convective system (moderate F_w), and (III) quasi-stationary and downstream propagating systems (large F_w). In the above, F_w may serve as the control parameter for these flow regimes, and it is defined as $F_w = U / N_w h$, where U is the basic flow speed, N_w (=$(g/\theta_v)\partial\theta_v/\partial z$; Emanuel 1994) the low-level buoyancy (Brunt-Vaisala) frequency for moist but unsaturated air, and h the mountain height.

In addition to F_w, the convective available potential energy (CAPE) might be used to characterize the flow regime. CAPE has been found to play important roles in producing orographic rainfall over the Alps. It has also been considered as one of the common ingredients for heavy orographic rainfall in different regions around the world (Lin et al. 2001). CAPE is well known as a significant control parameter for the strength of mesoscale convective systems generated in a conditionally unstable flow. When a conditionally unstable airstream reaches the mountain and being lifted above its LFC, the CAPE will be released to accelerate the air parcel upward, which can then generate and/or strengthen the convective systems over the mountain and its surrounding area. The circulation generated by the release of CAPE may interact with the mountain and then affect the evaporative cooling associated with the falling rainfall, and the propagation of the orographically generated convective system. In fact, preliminary numerical experiments of two-dimensional flow (Chu and Lin 1998) and three-dimensional flow (Chen and Lin 2001, 2003) have shown that flow regimes tend to shift from a higher number regime to a lower number regime when the CAPE is large. The dynamics of this flow behaviour needs to be explored further. The advantage of using CAPE as a control parameter is that it can be estimated easily from the upstream sounding. In this study, we will investigate the effects of CAPE on orographically generated convective system by analysing the governing equations and performing idealized numerical simulations.

2. Model description and experiment design

The Weather Research and Forecast (WRF) model is a next-generation mesoscale model, developed by a group of scientists from different institutions and research centers (Chen and Dudhia 2000; Michalakes *et al*. 2001; Skamarock *et al*. 2001). WRF is a fully compressible, three-dimensional nonhydrostatic model, and the governing equations are written in flux-form. This model adopts the Arakawa-C grid and a time splitting explicit scheme (Klemp and Wilhelmson 1978), and also has an option for a free-slip lower boundary condition, used for performing idealized simulations. Currently, two different prototypes of the WRF model are available (height coordinate and mass coordinate). The terrain-following height coordinate ($\sigma - z$) is used for this study. We also use the Runge-Kutta third-order time scheme, fifth and third order advection schemes in the horizontal and vertical directions, respectively, an open lateral boundary condition in the east-west direction, and a periodic boundary condition in the north-south direction. The Purdue Lin microphysics parameterization (Chen and Sun 2002) scheme is activated. Detailed information about this model may be found at: www.wrf-model.org.

The horizontally homogeneous initial conditions are from Schlesinger (1978) with specified wind fields, and the sounding has a CAPE of 3000 $J\,kg^{-1}$. The tropopause is located at about 12 km, and the atmosphere about this level is assumed to be isothermal up to the upper boundary. The unsaturated moist Brunt-Vaisala frequency (N_w) is approximately 0.0103 s^{-1}, which is roughly estimated from the surface to 3 km in a column away from the mountain ridge based on the following formula (Emanuel 1994)

$$N_w^2 = \frac{g}{\bar{\theta}_v}\frac{\partial \theta_v}{\partial z}, \qquad (1)$$

where θ_v is the virtual potential temperature and $\bar{\theta}_v$ is the mean virtual potential temperature in the layer considered. A uniform westerly flow, 10 $m\,s^{-1}$ is imposed across the entire model domain. However, different mean wind, as well as sheared flow, will be also tested.

In this study, idealized two-dimensional mountain geometry is used and the formula of terrain height is as follows,

$$h = \frac{h_m}{1+\left((x-x_o)/a\right)^2} \qquad (2)$$

The parameters for the mountain height (h_m), half-width (a), and grid spacing (Δx) in the x direction are 2 *km*, 30 *km*, and 1 *km*, respectively. The horizontal domain has 1001 grid points in the *x* direction, which covers an area of 1000 km. The vertical grid interval is stretched from 30 *m* at the lowest level to 500 *m* near the domain top. There are 50 vertical levels giving a physical domain height of 20 *km*. A 5-km deep sponge layer is added to the upper part of the physical domain to reduce wave reflection. The mountain is introduced impulsively into the basic flow at $t = 0$ *s*. For all cases, the model time step is 1 *s* and the model is integrated for 10 hours. To investigate effects of CAPE on the generation and propagation of orographically generated convective systems, we made one control (CNTL) and four sensitivity (CP1-4) simulations with varying CAPE amounts (Table 1). In Table 1, CAPE is in $J\,kg^{-1}$ and N_w is in $10^{-3}s^{-1}$

	CAPE	*Fw*	*Nw*
CNTL	3000	0.525	9.54
CP1	1372	0.513	9.75
CP2	1895	0.517	9.69
CP3	2438	0.52	9.62
CP4	3578	0.53	9.46

Table1: Characteristics of the numerical experiments.

3. Results

Figure 1 shows the time evolution of the accumulated rainfall and vertical velocity of the CNTL case for a uniform, conditionally unstable flow over a two-dimensional mountain ridge with $h = 2\,km$, $a = 30\,km$, and $U = 10\,ms^{-1}$. The moist Froude number (F_w) is 0.525. The heaviest rainfall is developed over the mountain peak ($x = 0\,km$) in the first hour and then propagates downslope with a linear speed in the next 2 hours ($t = 3\,h$). The rainfall then becomes quasi-stationary after $3h$ with the maximum rainfall located on the lee side at about $x = 80\,km$. In the vicinity of the mountain peak, one still can see an area of stationary rainfall, which is associated with the continuous orographic lifting.

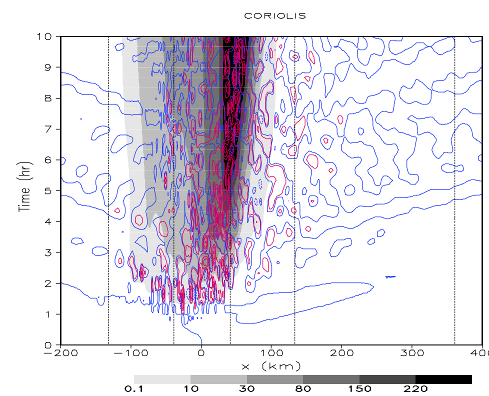

Fig.1: Time evolution of rainfall (shaded) and vertical velocity (contour lines) for the control case.

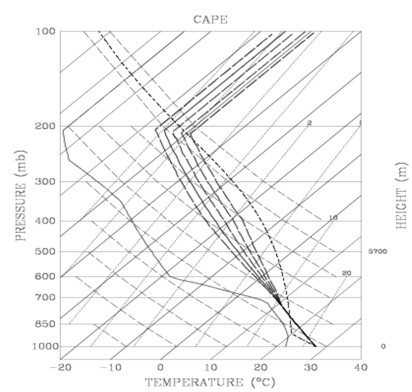

Fig.2: Soundings for five experiments. The long-dashed line from right to left are for CP1, 2,3, and 4, respectively. The solid line among the dashed lines is for the CNTL case.

The soundings for all experiments are shown in Fig. 2. The CAPE values for cases CP1 and CP4 are 1372 and 3578 $J\,kg^{-1}$, respectively (Table 1), and the CNTL case has a CAPE of

$3000\ J\,kg^{-1}$. Figure 3a shows the time evolution of surface vertical velocity and accumulated rainfall from Case CP1, which has a smaller CAPE ($1372\ J\,kg^{-1}$) than Case CNTL has. As anticipated, the accumulated rainfall is smaller than that from Case CNTL. It can also be seen clearly that the downstream propagating convective system continues to propagate downstream at later time, compared to that in Case CNTL. Based on CL's classification, both Cases CP1 and CNTL belong to Regime III. However, due to the propagation behaviour of the downstream convective system, the flow of CNTL may be called Regime III- since the downstream convective system becomes quasi-stationary. Figure 3b shows the time evolution of the surface vertical velocity and accumulated rainfall from Case CP4, which has a larger CAPE ($3578\ J\,kg^{-1}$) than Case CNTL has. As anticipated, the accumulated rainfall is larger than that from Case CNTL. The downstream convective system becomes quasi-stationary at later time, but is closer to the quasi-stationary convective system located in the vicinity of the mountain peak. Thus, the decrease of CAPE (e.g. CNTL to CP1) of the incoming airstream tends to shift the flow regime from a lower number (III-) to a higher number (III). On the other hand, the increase of CAPE may shift to a lower number regime. In order to prove this hypothesis, we simulate a case (MN5) of Regime II, which is identical to CNTL (CAPE=$3000 J\,kg^{-1}$) except with $U = 5ms^{-1}$. The evolution of the surface wind and accumulated rainwater for MN5 is shown in Fig. 3c, which indicates that the flow regime is shifted from Regime II to Regime I. It does prove the hypothesis and is also consistent with Chu and Lin's (1998) two-dimensional and Chen and Lin's (2001) three-dimensional preliminary results. Thus, we can conclude that *the increase (decrease) of CAPE tends to shift the flow to lower (higher) number regime*.

The effect of varying CAPE, as discussed above, may be restated as that *the propagating convective system tends to develop further upstream when the CAPE of the incoming airstream increases*. This, in a way, is analogous to the decrease of the incoming wind speed. Physically, this may be interpreted by that the decrease of the incoming wind speed tends to reduce the kinetic energy of the basic wind and then the advection of the convective system, which allows the convective system to develop further in vertical and stronger. Thus, it will produce stronger density current and allow more upstream development of the convective system. This idea may be expressed mathematically by the nondimensional control parameter, $\sqrt{CAPE/(U^2/2)}$, if it is held constant. This also infers that the increase of CAPE is also analogous to the decrease of the moist Froude number (F_w) because F_w is proportional to U.

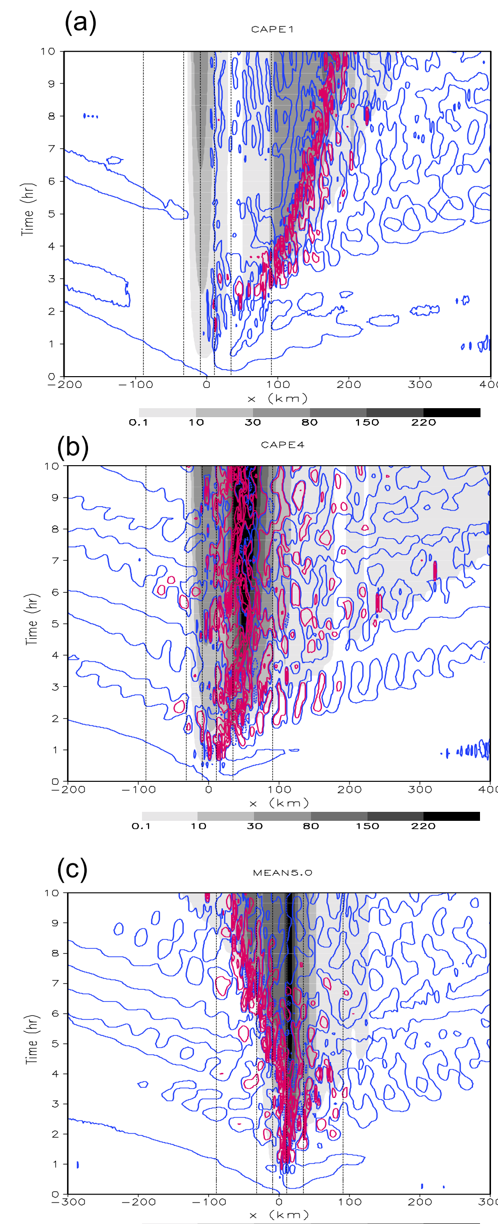

Fig.3: The vertical velocity (lines; m/s) and accumulated rainfall (shaded; mm) for (a) CP1, (b) CP4, and (c) MN5, respectively. The CAPE

in MN5 is the same as that in CNTL but with a mean wind of 5 m/s.

Figure 4 shows the maximum vertical velocities (the upper curve) predicted by the theory, which assumes no shear and all the CAPE is converted into kinetic energy of the air parcel in the cloud, $w_{max} = \sqrt{2CAPE}$. The maximum vertical velocities in the simulated convective clouds are much less (the lower curve in Fig. 4) than the theoretically predicted values, due to the adverse perturbation pressure gradient and to dilution of the buoyancy and momentum by mixing of environmental air (e.g. Emanuel 1994). In general, the simulated w_{max} increases as the $CAPE$ increases except when $CAPE$ increases from 2438 to 3000 $J\,kg^{-1}$. The increase of the simulated w_{max} with $CAPE$ do not increase as fast as that predicted by the theory.

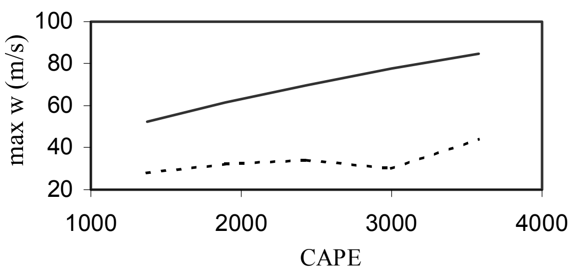

Fig. 4: The maximum vertical velocity (m/s) with respected to CAPE from the formula of (; solid line; $W = \sqrt{2CAPE}$) and from model simulations (dashed line).

4. Concluding Remarks

Effects of convective available potential energy (CAPE) on the generation and propagation of convective systems in a conditionally unstable airstream over a mesoscale mountain are studied using a two-dimensional, non-hydrostatic, nonlinear cloud model. In this study, we have found that the increase (decrease) of CAPE tends to shift the flow to lower (higher) number regime. This flow behaviour is explained by the nondimensional control parameter, $\sqrt{CAPE/(U^2/2)}$. In addition, we also found that in general, the simulated w_{max} does increase as the $CAPE$ increases, but not as fast as that predicted by theory.

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