Microphysical controls on the isotopic composition of wintertime orographic precipitation

M. Moore,¹ P. N. Blossey² and A. Muhlbauer²,³ Z. Kuang,¹,⁴

Key Points

- When orographic precipitation is sensitive to CDNC changes, its isotopic composition can be as well.
- Temperature and mountain height control precipitation isotopic composition more strongly than CDNC.
- Microphysical changes with CDNC, temperature and mountain height help fix isotopic composition.

Mary Moore, Department of Earth and Planetary Sciences, Harvard University, Cambridge, MA 02138, USA. (moore3@fas.harvard.edu)

¹Department of Earth and Planetary
Abstract. When precipitation from mixed-phase wintertime orographic clouds is sensitive to changes in cloud droplet number concentration (CDNC) within those clouds, the isotopic composition of the precipitation can be affected as well. The changes in isotopic composition due to CDNC changes is measurable, but small in comparison to changes due to temperature or mountain height. Microphysical changes, driven by CDNC changes, upstream temperature or mountain height, help determine the isotopic composition. In particular, the increasingly depleted precipitation — with increasing CDNC, increasing mountain height or decreasing temperature — is partly induced by the shift of precipitation production in these mixed-phase clouds from the accretion of cloud liquid by snow (riming) to vapor deposition onto precip-
inating ice, which has a more depleted isotopic signal. These sensitivities have
been explored in idealized simulations using the Weather Research and Fore-
casting Model (WRF). Stable water isotopologues (HDO and $\text{H}_2^{18}\text{O}$) are added
to the Thompson microphysics scheme by accounting for fractionating (e.g.
vapor deposition) and non-fractionating (e.g. melting/freezing) isotopic ex-
changes between water vapor, hydrometeors and precipitation. Using an ide-
alized setup, sensitivity studies are completed with varying mountain heights
and temperature profiles as well as varying CDNC. The largest sensitivities
to CDNC occur with the smallest mountain height and warm initial tem-
perature profile, while the CDNC dependence for other combinations of moun-
tain height and/or temperature is weaker and quite variable.
1. Introduction

The impact of aerosols on cloud microphysics and precipitation has been addressed by both observational and modeling studies. Certain types of hydrosopic atmospheric aerosols, such as sulfate, can act as cloud condensation nuclei (CCN), which in high concentrations can alter the microphysics of clouds by increasing the cloud droplet number concentration (CDNC) and (for same liquid water content), decreasing the average size of cloud droplets [e.g. Twomey, 1974; Twomey et al., 1984]. The resultant smaller droplet diameter reduces the efficiency of collision and coalescence processes, which can potentially suppress or delay precipitation [Albrecht, 1989; Ramanathan et al., 2001]. Despite this clear physical mechanism, there is no consensus of this cause and effect, because the overall impact of aerosols on precipitation will vary with environmental conditions and cloud type being considered [Khain, 2009]. For example, in some meteorological regimes (weak updrafts in some pyro-convective clouds), CDNC can be decoupled from CCN concentrations [Reutter et al., 2009]. In addition, precipitation from warm (liquid-only) orographic clouds can be insensitive to changes in CDNC [Miltonberger et al., 2015] for a range of dynamical/microphysical conditions. Despite these qualifications, the potential for a decrease in the amount or shift in location of precipitation has large implications for society and climate. In the case of orographic precipitation, the location and the amount of precipitation is of utmost importance, as rainfall and melting of the mountain snowpack feed into river basins and provide water to many populations. Therefore, the amount of precipitation and where it falls will determine how much water is available and into which basins it will flow.
There have been extensive regional studies regarding the formation and behavior of orographic precipitation [e.g. Hobbs, 1975; Smith et al., 2005; Smith and Evans, 2007; Zubler et al., 2011]. The total precipitation and its spatial distribution have been found to be dependent upon several variables, including the orientation and geometry of the terrain, atmospheric stability, orographic flow dynamics and microphysics [e.g. Colle, 2004; Galewsky, 2008, 2009; Muhlbauer and Lohmann, 2008]. For cloud microphysics, the motivating factor has often been to study the impact of anthropogenic aerosols on the growth pathways of orographic precipitation [e.g. Lynn et al., 2007; Rosenfeld et al., 2008; Muhlbauer et al., 2010; Saleeby et al., 2011; Xiao et al., 2014]. In mixed-phase orographic clouds, the growth and fallout of snow and graupel is enhanced by the "seeder-feeder" mechanism [Reinking et al., 2000], wherein ice crystals grow by vapor deposition in an ice cloud aloft before sedimenting to lower levels in the cloud where the ice continues to grow by collecting cloud droplets (riming). This enhanced low-level riming increases the fallspeed of snow and also the overall precipitation efficiency [Mitchell et al., 1990; Borys et al., 2003], thereby augmenting precipitation on the windward side of the mountain at the expense of the transport of hydrometeors (and the resulting precipitation) to the leeward slope in what is known as "spillover".

The addition of aerosols to air masses that form mixed-phase clouds can alter the pathway of precipitation growth. If these additional aerosols lead to higher CDNC and smaller cloud droplets, riming of cloud ice and snow can decrease. Pruppacher and Klett [1997] and Wang and Ji [2000] found that as droplet diameter fell below approximately 10 μm, the riming efficiency decreased. With little to no riming occurring, cloud ice and snow will grow mostly by vapor deposition, which will have smaller fallspeeds and
longer fallout timescales than heavily rimed snow [Locatelli and Hobbs, 1974]. In the specific case of orographic precipitation, the longer timescale for snow growth may lead to spillover, or decreased amounts of snow, as many of the hydrometeors might evaporate in the subsiding air on the leeward side of the mountain [Muhlbauer et al., 2010; Lowenthal et al., 2011; Saleeb et al., 2011]. Previous studies have attempted to address suppressed precipitation and leeward transport of winter snowfall in pollution events, but results have been inconclusive and appear to depend strongly on the environmental conditions of the region being considered [Borys et al., 2000, 2003; Khain and Pokrovsky, 2004; Lynn et al., 2007; Muhlbauer and Lohmann, 2008; Saleeb et al., 2013]. Though it focuses on warm (liquid-only) orographic clouds and precipitation, the study of Miltenberger et al. [2015] suggests one explanation: the interactions of dynamical and microphysical processes can lead to regimes where the precipitation, and the precipitation efficiency itself, is insensitive to changes in CDNC. In particular, orographic precipitation is most sensitive to changes in CDNC when the precipitation formation time is comparable to the transit time of air parcels within the cloud.

In order to provide a new perspective on this topic, we have performed idealized numerical experiments that use information and insights gained by the incorporation of stable water isotopes. Stable water isotopes (H$_2^{16}$O, HDO, H$_2^{18}$O) have been used in precipitation analysis dating back to the initial work of Dansgaard [1952]. Over the years, the general precipitation trends that emerged, as they relate to the isotopic signature of the precipitation, were that in the tropics and sub-tropics, the ratio of HDO and H$_2^{18}$O to the standard H$_2^{16}$O in rainfall was related to the total amount of precipitation [e.g. Dansgaard, 1964; Rozanski et al., 1993; Kurita et al., 2009]. In the mid-latitudes, however, the
depletion has been found to be temperature-dependent [e.g. Dansgaard, 1964; Noone and Simmonds, 2002; Jouzel, 2003; Lee et al., 2007], such that the ratio correlates positively with temperature.

In mountainous regions, the correlation of the isotopic composition of precipitation with temperature is expected to lead to a correlation with altitude. Air cools as it rises along the upslope on the windward side of a mountain, and the progressive removal of precipitation produces a gradient in the isotopic composition with altitude. This leads to richer precipitation (i.e., with a higher ratio of the heavier isotopologues) forming at lower altitudes and more depleted precipitation (i.e., with smaller isotopic ratios) at higher altitudes, as well as on the downslope in the lee of the mountain [Smith et al., 2005]. This isotopic gradient was connected to the fractional removal of water by a mountain barrier and its drying ratio by Smith et al. [2005], who used it to constrain a linear model of orographic precipitation. This relationship between isotopic composition and altitude has been used to relate paleoclimate proxies for the isotopic composition of precipitation to past mountain elevation [Poage and Chamberlain, 2001; Rowley et al., 2001]. However, as demonstrated by Galewsky [2009] and Lechler and Galewsky [2013], in different dynamical regimes, the airflow over a mountain can complicate this relationship between the isotopic composition of precipitation and the altitude of a mountain barrier. For example, more stable conditions may lead to blocking — where air flows mainly around, rather than over a mountain barrier — and modify the isotopic composition of precipitation both over the mountain [Galewsky, 2009] and downwind [Lechler and Galewsky, 2013].

In addition to the influence that dynamics has on orographic precipitation and its isotopic content, microphysical changes may also modify the isotopic composition.
observations of snowfall in the Sierra Nevada, Demoz et al. [1991] found that the isotopic composition of newly fallen snow was influenced by the degree of riming, and that the temperatures of snow formation implied by the isotopic composition of the snowfall at the surface and by its crystal structure did not agree, presumably due to riming. The extent of the deviation between these temperatures was dependent upon where the rimed supercooled liquid was located within the cloud. More recently, Lowenthal et al. [2011] sampled the isotopic and chemical composition of both snowfall and cloud droplets at a mountaintop site in Colorado, and related the degree of riming of snowfall there to the chemical composition of the snow and concurrently sampled cloud droplets. They found that snow mass formed mainly through riming was more enriched and had an isotopic signature that was similar to the cloud droplets. This relationship was then employed to make predictions about the altitude at which snow was formed through vapor deposition.

The goal in the present study is to characterize the isotopic composition of mixed phase precipitation over an idealized two-dimensional mountain and to identify regimes in which precipitation and its isotopic composition are sensitive to changes in cloud droplet number concentration, used here as a proxy for changes in aerosol concentrations, and to changes in temperature and mountain height. We seek to understand the pathways of precipitation formation and how these impact the isotopic content of precipitation. By tracking the isotopic content associated with various precipitation growth processes, we hope to determine if each microphysical process has a distinct isotopic signature, how each process contributes to the overall isotopic signal in the precipitation and if one could potentially work backward and determine the growth pathway of fallen precipitation from the isotopic content alone.
2. Model and Experiments

2.1. Model Setup

To conduct the orographic precipitation experiments, we use the Weather and Research Forecast (WRF) model version 3.5.1 [Skamarock and Klemp, 2008] provided by the Mesoscale and Microscale Meteorology Division of the National Center for Atmospheric Research. The model is configured to perform simulations of idealized 2D flow over a hill. The domain consists of 300 grid points with 2km spacing in the horizontal x-direction, 105 vertical levels spaced 25-200m in the lower 5km and constant grid spacing above, and outputs every 20 minutes for each 12 hour simulation. The Thompson microphysics scheme [Thompson et al., 2008] within WRF was chosen for this study because of its good representation of wintertime precipitation in mountainous areas [Liu et al., 2011] and because it includes a detailed treatment of the riming of cloud droplets by snow (as in Saleeby and Cotton [2008]), which has proven important for capturing the effects of pollution on riming in mixed-phase orographic clouds.

As a way to study the impact of aerosols on the cloud microphysics, we alter the value of the CDNC. This setup assumes that for high aerosol loading, there are more CCN and thus a higher CDNC. To represent the scenario with few aerosols and thus fewer CCN, a lower CDNC value is specified.

2.2. Isotopic implementation

For the present study, the microphysics scheme has been extended to include stable water isotope chemistry by incorporating phase changes of heavy water isotopes (HDO and H$_2^{18}$O). The isotopic composition of vapor and each hydrometeor species (cloud liquid, cloud ice, rain, snow and graupel) is tracked, and the exchanges of the heavy isotopologues
of water are accounted for during each microphysical process represented in the Thompson scheme. Fractionation — the unequal exchange of heavy and lighter isotopologues of water — is accounted for in processes that involve the deposition of vapor onto liquid or ice hydrometeors and those involving the evaporation of liquid phase hydrometeors (rain or cloud liquid). Other processes (e.g., freezing, melting, riming, sublimation of ice phase hydrometeors) are assumed to be non-fractionating. Beyond the assumption that the isotopic composition of each hydrometeor category in a given grid cell is uniform, there are few approximations involved in representing the isotopic exchanges, and the implementation follows Blossey et al. [2010]. The model uses time steps on the order of a few seconds, so that only cloud liquid and vapor are assumed to equilibrate within a single time step. Other processes are integrated explicitly by the model. A more detailed description of the water isotope physics is given in Appendix A.

For the isotopic analysis, our results are presented in terms of H$_2^{18}$O and specifically in delta-notation such that $\delta^{18}O = 1000 \left( \frac{R}{R_o} - 1 \right)$, where R is the isotopic ratio of H$_2^{18}$O in a specified water species and R$_o$ is the isotopic ratio of the standard. While HDO is also included in the microphysics scheme, the additional information that can be gained by considering both HDO and H$_2^{18}$O will be left to future work.

### 2.3. Experimental Setup

In order to address the robustness of an isotopic signal, several experiments are conducted that alter the initial temperature profile, mountain height (800m, 1500m and 3000m), and the CDNC (25, 100, 200, 400, and 800cm$^{-3}$). Two initial temperature profiles are used here and are referenced as warm or cold based on the surface temperature of the upstream sounding. Profiles with a surface temperature of 0°C are the cold cases,
while those with 7°C are referred to as warm. The different setups are referred to by abbreviations (e.g., W800m) which indicate the temperature (W=warm or C=cold) and mountain height. The setup and initial conditions, including the temperature profiles, are similar to those in Muhlbauer et al. [2010], with a mountain half-width of 20km and a horizontal wind profile that is a constant 15m s\(^{-1}\) below 10 km and linearly increases to 40m s\(^{-1}\) at the top model layer (30 km).

To generate the initial vapor conditions for H\(^2\)\(^{18}\)O and HDO, a Rayleigh distillation profile is generated assuming equilibrium with ocean water at 20°C, which is chosen to represent the average temperature of the ocean surface where the initial isotopic signature of the air mass will be set. The model’s initial conditions for the isotopic content of water vapor are interpolated from this Rayleigh profile based on the water vapor mass mixing ratio. Note that, as the cold sounding is drier than the warm sounding, it is also more depleted, and its \(\delta^{18}O\) of vapor at the surface is 8‰ more depleted than that of the warm sounding. Neither liquid nor ice condensate exists initially, and therefore their isotopic compositions do not need to be initialized.

### 2.4. Model Validation

The model’s ability to simulate orographic clouds and precipitation is on par with previous studies. As expected, the results are very similar to the WRF simulations in Muhlbauer et al. [2010], given the experimental setups are nearly identical. There are some small differences in the simulated orographic clouds and accumulated precipitation, typically that our experiments produce more precipitation, but these deviations can be attributed to the longer simulation time and different microphysical scheme used here, as well as the larger range of CDNC values. The changes in the orographic liquid cloud are also similar.
to results seen by Xiao et al. [2014], who also used the same idealized WRF setup, but coupled with a detailed bin microphysics scheme and a warmer initial temperature profile. For the comparison of the experiments using the cold initial temperature profile, the changes in the cloud liquid and microphysical processes as CDNC increases, are similar to those found by Saleeby et al. [2006], who used the Colorado State University - Regional Atmospheric Modeling System with a different microphysics scheme, to simulate realistic wintertime orographic clouds in northern Colorado.

The average $\delta^{18}O$ values of total precipitation compare well with observations in conditions similar to those used here for initial conditions. Anderson et al. [2015] calculated the average $\delta^{18}O$ of snowpack using the Isotopes in Rocky Mountain Snowpack (IRMS) database, and found that values ranged between -10‰ and -25‰, which matches up well with the range of $\delta^{18}O$ in the average precipitation for our experiments (approximately -10‰ to -16‰ in warm simulations and -19‰ to -26‰ in cold experiments). Comparable precipitation $\delta^{18}O$ values (-12‰ to -24‰) were measured during a 1985 March storm in Kingvale, CA, which is located upwind of the Sierra Nevada crest at an elevation of 1859m [Warburton et al., 1993]. Warburton and DeFelice [1986] analyzed samples in the Central Sierra Nevada, and found that snow formed through vapor deposition had a $\delta^{18}O$ signature that ranged from -18.4‰ to -22.9‰, which corresponds well with our cold temperature profile experiments (see further discussion in section 3.3). Snow samples from the same study, that indicated growth by a combination of riming and vapor deposition, were less depleted and ranged between -6.4‰ and -16.8‰, which resembles results in our warm simulations (see sections 3.1 and 3.3). Values very similar to Warburton and DeFelice [1986] were measured in Colorado by Lowenthal et al. [2011] for snow that had undergone little
riming. In the same study, snow that experienced more riming (as indicated by higher concentrations of sulfate), was less depleted and ranged between -15.6‰ and -20.4‰.

3. Model Results

3.1. Reference Simulations

The output from the 800m mountain experiments with the warmer sounding will serve as the reference simulation for all experiments conducted in this work. Fig. 1 shows the average simulated mass of cloud liquid along with the combined mass of cloud ice and snow for three cases of increasing CDNC values, which will illustrate potential changes in the orographic cloud as the CDNC changes. The figures combine ice and snow together as the setup of the Thompson scheme quickly leads to the production of snow, and as a result, produces little ice [Thompson et al., 2008]. Note that while a wave cloud exists aloft and downstream of the mountain in these simulations, our focus is on the cloud and precipitation over the mountain, which produces almost all of the precipitation in these simulations.

The orographic cloud produced in the 800m simulations does not extend higher than 4km in all CDNC cases and has a much higher mass of cloud liquid than the combined mass of snow and ice. As the CDNC value increases, the amount of cloud liquid increases, as well as the leeward region in which it spans. While there is already leeward spillover of cloud liquid in the 25cm$^{-3}$ case, cloud liquid extends an additional 15km down the leeward side in the 800cm$^{-3}$ experiment, with the region of maximum mass mixing ratio (red filled contours) also reaching approximately 5km further downstream. There is no change in the location of the cloud on the windward slope in all of the warm 800m experiments.
The isotopic values of vapor, cloud liquid, rain and ice/snow for the reference setup (with CDNC=200cm$^{-3}$) are presented in Fig. 2. Cloud liquid isotopic values range from approximately $-7\%$ near the mountain surface to $-22\%$ at cloud top (Fig. 2c). Isotopic equilibrium is enforced between cloud liquid and vapor, so that this decrease in the cloud liquid $\delta^{18}O$ with height is expected as the vapor $\delta^{18}O$ has the same trend (Figs. 2a and 2c). As the CDNC increases, there is no change in the simulated isotopic signature of cloud liquid (not shown). The cloud liquid that extends further leeward has roughly the same $\delta^{18}O$ value as the cloud liquid on the corresponding windward side, so there is no obvious $\delta^{18}O$ difference between the windward and leeward cloud liquid.

There is much less cloud ice and snow that is produced in these warm 800m mountain runs compared to cloud liquid. While Fig. 1 shows the combined snow and ice mass, looking at the two separately, emphasizes how little ice is actually present. Traces of ice are produced in the orographic cloud with a mixing ratio on the order of $10^{-7}$ kg kg$^{-1}$ located just over the mountain peak about 1-2km above the mountain surface. The rest of the frozen hydrometeor mass is composed of snow and occurs predominately upstream of the mountain peak with only a little spillover (~10km) to the downstream side. For the most part, the liquid and ice/snow regions of the cloud overlap, except on the leeward slope, where the glaciated cloud is located above the liquid one. As the CDNC increases, there is no change in location of the ice/snow; both the horizontal and vertical extent of the glaciated cloud remain the same. The mass mixing ratio, however, does decrease very slightly (note change in contours over the mountain peak region) as the CDNC increases, which is opposite to and of much smaller magnitude than the trend found for cloud liquid.
The cloud ice/snow $\delta^{18}O$ values range from $-10\%_\text{o}$ near cloud base to $-35\%_\text{o}$ at cloud top (Fig. 2d). The most depleted ice has a $\delta^{18}O$ value around $-20\%_\text{o}$, but the vertical extent of ice is not as high as snow, and so it is expected that the snow would be more depleted than the ice. Where the snow and ice coexist within the cloud, they have very similar $\delta^{18}O$ values. As in the case with the cloud liquid, there is no change in the isotopic signatures of the cloud ice/snow as the CDNC value is changed.

In the warm 800m runs, most of the precipitation falls as rain (see Tab. 1) with small, though similar amounts of snow and graupel produced. The third column in Tab. 1 indicates that the accumulated precipitation decreases as CDNC increases, and is reduced by more than half between the 25cm$^{-3}$ and 800cm$^{-3}$ experiments. Fig. 3a shows that the location of the maximum precipitation shifts leeward as CDNC increases, which has been seen in observations of wintertime orographic precipitation [Jirak and Cotton, 2006; Saleeby et al., 2011]. The magnitude of the shift is strongest and most obvious in these reference experiments, where there is a difference of approximately 10km between the precipitation peaks in the 25cm$^{-3}$ and 800cm$^{-3}$ simulations. This shift is also evident in the spillover calculations in Tab. 1, which calculates the ratio of the accumulated leeward precipitation to the total precipitation. While the precipitation transitions from being predominately windward to leeward, the horizontal extent of upstream precipitation shrinks slightly and falls closer to the mountain peak, though the location of downstream precipitation remains the same for all CDNC experiments as indicated by the overlap of accumulated precipitation in Fig. 3a.

Since most of the precipitation falls as rain, the isotopic signature of the rain dominates the $\delta^{18}O$ signal of the total precipitation. Windward rain is more enriched than that on the
leeward slope, though this difference changes with CDNC (Fig. 3b). The windward $\delta^{18}O$
values show a wider spread between CDNC experiments than the leeward side, such that
the difference between the 25cm$^{-3}$ and 800cm$^{-3}$ simulations on the upstream side is about
5\% and only 2\% downstream. The windward/leeward $\delta^{18}O$ difference also decreases with
increasing CDNC, leading to a difference between the most enriched windward rain and
most depleted leeward rain of about 5\% in the 25cm$^{-3}$ simulation and only about 2\% in
the 800cm$^{-3}$ case. Likely, this pattern is emerging due to the location and magnitude of
different precipitation growth processes.

Precipitation growth processes are the sources in an aggregate budget for total mass
of precipitation (rain, snow and graupel in the case of the Thompson microphysics) in
the model domain. Since isotopic composition is unchanged by exchanges between rain
and graupel by freezing, melting or aggregation, we focus on the sources which determine
the isotopic composition of the precipitation: autoconversion/accretion of cloud liquid
or cloud ice, riming of cloud liquid and exchanges with vapor by deposition or subli-
mation/evaporation. Note that, because the Thompson microphysical scheme produces
little cloud ice, much vapor deposition onto ice phase hydrometeors contributes directly
to snow growth. Other microphysical schemes would likely have stronger vapor deposition
onto cloud ice, so that the autoconversion/accretion of cloud ice would be relatively more
important and vapor deposition relatively less important in those schemes.

According to Fig. 4b, most precipitation is formed by riming of snow and graupel with
small contributions from accretion of liquid droplets and vapor deposition onto ice. In
the case of both accretion and riming of cloud droplets, these processes are predominant
on the windward side in the 25cm$^{-3}$ experiment, but decrease in magnitude and extend
further leeward as CDNC increases (not shown). As these processes involve cloud liquid and occur near the surface, the $\delta^{18}O$ signatures are more enriched than vapor deposition (Fig. 4d), which tends to occur higher above the mountain surface where temperatures are colder and the vapor is more depleted.

The increasingly leeward extent of accretion and riming with higher CDNC values explains why the windward/leeward $\delta^{18}O$ difference decreases. In the 25cm$^{-3}$ case, where the windward/leeward difference is largest, riming and accretion are confined mostly to the windward side of the mountain, which results in precipitation with a higher $\delta^{18}O$ value than precipitation on the leeward side, where vapor deposition predominately occurs. However, as riming and accretion shift further along the leeward side of the mountain, precipitation $\delta^{18}O$ values are higher than when formed through vapor deposition alone. Additionally, as the location of the precipitation shifts further downstream with higher CDNC, there is less rainout and removal of heavy isotopes upstream, leaving more to fall downstream, which would also help to explain the decreasing windward/leeward $\delta^{18}O$ difference [Smith et al., 2005]. Rimming and vapor deposition have been included in the discussion of rain $\delta^{18}O$ as temperatures near the mountain surface are above freezing, and much of the downstream rain results from melting of snow and graupel (not shown). As the snow and graupel will form through vapor deposition and then potentially rime some cloud liquid, the $\delta^{18}O$ signature of rain from snow and graupel melt is more depleted than rain formed from accretion of cloud droplets. This is another reason why leeward rain is more depleted in the lower CDNC cases, as more of this rain is from melt, while in the higher CDNC cases, rain results from a mix of accretion and melt. Leeward rain that
forms only through graupel and snow melt (i.e. rain that falls where no accretion occurs), has a $\delta^{18}O$ value similar to that of snow and graupel in that region.

Rain that falls furthest downstream seems to jump to higher $\delta^{18}O$ values in all CDNC experiments (Fig. 3b). This change occurs around 12km from the peak in the 25 cm$^{-3}$ case and increases to about 15km from the peak by the 800cm$^{-3}$ simulation. Throughout most of the domain, regions of rainfall typically overlap with cloud liquid, implying that there is likely little rain evaporation. However, on the lee slope, particularly where there is an increase in the $\delta^{18}O$ of precipitation, there is no rain/cloud overlap. Instead, the rain overlaps with a region of negative deposition, which indicates that sublimation and evaporation are occurring. Just as was seen in studies like Bony et al. [2008], equilibration between rain and environmental vapor acts to enrich the rain as the lighter H$_2$O will more quickly move from the liquid to the surrounding vapor, resulting in rain with higher concentrations of heavier isotopes and thus larger $\delta^{18}O$ values.

### 3.2. Additional Cases’ Sensitivities to CDNC

In the additional experiments, using both higher mountain heights as well as a colder initial temperature profile, resulted in sensitivity to changes in CDNC that were quite varied. While the orographic cloud produced in the 800m cold temperature experiment is quite similar to the reference in terms of vertical extent (Fig. 5b), the rest of the additional experiments with higher mountains (both warm and cold) have greater cloud depths, which reach altitudes $\geq$ 6km (Figs. 5c-f). All experiments also produce more ice/snow than the reference, and in the cold temperature experiments, there is more cloud ice/snow than liquid. The amount of cloud liquid is sensitive to changes in CDNC, and as in the reference simulation, the amount of cloud liquid increases with higher CDNC, though the
leeward shift present in the reference case does not occur in the additional experiments
(not shown). In the previous section, it was noted that there is a slight decrease in cloud
ice/snow as CDNC increases (Fig. 1), however, for most of the additional experiments,
the glaciated cloud appears insensitive to CDNC, except for the 800m cold temperature
case, which like the reference, shows a decreasing ice/snow content with increasing CDNC.
The glaciated cloud’s insensitivity to CDNC was also observed by Creamean et al. [2015],
who found that there was little change to ice production when an air mass contained a
higher concentration of aerosols that serve as CCN.

As indicated in Tab. 1, all of the additional simulations produce more total precipitation
than the reference experiment, though the reference has the greatest accumulation of
rain. Note that the total precipitation is computed as the domain-integrated precipitation
normalized by the mountain half-width (20 km in these simulations). The sensitivity
of accumulated precipitation to CDNC is inconsistent among the different experiments,
though there are some small relationships to be noted. For example, the total precipitation
decreases with increasing CDNC in many, though not all, of the experiments, with the
higher mountain heights showing less sensitivity to CDNC changes. This trend (or lack
thereof) can be seen more clearly in Fig. 6, which compares the total precipitation and
$\delta^{18}O$ of the accumulated precipitation for each of the CDNC simulations in all mountain
height experiments. In the reference experiment (labeled as W800m), there is a clear
decreasing trend in the total precipitation as the CDNC increases. This trend is also in
the cold 800m experiments and the warm 1500m and 3000m simulations, though only when
excluding the 25cm$^{-3}$ experiments in the latter two cases. There is not much difference in
the accumulated precipitation between CDNC experiments for the cold 1500m mountain,
and while there is a noticeable difference in the cold 3000m simulation, this is not related to changes in CDNC.

The spatial distribution of precipitation appears to be somewhat sensitive to changes in CDNC, though other than the reference and cold 800m experiment, this relationship is often quite weak. For the reference simulation, as CDNC increased, there was a clear shift in the location of maximum precipitation leeward in addition to the bulk of the precipitation falling downstream of the mountain peak (Fig. 3). This same shift in precipitation distribution is also clear in the cold 800m mountain experiments, which is evident when looking at the spillover values in Tab. 1. Though the spillover increase is not as large as the reference, except for the 25cm$^{-3}$ simulation, half or more of the accumulated precipitation falls leeward. In the rest of the experiments, the majority of the precipitation is falling upstream (Fig. 7), though there is a slight leeward shift associated with increasing CDNC, as indicated by the increasing spillover values (Tab. 1). This shift is typically small and less pronounced than the reference, but still present, except for the cold 3000m experiment. Overall, the location of the precipitation seems to have some dependence on the CDNC, though this relationship varies significantly with mountain height, but does not appear to depend on the initial temperature profile used.

Similar to the precipitation amount and distribution, the $\delta^{18}O$ of precipitation is generally unaffected by CDNC changes, and any variation measured is relatively small. Fig. 7d and Fig. 8c illustrate that the precipitation in the additional experiments is more depleted than the reference, however, unlike the reference there is typically little variation between the different cases. In Fig. 6, the $\delta^{18}O$ of precipitation in the reference case decreases by about 2% between the 25cm$^{-3}$ and 800cm$^{-3}$ simulations. The scatter in the
additional experiments is either quite small (less than 1%), as in the warm 1500m, warm 3000m and cold 800m simulations, or indistinguishable (cold 1500m and 3000m). Based on these results, sensitivity of the precipitation $\delta^{18}O$ to CDNC changes is smaller than the $\delta^{18}O$ variation related to mountain height and temperature soundings.

3.3. Contributions to Precipitation $\delta^{18}O$

The relative contribution to precipitation formation from the different microphysical sources is presented in Fig. 8b for all of the experiments. In the reference experiment (dark blue), riming is the biggest source of precipitation, with vapor deposition onto ice making the second largest contribution as well as measurable influence from accretion of cloud liquid. The $\delta^{18}O$ signature of riming and accretion are less depleted than that of vapor deposition onto ice as indicated in Fig. 8d. Since the average $\delta^{18}O$ of precipitation will reflect the composition of the various sources, precipitation that results mainly from riming will have an isotopic signature that resembles the $\delta^{18}O$ of riming, and will be more or less depleted from this $\delta^{18}O$ value depending on the relative contribution of other, more depleted processes, such as vapor deposition. In Fig. 8c, the $\delta^{18}O$ of total precipitation is the most enriched in the reference compared to all other simulations, and this is also the case where the source of riming and accretion are largest and vapor deposition smallest. The riming source is always greatest in the smallest mountain height experiments, and the contribution decreases as mountain height is increased (Fig. 8b). The opposite is true for the contribution of vapor deposition, which is smallest in the 800m experiments and increases with mountain height. In the cold temperature simulations, ice autoconversion becomes an increasingly important source of precipitation, such that in the 3000m experiment, this process contributes more than riming. The $\delta^{18}O$ signal
of ice autoconversion is very depleted, more so than vapor deposition, and thus as these
two processes contribute more to the formation of precipitation, the average $\delta^{18}O$ of
precipitation decreases and is lowest in this experiment (Fig. 8c). It should be noted that
the deposition measured in the bar charts of Fig. 8b includes some deposition from the
lee wave cloud (seen in Fig. 5). However, this cloud does not contribute much to the
accumulated precipitation other than trace amounts of rain far downstream ($\geq$50km from
the peak) likely forming from snow and graupel melt.

It is evident, particularly in the cold temperature experiments, that the role of the pre-
cipitation sources contribute significantly to the $\delta^{18}O$ signal and that the decreasing $\delta^{18}O$
signal with increasing mountain height is not simply a result of decreased temperature.
The solid line in Fig. 6 represents the regression of the warm temperature experiments’
precipitation and $\delta^{18}O$ values. The dashed line, however, is the same as the solid line,
but shifted down by 8%, which represents the surface vapor $\delta^{18}O$ difference between the
warm and cold temperature profiles (see Figs. 2a and 9a). The cold 800m simulations fall
on this dashed line in Fig. 6, but as the mountain height increases, the $\delta^{18}O$ values of the
cold temperature simulations fall well below this line, implying that precipitation is more
depleted than what is expected from the 8% offset in the upwind sounding in the cold
1500m and 3000m experiments. This deviation of the higher mountain simulations, can
be explained by the changing sources of precipitation. In these two cases in particular, the
riming source is smallest and deposition is greatest and the $\delta^{18}O$ values are more similar
to the deposition $\delta^{18}O$ signatures.

The spatial distribution of the $\delta^{18}O$ signal as seen in Fig. 7c-d, can also be attributed to
the sources of precipitation. In both the warm and cold temperature experiments, wind-
ward precipitation is less depleted than on the leeward slope. This windward/leeward $\delta^{18}O$ difference is smallest in the reference simulation and increases with mountain height. The spatial distribution of this signal relates both to the magnitude and the location of the microphysical processes that contribute to precipitation growth. In all experiments, riming and accretion of cloud liquid occur predominately windward with some spillover in the reference simulation, but little to no spillover in the other experiments. However, deposition, the larger source of precipitation in the higher mountain cases, occurs on both sides of the mountain, and is the dominant or only growth process that occurs leeward in all experiments (not shown). As riming becomes second to vapor disposition as a source term for precipitation, and the leeward extent of riming simultaneously decreases, precipitation forming and falling on the lee slope is more depleted, as it grows through vapor deposition, while windward precipitation forms through a combination of riming, deposition and accretion and is thus less depleted. Therefore, in the reference case where windward and leeward precipitation forms from a combination of riming, accretion (though small), and deposition, the windward/leeward $\delta^{18}O$ difference is smaller compared to other experiments (Figs. 7c-d). It is important to note that in addition to the microphysical processes, increased rainout would also contribute to the windward/leeward $\delta^{18}O$ difference [Smith et al., 2005]. Tab. 1 indicates that as the mountain height is increased, the accumulated precipitation more than doubles, which would lead to a greater removal of $\text{H}_2^{18}\text{O}$ windward in the higher mountain cases compared to the reference simulations.

4. Discussion

One of the main goals of this work has been to study the isotopic signatures of precipitation and cloud microphysical processes to determine if there is a distinct isotopic
signal associated with those processes. With the idealized setup using different mountain heights, warm and cold temperature profiles and increasing CDNC, the simulations show that there is a distinct difference in the $\delta^{18}O$ signatures of microphysical processes. Rimming and vapor deposition onto ice are the main pathways of precipitation growth among all of the experiments, with accretion of cloud droplets and ice autoconversion making notable contributions in some, but not all, simulations. Considering all of the warm temperature simulations (all mountain heights), the difference in the $\delta^{18}O$ of riming and vapor deposition ranges from 3-7‰, with that difference increasing with mountain height. A similar range of 4-8‰ was found in the cold temperature experiments. This result seems to imply that the difference is not dependent on the environmental temperature, but rather the temperature difference as a result of the altitude at which these processes occur. As deposition occurs both near the surface and especially at higher altitudes, and therefore at colder temperatures, the $\delta^{18}O$ is more depleted than riming that is predominately happening near the mountain surface. With increasing mountain height, the altitude of vapor deposition increases, and thus the larger difference results when comparing the $\delta^{18}O$ values.

Additionally, these experiments investigate the influence of aerosol loading on microphysical processes and precipitation, as represented by changes in the imposed CDNC. As can be seen Tab. 1, the biggest changes between the CDNC simulations occurs when rain is the dominant form of precipitation (e.g. warm 800m simulations) and/or when riming is the main process of precipitation growth as indicated in Figs. 4 and 8. Increasing the CDNC has the most influence on the location and horizontal extent of the liquid cloud, which is also associated with changes in the location of riming and accretion of cloud liq-
uid. Since the warm 800m simulations are the only example of producing almost entirely rain, an additional 1500m mountain simulation was conducted using a warmer initial temperature profile to determine if the precipitation trends associated with increased CDNC were also found to occur in a higher mountain case. The initial temperature profile for this simulation is 285.15K at the surface, which is chosen to get temperatures near the peak of the mountain that are similar to those near the peak in the 800m case. Most of the precipitation in this experiment is also rain, with small amounts of snow and graupel (Tab. 1). Results are similar to the reference simulations: as CDNC increases, the accumulated precipitation decreases and shifts downstream as indicated by the increase in the spillover calculation.

The impact of changing CDNC on the $\delta^{18}O$ of precipitation varies depending on the type of precipitation produced and what processes are responsible for precipitation growth. For example, it was found that — unlike cloud liquid — cloud ice/snow changes vary little as CDNC increases despite the expected impact of cloud droplet size on riming. Overall, different values of CDNC have little influence on ice phase physics, such as depositional growth of ice crystals and snow, except in instances where riming contributes significantly to snow growth (i.e. warm 800m and additional warmer 1500m experiments). Therefore, when considering the isotopic signature of snowfall, or even the total precipitation in cases where it mostly consists of snow, there is very little variation between CDNC experiments in terms of the $\delta^{18}O$ value. However, within those experiments, the $\delta^{18}O$ of graupel and rain increases with increasing CDNC. Both of these hydrometeors are dependent on liquid processes (e.g. accretion of cloud liquid and riming), and therefore are more susceptible to changes in the cloud liquid, which, as discussed previously, is found to occur with
increasing CDNC. However, the overall impact of aerosol loading on orographic clouds, and specifically ice and snow, might change more dramatically if one were to consider aerosols that are more likely to serve as ice nuclei, whereas varying the CDNC values in this study is geared toward the influence of aerosols that serve as CCN.

5. Conclusions

The microphysical controls on the isotopic composition of wintertime orographic precipitation have been explored in idealized simulations of flow over a two-dimensional mountain using a model that incorporates stable water isotopologue physics into its microphysical representation. Changes in precipitation and its isotopic content due to perturbations in CDNC, temperature and mountain height result in shifts between the different microphysical processes that contribute to precipitation formation, and the characteristic isotopic content of the precipitation formed by these different processes is reflected in the precipitation at the surface. In addition, the effects of decreasing temperature and increasing mountain height cause the precipitation to be formed in colder conditions from more depleted vapor and contribute to the trend in isotopic composition with mountain height and temperature.

The sensitivity of precipitation to CDNC changes is strongest for the lowest mountain height (800m) and weaker for colder conditions and larger mountain heights. This result agrees with the notion that, in some scenarios with efficient precipitation formation, precipitation is insensitive to aerosol (or in this case CDNC) perturbations, as in Miltenberger et al. [2015]. When precipitation is sensitive to CDNC, the isotopic composition of that precipitation and the difference in isotopic content across the mountain barrier is
also modified. These changes in $\delta^{18}O$ are measureable, but small in comparison to those
driven by temperature changes or different mountain heights.

Precipitation formation in these idealized wintertime orographic clouds is domi-
nated in most cases by riming, vapor deposition onto precipitating ice, and sublima-
tion/evaporation of falling precipitation, with contributions from the accretion of cloud
liquid and autoconversion of cloud ice in some cases. When increasing CDNC causes
changes in precipitation, an overall decrease in precipitation is accompanied by a shift in
precipitation production from riming to vapor deposition. As the precipitation formed
through vapor deposition tends to be formed aloft in more depleted conditions than that
formed through riming, the change in the microphysical sources of precipitation is reflected
in its isotopic content. Larger changes in isotopic content are seen with variations in the
temperature of the upstream sounding or mountain height, and the shift from riming to
vapor deposition plays a role in these changes, along with the increasingly depleted vapor
from which precipitation is formed at colder temperatures or increasing mountain heights.

The results from these simulations have potential implications for orographic snow and
observational campaigns. As liquid processes are most responsive to changes in CDNC,
locations where precipitation primarily forms through accretion of cloud liquid and/or
riming are likely to experience decreased accumulation of precipitation and a shift in
the location of precipitation, though the extent of this shift appears to depend on both
the mountain height and the makeup of the precipitation. The simulations indicate that
the degree of change in precipitation and $\delta^{18}O$ will vary with the specific location being
considered, however this could be beneficial to campaigns in determining sites that would
prove most useful for taking observations and collecting samples.
Appendix A: Incorporating Isotopologues Into the Microphysics

Stable water isotopologues were added to the Thompson microphysics scheme in the WRF model by duplicating all microphysical processes (e.g., freezing, melting, vapor deposition, evaporation) with additional process rates for the water isotopologues following Blossey et al. [2010, App. B]. Except for the sublimation of ice, which is assumed to be non-fractionating, all exchanges between vapor and condensate involve fractionation. The fractionation/equilibration of water isotopologues from rain is included, along with the fractionation of water vapor as it is deposited onto ice phase hydrometeors. Water vapor and cloud liquid are assumed to be in isotopic equilibrium. While a detailed description of the isotopic treatment including all of these processes can be found in appendix B of Blossey et al. [2010], we give a brief summary below that emphasizes those processes that play important roles in the cold and mixed-phase clouds central to this study.

For most processes, especially those in which whole hydrometeors are moved from one microphysical category to another (e.g., freezing of cloud droplets to form ice), the heavy isotopologues of water are transferred in proportion to their concentration in the source hydrometeor. For example, the freezing of cloud liquid droplets ($w_{fz}$) to form cloud ice transfers heavy isotopologues to cloud ice as follows:

$$\left. \frac{dr_i'}{dt} \right|_{w_{fz}} = \left. \frac{dr_c}{dt} \right|_{w_{fz}} R_c$$  \hspace{1cm} (A1)

where $R_c = r'_c/r_c$ is the isotopic ratio of cloud liquid, and $r_c$ and $r_i$ are the mass mixing ratios of cloud liquid and cloud ice, respectively. The mass mixing ratios of heavy isotopologues are denoted with primes, e.g. $r'_c$. Further, it is assumed that the isotopic ratio is uniform in each hydrometeor category, so that large and small raindrops have the same isotopic composition, for example. The latter assumption will not hold exactly in reality.
and is a source of error; the computation complexity of allowing such variation in a bulk scheme could be considerable.

For the few microphysical processes that result in fractionation (the unequal transfer of heavy and light isotopologues between phases), those processes are represented as described in Blossey et al. [2010, App. B]. In general, the lower vibrational energy of the heavier isotopologues of water cause them to prefer the condensed phases (liquid, ice) to the vapor phase, so that their concentrations in vapor are smaller than in the condensed phases. When comparing concentrations of isotopologues, the words “heavier” or “more enriched” are used to describe concentrations of heavy isotopologues that are higher, while “lighter” or “more depleted” are used for smaller concentrations of heavy isotopologues. We supply here a summary of how these processes might affect the isotopic composition of water in mixed-phase clouds.

The efficient exchange between small liquid water droplets in clouds and the surrounding water vapor leads many microphysical schemes (including Thompson) to assume that in-cloud water vapor mixing ratios are equal to the saturation mixing ratio when cloud liquid is present. The complementary condition for heavy isotopologues is that the isotopic ratios of cloud liquid and water vapor are in isotopic equilibrium:

\[ R_c = \alpha_l R_v \]  \hspace{1cm} (A2)

where \( R_v \) is the isotopic ratio of water vapor and \( \alpha_l \) is the equilibrium fractionation coefficient for liquid [Majoube, 1971]. As the equilibration time for isotopic composition of small liquid water droplets is on the order of a few seconds [Ciais and Jouzel, 1994], this is in general a good assumption and is included in our implementation.
Given the potentially large supersaturation with respect to ice, we need to consider the non-equilibrium processes driven by gradients of water vapor between the environment and ice particles which leads to vapor deposition onto those particles. The relatively smaller diffusivities of the heavy isotopologues modifies the transfer of water to the particle surface, so that the deposition of heavy isotopologues may be written as

\[
\frac{dr'_i}{dt_{dep}} = \alpha_s \alpha_k R_v \frac{dr_i}{dt_{dep}}
\]

[Clais and Jouzel, 1994], where \(\alpha_s\) is the equilibrium fractionation coefficient for ice [Majoube, 1970; Merlivat and Nief, 1967] and the kinetic fractionation coefficient, \(\alpha_k\), represents the effects of the relative diffusion of the heavy and light isotopologues [Jouzel and Merlivat, 1984].

The effects of these two processes on the isotopic composition of liquid and ice in mixed-phase clouds is depicted in Fig. A1. Here, the variation in the saturation ratios with respect to liquid and ice is depicted in the left panel as a function of temperature. In keeping with the assumption in the microphysical scheme, the saturation ratio with respect to liquid is one, while the ice saturation ratio grows with decreasing temperature. The isotopic content of the vapor, cloud liquid and ice formed through vapor deposition is shown in the right panel. Equations A2 and A3 have been used to compute isotopic composition, except for cloud liquid water, whose value is fixed to the relationship observed in mixed-phase orographic clouds by Lowenthal et al. [2011],

\[
\delta^{18}O_c = 0.9T - 10.12
\]

where \(T\) is temperature in degrees Celsius. While ice formed through vapor deposition is more enriched than cloud liquid at the same temperature, close to 0°C, increasing the
supersaturation with respect to ice and decreasing temperature causes a stronger kinetic
effect during deposition onto cloud ice. Note that vapor deposition onto ice in a liquid
class is most efficient at colder temperatures, peaking close to -15°C [Rogers and Yau,
1989, p. 161]. As a result, the typical isotopic composition of ice formed through vapor
deposition is often more depleted than that of cloud liquid closer to 0°C. Note that these
relationships will not hold once the liquid water is removed and the cloud is fully glaciated.

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cited and referred to in the reference list. Those interested in the isotope-enabled Thomp-
son microphysics scheme should contact co-author Peter Blossey (pblossey@uw.edu).

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Figure 1. Contoured temperature (black) and average mixing ratios of cloud liquid water (shaded) and ice (contoured) for the 800m warm temperature simulations. Cloud droplet number concentrations of (a) 25 cm$^{-3}$, (b) 200 cm$^{-3}$, and (c) 800 cm$^{-3}$ are shown to illustrate sensitivity to CDNC. Units are kg kg$^{-1}$ for hydrometeor mixing ratios and K for temperature. Note that cloud ice includes ice and snow hydrometeor categories.

Figure 2. Average $\delta^{18}O$ of (a) vapor, (b) rain, (c) cloud liquid and (d) ice/snow for reference experiment as represented by the 200cm$^{-3}$ simulation.

Figure 3. Reference simulation (a) accumulated precipitation and (b) $\delta^{18}O$ of precipitation for all CDNC experiments. The mountain peak is located at 300km.

Figure A1. [Left panel:] the saturation ratios with respect to cloud liquid (blue) and ice (green) as a function of temperature. [Right panel:] $\delta^{18}O$ of in-cloud water vapor (black), cloud liquid (blue) and vapor deposition onto ice particles (green) as a function of temperature.
Figure 4. Breakdown of the reference simulation (a) total precipitation, snow and graupel with (c) the respective average $\delta^{18}O$ values. The contributions from (b) precipitation sources normalized by the total precipitation and (d) the corresponding isotopic signatures. Sources from left to right are: autoconversion of cloud liquid, accretion of cloud liquid by rain, autoconversion of cloud ice, accretion of cloud ice, riming of cloud liquid, vapor deposition onto ice, and sublimation of ice.

Figure 5. Contoured temperature (black) and average mixing ratios of cloud liquid water (shaded) and ice (contoured) for the $200\text{cm}^{-3}$ simulations from the (a) 800m, (b) 1500m, (c) 3000m warm temperature experiments and (d) 800m, (e) 1500m, and (f) 3000m cold temperature experiments. Units are kg kg$^{-1}$ for hydrometeor mixing ratios and K for temperature. Note that cloud ice includes ice and snow hydrometeor categories.

Figure 6. Scatter plot of total precipitation vs $\delta^{18}O$ of precipitation for all mountain heights in both the warm (W) and cold (C) experiments as labeled on the figure. Different colored circles represent values for each CDNC experiment. Gray solid line is the regression for the warm temperature experiments and the dashed line is the solid line shifted down by 8‰.

Figure 7. Total precipitation distribution in (a) warm experiments and (b) cold experiments. Corresponding $\delta^{18}O$ of precipitation for (c) warm experiments and (d) cold experiments. All profiles are based on the $200\text{cm}^{-3}$ simulations. The mountain peak is located at 300km.
<table>
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<th>Rain</th>
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<tr>
<td>C 3000m</td>
<td>200</td>
<td>166.7mm</td>
<td>99%</td>
<td>0.3%</td>
<td>0.7%</td>
<td>22%</td>
</tr>
<tr>
<td>C 3000m</td>
<td>400</td>
<td>175.1mm</td>
<td>99.4%</td>
<td>0.3%</td>
<td>0.3%</td>
<td>22%</td>
</tr>
<tr>
<td>C 3000m</td>
<td>800</td>
<td>172.2mm</td>
<td>99.6%</td>
<td>0.3%</td>
<td>0.03%</td>
<td>22%</td>
</tr>
</tbody>
</table>

Table 1: Breakdown of the major statistics for all runs. Columns indicate the case (height and cold (C), warm (W) or warmer (W+) initial temperature profile), cloud droplet number concentration (CDNC) in cm$^{-3}$, normalized (by the mountain half-width) domain-integrated precipitation over 12 hour simulation, and the percent of snow, graupel and rain. Spillover is the ratio of the total leeward precipitation to the total mountain precipitation.
Figure 8. Breakdown of (a) total precipitation, snow and graupel and (c) corresponding \( \delta^{18}O \) values for all experiments. The contributions from (b) precipitation sources normalized by the total precipitation and (d) the corresponding isotopic signatures for all experiments. Sources from left to right are: autoconversion of cloud liquid, accretion of cloud liquid by rain, autoconversion of cloud ice, accretion of cloud ice, riming of cloud liquid, vapor deposition onto ice, and sublimation of ice. All values are based on the 200cm\(^{-3}\) simulations.

Figure 9. Average \( \delta^{18}O \) of (top) vapor, (middle) cloud liquid and (bottom) ice/snow for the (a,d,g) 800m, (b,e,h) 1500m, and (c,f,i) 3000m cold temperature experiments as represented by the 200cm\(^{-3}\) simulations.
Distance [km] | Height [km] | W800m |
--- | --- | --- |
230 | 240 | 250 |
260 | 270 | 290 |

(a) 25 cm\(^{-3}\)
(b) 200 cm\(^{-3}\)
(c) 800 cm\(^{-3}\)

\[0.01 \text{ kg kg}^{-1}\]

\[0.03 \text{ kg kg}^{-1}\]

\[0.1 \text{ kg kg}^{-1}\]

\[0.3 \text{ kg kg}^{-1}\]
Accum. Precip. [mm]

Distance [km]

δ$^{18}$O Precip. [%o]
Cloud Liquid and Ice Saturation Ratios

Temperature, deg C vs Saturation ratio

Ice, Vapor and Cloud Liquid $\delta^{18}$O

$\delta^{18}$O, per mil vs Saturation ratio

$S_{\text{liq}}$, $S_{\text{ice}}$, vapor, liquid, ice