Orographic Precipitation Response to Microphysical Parameter Perturbations for Idealized Moist Nearly Neutral Flow

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(Manuscript received 19 December 2017, in final form 23 February 2018)

ABSTRACT

This study explores the sensitivity of clouds and precipitation to microphysical parameter perturbations using idealized simulations of moist, nearly neutral flow over a bell-shaped mountain. Numerous parameters are perturbed within the Morrison microphysics scheme. The parameters that most affect cloud and precipitation characteristics are the snow fall speed coefficient A_s , snow particle density ρ_s , rain accretion (WRA), and ice-cloud water collection efficiency (ECI). Surface precipitation rates are affected by A_s and ρ_s through changes to the precipitation efficiency caused by direct and indirect impacts on snow fall speed, respectively. WRA and ECI both affect the amount of cloud water removed, but the precipitation sensitivity differs. Large WRA results in increased precipitation efficiency and cloud water removal below the freezing level, indirectly decreasing cloud condensation rates; the net result is little precipitation sensitivity. Large ECI removes cloud water above the freezing level but with little influence on overall condensation rates. Two environmental experiments are performed to test the robustness of the results: 1) reduction of the wind speed profile by 30% (LowU) and 2) decreasing the surface potential temperature to induce a freezing level below the mountain top (LowFL). Parameter perturbations within LowU result in similar mechanisms acting on precipitation, but a much weaker sensitivity compared to the control. The LowFL case shows ρ_s is no longer a dominant parameter and A_s now induces changes to cloud condensation, since more of the cloud depth is present above the freezing level. In general, perturbations to microphysical parameters affect the location of peak precipitation, while the total amount of precipitation is more sensitive to environmental parameter perturbations.

1. Introduction

Orographic precipitation during nonconvective events commonly occurs in environments characterized by moist, nearly neutral conditions. This type of flow allows for little resistance to orographic lifting, resulting in enhancement of precipitation over windward mountain slopes (Miglietta and Rotunno 2005, 2006, hereafter MR05 and MR06, respectively). Atmospheric rivers (ARs) have been observed to have moist, nearly neutral static stability in the lower levels of the troposphere, as well as high surface potential temperatures (~285–289 K), strong wind speeds (low-level jet $>20 \,\mathrm{m\,s^{-1}}$), high freezing level, and nearly saturated conditions (Ralph et al. 2005). When ARs interact with topography they can produce intense precipitation, triggering flooding and mudslides that result in devastating and costly impacts to local infrastructure and loss of life and property (Ralph et al. 2006; Neiman et al. 2008b; Leung and Qian 2009; Ralph et al. 2011).

In California, ARs are responsible for 20%–50% of the annual precipitation (Dettinger et al. 2011), often from only a few storms producing large amounts of snow, resulting in snowpack reaching "near-record

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DOI: 10.1175/JAS-D-17-0389.1

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levels of snow water equivalent" (Guan et al. 2010, 2013). Guan et al. (2013) found that lower-than-normal surface air temperatures during AR events favor increased snow accumulation over the Sierra Nevada. Colder AR events have lower amounts of integrated water vapor, yet they result in a higher snow-to-rain ratio and higher overall precipitation values (Guan et al. 2010). Freezing level can also play an important role in snowpack stability. High freezing levels can shift precipitation from snow to rain (Yuter and Houze 2003; Colle and Zeng 2004b). Rain falling on snowpack can promote early snowmelt (Kim et al. 2013) and create challenges for water resource managers. Past studies have also found that a high freezing level affects where precipitation will fall, resulting in less downwind transport and a peak in precipitation on windward slopes or near peak elevations (Colle 2004; MR06; Stoelinga et al. 2013).

Complex interactions between mountain geometry, thermodynamics, and cloud microphysics can also control precipitation type, amount, and its location over a mountain (Lin et al. 2001; Jiang and Smith 2003; Stoelinga et al. 2003; Colle 2004; MR05; MR06; Tushaus et al. 2015, hereafter T15). For example, a wide barrier $(\geq 30\text{-km half-width})$ will provide more time for snow growth aloft, while a narrower barrier will generate more graupel through the collection of supercooled cloud droplets (Colle and Zeng 2004b). Wider barriers are thus more sensitive to parameters such as snow fall speed over the windward slope. Hobbs et al. (1973) demonstrated the degree of ice particle riming influences where ice particles will precipitate over a mountain, that is, heavily rimed particles fall out faster and land on the windward slope.

Microphysical parameterizations are implemented in weather forecast models to represent the development of clouds and precipitation. Simulated orographic precipitation for both AR and non-AR events has been found to be sensitive to the choice of microphysics scheme (Jankov et al. 2007, 2009; Liu et al. 2011). Sensitivity studies exploring the effects of microphysical parameters on orographic precipitation show changes in these parameters can impact cloud and precipitation development. Colle and Mass (2000) found lower snow fall speeds shifted precipitation from the windward to the lee side of the barrier during an AR over the Pacific Northwest. Colle and Zeng (2004a) found condensation, snow deposition, and riming and melting of graupel contributed most to the development of surface precipitation on the upwind slope. Their results also showed surface precipitation and microphysical processes were most sensitive to parameters associated with snow/ graupel fall speeds, CCN concentrations, and snow size

distribution and less with ice initiation and autoconversion. A similar study performed over the Cascade Mountains (Colle et al. 2005) also found strong sensitivity to snow size distribution and fall speed parameters, as well as having cloud water accretion and melting of graupel and snow contribute to most of the surface precipitation production. Colle et al. (2005) found riming processes to be important for the amount and distribution of supercooled water within clouds, affecting snow growth upwind.

These microphysical parameters have both uncertainty and inherent variability. Uncertainty arises because some parameters have a measurable uncertainty (e.g., particle densities, fall speeds), while others are uncertain because of limited measuring capabilities (e.g., process rates) or because they do not correspond to any physical, observable quantity in nature (e.g., autoconversion thresholds). There are also parameters whose values are known to vary spatially and temporally, beyond the range of measurement uncertainty, but are constant in the model. Uncertainty and variability in these parameter values lead us to ask the question, what is the sensitivity of orographic precipitation to changes in microphysical parameters?

We hypothesize that microphysics exerts a strong influence on the amount and location of orographic precipitation within moist, nearly neutral conditions, such as those found within ARs. Changes in these microphysical parameters can then have secondary effects on the thermodynamics and dynamics through latent heating (Morales et al. 2015). Overall, we surmise microphysical parameter and process changes may result in nonlinear responses due to complex interactions between parameters within the microphysics scheme. While orographic precipitation may vary with changes to microphysical parameter and process, will these responses differ if the upstream environment is also perturbed? Past studies by MR05, MR06, and T15 have shown a nonlinear dependence of simulated precipitation to parameters controlling mountain geometry, mean zonal wind speed, moist static stability, relative humidity, and surface potential temperature. For example, increasing surface potential temperature did not result in higher rain rates on upwind slopes, since processes such as riming occur at lower temperatures and can contribute to surface precipitation (MR06). T15 performed a more systematic analysis of the parameter space for mountain geometry and upstream conditions, finding nonmonotonic responses for mountain halfwidth, moist static stability, and mean zonal wind speed and monotonic responses to mountain height, relative humidity, and surface potential temperature. Their study focused on liquid-only microphysics.

We extend this analysis to include ice microphysics and ask the following question: how will the sensitivities to changes in microphysical parameters change with different environments?

Following work by MR05, MR06, and T15, this study focuses on moist, nearly neutral flow over a twodimensional (2D) Gaussian-shaped mountain barrier and explores the sensitivities of orographic precipitation to changes in microphysical parameters in different environments. We utilize an idealized modeling framework, which allows us to reduce model complexity and isolate the most important physical controls on orographic precipitation within the Cloud Model 1 (CM1). Section 2 describes the idealized framework within CM1 and provides details on the experimental design. Section 3 describes the results for the microphysical and environmental perturbations, and section 4 provides a summary of the results and conclusions.

2. Methods

a. Model configuration

The idealized simulations in this study were performed using version 17 of CM1 (Bryan and Fritsch 2002). CM1 is a nonhydrostatic atmospheric model designed for modeling of cloud-scale processes. Following MR05, MR06, and T15, our configuration uses a 2D domain with a horizontally stretched x grid where the horizontal grid spacing is 2 km for the inner domain (1200 km in length) and stretches up to 6 km over 50 grid points on either side. The domain is 1600 km in length, and the total depth is 18 km with 55 vertical levels. The vertical grid spacing is 0.25 km from the surface to a height of 9 km, increases to 0.5 km between 9 and 10.5 km, and then remains constant at 0.5 km until 18 km.

Lateral boundary conditions are open radiative, with a no-slip bottom boundary condition and free-slip top boundary condition. A positive definite advection scheme is used, and a Rayleigh damper with damping coefficient of $0.0003 \,\mathrm{s}^{-2}$ is applied to the top 4 km to prevent reflection of vertically propagating gravity waves. Although interactions of radiation with the mountain surface can result in the forcing of mesoscale mountain circulations, for example, mountain-valley winds, our focus is on the interaction of microphysics and dynamics. Thus, radiative transfer and surface heat flux parameterizations are neglected. All simulations are run for 20h at a time step of 3s. The Morrison microphysics scheme version 3.4 (Morrison et al. 2005, 2009) is used for all microphysical parameter perturbation experiments. The Morrison scheme is a two-moment bulk microphysical parameterization that prognoses mass and number mixing ratios of four hydrometeor categories: rain, cloud ice, snow, and the choice of either graupel or hail. It also prognoses the mass mixing ratio of cloud liquid water, and in the version used here specifies a constant cloud droplet number concentration (set to $200 \,\mathrm{cm}^{-3}$). The rimed ice species in the experiments presented here is set to graupel. For rain, cloud ice, snow, and graupel, the particle size distributions (PSDs) are described by inverse exponential functions. For cloud liquid water, the PSD follows a gamma function with a shape parameter that depends on the specified cloud water number concentration. Analysis of the precipitation produced by each simulation utilizes 5-min output cadence during simulated hours 6-20, as the precipitation rate, cross-mountain flow, and spatial distribution of cloud are relatively steady during this time period.

b. Experimental design: Control

The environmental parameters for the idealized sounding are guided by observations of an AR event during the Olympic Mountains Experiment (OLYMPEX; Houze et al. 2017). The observed sounding (Fig. 1a) was taken at 0300 UTC 13 November 2015 at the base of the Quinault River Valley in Washington State (Zagrodnik et al. 2018). Our goal in idealizing this sounding was to create a realistic environment described by as few parameters as possible; thus, the observed profile was smoothed using a simple 1-2-1 filter and characteristic values for each parameter were determined. Table 1 lists the environmental parameters calculated from the observed sounding. Moist Brunt-Väisälä frequency (N_m^2) for the troposphere $(4 \times 10^{-5} \text{ s}^{-2})$ and stratosphere $(5 \times 10^{-4} \text{ s}^{-2})$ was calculated by averaging N_m^2 below and above the tropopause height (~12 km), respectively, using Eq. (36) of Durran and Klemp (1982). Within CM1, Eq. (13) is used for N_m^2 , along with Eq. (19) for the moist adiabatic lapse rate, found in Durran and Klemp (1982).

The surface potential temperature is set to 286 K. Figure 1b shows the original wind profile, which is smoothed and idealized with linearly increasing winds of 14 m s^{-1} at the surface to 42 m s^{-1} at a height of 12 km, and constant winds at 42 m s^{-1} above 12 km. The idealized wind profile contains no directional shear, that is, zonal winds are perpendicular to the barrier. The relative humidity (RH) profile (Fig. 1c) is divided into three layers: below 4.5 km, between 4.5 and 12 km, and above 12 km. This sounding has already likely undergone humidification relative to the upstream conditions through flow interaction with the barrier; thus, the RH below 4.5 km was decreased to 95% in order to have initial



FIG. 1. (a) Skew *T* diagram showing the observed sounding from OLYMPEX 13 Nov 2015 (black), idealized sounding (magenta), and LowFL sounding (blue). For all soundings, solid lines represent temperature and dashed lines represent dewpoint temperature, all in degrees Celsius. (b) Vertical profile of zonal wind speed ($m s^{-1}$) showing the observed winds (thin black line), smoothed profile (thick black line), idealized profile (magenta), and reduced wind profile (LowU, orange dot–dashed line). (c) Vertical profile of relative humidity (%) showing the observed values (thin black line), smoothed profile (thick black line), and idealized profile (magenta).

conditions representative of air not yet forced to rise. This subsaturated profile also helps limit cloud development far upstream before the flow approaches the barrier. The RH profile linearly decreases with height from 95% at 4.5 km to 20% at 12 km and remains constant at that value up to the model top (Fig. 1c).

 TABLE 1. List of parameters describing the environmental sounding and mountain geometry.

Parameter description (symbol)	Control	Units
Squared moist Brunt–Väisälä frequency for troposphere $(N^2 \pi)$	4×10^{-5}	s^{-2}
Squared moist Brunt–Väisälä frequency for stratosphere $(N_{w \text{ strat}}^2)$	5×10^{-4}	s^{-2}
Surface potential temperature ($\theta_{\rm sfc}$)	286	Κ
Mountain height (H_{mtn})	1×10^{3}	m
Mountain half-width (W_{mtn})	4×10^4	m

The idealized mountain barrier is created using the following equation from MR05, MR06, and T15:

$$h(x) = \frac{H_{\rm mtn}}{1 + \left(\frac{x - x_0}{W_{\rm mtn}}\right)^2},$$
 (1)

where mountain height *h* is a function of the horizontal position within the domain *x* and is described using the maximum mountain height H_{mtn} , the center of the mountain x_0 , and the mountain half-width W_{mtn} . The center of the mountain is shifted 400 km downstream of the center of the domain to reduce influence of mountain-generated, upstream-propagating waves on the flow entering at the upstream boundary.

Again, we use observations to guide the barrier characteristic parameters by averaging the elevation over a 0.6° -latitude-wide box through the center of the Olympic Mountains (Fig. 2). From the observations, the barrier is idealized to a maximum height of 1 km and a half-width of 40 km (Table 1). This mountain shape was chosen to match as closely as possible the upwind slope and mean height of the Olympic Mountains. Although the environmental sounding and mountain geometry are idealized from observations, our aim is *not* to simulate the exact evolution of the 13 November 2015 event, but to use these observations to anchor our simulations in reality.

c. Microphysical parameters

One-at-a-time parameter perturbation experiments were performed on the previously described control simulation using the microphysics parameters listed in Table 2. Some parameters are associated with physical characteristics of the frozen hydrometeors, for example, fall speed coefficient A and bulk particle density r of cloud ice i, snow s, and graupel g, and have an empirical range of values. Others are parameters within microphysics parameterizations that do not correspond to a real physical quantity, for example, autoconversion thresholds, and thus they generally have a greater range of uncertainty. Within the Morrison scheme, like other schemes, there exist "hard wired" thresholds



FIG. 2. Meridionally averaged elevation over a 0.6° -latitude-wide box including the center of the Olympic Mountains and the Quinault River Valley (black box in bottom map). Solid black line depicts mean topography, dashed black lines correspond to ± 1 standard deviation, and magenta line corresponds to the idealized mountain shape used in CM1 simulations (height of 1 km, halfwidth of 40 km).

for autoconversion rates and other processes; thus, we create parameters acting as multiplicative factors in order to vary the processes. Multiplicative factors are used to perturb the processes of cloud water autoconversion, rain accretion, and snow deposition to represent the uncertainty and inherent variability associated with these processes. The control values are the default settings for these parameters within the Morrison scheme. For parameters that describe observable cloud properties (e.g., fall speeds and particle densities), the range of values is set to a consistent range from observations. The other parameters are varied over a range of plausible values. Each one-parameter perturbation experiment has five points, that is, the control value, the minimum and maximum, and values between the control and minimum/maximum. Based on the complex interactions and feedbacks microphysics can have on the

Parameter description (symbol)	Control	Min	Max	Units
Cloud ice fall speed coefficient (A_i)	700	350	1050	$m^{(1-b)}s^{-1}$
Snow fall speed coefficient (A _s)	11.72	5.86	17.58	$m^{(1-b)} s^{-1}$
Graupel fall speed coefficient (A_g)	19.3	9.65	28.95	$m^{(1-b)}s^{-1}$
Cloud ice density (ρ_i)	500	250	750	$kg m^{-3}$
Snow density (ρ_s)	100	50	150	$kg m^{-3}$
Graupel density (ρ_{g})	400	200	600	$kg m^{-3}$
Cloud ice autoconversion	125	50	200	μm
threshold (DCS)				
Nucleated ice crystal radius (RMI0)	10	1	30	μm
Embryo graupel radius (RMG0)	45.71	35	1000	μm
Ice-ice collection efficiency (EII)	0.1	0.01	1	
Ice-cloud water collection efficiency (ECI)	0.7	0.3	1	_
Radius of splintered ice particle (RMMULT)	5	1	30	μm
Rain accretion factor (WRA)	1	0.5	2	
Cloud water autoconversion factor (WRC)	1	0.1	10	_
Snow deposition factor (SDEP)	1	0.5	2	—

 TABLE 2. List of microphysical parameters, including control values and range over which the parameters are perturbed. Parameters in bold are those that have the largest impact on surface precipitation in the control simulation.

system, we surmise that the responses to changes in the microphysical parameters will be nonlinear and test five values per parameter to explore this idea. The precipitation-rate responses to microphysical parameter changes are assessed over six regions on the mountain where spatial averaging is performed: the upwind foothills, slope, and top and the downwind foothills, slope, and top (Fig. 3c).

d. Experimental design: Environmental tests

Two additional experiments are performed to test if the sensitivities to changes in microphysical parameters



FIG. 3. (a) Vertical cross section over a portion of the domain of the control simulation showing temporally averaged (hours 6–20) streamlines colored with wind speed (m s⁻¹), cloud water and ice mixing ratios (gray shaded contours, g kg⁻¹), snow mixing ratio (0.05 g kg^{-1} contour, blue), rain mixing ratio (0.05 g kg^{-1} contour, orange), and freezing level (black line). (b) Hövmöller diagram of precipitation rate (mm h⁻¹) with simulation time (h) on the ordinate and distance from center of the mountain (km) on the abscissa. Dashed gray line corresponds to hour 6, start of temporal averaging. (c) Mountain height (m) profile with the six averaging regions used for the calculations: upwind foothills (UF), slope (US), top (UT), and downwind top (DT), slope (DS), and foothills (DF).

remain the same with a different environment. The first environmental test involves scaling down the wind profile. The control wind profile has high cross-barrier wind speeds ($\sim 16 \,\mathrm{m \, s^{-1}}$); the strong dynamical forcing in this case may exert a large influence on the precipitation response to microphysical parameter changes. In addition, T15 found that adjusting the environment can lead to large changes in precipitation for a similar idealized moist neutral environment. Their results found zonal wind speed to be a dominant control on precipitation, leading to a nonmonotonic upwind precipitation response to changes in wind speed. Thus, to explore how changes in the dynamical forcing impact the precipitation response to microphysical parameter perturbations, the first environmental test reduces the wind speeds by 30% throughout the entire profile (hereafter experiment LowU; Fig. 1b).

The second environmental test explores the impact of the freezing level height on the microphysical parameter perturbations. The freezing level in the control is at a height of approximately 2.6 km, which may result in warm-rain processes dominating the production of precipitation at the surface. To explore how changes to the most influential microphysical parameters are affected in an environment where ice processes play a larger role in precipitation production, the freezing level is lowered by reducing the surface potential temperature by 6K while retaining the same profile of N_m^2 (hereafter experiment LowFL; Fig. 1a). This new atmospheric profile has an initial freezing level at a height of ~ 1.3 km, which descends to below the mountain top after hour 3 of the simulation, likely through adiabatic ascent and melting of graupel and snow that cool air near the mountain top.

3. Results and discussion

a. Control simulation

A cloud forms on the upwind slope and top of the mountain with a depth of approximately 7 km within the first hour of our idealized simulation (not shown). Surface precipitation also occurs within the first hour of the simulation (Fig. 3b) through warm-rain processes and increases after the generation of snow and graupel. The high freezing level (approximately at 2.5 km height) provides sufficient distance for frozen hydrometeors to melt before reaching the surface. After hour 6, the midlevel cloud over the upwind slope begins to dissipate, seemingly due to the Wegener-Bergeron-Findeisen process as snow development extends upstream (not shown), and the upper-level cloud becomes detached from the parent cloud, persisting for the rest of the simulation (Fig. 3a). A shallow cloud forms between 2- and 3-km heights and extends across the domain, and a deep cloud containing high mass mixing ratios ($>0.5 \,\mathrm{g \, kg^{-1}}$) persists over the upwind mountain side (Fig. 3a). Also after hour 6, graupel no longer occurs and the orographic precipitation system becomes quasi steady. Thus, we neglect the initial transient graupel feature and perform our temporal averaging over hours 6–20. Melting of snow, in addition to warm-rain processes, provide the quasi-steady precipitation from hours 6 to 20, with the highest precipitation rates occurring over the upwind slope (maximum rate of 5.02 mm h⁻¹; Figs. 3b,c).

In general, the flow is laminar (i.e., there are no breaking waves, rotors, or strong lee waves) as it traverses the barrier. Downwind we see an elevated freezing level (Fig. 3a), likely due to adiabatic warming from descending winds on the lee side of the mountain. There are also low-amplitude lee waves, which trigger shallow cloud development downstream (Fig. 3a). The freezing level remains fairly constant at $z = 2.6 \,\mathrm{km}$ far upstream, while sloping downward near the upwind top of the mountain (Fig. 3a). This downward sloping of the freezing level was observed during OLYMPEX (Houze et al. 2017) and has been suggested to be a result of a combination of processes: latent cooling from snow melting, adiabatic cooling from upslope lifting, and the melting distance of hydrometeors, which is controlled by particle size and density, as well as ambient conditions (Marwitz 1987; Medina et al. 2005; Minder et al. 2011). The freezing level does not become low enough to reach the mountain top in our control simulation.

As mentioned in section 2, the goal of this study is not to simulate the events of 13 November 2015 exactly, but to use observations from this atmospheric river case to guide our upstream environmental conditions. Nevertheless, it is encouraging to see a similar magnitude of orographic enhancement in our idealized control simulation (precipitation rate over upwind top/precipitation rate over upwind foothills = 5.6) compared to the observed enhancement of 4-5 on 13 November 2015 (A. Rowe 2017, personal communication). Our idealized simulation also had the highest precipitation rates occur on the upwind slope and not on the highest peak, which was similarly observed during this OLYMPEX case (Zagrodnik et al. 2018). Thus, we are confident in our control simulation and its ability to provide understanding of this moist, nearly neutral environment.

b. Microphysical parameter perturbations

Perturbations were performed on the control simulation to explore the sensitivity of orographic precipitation and cloud development to changes in various parameters within the Morrison scheme. Table 2 lists all of the parameters tested, as well as the range of values over which the parameter perturbations were performed.

To understand the overall effect of these parameters on the control simulation, the relative change (%) within



FIG. 4. Percent relative change given by $100 \times (max - min)/average$ for each five-point parameter perturbation experiment for (a) PREC (mm h⁻¹), (b) LWP (includes cloud water and rain mixing ratios; kg m⁻²), and (c) IWP (includes cloud ice, snow, and graupel mixing ratios; kg m⁻²). Black denotes undefined values (i.e., infinity). Ordinate: list of parameter symbols defined in Table 2. Abscissa: averaging regions over the mountain shown in Fig. 3c.

each set of parameter perturbations was calculated by taking the difference between the maximum and minimum response values and dividing it by the average of the five response values. All values are temporally and spatially averaged over hours 6-20 for the six mountain regions described in Fig. 3c. Figure 4 shows the relative change for precipitation rate (PREC; $mm h^{-1}$) and liquid and ice water paths (LWP and IWP, respectively; $kg m^{-2}$). LWP is the total amount of cloud water and rain between the surface and model top, calculated by vertically integrating the cloud water and rain mixing ratios. Similarly, IWP represents vertically integrated cloud ice, snow, and graupel mixing ratios. Overall, some parameters have little influence (<20%) on our output metrics (PREC, LWP, and IWP), while others have a larger influence over specific regions or across the entire mountain. Downwind regions have the highest relative change, mostly because the PREC, LWP, and IWP there have small magnitudes, and hence small absolute changes will result in large relative changes. There are four parameters that stand out because of large relative responses or widespread impacts on PREC, LWP, and IWP: snow particle fall speed coefficient A_s , snow particle density ρ_s , ice-cloud water collection efficiency (ECI), and rain accretion multiplicative factor (WRA; Fig. 4). The quantities A_s and ρ_s describe hydrometeor characteristics, while ECI and WRA are associated with cloud water collection by ice species and rain, respectively. We focus on the responses to perturbing these four parameters in the rest of this section.

Hövmöller diagrams in Fig. 5 provide information on the precipitation location and amount compared to the control simulation for each parameter perturbation. Overall, the effect on precipitation amount is largest at lower perturbation values (left two columns in Fig. 5) for A_s , ρ_s , and ECI, and at the minimum and maximum values for WRA (Figs. 5m,p). Given that the control simulation's maximum precipitation value is $5 \,\mathrm{mm}\,\mathrm{h}^{-1}$, the parameter perturbations result in precipitation changes ranging from a 42% decrease (Fig. 5m) to a 30% increase (Fig. 5e). For low values of A_s , less precipitation falls on the upwind side (Figs. 5a–d), while ρ_s has the opposite effect, where less precipitation falls downwind for low ρ_s values (Figs. 5e-h). ECI precipitation responses are similar to A_s (less precipitation upwind for lower values), but with changes occurring over a narrower region (Figs. 5i-l). At the maximum value of WRA (WRA = 2; Fig. 5p), we see more surface precipitation over the upwind and downwind top, while the minimum value (WRA = 0.5; Fig. 5m) shows more precipitation over the upwind top initially but shifting with time to the upwind slope, in contrast to A_s , ρ_s , and ECI, which have a consistent spatial change in precipitation with time.

We next present temporally averaged (hours 6–20) precipitation rates for each region over the mountain and analyze the response functions to changes in A_s , ρ_s , ECI, and WRA. Precipitation rate (mm h⁻¹) is further analyzed using the following equation:

$$PREC = PE(COND), \qquad (2)$$

where surface precipitation rate (PREC) is a product of precipitation efficiency (PE) and vertically integrated total condensation rate ($\langle COND \rangle$, mm h⁻¹). Total condensation rate includes all processes that convert water vapor to condensate, that is, cloud ice, snow, and graupel deposition and cloud water condensation, which are



FIG. 5. Hövmöller diagrams of precipitation rate difference $(mm h^{-1})$ between microphysics parameter perturbation experiments and the control. Panels show perturbations for different values of (a)–(d) A_s [control = 11.72 m^(1-b)s⁻¹], (e)–(h) ρ_s (control = 100 kg m⁻³), (i)–(l) ECI (control = 0.7), and (m)–(p) WRA (control = 1). Blue colors represent larger precipitation rates for the parameter perturbation experiment. Horizontal dashed lines show hour 6, where temporal averaging begins.

then vertically integrated (indicated by $\langle \cdot \rangle$) from the surface to model top. In the model, $\langle \text{COND} \rangle$ depends mainly on the vertical air velocity in saturated conditions, and therefore changes to $\langle \text{COND} \rangle$ indicate a link to dynamical changes caused by parameter perturbations. PE is directly affected by microphysical parameter perturbations and describes how efficiently condensate is removed from the atmosphere through microphysical processes and sedimentation, resulting in surface precipitation. Since we analyze Eq. (2) over subregions of

limited spatial extent, PE is also affected by horizontal advection of condensate. Thus, the response of PE to parameter perturbations shows that a particular parameter has affected the conversion of condensate to vapor (from evaporation or sublimation) and/or transport (sedimentation and horizontal advection). A PE value less than one in the subregion corresponds to less precipitation reaching the surface compared to the amount of condensate produced within that subregion. If PE is greater than one, the increase in



FIG. 6. Snow fall speed coefficient $A_s[m^{(1-b)}s^{-1}]$ response functions for (a) precipitation rate (mm h⁻¹), (b) precipitation efficiency, (c) vertically integrated total condensation rate (mm h⁻¹), and (d) vertically integrated snow melting rate (mm h⁻¹). Values are temporally averaged over hours 6–20 and spatially averaged over the six mountain regions shown in Fig. 3c. Black vertical lines depict the control parameter value.

precipitation relative to the amount of condensate produced in the column is due to net convergence of the condensed water horizontal advective fluxes across the subregion. We use this analysis to determine how much condensate is generated ($\langle \text{COND} \rangle$), how efficient the model is in removing this condensate as surface precipitation, and how both are affected by microphysical parameter perturbations. In our calculations of PE from model output, we take the ratio of the average $\langle \text{COND} \rangle$ and PREC (spatial and temporal) to reduce noise, which can contaminate PE calculated as an average of the ratio.

Response functions for A_s show a general trend of increasing (decreasing) precipitation rate upwind (downwind) with increasing A_s (Fig. 6a). Using Eq. (2) to understand the precipitation response, we see PE has a similar response as precipitation rate, while $\langle \text{COND} \rangle$ is relatively unaffected (Figs. 6b,c). Perturbations to A_s influence the efficiency of condensate removal through the direct effect that A_s has on the snow particle fall speed in the Morrison scheme:

$$V_s = A_s D^{B_s}, (3)$$

where A_s is the snow fall speed coefficient $[m^{(1-b)} s^{-1}]$, D is the particle diameter (m), B_s is an exponent parameter (that we do not vary here), and V_s is the fall speed of snow. Decreasing A_s causes the mass-weighted mean fall speed of snow to decrease, allowing more snow produced upwind of the mountaintop to be advected downwind. Because of the change in horizontal advection of snow, the vertically integrated snow melting rate increases downwind with decreasing A_s (Fig. 6d). Thus, decreasing A_s reduces the snow terminal fall speed, leading to more snow being transported downwind of the mountain top and increasing the



FIG. 7. As in Fig. 6, but for ρ_s (kg m⁻³).

amount of snow that melts and reaches the surface as rain there. This is reflected by the increase of PE downwind with decreasing values of A_s (Fig. 6b).

Snow density (ρ_s) response functions show the opposite trend compared to A_s ; increased (decreased) precipitation rates occur downwind (upwind) with increasing values of ρ_s . Similarly, $\langle \text{COND} \rangle$ is relatively unaffected (Fig. 7c). From a physical standpoint, it would be expected that if the bulk density of snow particles is increased they should fall faster. However, in the Morrison scheme, as in almost all microphysics schemes, the snow particle fall speed is only determined through A_s , B_s , and D following (3). Nonetheless, the bulk particle density indirectly influences the fall speed through the particle size distribution. The slope parameter (λ), which describes the slope of the gamma particle size distribution, depends on the cubic root of the particle density ($\rho^{1/3}$):

$$\lambda = \left[\frac{\pi\rho N\Gamma\left(\mu+4\right)}{6q\Gamma\left(\mu+1\right)}\right]^{1/3},\tag{4}$$

where q is the mass mixing ratio, N is the number mixing ratio, Γ is the Euler gamma function, and μ is the shape parameter of the size distribution (equal to 0 for snow). Thus, increasing ρ_s for given values of q, N, and μ results in an increase in λ [(4)]. Increasing λ leads to a shift in the particle size distribution toward smaller diameters. Since the particle fall speed is a function of the diameter following (3), the shift of the distribution toward smaller particle sizes results in a decrease in the mass-weighted mean snow fall speed. This explains the trend in precipitation efficiency for ρ_s : more snow is advected downwind because of the slower mean fall speeds with increasing ρ_s (Fig. 7b), leading to more snow melting downwind (Fig. 7d). This relationship between ρ_s and snow fall speed causes an unphysical response in the precipitation rates. This unphysical behavior to changes in particle density has motivated the recent development of schemes that explicitly calculate fall speed as a function of the ratio of the particle mass and projected area, and hence incorporate the effects of particle density in a physically realistic way (e.g., Lin and



FIG. 8. WRA response functions for (a) precipitation rate $(mm h^{-1})$, (b) precipitation efficiency, (c) vertically integrated total condensation rate $(mm h^{-1})$, and (d) vertically integrated cloud water evaporation rate $(mm h^{-1})$. Values are temporally averaged over hours 6–20 and spatially averaged over the six regions shown in Fig. 3c. Black vertical lines depict the control parameter value.

Colle 2009, 2011; Harrington et al. 2013; Morrison and Milbrandt 2015).

Response functions for WRA differ from the other parameters, in that there is little effect on PREC (Fig. 8a). This occurs because of compensation between the effects of increasing PE and decreasing (COND) as WRA increases (Figs. 8b,c). The WRA parameter is a multiplicative factor that acts to increase or decrease the rain accretion process within the simulation. Higher WRA corresponds to more efficient removal of cloud water, while lower WRA results in more cloud water remaining within the atmosphere. Therefore, increasing the removal of cloud condensate through higher WRA results in higher PE (Fig. 8b). This rain accretion process has the largest effect on (COND) compared to the other parameters modified, with a decrease of over 70% on the upwind and downwind tops when WRA is increased from 0.5 to 2 (Fig. 8c). In all simulations, (COND) is dominated by cloud water condensation that occurs

through saturation adjustment in the Morrison scheme. That is, at every time step all excess vapor above water saturation is converted to cloud water, while evaporation is treated by converting cloud water to vapor each time step at a rate that ensures almost exactly saturated conditions are maintained inside liquid cloud. We emphasize that cloud condensation rate using saturation adjustment in the model does *not* depend directly on the number or size of existing cloud droplets.

Figures 9a and 9b show temporally averaged (hours 6–20) vertical cross sections of cloud mixing ratios, vertical air motion, cloud condensation rate, and cloud evaporation rate for the simulations with the smallest and largest values of WRA. We see updrafts increase in magnitude over the upwind slope with increasing WRA (Fig. 9b), which may be due to the reduction in condensate loading as more cloud water is effectively removed upwind. Stronger updrafts likely explain the



FIG. 9. Vertical cross sections over a portion of the domain of temporally averaged (hours 6–20) cloud mixing ratio (cloud water and ice; $g kg^{-1}$), positive vertical wind speed (black contours at 0.1 m s⁻¹ intervals starting at 0.2 m s⁻¹), cloud water condensation rate (red contours = $1 \times 10^{-3} g kg^{-1} s^{-1}$, yellow contours = $5 \times 10^{-4} g kg^{-1} s^{-1}$), cloud water evaporation rate (blue contours = $-1 \times 10^{-3} g kg^{-1} s^{-1}$, contours = $-5 \times 10^{-4} g kg^{-1} s^{-1}$), and freezing level (thick black line). (a) WRA = 0.5, (b) WRA = 2, (c) ECI = 0.3, (d) ECI = 1.

10% increase in (COND) over the upwind slope seen in Fig. 8c. For the other five regions, there is little change in vertical velocity for points containing liquid cloud, and the large decrease in (COND) with increasing WRA is primarily explained by an indirect effect of cloud water removal on the cloud base height (depicted here approximately by the 0.01 g kg⁻¹ cloud mixing ratio contour in Fig. 9). For small WRA, reduced conversion of cloud to rain means that more cloud water is available for evaporation in downdrafts compared to the simulations with large WRA. With more cloud water available, this implies a lowering of the cloud base and increasing vertical depth of the cloud layer, since the local evaporation rate calculated using saturation adjustment only depends on the pressure and tendencies of temperature (controlled mainly by the vertical velocity) and water vapor mixing ratio.

This point can be illustrated clearly for moist pseudoadiabatic descent; while the model has diabatic

forcing due to mixing and other microphysical processes, moist pseudoadiabatic descent can serve as a useful guide for explaining this behavior. For reversible moist pseudoadiabatic descent (and ascent), and assuming steady state, we can write the change in cloud mass mixing ratio (q_c) with height (z)as $dq_c/dz = -dq_v/dz = -dq_s/dz$ above the cloud base, where q_v is the water vapor mixing ratio and q_s is the saturation mixing ratio that is a function only of temperature (T) and pressure (p). Since dq_s/dz only depends on T and p, dq_c/dz also depends only on T and p. Thus, if q_c near cloud top is relatively large due to a reduction of WRA, cloud parcels must descend further in order to completely evaporate all of the cloud water they contain, compared to the situation with relative small q_c associated with an increase in WRA. Indeed, a lower cloud base and greater cloud depth are seen in the small WRA simulation compared to the large WRA simulation (cf. Figs. 8a and 8b). A lower cloud base in



FIG. 10. As in Fig. 8, but for ECI.

turn contributes directly to larger (COND) because the region of condensation extends over a greater depth. In addition to wider regions of condensation and evaporation associated with the lower cloud base, somewhat larger magnitudes of the condensation and evaporation rates occur for the small WRA simulation compared to the large WRA simulation (Fig. 8d). Since the small WRA simulation has somewhat higher relative humidity consistent with the lower cloud base, this is likely attributable to the mixing of relatively moist air into updrafts, which limits the effects of entrainment in reducing the condensation rate compared to the large WRA simulation.

To briefly summarize, perturbations to WRA demonstrate the nonlinear interactions between cloud water and surface precipitation. Reducing the rain accretion process initially reduces PE and precipitation rate and increases cloud water in the atmosphere, but the increased cloud water is associated with a deeper cloud layer (lower cloud base), leading to an increase in $\langle \text{COND} \rangle$ that compensates for the decrease in PE. The opposite occurs for large WRA, thus resulting in little sensitivity of PREC to WRA.

ECI also acts to remove cloud water from the atmosphere, but through the riming process occurring at temperatures below freezing. The response functions for ECI clearly show the importance of the altitude at which cloud water is being removed since, unlike WRA, changes to ECI do result in a change in precipitation rate; as ECI is increased there is an increase of PREC over the upwind slope and decrease over the downwind top (Fig. 10a). This is consistent with an increase in PE over the upwind slope and a decrease over the downwind top, but little change in (COND). The decrease in PE upwind by lowering ECI is expected since it corresponds to a decrease in the conversion of cloud water to precipitation through riming. Figures 9c and 9d show an impact on cloud water amount above the freezing level on the upwind top and slope as ECI is varied. Increased snow amounts with an increase in ECI and the fallout and subsequent melting of this snow leads to more precipitation on the upwind slope. The PREC and PE



FIG. 11. (top) As in Fig. 3a, but for (a) LowU and (b) LowFL. (bottom) Temporally averaged (hours 6–20) and spatially averaged (6 regions) values for (c) precipitation rate $(mm h^{-1})$, (d) vertically integrated total condensation rate $(mm h^{-1})$, and (e) precipitation efficiency. Solid black lines show values for the control simulation, dashed blue lines for the LowFL experiment, and dotted orange lines for the LowU experiment.

responses are similar to those for A_s (cf. Figs. 6a,b and Figs. 10a,b). In contrast to WRA, perturbations to ECI result in little change in $\langle \text{COND} \rangle$ even though it also affects the efficiency of conversion from cloud to precipitation. This occurs because ECI has little effect on cloud water below the freezing level. The amount of cloud water in descending air near cloud base is relatively unaffected, leading to little impact on the cloud base height. Thus, cloud depth is relatively unaffected with perturbations to ECI, in contrast to WRA (cf. Figs. 9c,d with Figs. 9a,b); correspondingly, there is little impact on overall evaporation and condensation rates whose magnitudes are much larger below the freezing level than above.

c. Environmental experiments

Experiments were performed to test how the sensitivities to microphysical parameter changes compared to those in an environment where the wind profile is reduced by 30% (LowU) or the temperature is reduced to produce a freezing level that reaches below the mountain peak (LowFL). The LowU experiment contains the same amount of precipitable water compared to the control simulation, but the reduction in horizontal wind speed causes a reduction in the vertical velocities (near the ground, vertical velocities are proportional to mean horizontal wind speed times mountain slope) resulting in a shallower cloud than in the control environment (Fig. 11a). Precipitation rates for LowU are a factor of 3 less than in the control (Fig. 11c), due to smaller $\langle \text{COND} \rangle$ and generally smaller PE (Figs. 11d,e). Although the melting layer remains deep enough for warm-rain processes to remain active, the reduced cloud depth substantially reduces the contribution of ice processes to surface precipitation, as seen from the reduced development of snow in Fig. 11a. The highest total condensation and precipitation rates are shifted toward the upwind top in LowU compared to the control.

The LowFL experiment results in a deep cloud with a detached upper-level ice cloud, similar to the control (Fig. 11b). In addition, high-amplitude and frequency lee waves are triggered downstream once the freezing level descends below the mountain top (not shown). These lee waves form clouds over the downwind foothills and beyond. Although the precipitable water available in the LowFL environment is 11.7 cm less than for the control, the peak precipitation rate over the upwind slope is closer to that of the control than LowU (Fig. 11c). This is because the condensate produced on the upwind slope is efficiently converted to precipitation (Figs. 11d,e). A secondary peak in (COND) is



FIG. 12. (top) Snow fall speed coefficient (A_s) and (bottom) snow particle density (ρ_s) response functions for (a),(d) precipitation rate (mm h⁻¹); (b),(e) precipitation efficiency; and (c),(f) vertically integrated total condensation rate (mm h⁻¹) for the LowU experiment. Black vertical lines depict the control parameter value.

located over the downwind slope, due to the lee waves previously mentioned, but because $PE \ll 1$ in this region, little of this condensate makes it to the surface as precipitation. The reduction in freezing level allows for more cloud area to exist above the freezing level, thus increasing the potential for ice processes to dominate in this environment, as shown by the spatial reduction of rain in Fig. 11b.

Overall, the LowU case results in a reduction in the magnitude of PREC responses to perturbations in A_s , ρ_s , ECI, and WRA than in the control environment (Figs. 12, 13), mainly due to less condensate being available (Figs. 11a,d, 12c,f, 13c,f). Compared to the control simulation, WRA does have a minor effect on PREC in this case, with PREC increasing with WRA over the upwind slope/top and downwind top associated with a combination of large increases in PE and smaller decreases in (COND) (Fig. 13). Additionally, PE responses to ECI are weaker in the LowU case compared to the control simulation (Fig. 13e). These responses are expected, as LowU results in weaker ascent, causing a reduction in the amount of cloud water available above the freezing level, thus allowing for warm-rain processes to dominate precipitation production. Although the ice processes play a smaller role in this case, the

mechanisms controlling PE and $\langle \text{COND} \rangle$ are similar to those in the control environment: 1) changes in PE primarily explain the PREC responses to A_s and ρ_s perturbations (Fig. 12), and 2) WRA affects $\langle \text{COND} \rangle$ through changes in the cloud base, while ECI does not (Fig. 13).

In contrast to the control environment and LowU environments, perturbations to A_s in the LowFL case produce notable impacts on (COND) (Fig. 14). With a lower freezing level in the LowFL case, more of the cloud depth is present above the freezing level, and thus more cloud condensation occurs, so that A_s has a noticeable effect on $\langle \text{COND} \rangle$ in addition to WRA (Fig. 15c). Figure 16 illustrates this effect. Along the upwind slope the horizontal extent of cloud water above the freezing level increases as A_s is decreased, meaning that condensation extends over a wider region, leading to greater (COND) on the upwind slope. The opposite occurs downwind, where (COND) increases as A_s is increased; this seems to be due to large A_s inducing higher-amplitude lee waves compared to small A_s , resulting in more condensation downwind (Fig. 16). As for WRA, the (COND) response functions are flatter than those in the control simulation (Fig. 15c); this is most likely due to a reduced impact of warm-rain processes



FIG. 13. As in Fig. 12, but for (top) WRA and (bottom) ECI.

on cloud and precipitation development because of the lower freezing level. For LowFL there is a minimal response of PREC, PE, and $\langle \text{COND} \rangle$ to changes in ρ_s (Figs. 14d–f), and PREC to changes in ECI (Fig. 15d), compared to the control simulation. These results reflect the complexity of precipitation sensitivities; although ice microphysical processes are generally expected to play a larger role in this case, this does not hold true for all of the associated parameter perturbations.

4. Summary and conclusions

Idealized 2D simulations of moist, nearly neutral flow over a bell-shaped mountain were performed using CM1. The idealized upwind sounding was guided by observations from an atmospheric river event during the 2015 OLYMPEX field campaign. The objective of these simulations was to explore the sensitivity of orographic precipitation to changes in microphysical parameters, as well as testing the robustness of these results in different environments. Fifteen microphysical parameters associated with both ice and warm-rain processes within the Morrison microphysics scheme were perturbed over a range of values encompassing the uncertainty and variability in these parameters. Two environmental sensitivity experiments were performed where the wind speed profile was reduced by 30% and the temperature was reduced enough to result in a freezing level below the mountain peak. These experiments allowed us to explore the precipitation sensitivity to microphysical parameters in environments where different precipitation processes were dominant.

Four parameters stood out as most influential to precipitation and cloud development: snow fall speed coefficient A_s , snow particle density ρ_s , rain accretion multiplicative factor (WRA), and ice-cloud water collection efficiency (ECI). Parameters A_s and ρ_s affect snow particle characteristics, while WRA and ECI are related to particle interaction (collection) processes. Precipitation rate responses to A_s and ρ_s perturbations are mainly a result of changes to precipitation efficiency (PE) through direct and indirect impacts, respectively, on snow fall speed. This results in a location-dependent orographic precipitation sensitivity with an increase of windward precipitation and decrease downwind with an increase of A_s , with the opposite occurring as ρ_s is increased. Increasing WRA leads to greater removal of cloud water below the freezing level from conversion of cloud to precipitation and reduced (COND)caused by drying and raising of the cloud base height. Increased PE compensates for the decrease in (COND)and results in little sensitivity on PREC. On the other



FIG. 14. As in Fig. 12, but for the LowFL experiment.

hand, increasing ECI also removes cloud water from the atmosphere yet *does* cause changes to PREC. Increasing ECI results in a decrease in cloud water above the freezing level, but little effect on updraft speeds or $\langle \text{COND} \rangle$ below the freezing level, where the largest condensation rates occur. An increase in riming efficiency through large ECI results in responses similar to those for increasing A_s . In conclusion, we see that



FIG. 15. As in Fig. 13, but for the LowFL experiment.



FIG. 16. As in Fig. 9, but with smaller values of cloud water condensation contours (dark orange contours $= 2 \times 10^{-4} \text{ g kg}^{-1} \text{ s}^{-1}$, orange contours $= 3.25 \times 10^{-4} \text{ g kg}^{-1} \text{ s}^{-1}$, yellow contours $= 5 \times 10^{-4} \text{ g kg}^{-1} \text{ s}^{-1}$) and cloud water evaporation rate (dark green contours $= -2 \times 10^{-3} \text{ g kg}^{-1} \text{ s}^{-1}$, green contours $= -3.25 \times 10^{-4} \text{ g kg}^{-1} \text{ s}^{-1}$, cyan contours $= -5 \times 10^{-4} \text{ g kg}^{-1} \text{ s}^{-1}$) for (a) $A_s = 5.86$ and (b) $A_s = 17.58$.

- microphysical parameters that directly or indirectly impact snow fall speed induce a large sensitivity on the location of orographic precipitation and
- 2) perturbations to processes that affect the conversion of cloud water to precipitation will result in different sensitivities to cloud development, orographic precipitation, and thermodynamics–dynamics, depending on where the cloud water is affected in the vertical.

The environmental experiments showed that perturbing the upwind environment results in changes to the dynamics, which has consequences for the precipitation distribution and amount, as found in MR06 and T15. Compared to perturbations of the microphysical parameters, the total amount of precipitation was more sensitive to environmental parameter perturbations. Overall, reducing wind speed had a strong effect on the amount of precipitation, while the changes in microphysical parameters in this environment had similar mechanisms responsible for changes in precipitation as in the control environment. However, because this case had much less condensate, the sensitivities were weaker. Lowering the freezing level through a reduction in the temperature profile resulted in A_s having an effect on (COND), while PREC and PE were insensitive to ρ_s perturbations. Changes to the freezing level also had an effect on the overall location of maximum precipitation, shifting it upstream to where the freezing level was above the surface and warm-rain processes could aid in precipitation production. In general, our results show complex responses of microphysical parameter perturbations to precipitation for environments where different cloud processes dominate, that is, warm-rain processes in LowU and ice processes in LowFL.

The orographic precipitation sensitivities to microphysical parameter perturbations found in this study have implications to forecasting and water management. Our results show these parameter perturbations have a strong influence on where precipitation falls, thus influencing which water basins may receive precipitation over a mountainous region. The environmental parameter changes had a larger effect on the total amount of precipitation and its distribution, and hence should continue to be explored in conjunction with microphysics. Evaluation of parameter sensitivity is also important for ensemble forecasting and data assimilation, as generation of ensembles increasingly relies on perturbations to model physics as well as initial conditions (Berner et al. 2017). It is generally not computationally feasible to perturb every parameter in the full set of model parameterization schemes. Our results suggest that a small subset of the total number of parameters are responsible for most of the microphysics-induced variability in orographic precipitation.

Although the results presented here focused on orographic precipitation, parameters such as WRA, ECI, and A_s also showed effects on the liquid and ice water paths. While cloud radiative effects are beyond the scope of this study, we suggest that future work should explore the impact of microphysical parameter perturbations on cloud optical properties and radiative transfer.

These results highlight the complexity of the orographic precipitation response to microphysical parameter changes, leading us to the following questions: 1) will changing various parameters simultaneously modify the precipitation response, and 2) how do simultaneous changes to microphysical, environmental, and mountain shape parameters affect the orographic precipitation sensitivity? To answer these questions, future work will involve performing multivariate perturbations using the parameters tested in this study as well as additional environmental and mountain geometry parameters. The goal will be to understand the orographic precipitation response to parameter covariation and the effects of interacting parameter perturbations.

Acknowledgments. The authors thank George Bryan for his assistance with the CM1 source code; Kristen Rasmussen for helping to get access to OLYMPEX sounding data; and Rich Rotunno, Dale Durran, and Marty Ralph for their helpful suggestions in setting up our control and environmental perturbation soundings. We would also like to thank three anonymous reviewers for their comments, which have improved the clarity and contents of this manuscript. A.M. is hosted by the National Center for Atmospheric Research (NCAR) and the Mesoscale and Microscale Meteorology lab. Part of this work was funded by NCAR's Advance Study Program's Graduate Student Fellowship and by NASA Interdisciplinary Science Grant NNX14AG68G. H.M. was partially supported by the U.S. Department of Energy Atmospheric System Research Grant DE-SC0016476. The National Center for Atmospheric Research is sponsored by the National Science Foundation. A portion of this research was carried out at the Jet Propulsion Laboratory, California Institute of Technology, under a contract with the National Aeronautics and Space Administration.

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