Parameters Controlling Precipitation Associated with a Conditionally Unsaturated, Unstable Flow over a Two-Dimensional Mesoscale Mountain

Shu-Hua Chen[1] and Yuh-Lang Lin[2]
[1] Land, Air, and Water Resources, UC Davis, California
[2] Marine, Earth, and Atmospheric Sciences, North Carolina State University, Raleigh, North Carolina

1. Introduction

Understanding the formation and propagation of precipitation systems in the vicinity of a mesoscale mountain is essential in helping forecast orographic rain and the damages caused by flooding associated with it. The propagation of orographic precipitation systems may be controlled by various factors, such as the basic wind speed (U), moist Brunt-Vaisala frequency (N_w), mountain height (h) and width, convective available potential energy (CAPE), atmospheric moisture content, and vertical wind shear. Based on idealized numerical simulations, Chu and Lin (2000; denoted as CL hereafter) identified three moist flow regimes for a two-dimensional conditionally unstable flow over a mesoscale mountain ridge: (I) flow with an upstream propagating convective system, (II) flow with a quasi-stationary convective system over the mountain peak, and (III) flow with both a quasi-stationary convective system over the mountain peak and a downstream propagating convective system. In this study, the control parameters of unsaturated moist Froude number, $F_w = U / N_w h$, and CAPE to the classification of flow regimes are examined.

2. Experiment design

The Weather Research and Forecast (WRF; Michalakes *et al*. 2001; Skamarock *et al*. 2001) model with terrain-following height coordinates is used. An open lateral boundary condition in the north-south direction, a free-slip lower boundary condition, and a periodic boundary condition in the east-west direction are also chosen. The horizontally homogeneous initial conditions are from Schlesinger (1978) with specified wind fields, and the sounding of the control case (CNTL/CP4F2) has a CAPE about 3000 $J\,kg^{-1}$. The *unsaturated moist Brunt-Vaisala frequency* (N_w) is approximately 0.0095 s^{-1}, which is estimated from the surface to approximately 3 *km*. A uniform southerly flow and temperature profile are imposed across the entire model domain. However, different basic wind speeds and temperature profiles are tested. The Purdue-Lin microphysics parameterization scheme (Chen and Sun, 2002) is activated in all simulations.

Corresponding author address: S.-H. Chen, Land, Air, and Water Resource, Davis, CA 95616. shachen@ucdavis.edu

An idealized bell-shaped mountain, $h = h_m a^2 / [(x - x_o)^2 + a^2]$, is used, where the mountain height (h_m), half-width (a), and horizontal grid spacing (Δx) are 2 *km*, 30 *km*, and 1 *km*, respectively. The horizontal domain has 1001 grid points. There are 50 stretched vertical levels, yielding 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 artificial wave reflection.

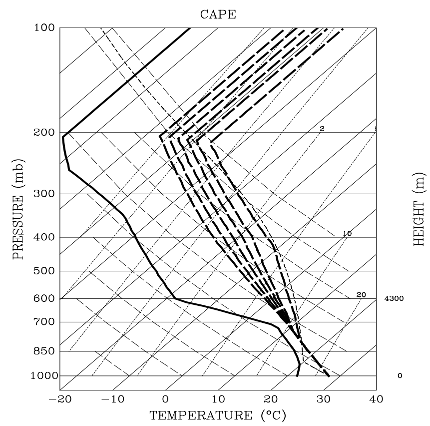


Fig. 1: Temperature profiles with different CAPEs. Long-dashed lines from right to left are soundings for CP0F1, CP1F2, CP2F2, CP3F2, CNTL (CP4F2), and CP5F2 cases, respectively. The CAPE values for them are 487, 1372, 1895, 2438, 3000, and 3578 $J\,kg^{-1}$, respectively. The basic wind speed is U = 5 ms^{-1}.

In order to investigate the effects of F_w and CAPE on a conditionally unstable flow over a mesoscale mountain, a matrix of numerical experiments is conducted, which is based on two control parameters, F_w and CAPE. Cases with different F_w are based on the variation of U and are denoted by (F1, F2, F3, F4, F5, F6) = (0.131, 0.262, 0.524, 0.786, 1.048, 1.572), which correspond to U = (2.5, 5, 10, 15, 20, 30 ms^{-1}), respectively. Cases with different CAPE are based on the variation of temperature profile above 2.0 *km* (Fig. 1), in order to keep the low-level CINH constant. This set of experiments is denoted by

CP0, CP1, CP2, CP3, CP4, and CP5, which correspond to CAPE = 487, 1372, 1895, 2438, 3000, and 3578 *Jkg*$^{-1}$, respectively. The mountain is introduced impulsively into the basic flow at the time the simulation is started, i.e. $t = 0$ *s*. For all cases, the model time step is 1 s and the model is integrated for 10 h.

3. Results

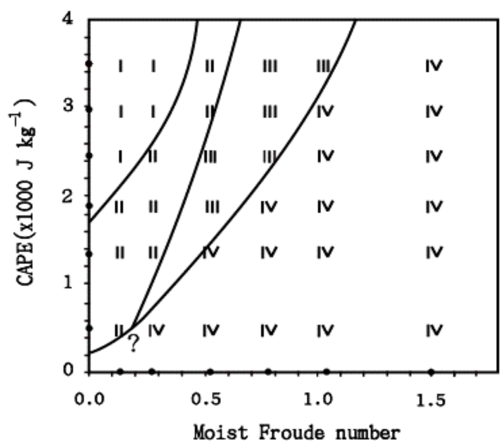
In this study, we identified four flow regimes. The first three are similar to those with CL with some modifications, while Regime Ⅳ is a new one (see Concluding Remarks for definition).

Based on F_w and CAPE, we have constructed a moist flow regime diagram presented in Fig. 2 which is analogous to the dry flow regime proposed in Smith (1989). From the figure, we can see that with a relatively large and fixed CAPE (e.g., greater than 1800 $J\,kg^{-1}$ in this study), the flow is shifted towards a larger number flow regime in a successive sequence as the F_w increases. However, when the CAPE is small (say, less than 400 $J\,kg^{-1}$), there exists a possible bifurcation point, which separates Regimes Ⅱ, Ⅲ, and Ⅳ of the moist flow regime when F_w is small (e.g. 0.16). Incidentally, based on the horizontal mountain scale aspect ratio and nondimensional mountain height, Smith (1989) found a bifurcation point separating four dry flow regimes: mountain wave, flow splitting, wave breaking, and flow splitting and wave breaking. Comparison of a case with a relatively larger CAPE and a case with a relatively smaller CAPE shows that two different modes of precipitation systems, the orographically forced *long-lasting precipitation system* in the vicinity of the mountain and *the lee side propagating precipitation system*, exist in both cases but behave very differently. When the CAPE is large, the difference in the rainfall amount produced by the long-lasting precipitation system over the mountain for different F_w (basic wind speeds) is relatively small, i.e. it is less sensitive to F_w (basic wind speed). On the other hand, with a small CAPE, the rainfall amount associated with the long-lasting precipitation system around the mountain is proportional to F_w (basic wind speed).

Physically and dynamically, the role of F_w and CAPE may be interpreted using the ingredient argument proposed by Lin et al. (2001), and the orographic rain forecasting models of Alpert (1986) and Smith (2003). The total precipitation (P) associated with an orographic precipitating system may be estimated as (see e.g. Lin et al. 2001),

$$P = (\rho / \rho_w)\, E (w_{oro} + w_{env}) qD\,, \qquad (1)$$

where ρ is the low-level air density, ρ_w the liquid water density, E the precipitation efficiency, w_{oro} the upward motion induced by the orography, w_{env} the upward motion induced by the environment (such as conditional instability, convective instability, etc.), q the low-level mixing ratio of water vapor, and D the duration of the precipitating system. Roughly, w_{oro} may be estimated by $U\partial h/\partial x$ for flow over a two-dimensional mountain ridge. For a conditionally unstable airstream, w_{env} may be estimated by the idealized equivalent vertical velocity, *W**, which is roughly equal to $\sqrt{2CAPE}$.



487 (CP 0), 1372 (CP 1), 1895 (CP 2), 2438 (CP 3), 3000 (CP4), to 3578 (CP5) $J\,kg^{-1}$. The last 5 thermodynamic profiles are those used for CP1F2, CP2F2, CP3F2, CP4F2 (CNTL), and CP5F2 in Section 3b. The basic wind increases from 2.5 *m s*$^{-1}$ to 30 *m s*$^{-1}$, which gives F_w = 0.131 to 1.572.

Consider the situation with relatively large CAPE. In this case, w_{env} will dominate the vertical motion, but a modest w_{oro} may be able to induce an orographic convective system associated with the long-lasting mode in the vicinity of the mountain, as long as the level of free convection (LFC) is reached. Hence the rainfall amount will be less sensitive to the magnitude of w_{oro}, or rather, the basic wind speed because the slope steepness ($\partial h/\partial x$) is fixed in this case. On the other hand, for the case with small CAPE, the vertical velocity induced by the conditional instability might be relatively small, and the orographic rainfall amount will be more sensitive to the basic wind since $w_{oro} = U\partial h / \partial x$. Thus, orographic rainfall amount is roughly proportional to the basic wind speed, or F_w, if the mountain slope remains unchanged. Implicitly, this indicates that conditional instability does not play an important role in producing orographic rainfall when the basic wind speed is large. Therefore, with small CAPE and large basic wind speed, *the long-lasting orographic precipitation system over the mountain belongs to the stratiform type, instead of the convective type,* due to the classical stable accent mechanism. This is evidenced by the vertical velocity, potential temperature, and shallow cloud fields for the case with $U = 20\, ms^{-1}$ and $CAPE = 1372\, J\, kg^{-1}$. As mentioned earlier, the stratiform cloud is defined to have a cloud depth less than 4 *km*. Under this condition, since the convective precipitation system is dominated by the orographically induced vertical motion, it should reflect the structure of the mountain shape. Here the vertical motion field reveals a hydrostatic mountain wave. The lee side propagating convective or cloud system is very weak, and therefore this flow regime may be referred to as a long-lasting *orographic stratiform precipitation system over the mountain and possibly a downstream propagating cloud system*. Note that this is a new flow regime (*Regime IV*), which was not discussed in CL and Chen and Lin (2004).

From a fundamental physical point of view, one might be curious to know why a deep, orographic convective system cannot develop when the basic wind is strong and the environmental CAPE is small. For the CP1F2 sounding, the LFC is located at about 820 hPa, which is about 1.8 *km*. With a 2 *km* high mountain, one would anticipate that the mountain is high enough to lift most of the low-level air parcels to their LFCs and release the instability even though the CAPE is not very large. Indeed, this is true. If one inspects case CP1F1, $F_w = 0.131$ and $CAPE = 1372\, J\, kg^{-1}$, the small CAPE is able to trigger a deep convective system in the vicinity of the mountain. This may be explained as follows: for weak basic flow, such as in CP1F1 (2.5 $m\ s^{-1}$), the advection time for airflow to cross the mountain is long enough for a deep cloud to develop over the mountain (Jiang and Smith 2003); due to the small CAPE (1372 $J\, kg^{-1}$). The kinetic energy (KE) associated with the cold pool produced by evaporative cooling is comparable to the KE associated with the weak basic flow. Thus, a quasi-steady, critical state is reached and convection is able to develop. In other words, *the flow is critical to the cold air outflow*. On the other hand, for stronger wind and smaller CAPE, the deep convective cloud has insufficient time to grow over the mountain (small advection time). Since the KE associated with the cold air outflow is much smaller than the KE associated with the basic wind, the precipitating system will be advected downstream by the basic wind and no strong, deep clouds can exist over the mountain area. In other words, *the flow is supercritical to the cold air outflow*.

A similar situation also occurs with variation of CAPE when F_w (basic flow speed) is kept constant. When the basic flow is weak (say, e.g., less than or equal to 5 ms^{-1} in this study), the flow is shifted to a smaller number flow regime (i.e., the system moves upstream) as the CAPE increases. This can be interpreted as follows: a stronger system on the lee side of the mountain can develop when an airstream has a larger CAPE and a weak basic flow (long advection time) and therefore, the cold pool produced by the system is relatively stronger. Thus, under this situation the convective system on the lee slope is able to propagate upstream against the basic flow and shifts the flow to a smaller number flow regime. This, in a way, is analogous to the decrease in F_w (incoming wind speed) for a fixed CAPE.

The above discussions may also be depicted by using the vertical velocity, potential temperature, and cloud fields and the temporal evolution of accumulated rainfall and the vertical velocity at 3.6 *km* height from four cases (CP1F1, CP5F1, CP1F5, and CP5F5).

4. Concluding Remarks

Based on idealized simulations of conditionally unstable flow passing over a mesoscale mountain, we found four moist flow regimes, which may be characterized as: (1) Regime I: flow with an upstream propagating convective system and a transient convective system existing in the vicinity of the mountain at an earlier time; (2) Regime II: flow with a long-lasting orographic convective system over the mountain peak, upslope or downslope; (3) Regime III: flow with a long-lasting orographic convective or mixed convective and stratiform precipitation system over the mountain peak and a downstream propagating convective system ; and (4) Regime IV: flow with a long-lasting orographic stratiform precipitation system over the mountain and possibly a downstream propagating cloud system. The first three flow regimes are the same as those found in CL and Chen and Lin (2004), but with modifications. The fourth flow regime is new, which was not discussed in CL and Chen and Lin (2004). These four moist flow regimes may be depicted schematically as shown in Fig. 3.

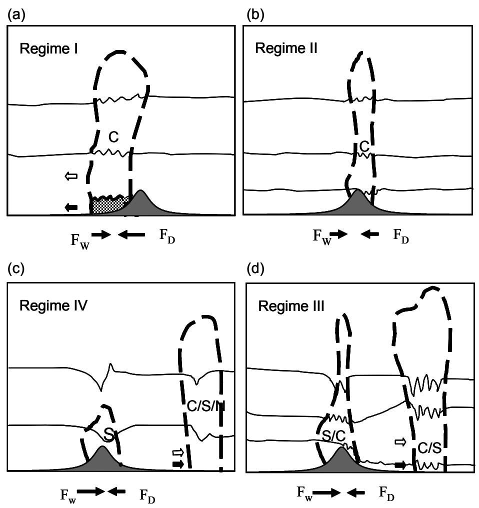


Fig. 3: Schematic of the flow regimes found in this study. (1) Regime I: flow with upstream propagating convective system and a transient convective system existing in the vicinity of the mountain at earlier time; (2) Regime II: long-lasting orographic convective system over the mountain peak; (3) Regime III: downstream propagating convective system and long-lasting orographic convective or mixed convective and stratiform precipitation system; and (4) Regime IV: a long-lasting orographic stratiform precipitation system over the mountain peak and possibly a downstream propagating cloud system. F_D is assumed to be a proxy of CAPE. Symbols C, S, and N denote convective, stratiform, and no cloud types, respectively. Outline (filled) arrow denotes the propagation direction of the precipitation system (cold-air outflow).

When the F_w (or basic wind speed) increases and the CAPE is fixed, the flow tends to shift to a higher number of the flow regime. Conversely, when the CAPE increases and the F_w (i.e., basic wind speed in this study) is fixed, the flow shifts to a lower regime. When the CAPE is large, the orographic rainfall amounts with different basic wind speeds are comparable (i.e., not sensitive to the basic wind speed) but the precipitation types can be different (i.e., convective vs. stratiform). However, when the CAPE is small, the orographic rainfall amount is strongly dependent on the strength of the basic wind speed – the stronger the wind, the larger the amount or rainfall.

Acknowledgments: This work is supported by US NSF Grant ATM-0096876. The authors would like to acknowledge the WRF model development team for their efforts in developing this model.

References

Alpert, P., 1986: Mesoscale indexing of the distribution of orographic precipitation over high mountains. *J. Clim. Appl. Meteor.*, **25**, 532-545.

Chen, S.-H., and Y.-L. Lin, 2004: Orographic effects on a conditionally unstable flow over an idealized three-dimensional mesoscale mountain. *Meteor. Atmos. Phys.*, in press.

Chen, S.-H., and W.-Y. Sun, 2002: A one-dimensional time-dependent cloud model, *J. Meteor. Soc. Japan*, **80**, 99-118.

Chu, C.-M., and Y.-L. Lin, 2000 (CL): Effects of orography on the generation and propagation of mesoscale convective systems in a two-dimensional conditionally unstable flow. *J. Atmos. Sci.*, **57**, 3817-3837.

Jiang, Q., and R. B. Smith, 2003: Cloud timescales and orographic precipitation. *J. Atmos. Sci.*, **60**, 1543-1559.

Lin, Y.-L., S. Chiao, T.-A. Wang, M. L. Kaplan, and R. P. Weglarz 2001: Some common ingredients for heavy orographic rainfall. *Wea. and Forec.*, **16**, 633-660.

Michalakes, J., S.-H. Chen, J. Dudhia, L. Hart, J. Klemp, J. Middlecoff, W. Skamarock 2001: "Development of a next generation regional weather research and forecast model" in developments in teracomputing. Proc. 9th ECMWF Workshop on the Use of High Performance Computing in Meteor. Eds. W. Zwieflhofer and N. Kreitz. World Scientific, Singapore. 269-276.

Schlesinger, R. E., 1978: A three-dimensional numerical model of an isolated thunderstorm: Part I. Comparative experiments for variable ambient wind shear. *J. Atmos. Sci.*, **35**, 690-713.

Skamarock, W. C., J. B. Klemp, and J. Dudhia, 2001: Prototypes for the WRF (Weather Research and Forecast) model. Ninth Conf. Meso. Processes, AMS, J11-J15 (also see http://www.wrf-model.org).

Smith, R. B., 1979: The influence of mountains on the atmosphere. *Adv. in Geophys.*, Vol. 21, B. Saltzman (Ed.), Academic Press, 87-230.

Smith, R. B., 2003: A linear upslope-time-delay model for orographic precipitation. *J. Hydrol.*, **282**, 2-9.