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Impacts of ice-nucleating particles from marine aerosols on mixed-phase orographic clouds during 2015 ACAPEX field campaign

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Abstract

A large fraction of annual precipitation over the western United States comes from wintertime orographic clouds associated with atmospheric rivers (ARs). The transported African and Asian dust and marine aerosols from the Pacific Ocean may act as ice-nucleating particles (INPs) to affect cloud and precipitation properties over the region. Here we explored the effects of INPs from marine aerosols on orographic mixed-phase clouds and precipitation at different AR stages for an AR event observed during the 2015 ACAPEX field campaign under low dust (< 0.02 cm^{-3}) conditions. Simulations were conducted using the chemistry version of the Weather Research and Forecasting model coupled with the spectral-bin microphysics at 1-km grid spacing, with ice nucleation connected with dust and marine aerosols. By comparing against airborne and ground-based observations, accounting for marine INP effects improves the simulation of cloud phase state and precipitation. The marine INPs enhance the formation of ice and snow, leading to less shallow warm clouds but more mixed-phase and deep clouds, and increased ice water path (over 5 times) and snow precipitation (over 40 times). The responses of cloud and precipitation to marine INPs vary with the AR stages with more significant effects before AR landfall and post-AR than after AR landfall, mainly because the moisture and temperature conditions change with the AR evolution. This work suggests weather and climate models need to consider the impacts of marine INPs since their contribution is notable under low dust conditions despite the much lower relative ice nucleation efficiency of marine INPs.

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1 Introduction

Atmospheric river (AR) events have great impacts on atmospheric and hydrological processes in the western United States during winter. They account for 30–50% of the total winter precipitation through their impacts on orographic clouds and associated heavy precipitation (Dettinger et al., 2011). Understanding the factors influencing different types of precipitation (rain vs. snow) associated with ARs is crucial for planning and managing regional water resources and hydrologic hazards and improving atmospheric and hydrologic forecasting in the western United States. Rain and snow precipitation produced by orographic clouds over the Sierra Nevada Mountains is closely related to the partitioning between cloud liquid and ice phases, which can be largely modified by aerosol particles (Rosenfeld et al., 2013; Fan et al., 2014, 2017b). However, aerosol-orography-precipitation relationships are complicated, depending on aerosol properties, mountain geometry, cloud phase, temperature, humidity, and flow patterns as reviewed in Chouldhury et al. (2019).

Over the western United States, understanding the roles of aerosols, particularly those capable of initiating ice crystal formation, in altering clouds and precipitation is still limited, which has motivated recent observational and modeling studies (Ault et al., 2011; Creamean et al., 2013, 2015; Rosenfeld et al., 2013; Fan et al., 2014, 2017b; Martin et al., 2019; Levin et al., 2019). While it has been found that long-range transported aerosols particularly dust particles as ice nucleating particles (INPs) influence clouds and precipitation in the mountainous western United States (Uno et al., 2009; Ault et al., 2011; Creamean et al., 2013), it is also clear from measurements that clouds occurring in and around ARs can also be influenced by INPs with apparent sources from the ocean (Levin et al., 2019).

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Where sufficient INPs are present, heterogeneous ice nucleation can occur at a higher temperature where liquid and ice can co-exist, in contrast to homogeneous freezing which normally requires -38 ˚C or colder environment (DeMott et al., 2010; Vali et al., 2015). Increasing INPs can lead to stronger ice formation and growth at the expense of supercooled liquid and increase precipitation. This concept is the basis of using cloud seeding to increase orographic precipitation (Reynolds, 1988; Geerts et al., 2010; French et al., 2018). For orographic clouds in the western United States, previous studies showed that INPs can increase total precipitation through the "seeder feeder" mechanism (Choularton and Perry, 1986; Creamean et al., 2013), in which ice crystals that form in the upper portions of orographic clouds can collect droplets and grow to a larger size as they fall through a supercooled liquid layer before reaching the ground. Fan et al. (2014, 2017b) found that INPs like dust particles can increase precipitation by enhancing riming and deposition processes in mixed-phase orographic clouds, consistent with other studies (e.g., Muhlbauer and Lohmann, 2009; Xiao et al., 2015; Hazra et al., 2016; Yang et al., 2020). Fan et al. (2017a) also noted that the relative importance of riming to deposition depends on the mixed-phase cloud temperatures. Despite the importance of INPs in cloud formation and precipitation, they typically have a low abundance and large variations in their nucleating characteristics, especially in terms of the temperatures over which they initiate ice crystal formation (Kanji et al., 2017; Levin et al., 2019). Hence, there is large uncertainty in evaluating INPs impacts on mixed-phase and ice clouds as well as precipitation.

It is known that dust particles are important INP sources, which can initiate freezing over a range of temperatures but most efficiently below −20 °C (Murray et al., 2012; Kanji et al., 2017). Another important type of INPs is terrestrially sourced biological particles, which cause freezing at temperatures as warm as −5 °C (Murray et al., 2012). During ARs, the long-range

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transport of dust or biological particles is highly episodic. Sea spray or marine aerosols consisting of sea salt and marine organic carbon resulting from wave breaking and bubble bursting at the ocean surface may also be a source of INPs (Burrows et al., 2013; Vergara-Temprado et al., 2017; McCluskey et al., 2018b; Levin et al., 2019). Recently, McCluskey et al. (2018a) derived an ice nucleation parameterization for INPs from sea spray aerosols based on observations collected at a North Atlantic coastal site and its relation to the marine aerosol surface area. Given the distinct physio-chemical characteristics and the different ice-nucleating efficiency, the impact of marine INPs on cloud and precipitation could be very different from dust or biological particles (DeMott et al., 2016; Kanji et al., 2017). However, studies of marine-sourced INP effects on clouds and associated precipitation are limited (Kanji et al., 2017; Levin et al., 2019). A few previous studies investigated the impacts of marine INPs on precipitation and radiation with global climate models (Hoose et al., 2010; Burrows et al., 2013; Yun and Penner, 2013), albeit without the advantage of direct data on their ice nucleation efficiencies. Further, a detailed, process-level understanding of how marine INPs affect mixed-phase cloud processes and precipitation is lacking.

Following the CalWater campaigns in 2009, 2011, 2014, an interagency sponsored study, CalWater 2015, utilized a larger suite of instruments and measurement platforms to study ARs and aerosol-cloud interactions in AR environments (Ralph et al., 2016). As part of CalWater 2015, the U.S. Department of Energy sponsored Atmospheric Radiation Measurement (ARM) Cloud Aerosol Precipitation Experiment (ACAPEX) field campaign aimed specifically at improving understanding and modeling of aerosol impacts on winter storms associated with landfalling ARs (Leung et al., 2016). The ACAPEX campaign conducted intensive sampling of clouds and aerosols using instruments on board the ARM Aerial Facility Gulfstream (G-1)

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aircraft and ARM Mobile Facility on board the research vessel Ron Brown. These measurements were made in conjunction with clouds and aerosols, meteorological, hydrological, and oceanic measurements collected by instruments on three other aircraft and Ron Brown and at a coastal surface station. Collectively, these data provide a unique opportunity to examine the complex interactions among aerosols, orographic clouds, and ARs.

A major AR event spanning over 5 - 9 February 2015 occurred during the ACAPEX campaign and made landfall on the coast of Northern California, producing heavy rainfall with some regions receiving up to 400 mm of total precipitation during the event (Ralph et al., 2016; Cordeira et al., 2017). This AR event was extensively sampled by the (G-1 aircraft (Schmid et al., 2014) for characterizing aerosol and cloud properties. During this event, marine aerosols were the main aerosol type and marine INPs were dominant at cloud activation temperatures. Aerosol sampled by G-1 indicated that dust and biological particles were rather scarce in and around ARs, which is in stark contrast to the dominance of dust INPs during the AR events in the CalWater 2011 campaign (Levin et al., 2019). Therefore, the AR event during the ACAPEX campaign provides a rather unique opportunity to explore the role of marine aerosols in the orographic clouds and precipitation associated with landfalling ARs in the western United States.

In our previous modeling studies (Fan et al., 2014, 2017b), we implemented an immersion freezing parameterization for dust particles (DeMott et al. 2015) in a spectral-bin microphysics (SBM) scheme to examine the long-range dust effects on AR-associated orographic mixed-phase clouds and precipitation during CalWater 2011. With marine INPs dominating in CALWATER 2015/ACAPEX, in this study we implemented the recently developed ice immersion nucleation parameterization for sea spray aerosols by McCluskey et al. (2018b) in the SBM scheme. To explicitly simulate various aerosol types, different from Fan et

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al. (2014, 2017a) who prescribed aerosols based on observations, a chemistry version of the Weather Research and Forecasting model (WRF-Chem) coupled with the SBM (Gao et al., 2016) was employed to predict aerosol properties and their interactions with clouds and radiation for the AR event on 6 - 9 February 2015. We focused on exploring the effects of INPs from sea spray aerosols, in competition with mineral dust INPs, on the orographic mixed-phase clouds and precipitation at different stages of the AR event as thermodynamic conditions evolved with the different AR stages.

2 Model configuration and experiment design

The WRF-Chem version 3.6 coupled with SBM as described in Gao et al. (2016) is employed for model simulations of this study, in which SBM is coupled with the Model for Simulating Aerosol Interactions and Chemistry (MOSAIC; Fast et al., 2006; Zaveri et al., 2008). The SBM scheme is a fast version in which ice crystal and snow (aggregates) in the full version are represented with a single size distribution (low-density ice) with a separation at 150 μm in radius, with graupel or hail populating larger sizes (Khain et al., 2009, 2010; Fan et al., 2012, 2017a). Here we choose the graupel version since hail is not one of the major cloud hydrometeors in the case we simulate. The WRF-Chem-SBM model is particularly designed to improve simulations of aerosol effects on clouds for complicated aerosol compositions and heterogeneous spatial distribution of aerosols. It has been applied in several studies including warm stratocumulus clouds (Gao et al., 2016), thunderstorms (Fan et al. 2020; Zhang et al., 2020), and supercell storms (Lin et al., 2020). Here WRF-Chem-SBM is employed, different from our previous studies in Fan et al. (2014, 2017a) which used WRF-SBM with prescribed aerosols, in order to explicitly simulate various aerosol types including marine aerosols and dust particles.

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156 The four-sector MOSAIC aerosol module is chosen for the simulations of aerosols and
157 the CBMZ (Carbon Bond Mechanism version Z) is used for gas-phase chemistry. The MOSAIC
158 module treats nine major aerosol species (sulfate, nitrate, chloride, ammonium, sodium, black
159 carbon, primary organics, other inorganics (OIN), and water). OIN is used as a surrogate of dust
160 and the production of dust is parameterized with the dust transport model DUSTRAN (Shaw et
161 al., 2008). Sea salt aerosol (the combination of sodium and chloride), as a surrogate for all SSA,
162 is parameterized as a function of sea-surface wind speed (Gong et al., 1997b, a). The dry
163 diameters of the particles over the four bins have a range of 0.039–0.156, 0.156–0.624, 0.624–
164 2.5, and 2.5–10.0 μm, respectively. For the total aerosol, aerosol size distribution over each
165 section is represented with a 2-moment approach that predicts aerosol mass and number
166 following a log-normal distribution [Simmel and Wurzler, 2006]. For each composition such as
167 dust and sea salt, only the mass mixing ratio in each section is predicted and outputted. The
168 aerosol number mixing ratio in each bin is only predicted for the total aerosol. Therefore, in this
169 study, the dust and sea salt number mixing ratios used for ice nucleation parameterizations are
170 derived based on their respective mass mixing ratio by assuming the same size and density of all
171 particles over each bin, that is,

172 $$N_{i,j} = \frac{m_j}{6\pi(D_j)^3\rho_i}$$

173 where i denotes the aerosol composition (sea salt or dust here), j denotes the j^{th} aerosol bin, m_j is
174 the total mass mixing ratio of the j^{th} bin, ρ_i is the assumed density (i.e., 2.6 g cm^{-3} for dust and
175 2.2 for sea salt), and D_j is the geometric mean diameter of the j^{th} bin. The approach for deriving
176 the number mixing ratio for each aerosol component has been used in the literature (i.e., Zhao et
177 al., 2013). We understand that the assumption that all particles have the same size over each bin

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may introduce some uncertainty. However, the size distribution of each aerosol component is unknown in the model and any assumption on the size distribution might introduce uncertainty.

2.1 Implementing an immersion freezing parameterization for marine INPs

In the original SBM model, the ice nucleation accounting for both deposition ice nucleation and condensation-freezing is parameterized based on Meyers et al. (1992) and Bigg (1953) is employed for the immersion and homogeneous drop freezing. Neither of the ice nucleation parameterizations is connected with aerosols. Bigg (1953) was formulated based on the stochastic hypothesis where the freezing probability is assumed proportional to drop mass and the freezing rate is as a function of temperature without involving INPs. Fan et al. (2014, 2017a) implemented DeMott et al. (2015) as an immersion freezing parameterization to investigate the effects of dust INPs on orographic mixed-phase clouds and precipitation during CalWater 2011. We adapted this implementation to WRF-Chem-SBM for this study to connect ice nucleation with dust particles. Developed based on both laboratory data and field measurements, DeMott et al. (2015) is an empirical parameterization for immersion freezing of natural mineral dust particles. INP concentrations are quantified as functions of temperature and the total number concentration of particles larger than 0.5 μm diameter. In our implementation, the dust number mixing ratio for each aerosol bin is derived from its mass as detailed in the section above. The total dust number mixing ratio inputted to DeMott et al. (2015) is the integration over 0.5 -10 μm.

To connect ice nucleation with sea spray aerosols, we implemented McCluskey et al. (2018a, thereafter MC2018), which was developed for quantifying ice nucleating activity by marine organics over the North Atlantic Ocean, in SBM following a similar approach as the implementation of DeMott et al. (2015). The nucleation site density in MC2018 is described as

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$$n_s = exp(-0.545(T - 273.15) + 1.012)$$

where n_s is the nucleation site density (m^{-2}) and T is the temperature (K). With n_s determined by MC2018, the nucleated ice particle concentration is obtained following Niemand et al. (2012) as

$$\sum_{j=1}^{n} N_j = \sum_{j=1}^{n} N_{\mathrm{tot},j}\{1 - \exp[-S_{\mathrm{ae},j} n_s(T)]\}$$

where $S_{ae,j}$ is the surface area of individual sea spray aerosol particles in the j^{th} bin which is calculated from $\pi D_j^2/4$ (D_j is the geometric-mean diameter), $N_{tot,j}$ is the total sea spray aerosol number in each bin which is derived from its mass as detailed in the section above, and N_j is the ice particle number in each bin. Sea salt particles are used as the surrogate of sea spray aerosols given that most marine organic aerosols exist with coating on the surface of sea salt particles in the size range that dominates surface area (e.g., Prather et al., 2013). MC2018 can have efficient ice nucleation at warm temperatures like -15 ˚C or warmer.

Bigg et al. (1953) is employed only for homogeneous drop freezing when the temperature is colder than -37 ˚C. As discussed in Fan et al. (2014), the deposition-condensation freezing is turned off because the simulation with deposition-condensation freezing produces a large number of small ice particles, which is not consistent with the observed mixed-phase cloud properties in the study region. Contact freezing is also turned off due to the negligible contribution (Fan et al., 2014).

2.2 Experiment design

Simulations are configured with two nested domains using the nesting down approach (i.e., the inner domain is run separately driven by the outer domain), covering most of the

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224 western US (Fig. 1). The outer domain consists of 399 × 399 grid points with a horizontal grid
225 spacing of 3 km and the inner domain consists of 498 × 390 grid points with a horizontal grid
226 spacing of 1 km. 50 vertical levels with stretched intervals are configured, with a grid spacing of
227 70 m at the lowest levels and ~400 m at the model top. The dynamics time step is 15 seconds for
228 the outer domain and 5 seconds for the inner domain.

229 The simulation for the outer domain starts at 00:00 UTC on February 3 and runs for 48
230 hours for chemistry spin-up using the WRF-Chem-SBM model, driven by global WRF-Chem
231 simulation as the initial and boundary conditions of gas-phase species and aerosols and the
232 Modern-Era Retrospective analysis for Research and Applications, Version 2 (MERRA2; spatial
233 resolution of 0.5 by 0.5 degree and temporal resolution of 6-hourly) as the initial and boundary
234 conditions of meteorological fields. Then the outer domain simulation is reinitialized at 00:00 UTC
235 on February 5 using the meteorological data from MERRA2 to avoid the large error growth in
236 meteorology associated with long-time model integration, although the chemistry simulations is a
237 continuation from the spin-up run, and runs until 23:00 UTC on February 8. Given that running
238 the WRF-Chem-SBM fully-coupled model is extremely computationally expensive for 1-km grid
239 spacing in the inner domain, we interpolate aerosol-related quantities such as aerosol composition,
240 hygroscopicity, and mass and number concentrations from the outer domain simulations using
241 bilinear interpolation for the inner-domain simulation to reduce computational cost. This means
242 we conduct the inner-domain simulation separately with chemistry turned off, and aerosol
243 information is updated hourly using data from the outer domain simulations. The inner-domain
244 simulation is run from 00:00 UTC on February 5 to 23:00 UTC on February 8, and the initial and
245 boundary meteorological conditions are from MERRA2. To validate this approach, we compare
246 the simulation with fully coupled WRF-Chem-SBM for the inner domain simulation and found

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that the two simulations resemble each other in terms of precipitation (Fig. S1). Therefore, it is a valid approach that saves computation time by about 40%.

For emissions data, the U. S. Environment Protection Agency (EPA) National Emission Inventory (NEI) with a 4 km by 4 km horizontal resolution based on the year 2011 rates (NEI2011) is commonly used for anthropogenic emissions in the United States. However, using NEI2011 predicts too large anthropogenic aerosol mass compared with observations. Since the emissions of gaseous species and particulate matter decreased significantly from 2011 to 2015 in California (Table S1), the California Air Resources Board emission inventory in 2015 (CARB2015) is used for anthropogenic emissions input for California, while NEI2011 is used for other states in the simulation domain. The use of NEI2011 for other states is acceptable since the lower and middle atmosphere in the simulation domain is dominated by southwesterly winds during the simulation period that transport air pollutants from coastal to inland regions. The use of CARB2015 reduces the simulation of aerosol number concentrations mainly below 2 km by 40% relative to the use of NEI2011, in better agreement with observations.

The Model of Emissions of Gases and Aerosols from Nature (MEGAN) with a monthly temporal and 1 km horizontal resolution (Guenther et al., 2012) is used for biogenic emissions. The Rapid Radiative Transfer Model for application to GCMs (RRTMG) is used for shortwave and longwave radiation schemes (Iacono et al., 2008), the Noah Land Surface Model for land surface physics (Chen and Dudhia, 2001), and the Mellor-Yamada-Janjic (MYJ) scheme for planetary boundary layer parameterization (Mellor and Yamada, 1982; Janjić and Prediction, 2001).

Three simulations were carried out over the inner domain for this study to investigate the impacts of marine INPs: (1) The reference case is Bigg, using the default immersion freezing

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parameterization of Bigg et al. (1953) in SBM which is temperature-dependent only; (2) DM15+MC18, in which both DeMott et al. (2015) and MC2018 parameterizations are used for ice nucleation from dust and marine aerosols, respectively; (3) DM15, using the parameterization of DeMott et al. (2015) for dust aerosols (diameter > 0.5 μm) with MC2018 turned off. The impacts of marine INPs are derived by comparing the DM15+MC18 and DM15 simulations.

3 Results

3.1 Model evaluation with observations

We evaluate the model simulations of aerosol and cloud properties and surface precipitation. Figure 2a shows a comparison of modeled aerosol properties including aerosol number concentration and chemical composition from the simulation of DM15+MC18 intended to represent the observed case, with the G-1 aircraft measurements on 7 February. Aerosol properties in all three simulations are similar, and thus only DM15+MC18 is shown. Overall, the simulated aerosol number concentration over the size range of 0.067 - 3 μm is comparable to the observations over the same size range estimated by combining data from the Ultra-High-Sensitivity Aerosol Spectrometer (UHSAS) and the Passive Cavity Aerosol Spectrometer Probe (PCASP) at below 2-km altitude. The simulation overestimates the total aerosol number concentrations up to ~ 2-times at altitudes of 2.2-3.2 km. For aerosol composition, the airborne Aerosol Time of Flight Mass Spectrometry (ATOFMS) measurements provided the mean fractional number contributions of aerosol source classifications (Levin et al., 2019), which is shown in Fig. 2b. For comparison with the model, the mean mass contributions of the corresponding aerosol source classifications are computed since the number concentrations of individual aerosol components are not predicted by WRF-Chem (Fig. 2c). Both the observed fractional number contributions and the simulated mass contributions show that marine aerosols

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293 are dominant during the AR event, accounting for more than 60% of the total aerosol number
294 based on ATOFMS measurements and total aerosol mass based on the simulation. Although the
295 simulated dust mass fraction is ~14%, the derived number concentration is very low (less than
296 0.02 cm^{-3} for sizes larger than 0.5 μm, shown in a later figure). This is because dust is mainly
297 from aerosol bins at larger sizes. The number concentrations of the sea salt aerosols are generally
298 three orders of magnitude higher than those of dust, and these numbers populate smaller bins of
299 the aerosol distribution (97% from the first two aerosol size bins) even though the sea salt mass
300 is predominately at larger sizes (96% from the last two size bins).

301 Figure 3 presents an evaluation of precipitation, showing the accumulated precipitation
302 during the AR event from 06:00 UTC 5 February to 09:00 UTC 8 February 2015 (Fig. 3a) and
303 the time-series of mean precipitation rates averaged over the observation stations (Fig. 3b). The
304 observed precipitation rates are from the rain gauge measurements, provided by the NOAA Earth
305 System Research Laboratory's Physical Sciences Division
306 (https://psl.noaa.gov/data/obs/datadisplay). The model generally captures the spatial pattern of
307 the observed accumulated precipitation (Fig. 3a) and reproduces the temporal evolution of the
308 precipitation rate (Fig. 3b). Two major precipitation periods in the observations including the
309 AR-induced orographic precipitation and the postfrontal precipitation are generally captured by
310 the simulations, although the simulated postfrontal precipitation occurs several hours later in the
311 simulations compared to the observations. The peak values of precipitation are overestimated by
312 all the simulations (25-50%) relative to the observations (Fig. 3b), but the values from
313 DM15+MC18 are lower than the other two, closer to the observations. From the spatial
314 distributions (Fig. 3a), the observed accumulated precipitation in the southern part of the
315 mountain range is generally less than 100 mm. DM15+MC18 simulates it well, whereas the

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316 other two simulations overestimate precipitation in that area. The lower precipitation by
317 DM15+MC18 is mainly because of the spillover effect caused by marine INPs (Fig. 3a, right).
318 That is, with marine INPs, more ice/snow formed over the windward side falls slower than rain
319 and more of them are transported to the lee side. In the northern part of the domain (> 40º N),
320 DM15+MC18 predicts more precipitation than the other two simulations. The simulated
321 precipitation between Bigg and DM15 are very similar, suggesting that in a low dust
322 environment, the temperature-dependent Bigg (1953) parameterization simulates similar ice
323 formation as DeMott et al. (2015).

324 Cloud phase is crucial to radiation and precipitation for mixed-phase clouds, and the
325 glaciation ratio is usually used to represent the cloud phase states. The glaciation ratio is defined
326 as (IWC/(IWC+LWC)), where LWC and IWC denote liquid and ice water content, respectively.
327 Values less than 0.1 and larger than 0.9 denote the liquid phase and ice phase, respectively, with
328 values between 0.1 and 0.9 for the mixed-phase (Korolev et al., 2003). The G-1 aircraft sampled
329 the postfrontal clouds on February 7 as shown in Fig. 4a. All three simulations cannot capture the
330 observed size of the precipitation cell (Figs. 4b and S2). In the simulations, precipitation is
331 dominated by a few heavy precipitation clusters instead of the observed wide precipitation area.
332 The simulated cells also do not reach the high altitudes found in the observations. The deviations
333 of the simulation from observations for the postfrontal clouds could be mainly because of the
334 long-time model integration time (the 4th day after model initiation). Anyhow, DM15+MC18
335 simulates the largest size of the precipitation cell, with the highest vertical extent among the
336 three simulations.

337 The LWC and IWC are derived from observations with the Water Content Monitor
338 (WCM), an instrument that uses the impact of water on several heated wires as the basis for

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measuring cloud total water content (TWC) and liquid water content (LWC) from which the ice water content (IWC) can be derived [Baumgardner et al., 2011; Matthews et al., 2015]. LWC and IWC along both horizontal and vertical flight segments are displayed in Fig. 5a-b. IWC is generally 2-4 times larger than LWC in the postfrontal clouds. To compare with observations, the model data are processed by: (a) selecting the grids at a distance from the simulated cell center similar to the distance of the airplane position from the observed postfrontal cell center, and sampling the data at a similar ambient temperature as observed by the aircraft (around -10 ºC along the horizontal segment shown in Fig. 5a); (b) accounting for the location mismatch and increasing the sample size in the simulation to be more representative by extending the sampling area to include 20 grids at the front and back of a selected grid along the flight track, mimicking approximately the distance traveled by the G-1 airplane in five minutes; (c) filtering out the sampled grids with values of (LWC + IWC) below the detection limit of WCM (i.e., 0.02 g m^{-3}, Thompson et al., 2016). Both horizontal and vertical flight segments are incorporated for comparison.

Figure 5 shows comparisons of LWC and IWC and the glaciation ratio of IWC/(IWC+LWC) between the simulations and aircraft measurements. LWC is overestimated in all three simulations, while IWC is underestimated in Bigg and DM15 but overestimated in DM15+MC18 (Fig. 5c). The mean glaciation ratios fall in the range of 0.1- 0.9 among the simulations (Fig. 5d), indicating that the observed mixed-phase cloud feature is simulated by the model. DM15+MC18 shows a mean ratio of ~ 0.70, similar to the observed value of 0.74. This shows that the mixed-phase state is well captured when the marine INP effect is considered. In contrast, in Bigg and DM15 with a glaciation ratio is 0.41 or less, the mixed-phase state is liquid-dominated. The inclusion of the marine INP effect improves the simulation of cloud phase states

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via enhancing heterogenous ice formation through immersion freezing. A detailed examination of how the marine INPs impacts ice nucleation and cloud properties will be discussed in the following section.

3.2 Marine INP effects under different AR stages

Impacts of the marine INPs transported from the Pacific Ocean on orographic clouds and precipitation are revealed by comparing the simulation of DM15+MC18 with the simulation of DM15.

The AR evolution has three distinct stages: before AR landfall (from 06:00 UTC 5 to 18:00 UTC 6 February), after AR landfall (from 18:00 UTC 6 to 12:00 UTC 7 February), and post-AR (from 12:00 UTC 7 to 09:00 UTC 8 February). The three stages can be identified from the change of the integrated water vapor (IWV) with time during the event (Fig. 6a). Before AR landfall, IWV in most of California was relatively low (Fig. 6a, left). IWV in northern California increased as the AR made landfall at about 18:00 UTC on 6 February and brought ample water vapor to California (Fig. 6a, middle). Heavy orographic precipitation along the Sierra Nevada Mountains occurred during this period (Fig. 7a). At 12:00 UTC 7 February, the AR started to retreat (Fig. 6a, right), and postfrontal cloud cells formed, with relatively small cloud fraction and precipitation (Fig. 7a).

Vertical profiles of the thermodynamic and kinematic environments at the three stages are shown in Fig. 6b. The thermodynamic and kinematic environments significantly varied with the AR stages. After AR landfall, water vapor increased significantly in the lower atmosphere (below 5 km), but the middle and upper levels became drier (dashed, Fig. 6b) compared with the stage before AR landfall (solid). The vertical motion also weakened after AR landfall (Fig. 6d),

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385 suggesting that the atmosphere became more stable. At the post-AR stage, moisture above 2-km
386 altitude was reduced compared to after AR landfall. Note that the temperature below 8 km was
387 colder by up to 6 ºC at the post-AR stage compared to the previous two stages (Fig. 6c). These
388 differences in the meteorological conditions among the different stages are very important to
389 understand the cloud and precipitation properties and their responses to marine INPs. The mean
390 water vapor and temperature profiles are similar between DM15 (blue) and DM15+MC18 (red),
391 as seen from the overlapping blue and red lines.

392 From the time series of average precipitation rates (Fig. 7a), the effect of marine INPs
393 varies with the different AR stages, from substantial increases of precipitation (up to 330%)
394 before AR landfall (the red dotted line, second y-axis) to no significant effects (a very small
395 increase) after AR landfall. In the first stage (before AR landfall), the total precipitation increases
396 by 36% on average due to the marine INP effect (Fig. 7a and Table 1). There is only a 4%
397 increase in the total precipitation after AR landfall. The total precipitation at the post-AR stage is
398 small and the change from DM15 and DM15+MC18 is negligible. Thus, the marine INP effect
399 only significantly increases the total precipitation over the domain at the stage before AR landfall
400 when a moderate amount of precipitation occurs in northern California (Fig. 8a). After AR
401 landfall, precipitation increases significantly. Although the total precipitation is not changed
402 much by the marine INPs, the marine INPs produce a spillover effect featuring reduced
403 precipitation on the windward slope of the mountains but increased precipitation over the lee side
404 (Fig. 8b and Fig. 9e). This is because with the marine INPs, the larger amount of ice/snow that
405 forms on the windward slope is transported to the lee side (Fig. 9d) and grows to a larger size
406 and precipitates as snow. This spillover effect is accompanied by a large reduction of cloud water
407 and rain over the windward side because of conversion of liquid to ice (Fig. 9b-c). Since the

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408 water vapor transport along the cross-section is very similar between DM15 and DM15+MC18
409 (Fig. 9a), the spillover effect by marine INPs is mainly the result of different cloud
410 microphysical properties instead of meteorological conditions.

411 Even though the total domain precipitation is not changed much by the marine INPs at
412 the later two stages, the cloud phase and the near-surface precipitation type (i.e., rain or snow)
413 are notably changed (Table 1). The mean glaciation ratio in the mixed-phase is very low in
414 DM15 (0.14, 0.16, and 0.001 for the 1st, 2nd, and 3rd stages, respectively) and is increased to 0.74,
415 0.59, and 0.36, respectively. We examine the ratio of snow/(rain+snow) in mass mixing ratio at
416 the lowest model level for the changes of the near-surface precipitation type (Fig. 7b). There is
417 negligible snow precipitation near the surface in DM15 and the ratios of snow precipitation are
418 very small during the entire AR event. The snow precipitation ratios increase in DM15+MC18
419 and the magnitudes vary significantly with different AR stages. On average, the ratio of snow
420 precipitation increases from 0.002, 0.001, <0.001 in DM15 to 0.08, 0.04, and 0.13 in
421 DM15+MC18 before AR landfall, after AR landfall, and post-AR, respectively (Table 1). This
422 shows that the marine INPs increase snow precipitation and the effect is particularly significant
423 before AR landfall and post-AR. Correspondingly, rain precipitation is reduced (Table 1). This
424 has an important implication for the regional hydrological resource since more snow
425 accumulation in winter increases freshwater resources in the summer while less rain reduces
426 flood risks.

427 The increased snow and reduced rain at the surface correspond to the increased ice water
428 path (IWP) and decreased liquid water path (LWP; Fig. 7c). The mean LWP in DM15+MC18 is
429 reduced by 66%, 46%, and 26% for the three stages relative to DM15, respectively (Table 1). We
430 showed an increased LWC from DM15 to DM15+MC18 in Figure 5c in the postfrontal cells.

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Here the decrease in LWC/LWP averaged over the entire post-AR stage is dominated by the strong decrease over the time before the postfrontal cloud formed. The postfrontal cells are invigorated (see section 3.3) by marine INPs so both LWC and IWC are increased as shown in Figure 5. IWP is greatly enhanced by about 8, 5, and 440 times at the three stages, respectively. Interestingly, the total condensate water path (TWP) is increased by the marine INPs (Fig. 7d). On average there are 45%, 29%, and 35% increases in TWP in DM15+MC18 at the three AR stages relative to DM15, respectively (Table 1). The increases in the total condensate water path and the increased surface precipitation (or no change) suggest that marine INPs enhance the conversion of water from the vapor phase to the condensate phase, which will be further discussed later. This is particularly the case before AR landfall with water vapor content notably reduced in DM15+MC18 compared with DM15 (Fig. S3a).

Cloud cover is slightly increased during the first two stages (4-5%) in the simulations including marine INPs, but the change at the post-AR stage is ~ 20% on average, which is very significant. Because both TWP and cloud cover are increased due to the marine INP effect, the cloud radiative forcing (CRF) at TOA gets stronger by 15%, 13%, and 10% for the three AR stages, respectively. Although the cloud phase, precipitation type, and cloud fraction at the post-AR stage have the largest changes among the three stages by the marine INP effect (Table 1), the CRF does not change drastically probably because of the offset between the increase resulting from the increased cloud fraction and TWP and the decrease from the reduced cloud liquid is the largest.

Overall, the marine INP effects on TWP, IWP, and snow precipitation are more significant at the first and third stages (i.e., before AR landfall and post-AR) than the stage after AR landfall. But a notable spillover effect is seen after AR landfall. The cloud and precipitation

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454 quantities are more sensitive to marine INPs before AR landfall than after AR landfall, and the
455 responses of TWP/IWP and snow precipitation are particularly drastic at the post-AR stage
456 (Table 1). The reasons leading to the different responses at different AR stages are now
457 examined.
458
459 **3.3 Explaining different marine INP effects at different AR stages**
460 We first examine the temporal evolution of dust and marine aerosol number
461 concentrations, which are derived based on the predicted mass mixing ratios as described in
462 Section 2 and used as input to the DeMott et al. (2015) and MC2018 parameterizations (Fig. 10a,
463 b), as well as their corresponding immersion freezing (i.e., ice nucleation) rates (Fig. 10c, d). The
464 dust concentrations and the corresponding ice nucleation rates (Fig. 10a, c) are about three orders
465 of magnitude lower than those of the marine aerosols (Fig. 10b, d) during the AR events. This is
466 driven both by the activation temperature spectrum of dust and its very low mass/number
467 concentrations in this case. Ice nucleation from dust is negligible at temperatures warmer than -
468 15 ºC but the ice nucleation from marine aerosols is notable. This is mainly because of three
469 orders of magnitude higher marine aerosol number concentrations from the surface up to 8 km.
470 The deep marine aerosol layer during the AR allows notable ice nucleation at temperatures
471 higher than -15 ºC. The clear-sky marine aerosol number concentrations increase from before
472 AR landfall to post-AR as the AR evolved (Fig. 10b). After the AR makes landfall, marine
473 aerosols increase significantly as AR strong winds near the ocean surface produce them more
474 and also transport more to the Sierra Nevada Mountains (Fig. 10b). Despite the significant
475 increase in marine aerosols after AR landfall, the marine INP effects on clouds and precipitation
476 are small at this stage, because the increase of marine aerosols does not increase ice nucleation

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477 rates (Fig. 10d). However, at the post-AR stage, the ice nucleation rates from the marine INPs
478 are up to a few times larger than the earlier two stages (Fig. 10d), explaining why the effects on
479 IWP and snow precipitation at the post-AR stage are largest among the three stages.
480 To further understand how and why the cloud and precipitation responses to marine INPs
481 are different at different AR stages, we separate clouds into three cloud regimes: a shallow warm
482 cloud regime with cloud top temperature (CTT) warmer than 0 ºC, a mixed-phase cloud regime
483 with CTT between -30 and 0 ºC, and a deep cloud regime having CTT colder than -30 ºC and
484 cloud base temperatures above 0 ºC. Figure 11 shows that the marine INP effect consistently
485 shifts the cloud occurrences from the shallow warm cloud regime to mixed-phase and/or deep
486 cloud regimes among the three AR stages. It is noted that the deep cloud regime is enhanced
487 much more at the first and third stages than the second stage, i.e., 22% before AR landfall and
488 235% at the post-AR stage but only 8% after AR landfall. The post-AR stage also has the largest
489 increase in mixed-phase cloud occurrences.
490 Accordingly, the mean cloud depth for each cloud regime is changed by marine INPs,
491 with a decrease for the shallow warm clouds and an increase for the mixed-phase and deep
492 clouds (Fig. 11b). Before AR landfall, the increase in the deep cloud depth is largest while at the
493 post-AR stage, the increase in the mixed-phase cloud depth is the largest. Consistent with a shift
494 in cloud regimes, the total precipitation produced by shallow warm clouds is reduced by 9%,
495 22%, and 16% while the total precipitation produced by deep clouds is increased by 66%, 4%,
496 and 350%, respectively, at the three AR stages (Fig. 11c). Therefore, the large increase in the
497 surface accumulated precipitation by marine INPs before AR landfall (36%) is mainly because of
498 the increase in deep cloud precipitation. The larger occurrence of deep clouds at this stage is
499 consistent with a larger increase in TWP and reduction in moisture. Although the relative

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500 increases in deep cloud occurrences and precipitation by marine INPs are very large at the post-
501 AR stage, their occurrences are so small that their contribution to the total precipitation is
502 negligible.

503 How do marine INPs reduce shallow warm clouds but invigorate mixed-phase and deep
504 clouds and why is this effect larger at the first and third stages? Marine INPs greatly enhance ice
505 and snow number concentrations and mass mixing ratios through immersion freezing, which
506 converts drops to ice or snow particles (Figs. 12a and 13a). The mean number concentrations
507 and mass mixing ratios of ice particles (ice +snow) in mixed-phase and deep cloud regimes are
508 several orders of magnitude higher in DM15+MC18 than in DM15. As detailed in Fan et al.
509 (2017a) which studied the same type of mixed-phase clouds in the same region, more ice/snow
510 particles forming from the immersion freezing enhance the Wegener–Bergeron–Findeisen
511 (WBF) and riming processes (Table 2), converting supercooled drops to ice or snow and leading
512 to more ice/snow but fewer cloud droplets and raindrops (Figs. 12b, c and 13b, c). The
513 reductions of cloud droplet and raindrop number concentrations and mass mixing ratios from
514 DM15 to DM15+MC18 are larger before AR landfall and during post-AR relative to the stage
515 after AR landfall, corresponding to a larger shift to the mixed-phase and deep clouds. Thus, the
516 larger increases in deposition/WBF and riming rates are seen (Table 2).

517 As discussed earlier, the largest ice nucleation rates at the post-AR stage explain the
518 largest marine INP effects among the three stages. The postfrontal clouds have the lower cloud
519 top heights (warmer than -25 ºC, i.e., shallower clouds) compared with the clouds at the first two
520 stages, thus the dust INP nucleation rates are smaller (negligible) as shown in Fig. 10c but the
521 deep marine aerosol layer and its action as INPs adds significantly to ice nucleation. In addition,
522 with the ~ 6 ºC colder temperatures below 8-km altitudes during the post-AR stage compared to

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the other two stages, ice nucleation from marine aerosols becomes more efficient (Fig. 10d). The mostly significantly invigorated postfrontal cloud cells by the marine INP effect (i.e., the increase in both LWC and IWC and a large increase in cloud fraction) might also be related to small scale thermodynamic changes through the feedback of microphysical changes over the first two AR stages.

As for why increases of deep cloud occurrence and precipitation are less significant after AR landfall compared to before AR landfall, first, we see the ice nucleation rates from dust INPs is larger after AR landfall (the largest among the three stages; Fig. 10c), because of the increased dust loading due to stronger transport (Fig. 10a). Stronger dust INP effects would limit the marine INP effects since they compete for liquid drops. Second, the moisture increase after AR landfall occurs in the lower atmosphere while the middle- and upper-level atmosphere are much drier than before AR landfall (Fig. 6d), which favors more warm clouds and rain but is less favorable to ice cloud development as indicated by the smallest ratio of snow precipitation (Fig. 7b). For more warm clouds/rain-dominated situations, the enhancement of ice formation would have less influence. Furthermore, in the drier condition aloft, more ice formation means less efficient growth, thus the impacts on IWC/IWP and precipitation would be smaller. Cloud dynamics (vertical velocity) is not changed much by the marine INP effect at all three stages, indicating that the feedback from the increased latent heating resulting from enhanced deposition and riming does not play an important role here, likely because this is not a convective environment.

4 Conclusion and discussion

We have explored the effects of INPs from sea spray aerosols transported from the Pacific Ocean on wintertime mixed-phase stratiform cloud properties and precipitation

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546 associated with atmospheric river (AR) events. This is done by carrying out simulations at a
547 cloud-resolving scale (1 km) using WRF-Chem coupled with the spectral-bin microphysics
548 (SBM) scheme for an AR event observed during the 2015 Atmospheric Radiation Measurement
549 Cloud Aerosol Precipitation Experiment (ACAPEX). We have implemented the ice nucleation
550 parameterization for marine aerosols (McCluskey et al. 2018a) into SBM to account for the
551 marine INP effect. By comparing with available airborne and ground-based observations, we
552 show that considering the marine INP effect in the model improves the simulation of cloud phase
553 states (i.e., increased glaciation ratio) and precipitation.

554 Through enhancing ice and snow formation, marine INPs greatly enhance WBF and
555 riming processes, which convert liquid clouds to mixed-phase and ice clouds. There is a notable
556 shift in cloud occurrences with reduced shallow warm clouds (44%, 26%, and 7% for before and
557 after AR landfall and the post-AR stages, respectively) and increased mixed-phase (10%, 7%,
558 and 38%) and/or deep cloud regimes (~ 22%, 8%, and 230%) because of the marine INP effect.
559 As a result, large increases in the ice water path (5 times or more), total condensate water path
560 (29% or more), and the ratio of snow precipitation (40 times or more) are seen. There is an
561 enhanced conversion of water from the vapor phase to the condensate phase so the water vapor is
562 generally reduced with the marine INP effect considered.

563 The significance of the above-described marine INP effects varies with the AR stages,
564 with a larger effect before AR landfall and post-AR than after AR landfall that has the dominant
565 precipitation. Note that the marine INP effects on cloud properties and snow precipitation are
566 still notable even at the stage after AR landfall. Although the total precipitation increases only by
567 4%, the drastic increase of snow precipitation and reduced rain precipitation at the surface have
568 an important implication for the regional water resources and flood risks since more snow

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569 increases freshwater resources while less rain reduces flash flood risks. In addition, at this stage,
570 the marine INPs produce a notable spillover effect with a precipitation decrease (up to 30%) over
571 the windward slope of the mountains but precipitation (snow) over the lee side is doubled,
572 because more ice/snow formed over the windward side falls slower than rain and more easily
573 transported to the lee side.

574 Several reasons can be responsible for the smaller marine INP effects on cloud properties
575 (particularly reduction of shallow warm clouds and increased mixed-phase and deep clouds) and
576 snow precipitation after AR landfall compared with before AR landfall. First, the dust INP
577 effects are larger at this stage, which would limit the marine INP effect since they compete for
578 liquid drops. Second, after AR landfall, the moisture is heavily concentrated at the lower
579 atmosphere while the middle- and upper-level atmosphere are much drier than before AR
580 landfall. Therefore, the environment is more warm cloud and rain dominated, limiting the effects
581 of enhanced ice formation. Furthermore, in the drier condition, more ice formation means less
582 efficient growth, thus the impacts on IWC/IWP and precipitation would be smaller.

583 The post AR stage has the largest response of the cloud regime shift and snow
584 precipitation among the three stages, because of the largest ice nucleation rates from the marine
585 aerosols. The larger ice nucleation rates compared with the other two stages are probably
586 because the lower cloud top heights (warmer than -25 ºC, i.e., shallower clouds) limit dust INP
587 nucleation, and with ~ 6 ºC colder temperatures below 8-km altitudes than the other two stages,
588 ice nucleation from the deep marine aerosol layer is more efficient.

589 This study suggests that the inclusion of marine INPs enhances orographic precipitation
590 mainly through more efficient growth (deposition and riming) of a larger number of ice particles
591 than liquid droplets, which is consistent with literature studies (Mühlbauer and Lohmann, 2009;

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592 Fan et al., 2014, 2017; Xiao et al., 2015). The spillover effect by the increase of CCN has been
593 presented in several previous studies (e.g., Mühlbauer and Lohmann, 2008, 2009; Saleeby et al.,
594 2011, 2013; Carrio and Cotton, 2014; Letcher and Cotton, 2014). To our knowledge, this study is
595 the first to show the spillover effect associated with the INP effect. The prominent spillover
596 effect by the marine INP is different from Fan et al. (2014, 2017) that did not find such an effect
597 by dust INPs. There are a couple of factors that might be responsible for the difference. First,
598 marine INPs are mainly brought by ARs so the windward side gets INP first while dust INPs are
599 not associated with AR so there is no temporal sequence to have dust between the windward and
600 Lee sides. Second, the AR event is different with a different wind direction and speed, which
601 makes the transport of ice/snow to the lee side easier.
602 The marine INP effect revealed in this study is clearly emphasized due to the very low
603 dust INP concentrations for this particular situation and the deep marine aerosol layer during the
604 AR which allows notable ice nucleation at temperatures higher than -15 ºC. With high dust INPs,
605 the effects of marine INPs might not be as significant since they compete for supercooled liquid
606 drops. Although this is a single case study, the AR event and its evolution are representative.
607 Thus, the study suggests the importance of accounting for marine aerosols as INPs, in addition to
608 long-range transported mineral dusts, to simulate winter clouds and precipitation in the western
609 United States in regional and global climate models. We employ an empirical parameterization
610 for marine INPs developed from the data collected over the northern Atlantic Ocean and use sea
611 salt aerosols as a surrogate of marine organics, which might produce some uncertainties.
612 Nevertheless, the marine INP parameterization appears representative for this region based on
613 Levin et al. (2019). More observational data are needed in the western U.S. for (a) developing ice
614 nucleation parameterizations for potentially variable marine organics and (b) understanding

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615 marine organics emission and chemical mechanisms and accurately simulating marine organics
616 in the model. As discussed earlier, the conversion of mass to number concentrations over each
617 aerosol bin might introduce some uncertainty to this study, which calls for model developments
618 of predicting the number concentration of each aerosol component.
619
620 **Data availability.**
621 The observational data can be accessed from the ARM data archive,
622 https://www.arm.gov/research/campaigns/amf2015apex. The model simulation data will be
623 available through the NERSC data repository after the paper is accepted.
624
625 **Supplement.**
626 The supplement related to this article is available online.
627
628 **Author contributions.**
629 JF designed the study and model experiments. YL, JF, and PL performed numerical simulations
630 and analyses. JF and YL wrote the paper and other authors commented on it. LRL, PJD, LG, JF,
631 JT, YL, and JHJ contributed by either processing data including model input and observational
632 data or participating in the discussion of results.
633
634 **Competing interests.**
635 The authors declare that they have no conflict of interest.

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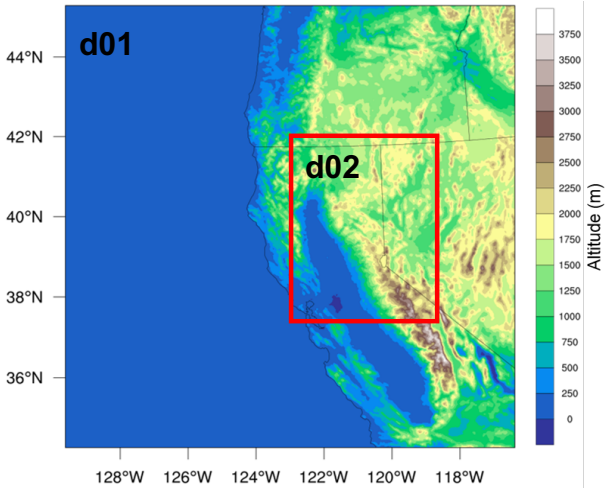
871 **Figures**



872

873 **Figure 1.** Two nested simulation domains: d01 and d02 centering over California. The color
874 shading denotes the terrain elevation.

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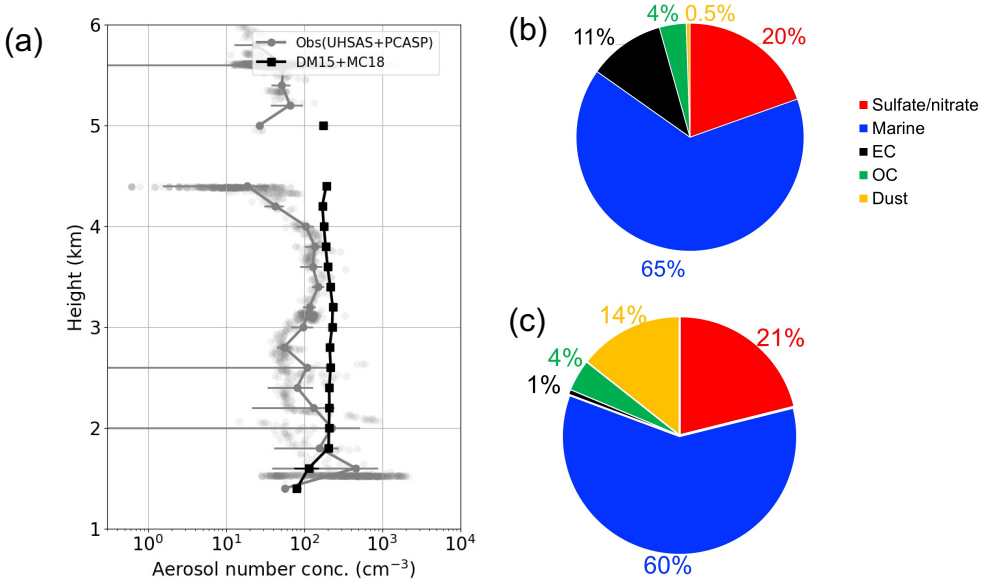

Figure 2. (a) Vertical distributions of aerosol number concentrations from aircraft observations (Obs, grey) and DM15+MC18 (black) for particles with a dry diameter over a range of 0.067~3 μm, (b) mean fractional number contributions of aerosol classifications based on measurements of single-particle mass spectra of aerosols and cloud particle residuals reported in Levin et al. (2019), and (c) mean fractional mass contributions of aerosols in DM15+MC18 (number concentration for each aerosol component is not predicted by WRF-Chem). The aerosol number concentration from aircraft observations in (a) consists of both measurements from UHSAS and PCASP. The modeled data in (a) and (c) are sampled along the aircraft route on 7 February 2015.

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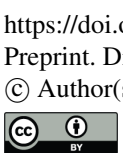

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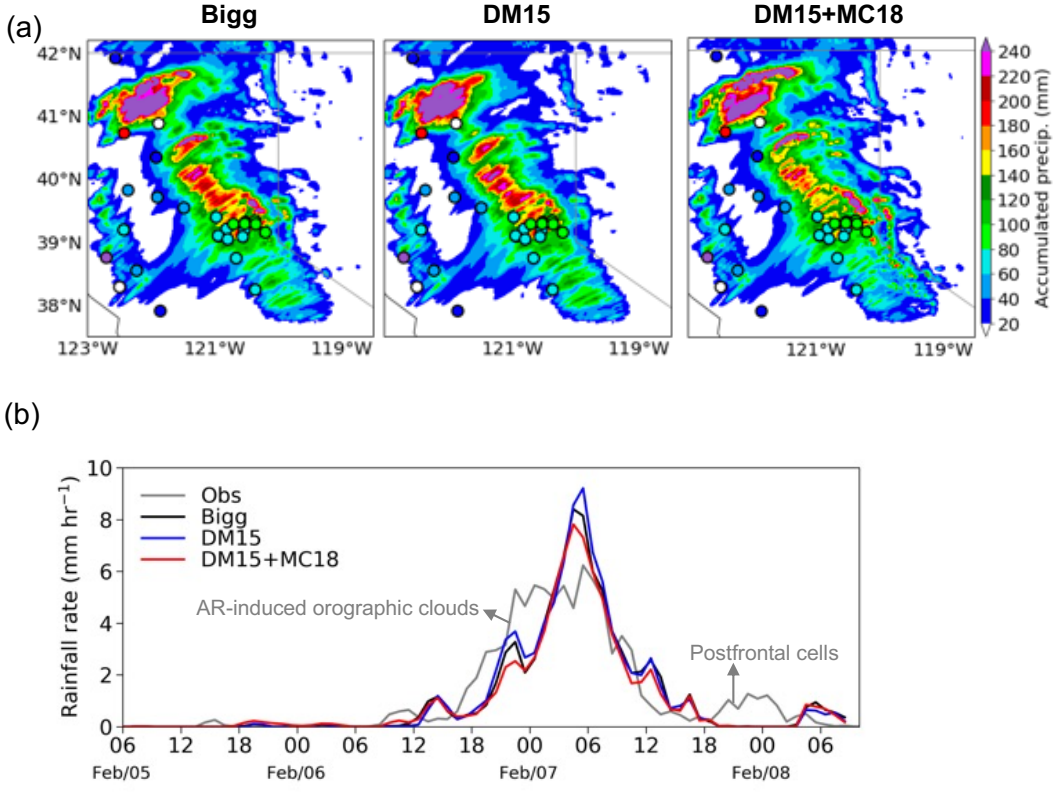

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888 **Figure 3.** (a) Spatial distributions of accumulated precipitation during the AR event (06:00 UTC
889 5– 09:00 UTC 8 February). The color shading is for simulations and the circles denote the rain
890 gauge measurements provided by NOAA's Physical Sciences Laboratory. (b) Time series of
891 precipitation rates during the entire AR event for rain gauge observations (grey), and the
892 simulations of Bigg (black), DM15 (blue), and DM15+MC18 (red). The precipitation rates are
893 averaged over all the rain gauge sites shown in (a) for both observations and simulations. The
894 observed AR-induced orographic clouds and postfrontal cells are marked in (b).

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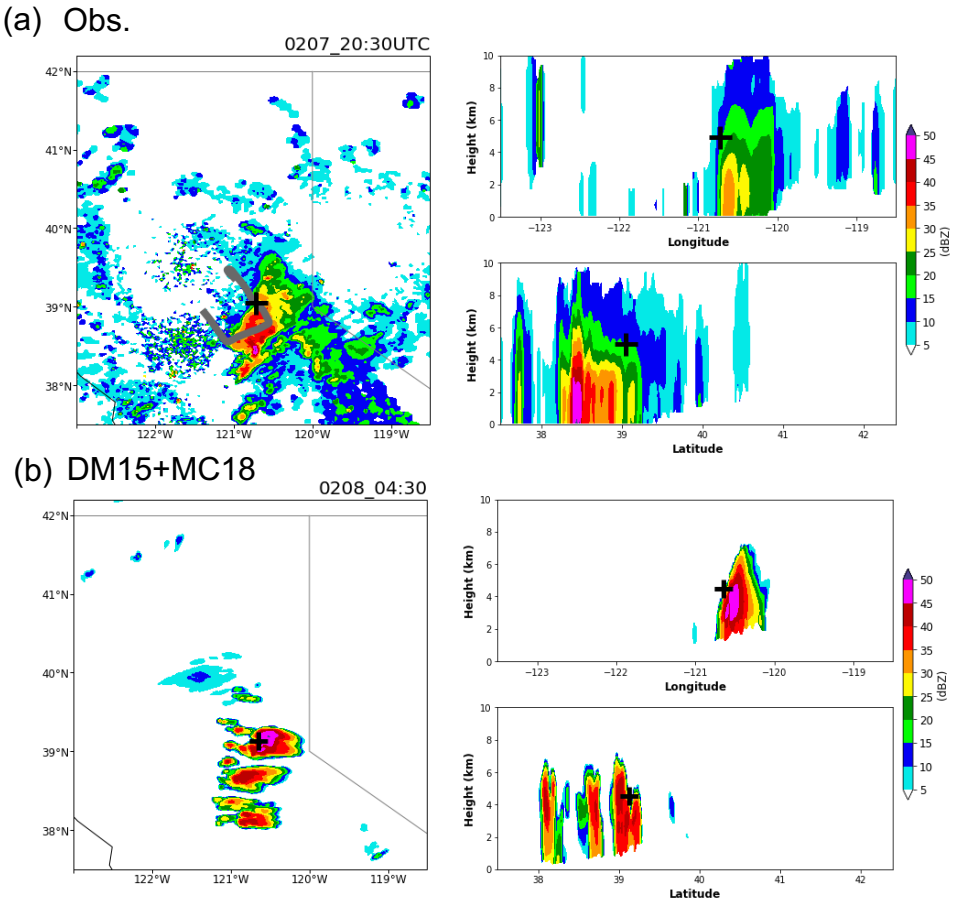

Figure 4. (a) Composite reflectivity of NEXRAD for the postfrontal clouds that the G-1 aircraft sampled, (b) composite reflectivity from the simulation of DM15+MC18 for the postfrontal clouds. The observation and simulation are compared at the peak reflectivity time which is 20:30 UTC 7 February for the observed clouds and 04:30 UTC 8 February for the simulated clouds. The black crosses in the left two panels denote the positions where the longitude-height and latitude-height cross-sections in the right panels are plotted. The grey line in the left panel of (a) shows the flight track of the G-1 aircraft.

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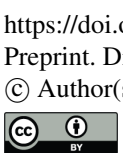

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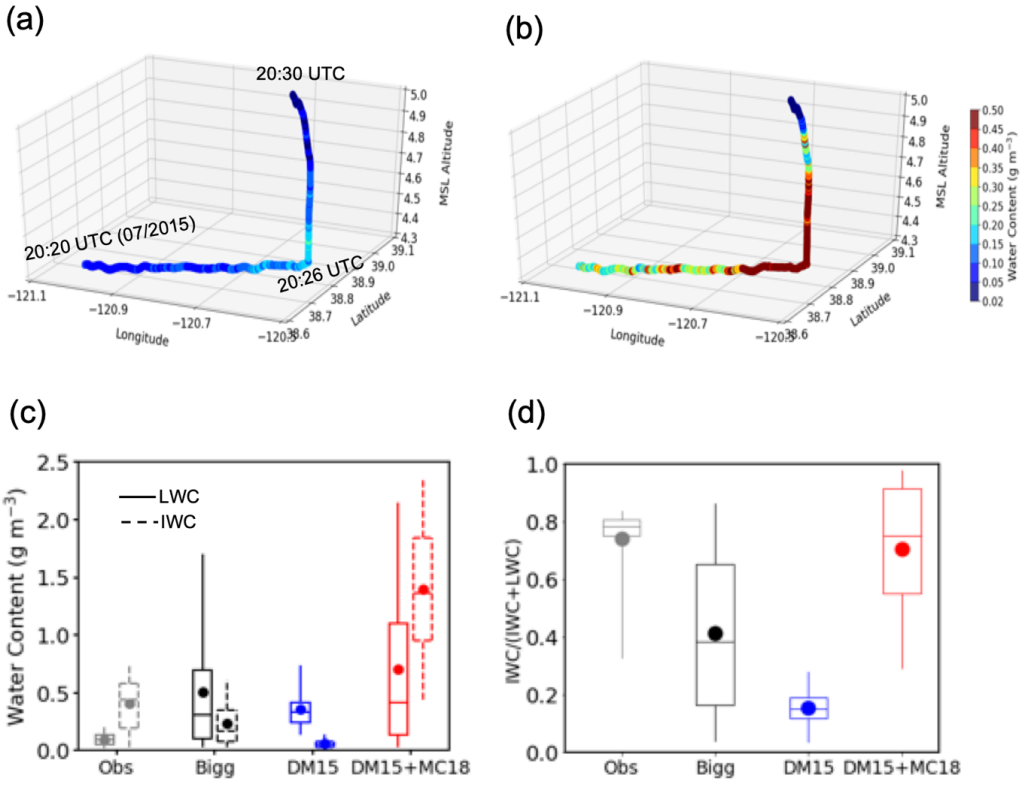

Figure 5. Comparisons of the simulations with aircraft observations. The observed (a) LWC and (b) IWC along the flight track during 20:20 – 20:30 on 7 February when the aircraft flew through the mixed-phase regime of the postfrontal clouds. (c) LWC (solid) and IWC (dashed) and (d) the glaciation ratios of IWC/(IWC+LWC) from the aircraft measurements (Obs, grey) and simulations of Bigg (black), DM15 (blue), and DM15+MC18 (red). The boxes show the 25th, median (horizontal lines in the box), and 75th percentiles of the data. The upper and lower whiskers show the 95th and 5th percentiles, respectively. The mean values are denoted by circles.

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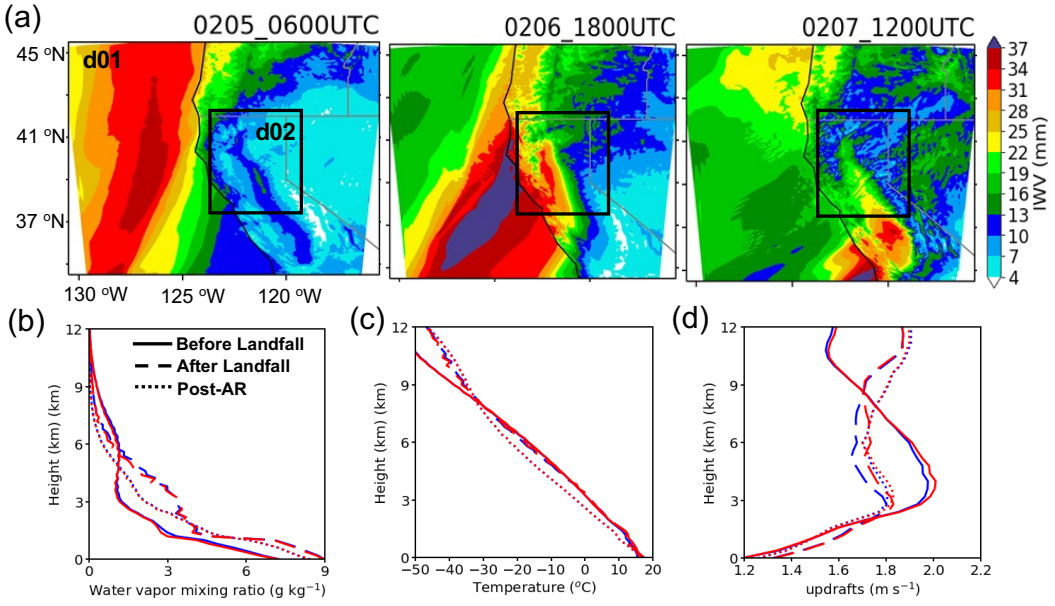

Figure 6. (a) Evolution of integrated water vapor (IWV) at 06:00 UTC 5 February (before AR landfall), 18:00 UTC 6 February (after AR landfall), and 12:00 UTC 7 February (post-AR). The black box (i.e., d02) in (a) is the domain of this study with the 5 lateral boundary grids excluded for analysis at each side. (b-d) show the mean vertical profiles of (b) water vapor mixing ratio, (c) temperature, and (d) updraft velocity at the three AR stages, i.e., before (solid lines) and after (dashed lines) AR landfall and post-AR stages (dotted lines), for the simulations of DM15 (blue) and DM15+MC18 (red). The water vapor mixing ratio and temperature are averaged for cloud-free grids, and updraft velocity is averaged over the grids with a vertical velocity greater than 1 m s^{-1}.

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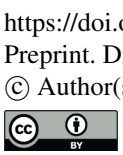

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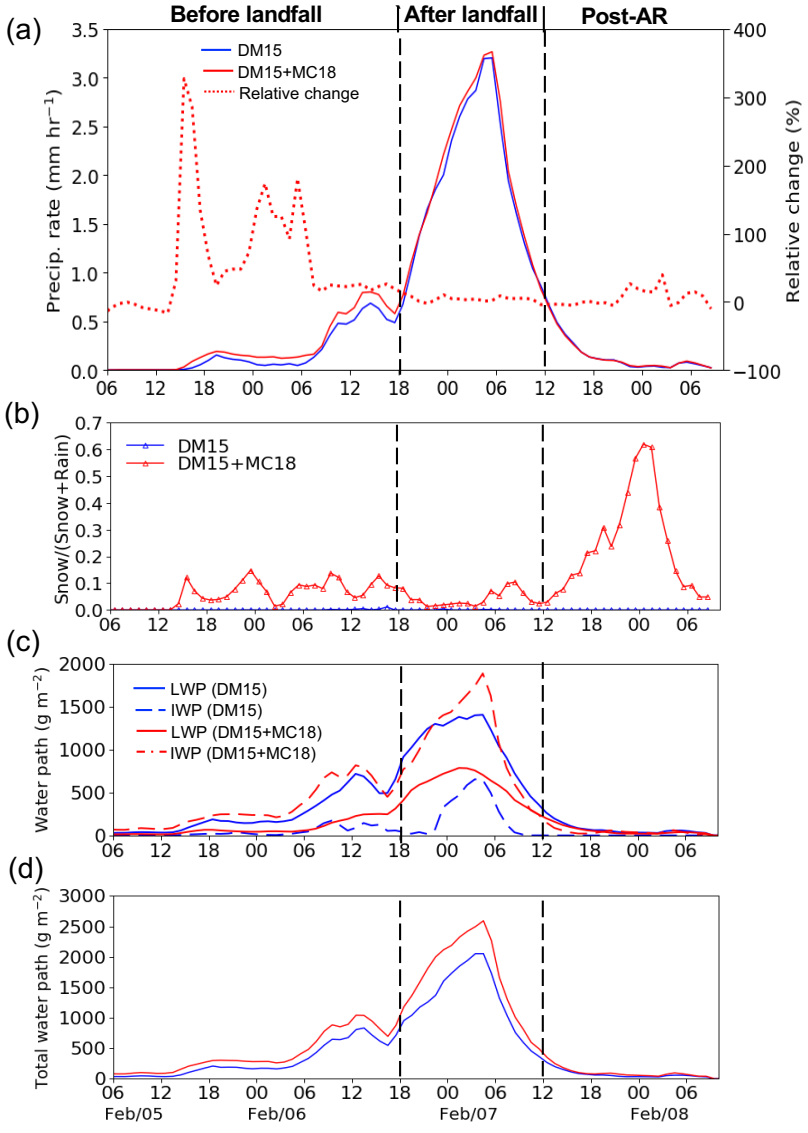

Figure 7. Time series of (a) precipitation rate (solid lines, left y-axis), (b) ratio of snow precipitation (i.e., snow/(snow+rain) in mass mixing ratio) at the lowest model level, (c) LWP (solid) and IWP (dashed) for DM15 (blue) and DM15+MC18 (red), and (d) total condensate water path (TWP). The relative changes in precipitation rate from DM15 to DM15+MC18 are shown in the red dotted line in (a) with values shown on the right y-axis. The vertical dashed lines divide the three AR stages.

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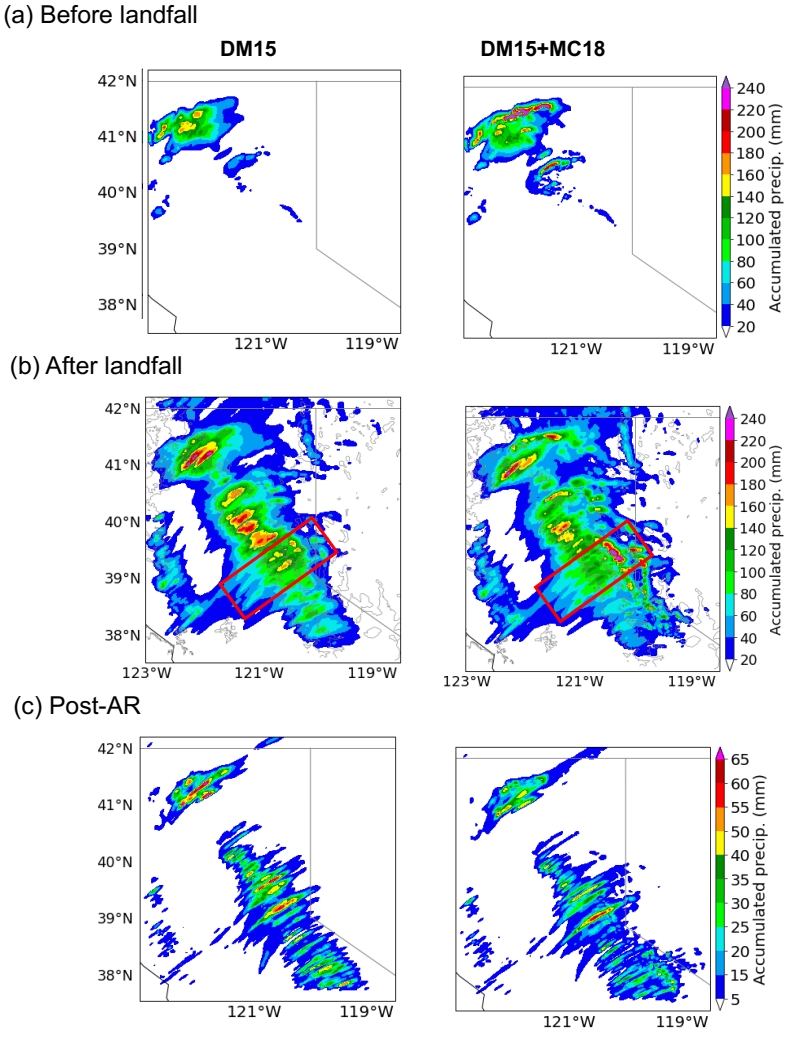

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931 **Figure 8.** Spatial distribution of accumulated precipitation during the stages of (a) before AR
932 landfall, (b) after AR landfall, and (c) post-AR in DM15 (left) and DM15+MC18 (right). The
933 parallelograms marked in (b) denotes the area for the east-west cross-section analysis shown in
934 Figure 9.

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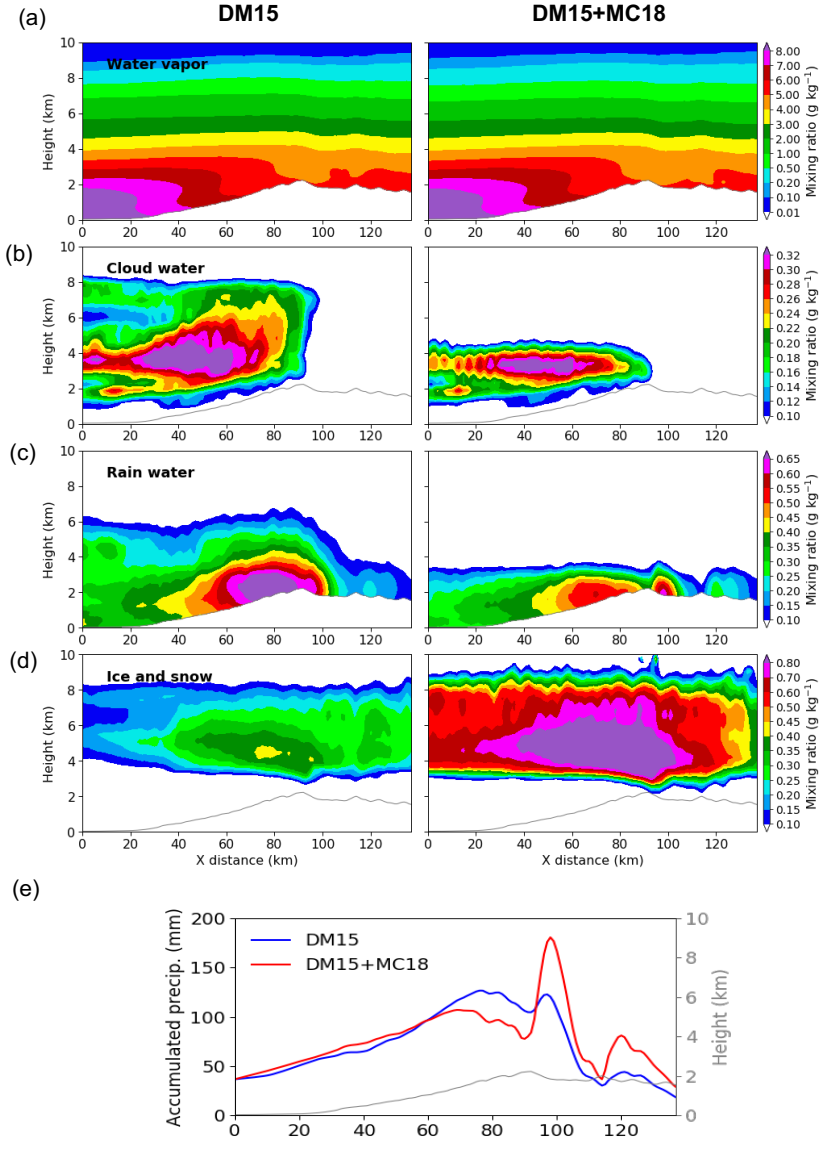

Figure 9. Mean mixing ratios of (a) water vapor, (b) cloud water, (c) rainwater, (d) ice + snow, and (e) surface precipitation at the stage after AR landfall in DM15 and DM15+MC18. The vertical cross-sections are averaged over the red boxes marked in Fig. 8b and the entire stage.

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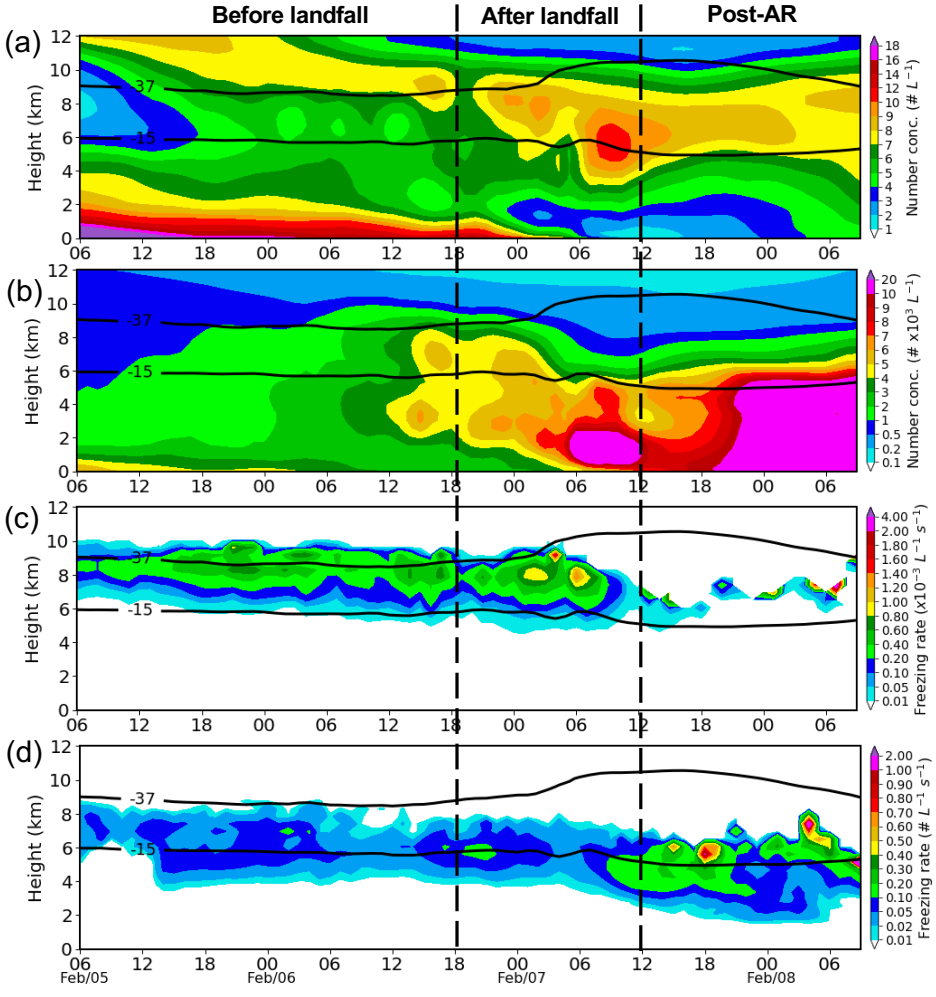

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945 **Figure 10.** Time-height cross-sections of (a) dust particle (>0.5 μm) number concentration, (b)
946 marine aerosol number concentration, (c) the freezing rate in DM15, and (d) the freezing rate in
947 DM15+MC18. The number concentrations in (a) and (b) are derived from their corresponding
948 mass mixing ratios under the clear-sky condition only. The freezing rates in (b) and (d) are the
949 ice nucleation rates via immersion freezing at T > - 37 ºC and the drop homogenous freezing
950 rates at T < - 37 ºC, and the values are for cloudy-points only. The black contour lines in each
951 panel mark the temperature levels of -15 and -37 ºC, representing the efficient immersion
952 freezing temperature in DM15+MC18 and the homogeneous freezing temperature in the model,
953 respectively.

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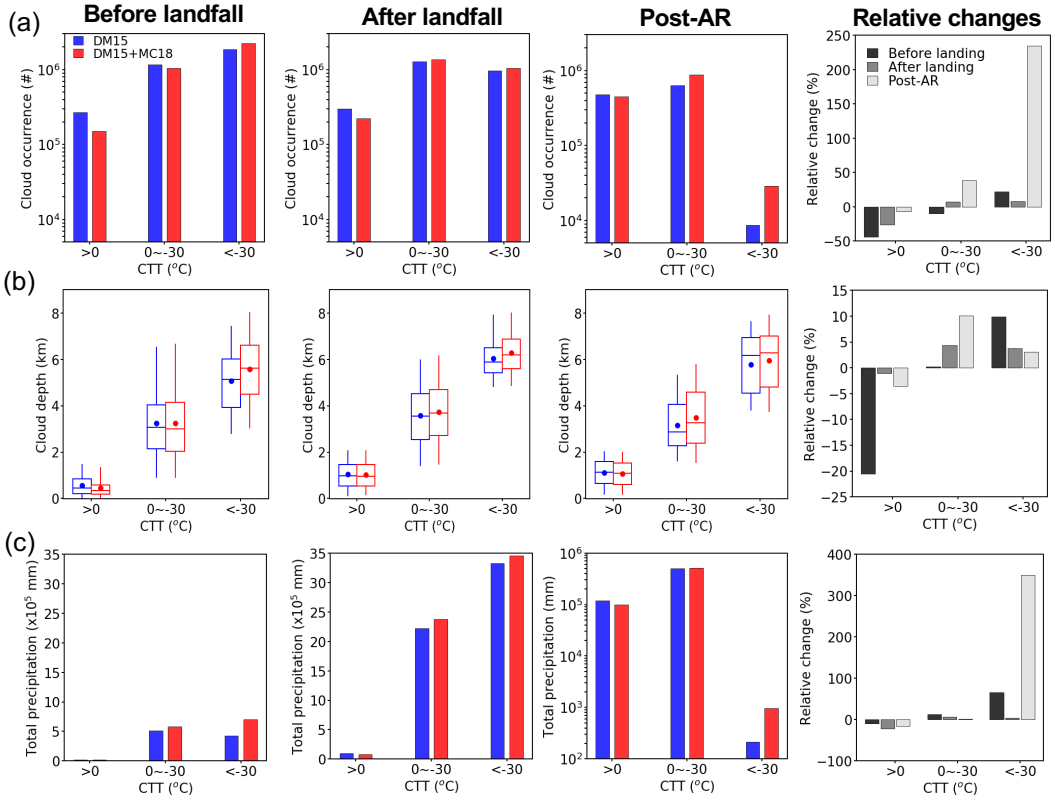

Figure 11. (a) cloud occurrences, (b) cloud depth, and (c) total precipitation for three cloud regimes in DM15 (blue) and DM15+MC18 (red) at three AR stages from left to right: before AR landfall, after AR landfall, post-AR. The last column shows the relative changes caused by the marine INP effect, which are calculated as [(DM15+MC18) – DM15]/DM15*100%. Note that the total precipitation at the post-AR stage uses a log scale for the y-axis. The box-whisker plots follow the description in Figure 5c.

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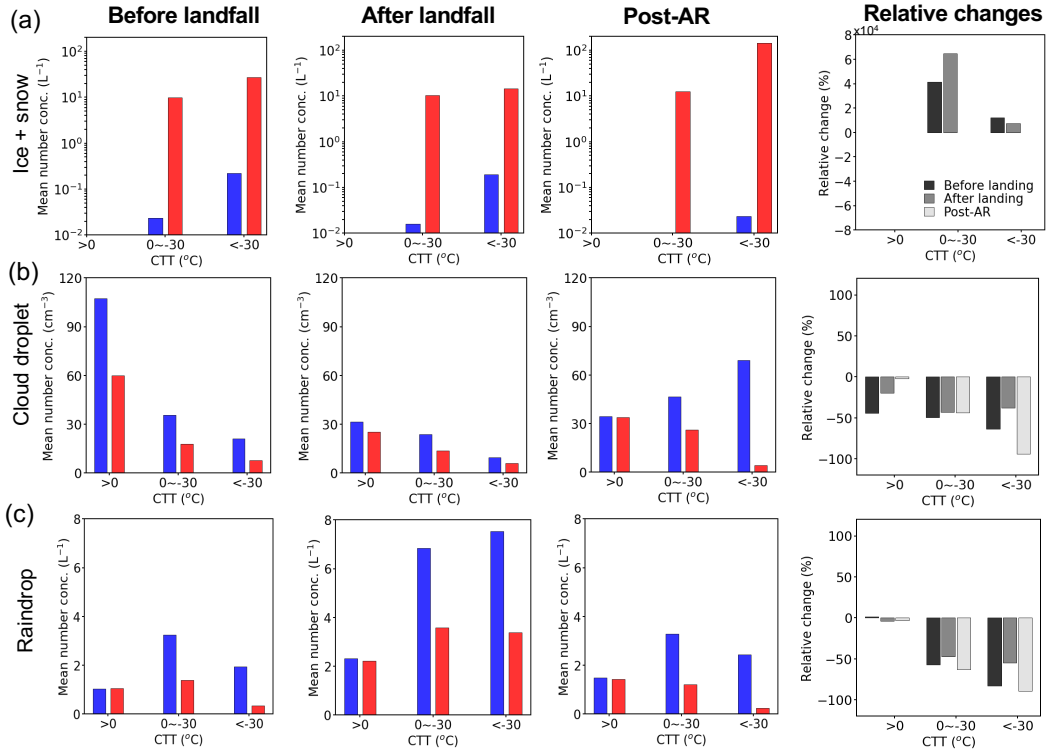

Figure 12. Hydrometeor number concentrations and their relative changes in three cloud regimes in DM15 (blue) and DM15+MC18 (red) at the three AR stages for (a) ice particles (sum of ice and snow), (b) cloud droplets, and (c) raindrops. The last column shows the relative changes caused by the marine INP effect, which are calculated as [(DM15+MC18) – DM15]/DM15*100%. Since ice particles are very limited at the post-AR stage in DM15, the percentage changes of ice particles from DM15 to DM15+MC18 are huge numbers that are omitted from the plots.

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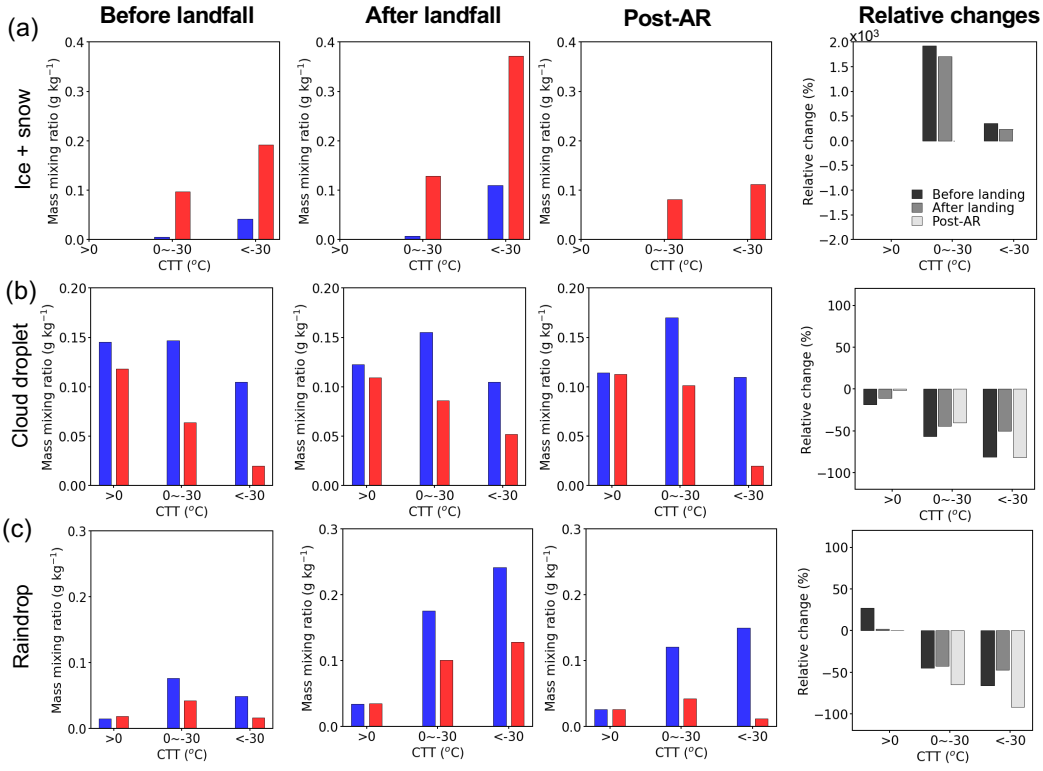

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971 **Figure 13.** Same as Figure 12, except for the mass mixing ratios of (a) ice particles (sum of ice
972 and snow), (b) cloud droplets, and (c) raindrops.
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983 **Table 1.** The changes in total precipitation, total condensate water path (TWP), liquid water path
984 (LWP), and ice water path (IWP), and cloud fractions (CF), net cloud radiative forcing (CRF) at
985 TOA from DM15 to DM15+MC18 (i.e., the marine INP effect), as well as the glaciation ratio,
986 i.e., IWC/(LWC+IWC), and the ratios of snow precipitation, i.e., snow/(rain+snow) in mass
987 mixing ratio at the lowest model level from DM15 to DM15+MC18, at the three AR stages. The
988 percentage changes are calculated following ((DM15+MC18)- DM15)/DM15*100.

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AR stages		**Before landfall**	**After landfall**	**Post-AR**
Total precipitation		36%	4%	-1%
TWP		45%	29%	35%
LWP		-66%	-46%	-26%
IWP		8 times	5 times	440 times
CF		5%	4%	20%
Net CRF at TOA		15%	13%	10%
IWC/(LWC+IWC)	DM15	0.14	0.16	0.001
	DM15+MC18	0.74	0.59	0.36
Snow/(Rain+Snow)	DM15	0.002	0.001	<0.001
	DM15+MC18	0.085	0.042	0.131

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992 **Table 2.** The domain-mean mass rates of deposition and riming in the mixed-phase and deep
993 cloud regimes in DM15 and DM15+MC18 at the three AR stages.

AR stages		**Before landfall**		**After landfall**		**Post-AR**	
		Mixed-phase clouds	Deep clouds	Mixed-phase clouds	Deep clouds	Mixed-phase clouds	Deep clouds
Deposition ($mg\ kg^{-1}\ h^{-1}$)	DM15	44	171	81	388	7	8
	DM15+MC18	846	780	1128	1397	781	1013
Riming ($mg\ kg^{-1}\ h^{-1}$)	DM15	27	89	57	297	25	34
	DM15+MC18	377	228	575	858	505	361

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