Stable Water Isotopes as Tracers in Global Precipitation

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Stable water isotopes as tracers in global precipitation

A dissertation presented
by
Mary Moore
to
Department of Earth and Planetary Science
in partial fulfillment of the requirements
for the degree of
Doctor of Philosophy
in the subject of
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Stable water isotopes as tracers in global precipitation

Abstract

Stable water isotopes (H$_2$O, H$_2^{18}$O, and HDO) are incorporated into the microphysics schemes of two different atmospheric models. This thesis describes the use of these molecules as tracers in precipitation budgets to assess the processes controlling the isotopic signatures of precipitation in the tropics and orographic snow in the mid-latitudes.

The idealized simulations of seasonal precipitation budgets in the tropics determine that increased vapor convergence during intense precipitation is most important for setting the isotopic composition of the convective precipitation. The isotopic signal of the converged vapor is more important than the local evaporation and smaller scale post-condensational processes.

Flow over a 2D-mountain and realistic simulations of orographic clouds show that the isotopic signature of precipitation is more sensitive to changes in mountain height and initial temperature profiles than to the cloud droplet number concentration. Rimming of cloud liquid and vapor deposition onto ice are the largest source terms for orographic precipitation, and have distinct isotopic signatures that are altitude-dependent. When riming is the larger source term, precipitation tends to be more enriched than when vapor deposition dominates.
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1

Introduction

Stable water isotopologues, namely the molecules H$_2^{16}$O, HD$^{16}$O, H$_2^{18}$O and H$_2^{17}$O (the latter is generally not included because of its extremely low abundance, 0.037%, in the atmosphere (Mook (66)), have become an important part of studying climate and weather due to their use as tracers of water through the hydrologic cycle. The work in this thesis uses stable water isotope-enabled cloud resolving models to determine how large-scale factors (such as moisture convergence) as well as cloud microphysics influence isotopic signals in precipitation.

The slight difference in the masses of each of these molecules changes the saturation vapor pressure, leading to variations in the partitioning of each molecule during phase changes. As such, the lighter isotopologue, H$_2^{16}$O, has the highest saturation vapor pressure and tends to collect in the vapor phase, while the heavier isotopologues (HD$^{16}$O, H$_2^{18}$O) have lower saturation vapor pressures and thus accumulate in the condensed phases. The partitioning of the different water molecules is called fractionation, and the conditions under which it occurs will lead to different signals in precipitation (Dansgaard (21), Rozanski et. al. (88)). For example, fractionation that occurs at equilibrium (such as evaporation of ocean water), sometimes called thermodynamic fractionation, is temperature dependent (Mook (66)). As the rate of fractionation is dependent on the bond energies of the molecules, when temperature increases, the rate of fractionation decreases (Gat (27)). Non-equilibrium (or kinetic) fractionation is the result of each molecule having a different diffusivity, which is also weakly dependent on temperature by a factor of $\sqrt{T}$ (Mook (66)).

When measuring and referencing isotopic signals, the common practice is to consider the heavy isotopologues as a ratio to the lighter, standard H$_2^{16}$O molecule. This ratio is
often expressed in per mil (‰) using delta notation, which is calculated from the following equation:

$$
\delta = \left( \frac{R_{\text{sample}}}{R_{\text{standard}}} - 1 \right) \times 1000, \quad (1.1)
$$

where $R_{\text{sample}}$ is the measured ratio in the sample and $R_{\text{standard}}$ is the ratio from the Vienna Standard Mean Ocean Water, which serves as a universal standard for isotopic measurements (Mook (66)). Measurements with a high ratio of heavier isotopologues (less negative $\delta$-values) are referred to as ‘enriched,’ while small ratios and more negative $\delta$-values are referred to as ‘depleted.’

The work in this thesis investigates the isotopic signals of precipitation in different regions of the world. In Chapter 2, analysis focuses on the tropics, where due to the convective activity, the relationship between the isotopic composition and temperature is weaker (Bowen (11)). Instead, the extent of depletion is positively correlated with the amount of precipitation on monthly or longer timescales (amount effect) and is caused by the continued and preferential removal of the heavy isotopologues from an air mass (e.g. Dansgaard (21), Lawrence and White (46), Rozanski et. al. (88)). However, additional research has argued that the depleted precipitation is caused by post-condensational effects within the unsaturated downdrafts (e.g. Bony et al. (7), Kurita et. al. (42), Risi et al. (82), Risi et. al. (83)). Lower isotopic signatures in precipitation may be caused by the injection of depleted vapor from downdrafts into the boundary layer (Kurita et. al. (42), Risi et al. (82, 84)), decreased rates of equilibration between raindrops and the environmental vapor (Ciais and Jouzel (14), Field (24), Lawrence et. al. (45), Lee and Fung (51)), or a combination of the two processes. An investigation of the precipitation budget in the tropics is described in Chapter 2 to present an additional explanation for the amount effect as observed on seasonal or longer timescales.

At higher latitudes outside of the tropics, the isotopic signature is more temperature-dependent (e.g. Dansgaard (21), Jouzel (36), Lee et al. (48), Noone and Simmonds (73)) and in mountainous regions is further linked to altitude (Dansgaard (21)). Precipitation that forms from air forced over a mountain barrier is referred to as orographic precipitation and is an important contributor to surface water resources. The amount and distribution of orographic precipitation can be sensitive to several variables, such as the orientation and geometry of the terrain, atmospheric stability, orographic flow dynamics and cloud microphysics (e.g. Colle (16), Galewsky (25), Muhlbauer and Lohmann (70)). In recent years, research has found that increased aerosols have the potential to alter the growth processes of
orographic precipitation (Albrecht (2), Pruppacher and Klett (77), Ramanathan et al. (78), Wang and Ji (103)). Such changes may decrease the total amount of precipitation and/or where it falls over the mountain, though such impacts depend strongly on the environmental conditions of the region being considered (Borys et al. (9, 10), Khain and Pokrovsky (39), Lynn et al. (57), Muhlbauer and Lohmann (70), Saleeby et al. (93)).

Given the importance of orographic precipitation as a surface water resource, the work in this thesis uses an isotope-enabled model to study the microphysical processes associated with orographic precipitation. Idealized model simulations in Chapter 3 establish the source terms and relative contributions to precipitation growth from each of the microphysical processes and how sensitive these processes are in different atmospheric regimes. This work is then expanded in Chapter 4, when realistic model simulations are utilized to investigate the microphysics in orographic clouds during a wintertime observational period in Colorado.
A Moisture Budget Perspective of the Amount Effect

This chapter originally appeared as a research article in Geophysical Research Letters. For the original article, please refer to Moore et al. (67).

A stable water isotopologue-enabled cloud-resolving model was used to investigate the cause of the amount effect on the seasonal (or longer) timescales. When the total water (vapor and condensed phase) budget of the precipitating column of air is considered, our results indicate that, as convection becomes stronger and the precipitation rate increases, the \( \delta D \) of precipitation (\( \delta D_p \)) depends on the isotopic composition of the converged vapor more than that of surface evaporation. Tests with disabled fractionation from rain evaporation demonstrate that this mechanism does not account for the amount effect as has been previously suggested. If the isotopic content of converged vapor is made uniform with height with a value characteristic of surface evaporation, the amount effect largely disappears, further supporting the dominance of converged vapor in changes to the \( \delta D_p \) signal with increasing precipitation. \( \delta D_p \) values were compared to the water budget term \( \frac{E}{P} \), where P is precipitation and E is evaporation. Results from this comparison support the overall conclusion that moisture convergence is central in determining the value of \( \delta D_p \) and the strength of the amount effect in steady state.
2.1 Introduction

Stable water isotopologues (H$_{16}$O, HDO, H$_{17}$O, H$_{18}$O) are useful as climate proxies and tracers of the hydrologic cycle. Slight differences in the mass of each water isotopologue changes the saturation vapor pressure of the molecule such that the heavier isotopologues tend to collect in condensed phases, while the light isotopologue accumulates in the vapor phase. This fractionation of the heavy and light molecules leads to variations in their respective ratios in a particular phase of water. Changes in these ratios can provide information about temperature (Jouzel (36)), precipitation amount (Dansgaard (21), Rozanski et. al. (88)) and moisture source region (Vuille et al. (102)).

Isotopologue ratios in rainwater samples collected from stations included in the Global Network of Isotopes in Precipitation (GNIP) (IAEA/WMO (33)) were analyzed by Dansgaard (21) who noted that in the sub-tropics and tropics, isotopically depleted rainfall coincided with higher precipitation amounts. This relationship was termed the ‘amount effect’ and was argued to be caused by the increased removal of heavy isotopologues by rainout as clouds cooled. This relationship was later confirmed by Rozanski et. al. (88), who repeated the analysis by Dansgaard (21), but with an additional 30 years of data and specifically focused on tropical marine locations.

More recent studies emphasize the role of the unsaturated downdrafts that both directly and indirectly lead to isotopically depleted rainfall (e.g. Bony et al. (7), Kurita et. al. (42), Risi et al. (82), Risi et. al. (83)). Processes occurring in the unsaturated downdrafts that alter the isotopic composition of the falling precipitation, namely evaporation and equilibration, are referred to as direct effects. As convection strength intensifies, the precipitation rate, average drop size and the relative humidity in the downdrafts increase. Larger drops do not evaporate as significantly as smaller drops (Stewart (99)) and overall evaporation decreases as the relative humidity increases. Since evaporation acts to enrich the drops by preferentially removing of the lighter isotopologue, decreased evaporation leads to lighter, more depleted rain and thus the amount effect (e.g. Bony et al. (7), Dansgaard (21), Risi et al. (82)).

As the relative humidity increases in the unsaturated downdrafts, equilibration becomes the dominant mechanism for isotopologue exchange and acts to bring the rain and the surrounding vapor into equilibrium. When the drop size increases in strong convection,
the equilibration time of the drop increases while the residence time of the drop in the boundary layer (BL) decreases leading to rain that does not equilibrate completely with the surrounding vapor (Ciais and Jouzel (14), Field (24), Lawrence et. al. (45), Lee and Fung (51)). These partially equilibrated drops do not become as enriched as drops that reach equilibrium and instead maintain the depleted isotopic signal at the cloud base, producing the observed amount effect.

The indirect influence of unsaturated downdrafts is referred to as downdraft recycling, whereby depleted vapor from the downdrafts is injected into the BL, which feeds the convective system. Downdraft vapor is depleted through efficient equilibration of the precipitation with the surrounding vapor in addition to a decreased fraction of evaporation (Risi et al. (82)) as well as by subsidence of more depleted vapor from the environment (Kurita et. al. (42), Risi et al. (82, 84)). The more efficient the downdraft recycling, the more depleted the precipitation that is produced by the convective system becomes.

As illustrated above, much work has focused on how the boundary layer, convection and microphysical processes conspire to produce the observed amount effect as precipitation rates increase. In contrast, Lee et al. (48) and Kurita et. al. (41) used Global Climate Models (GCM) and found that the isotopic content of the precipitation is a consequence of moisture source region and transport patterns, which is also supported in a recent study by Aggarwal et. al. (1). In this work, the authors define a residence time parameter that is implicitly dependent on both temperature and circulation patterns and find that it is positively correlated with $\delta^{18}O$ in precipitation collected from 12 global locations. A second recent study (Kurita (43)) suggests that the amount effect is related to the degree of organization of convection, with relatively enriched precipitation associated with disorganized convection and more depleted precipitation arising from mesoscale convective systems.

The goal of this work is to present a different interpretation as to the cause of the amount effect. In contrast to previous work that studied separate budgets of vapor and condensate (e.g. Kurita (43), Risi et al. (82)), we focus on the total column water (vapor plus condensate) budget of the precipitating column. In regions of deep convection where $P \geq E$, this budget has two sources, surface evaporation and moisture convergence, and a single sink, precipitation. We propose that the anti-correlation between the $\delta D$ in precipitation ($\delta D_p$) and precipitation amount is largely a result of isotopically depleted vapor converging in the lower- and mid-troposphere with smaller contributions from surface evaporation. This
approach is similar to that of Lee et al. (48) who clearly identify the relationship between P-E and the isotopic composition of precipitation in simulations with an isotope-enabled GCM. The paper is organized so as to introduce the model and experimental setups in section 2. Model results are presented in section 3, followed by discussion of the results and the water budget terms in section 4. Finally, conclusions to this work are in section 5.

## 2.2 Model and Experiments

### 2.2.1 Model Setup

The model used for this study is an isotope-enabled version of the System for Atmospheric Modeling (SAM), version 6.7 (Khairoutdinov and Randall (38)) and will hereafter be referred to as IsoSAM. Isotopologue physics is implemented in the model by incorporating phase changes of the heavy water isotopologues into the Lin microphysics scheme of the Weather and Research Forecasting model, version 3.1 (Skamarock and Klemp (94)). This version of the Lin microphysics scheme closely resembles that of Lin et. al. (52) and is a single-moment bulk scheme that performs mixed-phase saturation adjustment and does not allow for supersaturation with respect to ice at temperatures below -40°C (Blossey et al. (6)). IsoSAM incorporates the heavy isotopologues (HDO and $\text{H}^{18}\text{O}$) by replicating the transformation processes of $\text{H}^{16}\text{O}$. Fractionation is set to occur during water phase changes, which is described in detail in Appendix B of Blossey et al. (6).

For all of the experiments, IsoSAM is configured as a three-dimensional cloud-resolving model using a radiative-convective equilibrium (RCE) framework over an ocean surface. The simulations run for 500 days with an initial sea surface temperature (SST) of 301.15K on a domain that is 128 km x 128 km with periodic boundary conditions and 64 vertical levels. The weak temperature gradient approximation (WTG) is implemented once the model reaches equilibrium, making it possible to diagnose the vertical velocity and precipitation associated with the changes in SST and prescribed temperature profiles (Sobel and Bretherton (98)). Results presented here use the damped gravity wave approach of Kuang (40) and Blossey et. al. (5) to compute the pressure velocity perturbation. Simulations using the approach of Sobel and Bretherton (98) produce similar results and will not be presented here for brevity.
2.2 Model and Experiments

Bony et al. (7) found using WTG to diagnose the vertical velocity in a RCE framework capable of reproducing $\delta D_p$ values close to those in observations.

Tendencies of moisture and temperature (represented here by $\chi$) due to large-scale circulations that cannot be represented in our small periodic domains are calculated using equation 1:

$$
\left( \frac{\partial \chi}{\partial t} \right)_{ls} = \overline{\chi} \frac{\partial \rho w}{\partial z} - \frac{1}{\overline{\rho}} \frac{\partial}{\partial z} (\overline{\rho w \chi}) ,
$$

as in Daleu et. al. (19) and Raymond and Zeng (80) (see Appendix Fig. A.1 for a depiction of these terms). The over bar indicates the horizontal average of the variables and it is assumed that the horizontal wind is either into or out of the column, as represented by the upwind value $\chi^*$. In regions of large-scale divergence ($\frac{d(\rho w)}{dz} < 0$), the upwind value of $\chi^* = \overline{\chi}$, while in regions of large-scale convergence ($\frac{d(\rho w)}{dz} > 0$), $\chi^* = \chi^{ref}$. The model reaches equilibrium (~ day 60) before reference profiles are calculated (days 100-120), after which point WTG is initiated. The SST is increased by 0.5K increments approximately every 45 days starting at day 255, providing 6 SST regimes ranging from 301.15K to 303.65K for our analysis. In all of the experiments, heavy isotopologues in the atmosphere are initialized at 0 such that the ocean is assumed to serve as an infinite source of all water isotopologues and heat.

2.2.2 Experimental Setup

For this study, three different test cases are created to examine the different theories relating to the causes of the observed amount effect. It is first essential to see that the model is able to reproduce this relationship between the precipitation amount and the isotopic content of the precipitation. In steady state, precipitation is the sum of the surface evaporation and the horizontal convergence of moisture:

$$
P = E + \left< \frac{q^* d(\rho w)}{\partial z} \right> ,
$$

where $P$ is the rate of precipitation and $E$ is the evaporation. The angled brackets indicate the mass-weighted vertical integral and in cases of large-scale convergence the value of $q^* = q^{ref}$. The second term comes from equation (1) above, noting that the mass-weighted vertical integral of the second term on the RHS of equation (1) is zero. By using WTG, increasing
SST allows for the examination of increasing precipitation regimes associated with changes in \( \delta D_p \) indicating the presence and strength of the amount effect. While IsoSAM is configured to simulate the microphysical processes of both HDO and H\(_2\)\(^{18}\)O, our results will be shown in terms of HDO, since \( \delta^{18}\)O and \( \delta D \) values are generally proportional, and showing results for both isotopologues would be redundant.

Two sensitivity studies are performed to identify the key contributor to changes in \( \delta D_p \) with increasing precipitation. First, an evaporation test disables fractionation associated with rain evaporation and equilibration. The vapor produced by rain evaporation has the same isotopic composition as the rain itself and thus ensures that there is no preferential removal of the lighter isotopologue. This is done by setting the isotopic content of the evaporation flux to be equal to the ratio of HDO rain to H\(_2\)O rain multiplied by the amount of H\(_2\)O water evaporated from the raindrop. Prohibiting fractionation during rain evaporation can be expected to prevent the rain from becoming increasingly heavier as the rain evaporation takes place in relatively dry air and instead should cause it to become lighter compared to the control case.

Second, we performed an experiment with a reference water vapor profile that has a uniform isotopic composition with height (\( \delta D_{v}^{\text{ref}} \)). The uniform reference vapor profile is set to be -23‰, representing the \( \delta D_{v} \) of the evaporated surface flux in the control case. As vapor generally becomes more depleted with height, using the uniform \( \delta D_{v}^{\text{ref}} \) will tend to make the isotopic composition of converged vapor and the resulting precipitation isotopically more enriched relative to the control simulation. However, in simulations where \( P \sim E \) and there is no net import of vapor to the column, the effect of changing \( \delta D_{v}^{\text{ref}} \) will likely be small.

2.3 Model Results

2.3.1 Control Case

In the control run, the output produces an amount effect comparable to observations, though more depleted by as much as 40‰ for the highest precipitation rates (see Fig. 2.1 (left) and Table 2.1).

The results are also in agreement with those of previous work using a similar setup but with a single column model (Bony et al. (7)). Since the simulation is run in RCE
Figure 2.1: Correlations Between Precipitation and $\delta D_p$ - Showing (left) the correlation between precipitation and $\delta D_p$ for the control (red), evaporation test (blue) and uniform $\delta D_{ref}$ profile test (green). Each of the six points per model run represent the average of the two parameters for each SST regime. The error bars in both the x- and y-directions indicate 1 standard deviation for precipitation rate and $\delta D_p$, respectively. $\delta D_p$ is compared to average values of E for each SST regime (colors) of the control run (right).
2.3 Model Results

Table 2.1: Control Run and Uniform $\delta D_v^{ref}$ Test Values

<table>
<thead>
<tr>
<th>SST</th>
<th>$\delta D_{SfcEvap}$</th>
<th>Evap</th>
<th>BL</th>
<th>$\delta D_p$</th>
<th>Prec</th>
<th>Conv Vap</th>
<th>$\alpha_{eff}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>301.15K</td>
<td>-25%</td>
<td>3.1 mm/day</td>
<td>-93%</td>
<td>-23%</td>
<td>3.0 mm/day</td>
<td>—</td>
<td>1.08</td>
</tr>
<tr>
<td>301.65K</td>
<td>12%</td>
<td>3.6 mm/day</td>
<td>-116%</td>
<td>-50%</td>
<td>6.1 mm/day</td>
<td>-160%</td>
<td>1.07</td>
</tr>
<tr>
<td>302.15K</td>
<td>29%</td>
<td>4.0 mm/day</td>
<td>-130%</td>
<td>-66%</td>
<td>8.9 mm/day</td>
<td>-152%</td>
<td>1.07</td>
</tr>
<tr>
<td>302.65K</td>
<td>33%</td>
<td>4.4 mm/day</td>
<td>-138%</td>
<td>-74%</td>
<td>11.4 mm/day</td>
<td>-147%</td>
<td>1.07</td>
</tr>
<tr>
<td>303.15K</td>
<td>35%</td>
<td>4.8 mm/day</td>
<td>-143%</td>
<td>-79%</td>
<td>14.2 mm/day</td>
<td>-144%</td>
<td>1.07</td>
</tr>
<tr>
<td>303.65K</td>
<td>32%</td>
<td>5.1 mm/day</td>
<td>-146%</td>
<td>-83%</td>
<td>16.1 mm/day</td>
<td>-142%</td>
<td>1.07</td>
</tr>
<tr>
<td>301.15K</td>
<td>-29%</td>
<td>3.1 mm/day</td>
<td>-91%</td>
<td>-20%</td>
<td>3.0 mm/day</td>
<td>—</td>
<td>1.08</td>
</tr>
<tr>
<td>301.65K</td>
<td>-27%</td>
<td>3.6 mm/day</td>
<td>-96%</td>
<td>-25%</td>
<td>6.1 mm/day</td>
<td>-21%</td>
<td>1.08</td>
</tr>
<tr>
<td>302.15K</td>
<td>-27%</td>
<td>4.0 mm/day</td>
<td>-98%</td>
<td>-26%</td>
<td>8.9 mm/day</td>
<td>-22%</td>
<td>1.08</td>
</tr>
<tr>
<td>302.65K</td>
<td>-29%</td>
<td>4.4 mm/day</td>
<td>-98%</td>
<td>-27%</td>
<td>11.4 mm/day</td>
<td>-22%</td>
<td>1.08</td>
</tr>
<tr>
<td>303.15K</td>
<td>-31%</td>
<td>4.8 mm/day</td>
<td>-99%</td>
<td>-27%</td>
<td>14.2 mm/day</td>
<td>-22%</td>
<td>1.08</td>
</tr>
<tr>
<td>303.65K</td>
<td>-31%</td>
<td>5.1 mm/day</td>
<td>-99%</td>
<td>-28%</td>
<td>16.1 mm/day</td>
<td>-22%</td>
<td>1.08</td>
</tr>
</tbody>
</table>

Table 2.1 Table includes $\delta D_v$ of surface evaporation, surface evaporation rate, $\delta D_v$ of boundary layer (lower 50mb), $\delta D_p$, precipitation rate, and $\delta D$ of converged vapor as well as the effective fractionation factor ($\alpha_{eff} = R_{\delta D_p}/R_{BL}$) for the control (rows 1-6) and uniform $\delta D_v^{ref}$ test (rows 7-12) averaged for each SST regime. Converged vapor is near-zero for the first SST regime, and thus no $\delta$-value has been calculated.
with an idealized representation of the large-scale circulation, one cannot expect to replicate observations exactly.

### 2.3.2 Evaporation Test

Decreased rain evaporation has been argued to be a key mechanism of the amount effect (e.g. Bony et al. (7), Dansgaard (21), Lee et al. (49), Risi et al. (82)) and here we test this theory in our model framework by disabling fractionation during rain evaporation (i.e., vapor evaporated from rain has the same isotopic content as the rain itself).

The blue points in Fig. 2.1 (left) show the output from this experiment and due to the restriction on the rain evaporation, the rain becomes slightly more depleted than the control run. We note that the change is small between the two runs and the difference decreases as the SST increases, which is expected with increased precipitation rates and thus decreased evaporation. This leads to a smaller difference in the $\delta D_p$, while the opposite is true in lower precipitation rate cases, where rain evaporation will be greater.

These small changes of $\delta D_p$ values from the control run as well as the fact that the amount effect is still present indicates that rain evaporation does not appear to contribute as greatly to the amount effect as has been previously assumed. Had the evaporation been an essential part of the amount effect, disabling this fractionation process would have more of an impact on the amount effect.

### 2.3.3 Uniform $\delta D^\text{ref}_v$ Profile Test

This test demonstrates how the observed $\delta D_p$ signal results from the relative contributions of the various vapor sources of distinct isotopic contents. Here, the isotopic content of the environment, and thus converged vapor, is set to have the same $\delta D$ value as surface evaporation. The output from this test (green points in Fig. 2.1 (left)) indicates, as expected, that the amount effect essentially disappears, with $\delta D_p$ values oscillating around -24‰ despite increases in the precipitation rate.
2.4 Discussion

2.4.1 Amount Effect and Moisture Convergence

We present an interpretation that explains the amount effect when considering the total water (vapor plus condensate) budget of the precipitating column. Here, surface evaporation and large-scale moisture convergence are the two sources of atmospheric water, and convergence of vapor is most important in determining the value of $\delta D_p$ as convection strengthens. This interpretation is in accordance with Kurita (43) who finds that an increasing fraction of stratiform (as opposed to convective) precipitation (e.g. Houze (32)) is associated with more depleted rainfall. As the inflow to stratiform regions occurs mainly in the mid-troposphere (Mapes and Houze (61)) where the vapor is more depleted than in the boundary layer, moisture import to these regions will reinforce the amount effect.

The moisture import could potentially alter the influence of the BL vapor because of variations in the isotopic content of the converged vapor, due to changes in the structure of the vertical velocity profile. Back and Bretherton (4) analyzed data from ERA40 and found that vertical velocity profiles in the East Pacific tend to be bottom heavy while in the West Pacific they are top heavy. In the present study, the large-scale vertical velocity $\bar{w}$ peaks in the upper troposphere for all but the coldest SST (Appendix Fig. A.2), so that moisture convergence occurs over a deep layer. Since the profile of $\delta D_v$ becomes more depleted with height, the level of strongest convergence will influence the composition of the converged vapor; however comparing how $\delta D_p$ changes according to different vertical velocity profiles is outside the scope of this paper.

2.4.2 Amount Effect and Moisture Residence Time

We define a budget parameter, $\frac{E}{P}$, to correlate with $\delta D_p$, similar to work by Lee et al. (48). The distinction between $\frac{E}{P}$ and residence time (Aggarwal et al. (1)) is important for paleoclimate studies that use the correlation between $\delta D_p$ and such parameters to estimate regional and local precipitation patterns. Held and Soden (29) found that total precipitable water scales with the Clausius Clapeyron (CC) equation and varies by 7% per degree Kelvin change, while P (and therefore E) is much weaker than the CC scaling and only changes about 2% per degree Kelvin.
The value of $1-\frac{E}{P}$ will indicate the relative contribution of moisture convergence (and horizontal advection of upstream air into the domain, which is neglected here) in the steady state system. When $P=E$, converged vapor makes no net contribution to surface precipitation, which appears to be the case for the first SST regime (blue point in Fig. 2.1 (right)). If $P \gg E$, this indicates converged vapor is the primary source for precipitation. Performing the calculations with the output from the control run, we find that as the SST increases and convection becomes stronger, $\frac{E}{P}$ decreases, implying that converged vapor is making up the bulk of the precipitation.

### 2.4.3 Amount Effect and Boundary Layer Vapor

Table 1 includes the $\delta D$ values of the various vapor sources and the precipitation for the control run and uniform $\delta D^{\text{ref}}$ test, respectively. Across the range of SSTs and precipitation rates in our experiments, the isotopic ratios of the boundary layer vapor and precipitation change together, with an effective fractionation factor $\alpha_{\text{eff}} = \frac{R_p}{R_{bd}} = 1.07 - 1.08$ that is roughly fixed and close to the equilibrium fractionation factor $\alpha_{\text{equil}} = 1.08 - 1.09$ (Majoube (60)) at temperatures within the boundary layer. Appendix Fig. A.3 depicts the relationship between $\delta D_{bd}$ and $\delta D_p$ across the range of SSTs in our experiment. Kurita (43) found a similar relationship in observations of the isotopic content of surface vapor and precipitation, as the fit in his Fig. 8 corresponds to $\alpha_{\text{eff}} = \frac{R_p}{R_{vap}} = 1.07 - 0.25 \times (R_{vap} - 0.9) \sim 1.05 - 1.08$ for the value of $R_{vap}$ in that figure. Note that in our experiment, there is more scatter and a different slope in the relationship between $R_{vap}$ and $R_{bd}$ for a given SST and precipitation regime than is found across the whole range of SSTs and precipitation rates.

The relative roles of post-condensation exchange, downdraft recycling and environmental subsidence in maintaining the relationship between $R_{bd}$ and $R_p$ across the wide range of precipitation rates is secondary to the role of converged vapor in maintaining steady state and is thus beyond the scope of this study but is planned for future work.

### 2.5 Conclusions

The goal of this study has been to demonstrate that increasing convergence of water vapor is the key contributor to the observed amount effect in steady state. While a range
of processes, such as entrainment, condensation, evaporation, equilibration, are involved in producing the isotopic composition of rain, the moisture budget perspective offers a convenient overall constraint that the system must adjust to satisfy when in steady state and that offers a simple interpretation of the amount effect. We have shown that the decrease in rain evaporation and equilibration, will contribute to the depletion, but the role of these processes is small compared to convergence of vapor in a steady state scenario. Results do suggest a secondary role of the boundary layer vapor, whose isotopic composition is found to be well-correlated with that of precipitation across the wide range of precipitation rates in this study. However, the precise role of the boundary layer and convective processes in maintaining this relationship is outside the scope of this study.

We have also proposed the parameter $\frac{E}{P}$ for comparison with $\delta D_p$. This variable indicates the strength of the hydrological cycle as well as provides information about the local water budget, making it possible to diagnose the relative contributions of vapor sources for precipitation in steady state. Knowing where moisture is coming from is important since a change in the moisture sources of convection could drastically alter the value of $\delta D_p$.

The results presented support our hypothesis, though there are some caveats that should be addressed further in future work. As horizontal advection was neglected during our budget calculations, we intend to include this term in future work with an isotope-enabled version of the Super-Parameterized Community Atmosphere Model (SP-CAM). Also, our results only address steady state, and therefore represent the amount effect on monthly and seasonal timescales. In some instances, the amount effect has been observed on the timescale of single storms (e.g. Dansgaard (21), Njitchoua et. al. (72), Yoshimura et. al. (108)), which the current results do not address. The arguments presented here could be applicable to individual storms, but further testing is required to confirm this.
Microphysical controls on the isotopic composition of wintertime orographic precipitation

This chapter has been submitted as a research article to the Journal of Geophysical Research and is currently in review.

The sensitivity of mixed-phase orographic clouds, precipitation and their isotopic content to changes in dynamics, thermodynamics and microphysics is explored in idealized simulations of two-dimensional flow over a mountain barrier. These simulations use the Weather Research and Forecasting (WRF) Model with stable water isotopologues (HDO and H$_2^{18}$O), which have been integrated into the Thompson microphysics scheme within WRF as part of the present project. In order to understand how the isotopic composition of precipitation ($\delta^{18}O_{\text{precip}}$) is fixed, the mountain height, temperature, and the prescribed cloud droplet number concentration (CDNC) have been varied in a series of simulations. For the given range of values explored in this work, changes in mountain height and temperature induce stronger changes in domain-averaged $\delta^{18}O_{\text{precip}}$ than to changes in CDNC by a factor of approximately 10. The strongest response to changing CDNC in the present study leads to local variations of $\delta^{18}O_{\text{precip}}$ of about $3\%_v$, though those occur in regions of weak precipitation ($<0.1$ mm hr$^{-1}$) and would likely not be visible in snow accumulated over multiple storms. Changes in $\delta^{18}O_{\text{precip}}$ can be understood through the microphysical pathways by
which precipitable hydrometeors are formed and by the isotopic signature associated with each pathway. The decrease in $\delta^{18}O_{\text{precip}}$ with increasing mountain height is not a simple function of decreasing temperature, but also reflects the changing contributions from riming of cloud liquid and vapor deposition onto snow, the leading sources of precipitating hydrometeors in these simulations. These two processes have distinct isotopic signatures with a $\delta^{18}O$ difference of 3–8‰.

3.1 Introduction

Precipitation that forms due to interaction with mountain barriers, or orographic precipitation, is an important contributor to surface water resources. In particular, runoff from rainfall and, more notably, melting of the mountain snowpack feed into river basins that provide water to a number of heavily populated regions. As the amount and location of precipitation on the mountain barrier will determine the volume of runoff and into which drainage basin it flows, understanding all of the factors that influence and control orographic precipitation is essential for current and future forecasts of this necessary resource.

There have been extensive regional studies regarding the formation and behavior of orographic precipitation (e.g. Hobbs (30), Smith et al. (96), Smith and Evans (97), Zubler et al. (109)). 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 cloud microphysics (e.g. Colle (16), Galewsky (25), Muhlbauer and Lohmann (70)). In terms of cloud microphysics, the different pathways through which precipitating hydrometeors grow, can be more or less efficient and thus greatly influence the amount of precipitation. For example, in mixed-phase orographic clouds, the growth and fallout of snow and graupel may be enhanced by the “seeder-feeder” mechanism (Reinking et al. (81)), 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 of the cloud (Borys et al. (10), Mitchell et al. (65)), thereby augmenting precipitation on the windward side of the mountain at the expense of the transport of hydrometeors to the leeward slope (and the resulting precipitation there, which is known as “spillover”).
The stable isotopologues of water (H$_2^{16}$O, HDO, H$_2^{18}$O) have been used in precipitation analysis dating back to the initial work of Dansgaard (20). In the mid-latitudes, where westerlies impinging on north-south oriented mountain ranges form the motivation for our idealized simulations, the isotopic composition of precipitation is primarily temperature-dependent (e.g. Dansgaard (21), Jouzel (36), Lee et al. (48), Noone and Simmonds (73)), such that the ratio of the heavy (e.g., H$_2^{18}$O) to light (H$_2^{16}$O) isotopes correlates positively with temperature. In mountainous regions, the relationship between the isotopic composition of precipitation and temperature is additionally linked with altitude (Dansgaard (21)). 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 precipitation enriched in heavy isotopes 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 peak (Smith et al. (96)). This isotopic gradient was connected to the fractional removal of water by a mountain barrier and its drying ratio by Smith et al. (96). The relationship between isotopic composition and altitude has also been used to relate paleoclimate proxies for the isotopic composition of precipitation to past mountain elevation (Poage and Chamberlain (75), Rowley et al. (87)). However, as demonstrated by Galewsky (26) and Lechler and Galewsky (47), in different dynamical regimes, the airflow over the mountain can complicate the relationship between the isotopic composition of precipitation and the altitude of a mountain barrier.

In addition to the dynamical influences on orographic precipitation and its isotopic content, microphysical processes can also modify the isotopic signature of precipitation. Coplen et al. (17) connect variations in the isotopic content of precipitation in landfalling extratropical cyclones with changes in the storm structure and different pathways of precipitation formation. Observations of snowfall from the Sierra Nevada (Demoz et al. (22)) and snowfall and cloud liquid in Colorado (Lowenthal et al. (55)) suggested that the isotopic composition of snowfall is influenced by the degree of riming. By sampling both the isotopic and chemical composition of both snowfall and cloud droplets at a mountaintop site in Colorado, Lowenthal et al. (55) related the degree of riming of snowfall 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
3.1 Introduction

to the cloud droplets. This relationship was then employed to make predictions about the altitude at which snow formed through vapor deposition.

The relative role of riming in mixed-phase orographic precipitation can be reduced by decreasing temperature, through the glaciation of liquid clouds, and by increasing aerosol concentrations, which tend to lead to more numerous and smaller cloud droplets that are less likely to be collected by falling snow (Pruppacher and Klett (77), Wang and Ji (103)). Increased aerosol concentrations can also suppress or delay the formation of precipitation in liquid-only clouds by reducing the efficiency of collision and coalescence processes (Albrecht (2), Ramanathan et al. (78)). While aerosols have the potential to impact individual microphysical processes that contribute to precipitation, their influence on the amount and distribution of orographic precipitation has not been definitely established and appears to depend strongly on the environmental conditions of the region being considered (Borys et al. (9, 10), Khain and Pokrovsky (39), Lynn et al. (57), Muhlbauer and Lohmann (70), Saleeby et al. (93)). In a study of warm (liquid-only) orographic clouds and precipitation, Miltenberger et al. (64) suggest that interactions between dynamical and microphysical processes can lead to regimes where the precipitation and the precipitation efficiency are insensitive to changes in cloud droplet number concentration (CDNC), which is used as a proxy for aerosol concentrations.

To explore how such changes in microphysical processes influence mixed-phase orographic precipitation and its isotopic content in idealized two-dimensional simulations, we vary mountain height, temperature and CDNC. Particular attention is paid to changes in the microphysical processes that contribute to the growth of precipitating hydrometeors and how those processes and their isotopic signatures help control the amount, distribution and isotopic content of precipitation in these experiments. By tracking the isotopic content associated with precipitation growth processes within these simulations, we determine if each microphysical process has a distinct isotopic signature and how each process contributes to the overall isotopic signal of precipitation in different regimes.

Although the results presented in this paper are based on idealized simulations, they represent a step towards constructing an isotope-enabled regional modeling capability for WRF. Previous work with isotope-enabled global models (e.g. Field (24), Lee et al. (48), Noone and Simmonds (73), Vuille et al. (102)) has advanced our knowledge of how large-scale processes affect isotopic composition. However, the limitations of global models in
representing topography and cloud-scale processes leaves room for higher-resolution regional models that more faithfully represent such fine-scale phenomena (Prein et al. (76)). The work of Pfahl et al. (74) provides an example of how a fine-scale regional model can improve the representation of isotopic signals over that of a global model, and the present work aims to understand orographic precipitation by focusing on scales much finer than those represented in a typical isotope-enabled global model.

3.2 Model and Experiments

3.2.1 Model Setup

To conduct the orographic precipitation experiments, we use the WRF model version 3.5.1 (Skamarock and Klemp (94)) 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 2 km spacing in the horizontal direction and 105 vertical levels spaced 25-200 m in the lower 5 km and constant grid spacing above. The duration of each simulation is twelve hours. The Thompson microphysics scheme (Thompson et al. (100)) within WRF was chosen for this work because a study of wintertime precipitation in a mountainous region of the western United States found that, along with one other scheme, the Thompson microphysics scheme “clearly outperformed” the other microphysics schemes considered (Liu et al. (53)). In addition, the Thompson scheme includes a detailed treatment of the riming of cloud droplets by snow (as in Saleeby and Cotton (90)), which has proved important for realistic simulation of the effects of pollution on riming in mixed-phase orographic clouds (Lohmann (54), Saleeby and Cotton (90), Saleeby et al. (92)).

As noted previously, different atmospheric regimes are created in order to study the response of model microphysics. The Thompson microphysics scheme allows the user to specify the CDNC value, which is utilized in this work as a proxy for aerosols. The chosen CDNC values represent conditions that range from pristine to polluted. The working assumption is that for high aerosol loading, there are more cloud condensation nuclei (CCN) and thus a higher CDNC value, while a lower CDNC value indicates a scenario with few aerosols and therefore fewer CCN. More specifics about the experimental setup are given in section 2.3.
3.2 Model and Experiments

3.2.2 Isotopic implementation

In its default configuration in the WRF model, the Thompson scheme only treats microphysical transfers of the standard isotopologue of water (H$_{16}$O) among water vapor and the different hydrometeors included in the scheme: cloud liquid, rain, cloud ice, snow and graupel. As part of the present project, we have extended the Thompson scheme so that the microphysical transfers of the stable isotopologues of water (HDO and H$_{18}$O) are also included. The isotopic composition of water vapor and each hydrometeor is tracked, and the exchanges of the heavy isotopologues of water are accounted for during each microphysical process represented in the Thompson scheme. Isotopic 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 that involve the transfer of whole hydrometeors from one category to another (e.g., freezing, melting, riming), occur without fractionation. As in Bony et al. (7), Blossey et al. (6) and Pfahl et al. (74), the sublimation of ice phase hydrometeors (snow, cloud ice, graupel) is also assumed to occur without fractionation, so that the vapor produced by sublimation of snow, for example, has the same isotopic composition as the snow. While sublimation is expected to produce vapor from the outer shell of an ice phase hydrometeor, and this layer may not have the same isotopic composition as the particle as a whole, tracking the composition of individual layers within the crystals is judged to be too complicated and expensive to include in the present implementation. Note that the implementation, which follows Blossey et al. (6), includes few approximations in its representation of isotopic exchanges beyond the assumption that the isotopic composition of each hydrometeor category in a given grid cell is uniform and that no fractionation occurs during sublimation. 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 in time explicitly by the model. A more detailed description of the water isotope physics is given in Appendix B.

The quality of the isotopic simulation depends strongly on the representation of the standard isotopologue of water. If the microphysics scheme and the broader model do a poor job in representing the amount and distribution of precipitation of the standard isotopologue of water, this will be reflected in the isotopic composition as well. Encouraged by the
performance of the Thompson scheme within WRF on wintertime orographic precipitation \cite{Liu2021} and by the representation of isotopic composition in tropical convection in a similar implementation of water isotopologues in \cite{Blossey2018}, we proceed with the simulations here.

For the isotopic analysis, our results on H$_2^{18}$O are presented 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.

### 3.2.3 Experimental Setup

In order to address the robustness of an isotopic signal, several experiments are conducted that alter the initial temperature profile, mountain height (800, 1500 and 3000 m), and the CDNC (25, 100, 200, 400, and 800 cm$^{-3}$). Two initial temperature profiles are used here and are referenced as warm or cold based on the surface temperature of the upstream sounding ($T_{\text{sfc}} = 7$°C and 0°C, respectively). Experiments are referenced by abbreviations (e.g., W800m), which indicate the temperature sounding (W=warm or C=cold) and mountain height settings. The setup and initial conditions, including the temperature profiles, are similar to those in \cite{Muhlbauer2020}, with a mountain half-width of 20 km and a horizontal wind profile that is a constant 15 m s$^{-1}$ below 10 km and linearly increases to 40 m 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 represents 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. As the cold sounding is drier than the warm sounding, it is also more depleted, such that the $\delta^{18}O$ of vapor at the surface is 8‰ less than that of the warm sounding. Neither liquid nor ice condensate exists initially, and therefore their isotopic compositions do not need to be initialized.
3.2 Model and Experiments

3.2.4 Model Validation

The model’s ability to simulate orographic clouds and precipitation is on par with previous studies. The results are very similar to the WRF simulations in Muhlbauer et al. (71), despite the use of a different microphysical scheme. There are some small differences in the simulated orographic clouds, and our experiments produce more accumulated precipitation. However, these deviations can be attributed to our implementation of a longer simulation time and a larger range of CDNC values in addition to the choice of microphysical scheme. The changes in the liquid orographic cloud are also similar to results seen by Xiao et al. (107), who also used the same idealized WRF setup, but coupled with a detailed bin microphysics scheme and a warmer initial temperature profile. The evolution of cloud liquid and microphysical processes as CDNC increases in our cold experiments, is similar to that of Saleeby et al. (89), who used the Colorado State University - Regional Atmospheric Modeling System with a different microphysics scheme, to simulate realistic wintertime orographic clouds in northern Colorado.

This project represents the first use of this isotope-enabled version of the Thompson scheme within WRF. As the present modeling study is idealized and the incorporation of water isotopologues into the real-case forecasting capability of WRF is not complete, we focus on the performance of the scheme within the present simulations. First, the isotopic composition of precipitation along the upslope of the mountain approximately conforms to that of a Rayleigh process and is slightly more depleted than the Rayleigh process due to dynamical effects of the mountain (Galewsky (26)) and the formation of precipitation from more depleted vapor above the surface of the mountain (Fig. 3.1). Second, in section 3.3.3, closed budgets for the surface precipitation are constructed that explain the precipitation itself and its isotopic composition of precipitation in terms of the various microphysical processes that contribute to the formation and growth of precipitating hydrometeors. Last, the isotopic composition of water vapor and hydrometeors described in sections 3.3.1 and 3.3.4 shows the expected influence of microphysical processes on isotopic composition, from isotopic equilibration of cloud liquid and water vapor to vapor deposition onto ice and the evaporation of rain in sub-saturated conditions.

It should also be noted that the average $\delta^{18}O$ values of total precipitation compare well with observations in conditions similar to those used here for initial conditions. Anderson
Figure 3.1: $\delta^{18}O_p$ Mountain and Rayleigh Profiles - Profiles comparing the $\delta^{18}O$ of precipitation for all three mountain heights (800, 1500, and 3000 m) in the warm (left) and cold (right) cases to the Rayleigh distillation model (blue). The 100 cm$^{-3}$ simulations were used in these calculations, though the results were similar with other cloud droplet number concentration cases.
et al. (3) calculated the average $\delta^{18}O$ of snowpack using the Isotopes in Rocky Mountain Snowpack (IRMS) database, and found that values ranged between $-10^\circ$ and $-25^\circ$, which compares well with the range of $\delta^{18}O$ in the average precipitation for our experiments (approximately $-10^\circ$ to $-16^\circ$ in warm simulations and $-19^\circ$ to $-26^\circ$ in cold experiments). Comparable $\delta^{18}O_{\text{precip}}$ values ($-12^\circ$ to $-24^\circ$) were measured during a 1985 March storm in Kingvale, CA, which is located upwind of the Sierra Nevada crest at an elevation of 1859 m (Warburton et al. (105)). Warburton and DeFelice (104) analyzed samples in the Central Sierra Nevada, and found that snow formed through vapor deposition had a $\delta^{18}O$ signature that ranged from $-18.4^\circ$ to $-22.9^\circ$, which corresponds well with our cold temperature profile experiments (see further discussion in section 3). The snow samples from the same study that indicated growth by a combination of riming and vapor deposition, were less depleted and ranged between $-6.4^\circ$ and $-16.8^\circ$, which resembles results in our warm simulations (see section 3). Values similar to Warburton and DeFelice (104) were measured in Colorado by Lowenthal et al. (55) 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 $-15.6^\circ$ and $-20.4^\circ$.

3.3 Model Results

3.3.1 Reference Simulation

To outline the general characteristics of the cloud and precipitation in these simulations, the simulation with the warmer sounding ($T_{\text{sfc}} = 7^\circ$C), a 800 m high mountain, and a cloud droplet number concentration (CDNC) of 200 cm$^{-3}$ is chosen as the reference simulation. Fig. 3.2b shows the average simulated mass of cloud liquid along with the combined mass of cloud ice and snow for the reference simulation. (Figs. 3.2a and c will be discussed in section 3.3.2.) The figure combines cloud ice and snow together, as the setup of the Thompson scheme quickly leads to the conversion of cloud ice to snow, and as a result, produces little cloud ice (Thompson et al. (100)). 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, where almost all precipitation is produced.
3.3 Model Results

Figure 3.2: Orographic Liquid and Ice Clouds - Contoured temperature (red) and mixing ratios of cloud liquid water (shaded) and combined cloud ice/snow (color contours) for the 800 m warm temperature simulations, averaged over the last four hours of each simulation. 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.

The orographic cloud in the reference simulation does not extend higher than 4 km and has a much higher mass of cloud liquid than a combined mass of snow and ice (Fig. 3.2b). Further, there is much less combined cloud ice/snow than cloud liquid in the reference simulation. The frozen hydrometeor mass is composed mainly of snow and occurs predominately upstream of the mountain peak with only a little spillover (~10 km) 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.

The isotopic values of vapor, cloud liquid, rain and ice/snow for the reference simulation are presented in Fig. 3.3. Cloud liquid isotopic values range from approximately -7‰ near the mountain surface to -22‰ at cloud top (Fig. 3.3c). Isotopic equilibrium is enforced between cloud liquid and vapor, so that decreasing $\delta^{18}O$ of cloud liquid with height is expected given that the vapor $\delta^{18}O$ shows the same trend (Figs. 3.3a and c). 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. As expected from rainout, the progressive removal of heavy isotopologues by precipitation across the mountain barrier (Clark and Fritz (15), Smith et al. (96)), water vapor does show asymmetry about the mountain and is more depleted at low levels further downstream of the mountain. The cloud ice/snow $\delta^{18}O$ values range from -10‰ near cloud base to -35‰ at cloud top (Fig. 3.3d). The most depleted ice has a $\delta^{18}O$ value around -20‰, 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.

Figure 3.3: Reference Case $\delta^{18}O$ of Hydrometeors and Vapor - Average over the last four hours of $\delta^{18}O$ of (a) vapor, (b) rain, (c) cloud liquid and (d) ice/snow for reference simulation with a CDNC of 200 cm$^{-3}$.

Fig. 3.4a shows the profile of accumulated precipitation across the mountain for the reference simulation along with a number of different CDNC concentrations. (The sensitivity to CDNC will be discussed in the following section.) The precipitation for the reference simulation (CDNC = 200 cm$^{-3}$, red line) peaks over the mountain top and is nearly symmetric, with slightly more precipitation falling downwind of the peak and a spillover ratio of 0.56 (Tab. 1). Most of the precipitation falls as rain at the surface, with similar, smaller amounts of snow and graupel (Tab. 1). The isotopic composition of the accumulated precipitation $\delta^{18}O_{\text{precip}}$ in the reference simulation (Fig. 3.4b, red line) becomes more enriched as one ascends the lower slope on the upwind side of the mountain. This is also seen in the isotopic composition of rain in Fig. 3.3b, and is associated with a shift from rain resulting from the melting of snow that was formed aloft through vapor deposition, to rain and snow that grew through the conversion and accretion of cloud liquid. Such changes in the microphysical pathways through which precipitating hydrometeors are formed, and their impact on
\( \delta^{18}O_{\text{precip}} \) will be discussed in greater detail in section 3.3.3. Following this peak in \( \delta^{18}O_{\text{precip}} \) at \( x=280 \) km, the isotopic composition of precipitation falls off across the mountain as the heavier isotopes are removed preferentially through fallout. As noted in the introduction, this may be modeled approximately as a Rayleigh process (Smith et al. (96)), though there are some complications due to dynamical response to topography (Galewsky (26)) and microphysical effects. The increase in \( \delta^{18}O_{\text{precip}} \) on the downslope at \( x=315-320 \) km is associated with the fractionation of evaporating rain once it passes downstream of the orographic cloud (see also Fig. 3.3b-c.) As shown in previous studies (e.g. Bony et al. (7), Lawrence et al. (44), Risi et al. (82), Stewart (99)), evaporation in subsaturated conditions tends to enrich the rain and deplete the vapor, as the lighter H\(_2\)O will more quickly move from the liquid to the surrounding vapor.

### 3.3.2 Sensitivity to CDNC

Next, the sensitivity of the reference simulation to changes in CDNC (as a proxy for aerosol variations) is shown. This is interesting both as a way to understand whether aerosol impacts on orographic precipitation (e.g. Rosenfeld et al. (86)) could impact the isotopic composition as well, and as an example of how changing the microphysical processes which contribute to precipitation, could impact the amount, distribution and isotopic composition of orographic precipitation.

Three cases with increasing values of CDNC are shown in Fig. 3.2, which illustrate potential changes in the orographic cloud with CDNC. As the CDNC value increases, the conversion of cloud to rain and the riming of cloud liquid by snow become less efficient, resulting in an increase in both the amount of cloud liquid and the leeward region it spans. While there is already leeward spillover of cloud liquid in the 25 cm\(^{-3}\) case, cloud liquid extends an additional 15 km down the leeward side in the 800 cm\(^{-3}\) experiment, with the region of maximum mass mixing ratio (red filled contours) also reaching approximately 5 km further downstream. This shift in the leeward extent of cloud liquid is mirrored in the isotopic composition of precipitation in Fig. 3.4b, where the increase in \( \delta^{18}O_{\text{precip}} \) due to rain evaporation occurs farther downstream as CDNC increases. On the windward slope, the location of the leading edge of the cloud does not change in all of the warm 800 m experiments. Increases in CDNC have little impact on the location of snow and cloud ice:
Figure 3.4: Accumulated Precipitation and $\delta^{18}O_{\text{precip}}$ - (a) Precipitation accumulated over the 12 hours of the simulation and (b) the associated $\delta^{18}O_{\text{precip}}$ for the reference simulation and its sensitivity to CDNC changes. The mountain peak is located at 300 km.
### Table 3.1: Precipitation Breakdown of Reference and Sensitivity Experiments

<table>
<thead>
<tr>
<th>Case</th>
<th>CDNC</th>
<th>Total Prec</th>
<th>Snow</th>
<th>Graupel</th>
<th>Rain</th>
<th>Spillover</th>
</tr>
</thead>
<tbody>
<tr>
<td>W800m</td>
<td>25</td>
<td>38.6mm</td>
<td>7.8%</td>
<td>7%</td>
<td>85.2%</td>
<td>37%</td>
</tr>
<tr>
<td>W800m</td>
<td>100</td>
<td>32.1mm</td>
<td>13.5%</td>
<td>12.9%</td>
<td>73.6%</td>
<td>59%</td>
</tr>
<tr>
<td>W800m</td>
<td>200</td>
<td>28.5mm</td>
<td>14.3%</td>
<td>13.7%</td>
<td>72%</td>
<td>56%</td>
</tr>
<tr>
<td>W800m</td>
<td>400</td>
<td>23.7mm</td>
<td>14.8%</td>
<td>14.2%</td>
<td>70.8%</td>
<td>62%</td>
</tr>
<tr>
<td>W800m</td>
<td>800</td>
<td>18.3mm</td>
<td>13.8%</td>
<td>13%</td>
<td>73.1%</td>
<td>68%</td>
</tr>
<tr>
<td>C800m</td>
<td>25</td>
<td>34.2mm</td>
<td>92.2%</td>
<td>2.6%</td>
<td>5.2%</td>
<td>45%</td>
</tr>
<tr>
<td>C800m</td>
<td>100</td>
<td>32.2mm</td>
<td>96.3%</td>
<td>3%</td>
<td>0.7%</td>
<td>50%</td>
</tr>
<tr>
<td>C800m</td>
<td>200</td>
<td>30.9mm</td>
<td>96.9%</td>
<td>2.9%</td>
<td>0.2%</td>
<td>53%</td>
</tr>
<tr>
<td>C800m</td>
<td>400</td>
<td>29.3mm</td>
<td>97.5%</td>
<td>2.5%</td>
<td>–</td>
<td>56%</td>
</tr>
<tr>
<td>C800m</td>
<td>800</td>
<td>26.6mm</td>
<td>98.1%</td>
<td>1.9%</td>
<td>–</td>
<td>61%</td>
</tr>
<tr>
<td>W1500m</td>
<td>25</td>
<td>113.6mm</td>
<td>60.9%</td>
<td>7%</td>
<td>32.1%</td>
<td>31%</td>
</tr>
<tr>
<td>W1500m</td>
<td>100</td>
<td>110.5mm</td>
<td>64.1%</td>
<td>11%</td>
<td>24.9%</td>
<td>33%</td>
</tr>
<tr>
<td>W1500m</td>
<td>200</td>
<td>106.6mm</td>
<td>65.8%</td>
<td>13%</td>
<td>21.2%</td>
<td>35%</td>
</tr>
<tr>
<td>W1500m</td>
<td>400</td>
<td>104.8mm</td>
<td>67%</td>
<td>13.7%</td>
<td>19.3%</td>
<td>36%</td>
</tr>
<tr>
<td>W1500m</td>
<td>800</td>
<td>105.3mm</td>
<td>67%</td>
<td>14%</td>
<td>18.9%</td>
<td>37%</td>
</tr>
<tr>
<td>C1500m</td>
<td>25</td>
<td>91.3mm</td>
<td>96.1%</td>
<td>0.5%</td>
<td>3.4%</td>
<td>32%</td>
</tr>
<tr>
<td>C1500m</td>
<td>100</td>
<td>92.6mm</td>
<td>98.3%</td>
<td>0.6%</td>
<td>1.1%</td>
<td>33%</td>
</tr>
<tr>
<td>C1500m</td>
<td>200</td>
<td>90.7mm</td>
<td>98.7%</td>
<td>0.7%</td>
<td>0.6%</td>
<td>34%</td>
</tr>
<tr>
<td>C1500m</td>
<td>400</td>
<td>91.7 mm</td>
<td>99.1%</td>
<td>0.7%</td>
<td>0.2%</td>
<td>35%</td>
</tr>
<tr>
<td>C1500m</td>
<td>800</td>
<td>87.3mm</td>
<td>99.4%</td>
<td>0.6%</td>
<td>–</td>
<td>38%</td>
</tr>
<tr>
<td>W3000m</td>
<td>25</td>
<td>235.8mm</td>
<td>67.8%</td>
<td>3.2%</td>
<td>29%</td>
<td>27%</td>
</tr>
<tr>
<td>W3000m</td>
<td>100</td>
<td>237.8mm</td>
<td>69.1%</td>
<td>3.8%</td>
<td>27.1%</td>
<td>27%</td>
</tr>
<tr>
<td>W3000m</td>
<td>200</td>
<td>233.9mm</td>
<td>71%</td>
<td>4.3%</td>
<td>24.7%</td>
<td>28%</td>
</tr>
<tr>
<td>W3000m</td>
<td>400</td>
<td>232.1mm</td>
<td>72.1%</td>
<td>4.8%</td>
<td>23.1%</td>
<td>29%</td>
</tr>
<tr>
<td>W3000m</td>
<td>800</td>
<td>227.9mm</td>
<td>74%</td>
<td>5.3%</td>
<td>20.7%</td>
<td>31%</td>
</tr>
<tr>
<td>C3000m</td>
<td>25</td>
<td>182.3mm</td>
<td>97.3%</td>
<td>0.5%</td>
<td>2.2%</td>
<td>20%</td>
</tr>
<tr>
<td>C3000m</td>
<td>100</td>
<td>183.2mm</td>
<td>98.7%</td>
<td>0.3%</td>
<td>1%</td>
<td>21%</td>
</tr>
<tr>
<td>C3000m</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>C3000m</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>C3000m</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 3.1:** Breakdown of the major statistics for all runs. Columns indicate the case (height and cold (C) or warm (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.
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 isotopic signatures of cloud liquid and combined cloud ice/snow are similar to those of the reference simulation (Fig. 3.3) and are not shown.

As in the reference simulation, most of the precipitation in the simulations with varying CDNC falls as rain (see Tab. 1) with small, similar amounts of accumulated snow and graupel. The third column in Tab. 1 indicates that the accumulated precipitation decreases as CDNC increases, and is reduced by more than half between the 25 cm$^{-3}$ and 800 cm$^{-3}$ experiments. Fig. 3.4a shows that the location of the maximum precipitation shifts leeward as CDNC increases, which has been previously observed in wintertime orographic precipitation (Jirak and Cotton (35), Saleeby et al. (92)). Among the different mountain heights and temperatures considered here, the magnitude of the shift is strongest and most obvious for the W800m experiments, where there is a difference of approximately 10 km between the precipitation peaks in the 25 cm$^{-3}$ and 800 cm$^{-3}$ 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.

The domain can be broken down into three smaller regions: upstream of the peak (up to 290 km into the domain), around the peak (290–310 km) and downstream of the peak (310 km onwards). As Fig. 3.4a indicates, most of the precipitation falls in the first and second regions. The influence of CDNC on precipitation is also most pronounced in these regions. However, the CDNC impact on the $\delta^{18}$O$\text{precip}$ is slightly different. Fig. 3.4b illustrates that the largest $\delta^{18}$O$\text{precip}$ difference between simulations occurs in the first region, but over the second region, variation in the isotopic signal is small ($\leq 1\%$). In the third region, downstream of the mountain peak, the accumulated precipitation is relatively unchanged between simulations, but there is some separation in the $\delta^{18}$O$\text{precip}$ of approximately 2$\%$ at $x=310$ km before the effects of rain evaporation enter further down the lee slope. The slopes of $\delta^{18}$O$\text{precip}$ across the peak differ, with the steepest change in $\delta^{18}$O$\text{precip}$ across the peak in the simulation with the largest precipitation (CDNC=25 cm$^{-3}$) as one would expect due to the effect of rainout (Smith et al. (96)). This leads the 25 cm$^{-3}$ simulation to have the largest upstream-downstream difference in $\delta^{18}$O$\text{precip}$ around the peak ($x=290–310$ km).
3.3 Model Results

3.3.3 Microphysical pathways

To better understand the changes in precipitation and its isotopic composition across the mountain, we consider the budget for the total mass of precipitating hydrometeors (rain, snow and graupel combined) in these simulations, integrated in time and over the whole domain or a sub-region of the domain. Since isotopic composition is unchanged by exchanges between rain, snow 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 sublimation/evaporation. In this budget, surface precipitation, $P$, is a sink of hydrometeor mass and is balanced by various microphysical sources of rain, snow and graupel as well as advection and storage of these hydrometeors:

$$ P = Q_{LAUT} + Q_{LACC} + Q_{IAUT} + Q_{IACC} + Q_{RIM} + Q_{DEP} + Q_{SUB} + Q_{ADV} - Q_{STOR}. $$

Here, the sources of hydrometeor mass include microphysical processes, such as autoconversion of cloud liquid (LAUT), accretion of cloud liquid (LACC), autoconversion of cloud ice (IAUT), accretion of cloud ice (IACC), riming of cloud liquid (RIM), vapor deposition onto ice (DEP), sublimation of ice (SUB), along with those associated with advection (ADV) and storage (STOR). The storage term is negative because increases in hydrometeors in the domain over time come at the expense of surface precipitation. Each of these terms are integrated over the duration of the simulations and over the domain or a subset of the domain in the horizontal direction and then normalized by the mountain half-width (20 km). Note that, because the Thompson microphysical scheme produces little cloud ice, much of the vapor deposition onto ice phase hydrometeors that occurs in the domain 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. Also, note that the net tendency of vapor deposition (including deposition, sublimation and rain evaporation) has been averaged over the simulation and then partitioned into regions of deposition and sublimation/evaporation according to the sign of the mean tendency.

A similar budget can be written for the mass of the heavy isotopologues, and the isotopic
composition of those contributions can be computed from the ratio of the contribution to heavy isotope mass, e.g., \( H_2^{18}O \), to that for the standard isotope, \( H_2^{16}O \). The \( \delta^{18}O \) of hydrometeor mass generated by each process may then be computed as for precipitation itself, \( \delta^{18}O_{\text{precip}} \).

In Fig. 3.5, the total precipitation and the contributions of the dominant microphysical processes to precipitation and its isotopic composition are shown in three regions: the peak (x=295–305 km) and the regions upwind and downwind of the peak. In the following, the sources of precipitation in each region are analyzed. Note that the precipitation produced in each region may fall to the surface there or be transported downstream. In Figs. 3.5b–d, the contribution of each process in each region has been normalized by the total precipitation in the domain for each case. These normalized contributions can be interpreted as weights, which can be applied to the characteristic isotopic composition from each process to determine \( \delta^{18}O_{\text{precip}} \).

**Figure 3.5: Regional Breakdown of Accumulated Precipitation and \( \delta^{18}O_{\text{precip}} \)** - Breakdown of the reference simulation (a) accumulated precipitation over the whole domain and the regions upwind of the peak, over the peak and downwind of the peak, as well as the (e) the respective average \( \delta^{18}O \) values. The contributions from precipitation sources normalized by the total precipitation in the three subregions are shown in (b–d) and the corresponding isotopic signatures in (e–g). The accumulated precipitation and its sources are integrated over the domain and normalized by the mountain half-width (20 km). Sources are: autoconversion of cloud liquid (laut), accretion of cloud liquid by rain (lacc), autoconversion of cloud ice (iaut), riming of cloud liquid (rim), vapor deposition onto ice (dep), and sublimation of ice/evaporation of rain (sub).
In the upwind region (Fig. 3.5b), riming of cloud droplets contributes most to the growth of precipitating hydrometeors, with vapor deposition onto ice making the second largest contribution in most cases. The 25 cm\(^{-3}\) simulation differs in the importance of autoconversion and accretion of cloud liquid. Autoconversion of cloud liquid and riming both have a direct dependence on the size of cloud droplets and therefore on CDNC. Accretion of cloud liquid may also depend indirectly on CDNC if less rain is generated through autoconversion as CDNC increases. As autoconversion and accretion of cloud liquid fall off with increasing CDNC, the contribution from vapor deposition increases. The isotopic signatures of the liquid processes are more enriched than the vapor deposition by approximately 4-9‰ (Fig. 3.5f), and thus the precipitation in the lower CDNC simulations is more enriched than the higher CDNC simulations. Overall, the microphysics explain the decrease in both the accumulated precipitation and the isotopic content of the precipitation as seen in Fig. 3.4.

Riming of cloud liquid over the mountain peak is the largest source of precipitation in the three regions and itself produces enough hydrometeor mass to account for half of the surface precipitation in all cases except CDNC=25 cm\(^{-3}\) (Fig. 3.5c). Accretion of cloud liquid and vapor deposition onto ice also contribute to precipitation over the peak. Similar to the upwind region, accretion of cloud liquid decreases with increasing CDNC, though more modestly, but riming actually increases. In this region, vapor deposition onto ice is still the most depleted source term. However, the ice produced by vapor deposition is more enriched above the peak than in the upstream region. The average \(\delta^{18}O\) differences between the ice produced by riming and vapor deposition over the peak range between 2-4‰ (Fig. 3.5g). Therefore, the variation in the source terms of precipitation with CDNC over the peak produce little change in the \(\delta^{18}O_{\text{precip}}\) formed there, in part because the isotopic composition of the sources are more similar.

Precipitation production on the leeward slope derives predominantly from vapor deposition onto ice in addition to relatively small contributions from ice autoconversion and accretion of cloud liquid (Fig. 3.5d). One significant difference in this downwind region compared to the other two regions, is the presence of a large sink of precipitation mass caused by sublimation. This pocket of sublimation is expected due to subsidence and thus warming of air as it flows over the mountain peak. Though all of the microphysical source terms increase with CDNC in this region, removal of precipitation due to sublimation essentially balances out the source terms, and the accumulated precipitation remains relatively constant in the
different simulations (Fig. 3.4a). The smaller range of the $\delta^{18}O$ in the accumulated precipitation can be attributed to the similarity of the source terms in each simulation, except for that due to sublimation/evaporation which becomes more depleted with increasing CDNC.

### 3.3.4 Sensitivity to mountain height and temperature

The sensitivity of domain-integrated precipitation amount and its isotopic composition to CDNC changes was also studied for a number of mountain heights (800 m, 1500 m and 3000 m) and two temperature profiles (with $T_{sfc} = 0^\circ C$ and $7^\circ C$). To understand how precipitation and its isotopic content are related across these simulations, Fig. 3.6 shows their relationship when integrated over the whole domain (Fig. 3.6a) and over the regions upstream of the peak, over the peak and downstream of the peak (Figs. 3.6b–d, respectively). These regions are defined as above in section 3.3.3. As seen in Fig. 3.4, the response of total (domain-integrated) precipitation and its isotopic content to CDNC changes (where it exists) is modest in comparison to that due to mountain height and temperature. The only significant response of $\delta^{18}O_{\text{precip}}$ to CDNC occurs for small precipitation amounts ($<5$ mm) in the upwind region of W800m (Fig. 3.6b). Otherwise, the change in isotopic content due to temperature exceeds that due to CDNC by a factor of approximately 10 for the ranges of temperature and CDNC explored here. The weaker sensitivity of precipitation to CDNC changes with increasing precipitation is reminiscent of the work of Muhlbauer et al. (71) for mixed-phase clouds and Miltenberger et al. (64) for warm clouds. The possibility remains that a model setup that yields weaker precipitation might show a stronger sensitivity of precipitation to CDNC changes, as in Miltenberger et al. (64). However, the change in $\delta^{18}O_{\text{precip}}$ due to CDNC is unlikely to increase far beyond the range seen in the reference case (W800m).

Since the response to CDNC is weak in many cases, the present section focuses on a single CDNC value ($200$ cm$^{-3}$) across the range of mountain heights and temperatures to understand the responses to mountain height and temperature seen in Fig. 3.6. The changing configuration of the orographic cloud with mountain height and temperature is shown in Fig. 3.7. The orographic cloud produced in the C800m experiment is quite similar to that

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1 Each simulation is labeled according to its temperature and mountain height. For example, W800m denotes the reference case with “W” denoting the warmer sounding with $T_{sfc} = 7^\circ C$ and, in other runs, “C” the colder sounding with $T_{sfc} = 0^\circ C$. 

Figure 3.6: Scatter Plot of Precipitation vs. $\delta^{18}O_{\text{precip}}$ - Scatter plot of area-integrated precipitation vs $\delta^{18}O_{\text{precip}}$ for all mountain heights, temperatures and CDNC values. These quantities are presented both for the whole domain (a) and for regions upstream of the peak (b), over the peak (c) and to the lee of the peak (d). For each experiment, the CDNC value is indicated by the color and size of the symbol, while the mountain height and temperature are shown by the shape of the symbol. The grey solid line in (a) and (b) is the regression of the domain averages in the warm temperature experiments and the dashed line is the solid line shifted down by 8%. The regression is not shown in (c) and (d) because the isotopic composition over and downwind of the peak depends on the precipitation amount upstream.
3.3 Model Results

of the reference simulation in terms of vertical extent (Fig. 3.7d). However, the extent of the cloud and snow on the lee slope changes with cloud liquid ending closer to the peak and the cloud ice/snow reaching farther down the slope. For the higher mountain heights (Figs. 3.7b-c, e-f), the liquid cloud is shallower in the colder simulations while the snow has a similar vertical extent. These higher mountain heights also produce more ice/snow than the reference simulation, and in the cold temperature experiments, there is more cloud ice/snow than liquid. Note that the 3000 m mountain wave response depends on temperature, with the isotherms downstream of the mountain suggesting a stronger downslope flow in the colder simulation.

Figure 3.7: Orographic Clouds in Sensitivity Experiments - As in Fig. 3.2, but for the 200 cm$^{-3}$ simulations from the (a) 800 m, (b) 1500 m, (c) 3000 m warm temperature experiments and (d) 800 m, (e) 1500 m, and (f) 3000 m cold temperature experiments.

Fig. 3.8 shows the isotopic composition of water vapor, cloud liquid and combined cloud ice/snow for the three mountain heights with the colder temperature profile ($T_{sfc} = 0^\circ$C). Unlike in the reference simulation, these simulations have little rain, and its isotopic composition is not included. As noted in section 3.2.3, the water vapor at the surface upwind of the mountain (Figs. 3.8a–c) is 8% more depleted than that of the reference simulation (Fig. 3.3a). The water vapor isotopic composition becomes increasingly asymmetric for the higher mountains due to rainout (Smith et al. (96)), and the thin layer of downslope flow is visible in the water vapor isotopic composition for the 3000 m mountain (Fig. 3.8c). The
3.3 Model Results

cloud liquid is almost entirely confined to the upstream side of the mountain, and its isotopic content (Figs. 3.8d–f) is tied to the water vapor through the assumption of vapor-liquid isotopic equilibrium. The combined cloud ice/snow (Figs. 3.8g–i) is more depleted than cloud liquid at the same altitude, and this difference increases with mountain height. On the lee side of the mountain, the snow reaches to the base of the mountain in each case and becomes more depleted with mountain height, as the snow has formed from vapor that either originates at higher altitudes or has been depleted through precipitation.

As suggested by the sensitivity of total precipitation shown in Fig. 3.6, the distribution of precipitation and its isotopic content across the mountain changes much more substantially with mountain height and temperature than with CDNC (Fig. 3.9). The precipitation amount increases and shifts upstream with increasing mountain height, and δ¹⁸O of precipitation on the lee slope becomes more depleted with mountain height in agreement with the snow isotopic composition shown in Figs. 3.8g–i. The lee slope difference in δ¹⁸O of precipitation between the 800 m

Figure 3.8: Sensitivity Experiments’ δ¹⁸O of Hydrometeors and Vapor - Based on averages over the last four hours of each simulation, δ¹⁸O of (top) vapor, (middle) cloud liquid and (bottom) ice/snow for the (a,d,g) 800 m, (b,e,h) 1500 m, and (c,f,i) 3000 m cold temperature experiments with a CDNC of 200 cm⁻³.
and 3000 m mountain heights at x=315 km reaches 11% and 13% in the warm and cold simulations, respectively. Similar differences are seen in the precipitation integrated in the lee of the peak in Fig. 3.6d. The stronger dependence of $\delta^{18}O_{\text{precip}}$ on mountain height in the cold simulations, mirrors that seen in total precipitation and its isotopic content in Fig. 3.6, and suggest that the isotopic lapse rate, the change in $\delta^{18}O_{\text{precip}}$ with altitude, itself depends on temperature.

![Figure 3.9: Sensitivity Experiments’ Accumulated Precipitation and $\delta^{18}O_{\text{precip}}$ - Distribution of precipitation accumulated over the 12 hour simulations in (a) warm experiments and (b) cold experiments. Corresponding $\delta^{18}O_{\text{precip}}$ for (c) warm experiments and (d) cold experiments. All profiles are based on the 200 cm$^{-3}$ simulations. The mountain peak is located at 300 km. Note that the axis limits for $\delta^{18}O_{\text{precip}}$ have been shifted by 8% from the warm to cold simulations to account for the difference in the isotopic composition of vapor at the surface in the two cases.](image)

As in section 3.3.3, the relative contributions of different microphysical pathways to the formation of precipitating hydrometeors are shown in Fig. 3.11b to understand better the influence of mountain height and temperature on isotopic composition, which was seen in Fig. 3.6. Most of the precipitation occurs windward of the peak in the sensitivity simulations, and thus the source terms plotted in Fig. 3.11b, also represent closely the breakdown of upwind precipitation. Fig. 3.10 shows the full breakdown of precipitation sources by region as in Fig. 3.5.
3.3 Model Results

Figure 3.10: Regional Breakdown of Precipitation Source Terms - Breakdown of all simulations (for CDNC = 200 cm$^{-3}$) (a) accumulated precipitation over the whole domain and the regions upwind of the peak, over the peak and downwind of the peak, as well as the (e) the respective average $\delta^{18}O$ values. The contributions from precipitation sources normalized by the total precipitation in the three subregions are shown in (b–d) and the corresponding isotopic signatures in (e–g). The accumulated precipitation and its sources are integrated over the domain and normalized by the mountain half-width (20 km). Sources are: autoconversion of cloud liquid (laut), accretion of cloud liquid by rain (lacc), autoconversion of cloud ice (iaut), riming of cloud liquid (rim), vapor deposition onto ice (dep), and sublimation of ice/evaporation of rain (sub).
The $\delta^{18}O_{\text{precip}}$ is the most enriched in the reference simulation (W800m) compared to all other simulations (Fig. 3.11c), and this is also the case where the sources of riming and accretion are largest and vapor deposition smallest. The contribution to precipitation from riming decreases with increasing mountain height and decreasing temperature (Fig. 3.11b), while the contribution of vapor deposition increases. (Remember that these contributions are normalized by total precipitation, which itself increases with mountain height.) While there is considerable variation in the isotopic composition of the precipitation sources with mountain height and temperature, this variation is systematic in the most important contributors to precipitation: riming, vapor deposition and sublimation. As the mountain height increases or the temperature falls, these processes form precipitating hydrometeors from more depleted water vapor in the drier air found at colder temperatures and/or further aloft. Despite the variation with mountain height and temperature seen in Fig. 3.11d, a clear separation exists between the isotopic compositions contributed by riming and vapor deposition to precipitation, and the shift towards the formation of snow by vapor deposition at colder temperatures and higher mountains is reflected in the more depleted isotopic compositions in those experiments.

It is evident, particularly in the cold temperature experiments, that precipitation source significantly influences the $\delta^{18}O_{\text{precip}}$ signal, and that the decreasing $\delta^{18}O_{\text{precip}}$ signal with increasing mountain height is not a simple reflection of temperature. The solid line in Fig. 3.6a–b represents the regression of the warm temperature experiments’ domain-integrated precipitation and $\delta^{18}O_{\text{precip}}$. The dashed line in Figs. 3.6a–b 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 3.8a). The cold 800 m simulations fall on this dashed line in Fig. 3.6, but as the mountain height increases, the $\delta^{18}O_{\text{precip}}$ 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 1500 m and 3000 m experiments. As noted above, this suggests that the dependence of $\delta^{18}O_{\text{precip}}$ on altitude is itself a function of temperature. This can be explained by the combination of two effects. The nonlinearity in the relationship between isotopic composition and height plays a role here, as the gap between $\delta^{18}O$ for the warm and cold simulations increases with height due to the curvature of the Rayleigh curve (Fig. 3.1). The changing sources of precipitation also contribute with a shift from riming to vapor deposition with decreasing temperature.
Figure 3.11: Precipitation Source Breakdown for Sensitivity Experiments - 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 (LAUT), accretion of cloud liquid by rain (LACC), autoconversion of cloud ice (IAUT), accretion of cloud ice (IACC), riming of cloud liquid (RIM), vapor deposition onto ice (DEP), and sublimation of ice/evaporation of rain (SUB). All values are based on the 200 cm$^{-3}$ simulations.
3.4 Discussion

For the C3000m case, the domain-averaged $\delta^{18}O_{\text{precip}}$ is close to the $\delta^{18}O$ signatures of vapor deposition itself.

3.4 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 an idealized setup using different mountain heights, warm and cold temperature profiles, and increasing CDNC, our simulations show that there is a distinct difference in the $\delta^{18}O$ signatures of microphysical processes. Riming 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%; the difference correlating positively with mountain height. A similar range of 4-8% was found in the cold temperature experiments. This result seems to imply that the riming and vapor deposition $\delta^{18}O$ difference is independent of the environmental temperature. Instead, the distinctness of the isotopic signals depends on the temperature difference relative to the altitudes at which these two processes occur. As vapor deposition occurs both near the surface and in air with more depleted water vapor aloft, the $\delta^{18}O$ is more depleted than riming that is predominately happening near the mountain surface.

The sensitivity of $\delta^{18}O_{\text{precip}}$ to mountain height and temperature reflects, in part, the changing sources of precipitating hydrometeors. The dominant source of precipitation shifts from riming for smaller mountains and the warmer temperature profile, to vapor deposition for higher mountains and colder temperatures. The more depleted isotopic composition of the precipitating hydrometeors generated by vapor deposition contributes to the decrease of $\delta^{18}O_{\text{precip}}$ with mountain height and temperature. It is notable that the relationship between precipitation amount and $\delta^{18}O_{\text{precip}}$ driven by increasing mountain height differs with temperature, and that the difference in $\delta^{18}O_{\text{precip}}$ between the warm and cold simulations increases with mountain height. This suggests that the isotopic lapse rate is itself a function of temperature, and that this temperature dependence partly results in a shift in the microphysical pathways through which precipitating hydrometeors grow.
3.5 Conclusions

We have attempted to illuminate how $\delta^{18}O_{\text{precip}}$ depends on the processes responsible for the growth of precipitating hydrometeors. The decrease in $\delta^{18}O_{\text{precip}}$ with increasing mountain height and colder temperature profiles is largely driven by the formation of hydrometeors from more depleted water vapor in the drier air further aloft or at colder temperatures. However, the pathways through which precipitating hydrometeors are formed also plays a role, as the more enriched precipitating hydrometers produced by riming contribute less to surface precipitation and the more depleted hydrometeors produced by vapor deposition onto ice contribute more. The weaker dependence of $\delta^{18}O_{\text{precip}}$ on CDNC, where it exists, can be explained in a similar manner. While the weak dependence on CDNC suggests a similarly weak dependence on aerosol concentrations, the domain-integrated signal in the strongest case is roughly equivalent to a 1°C shift in temperature along the Rayleigh curve. Such a change might be visible in paleoclimate records of precipitation if there were systematic shifts in aerosol concentrations on longer timescales. However, if the majority of precipitation was produced in colder conditions, the sensitivity to CDNC might not be visible, as is the case in the more strongly precipitating cases here. While not considered in the present study, mixed-phase orographic precipitation does respond to changes in ice nuclei concentrations (e.g. Fan et al. (23)), and the associated shifts in microphysical processes could also impact the isotopic composition of precipitation.

3.5 Conclusions

Orographic precipitation is an important water resource, and in this work we have attempted to provide new perspective on how different atmospheric regimes may influence the formation of precipitation. The isotopic composition of orographic precipitation also provides additional information about the sources of water vapor and the microphysical processes that produce this precipitation. In the present study, 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 an isotope-enabled version of WRF. One reference simulation was performed along with sensitivity experiments that varied CDNC, temperature and mountain height to study the responses in the microphysical processes, their respective isotopic composition and the $\delta^{18}O_{\text{precip}}$. 
3.5 Conclusions

We find that in all of the simulations, precipitation grows mainly by riming of cloud liquid, vapor deposition onto ice, or a combination of the two processes. As mountain height is increased, autoconversion of ice becomes a non-negligible contributor to snow formation; however, the contribution from this process remains 4 to 5 times smaller than the largest source (i.e., vapor deposition or riming). The $\delta^{18}O_{\text{precip}}$ reflects the relative contributions from each of the sources, and thus hydrometeors that form from isotopically lighter sources lead to more depleted precipitation. The $\delta^{18}O$ difference between riming and vapor deposition ranges between 3-8‰ in all simulations and is independent of the initial temperature profile. Instead, the isotopic distinctness of the two sources persists despite wide variation in the isotopic composition of these sources with mountain height and temperature, and is related mainly to the relative altitude of the growth processes within the cloud.

The sensitivity of precipitation amount and location to CDNC changes is strongest for the lowest mountain height (800 m). However, as the mountain height is increased, or if the colder initial temperature profile is used, increasing CDNC has little to no influence on the precipitation. As seen in observations (Creamean et al. (18)) and modeling studies (e.g. Saleeby et al. (93)), in regimes where precipitation growth is dependent on liquid sources (e.g. riming, accretion of cloud liquid), higher CDNC values lead to a leeward shift and decrease in precipitation, whereas when ice-based processes (e.g. vapor deposition, ice autoconversion) dominate precipitation growth, precipitation location and amount are insensitive to CDNC. Similarly, with the exception of the reference simulation, the $\delta^{18}O_{\text{precip}}$ is insensitive to CDNC. In the sensitivity experiments, the average $\delta^{18}O_{\text{precip}}$ changes by $\leq 1\%$ between the highest and lowest CDNC simulations compared to $\sim 1\%$ in the domain average for the reference simulation. Experiments with a distinct (though small) average $\delta^{18}O_{\text{precip}}$ difference between the highest and lowest CDNC simulations, also have non-negligible contributions to precipitation from accretion of liquid droplets, a process that falls off substantially with increasing CDNC.

The results have potential implications for research and field campaigns looking to study the influence of different atmospheric regimes on orographic precipitation (e.g. IFRACS (Lowenthal et al. (56)), ISPA-III ((106)), StormVEx (Mace et al. (58)), etc.). As liquid processes are most responsive to CDNC, locations where precipitation primarily forms through accretion of cloud liquid and/or riming are likely to experience decreased accumulation, a shift in the location of, and a decrease in the $\delta^{18}O_{\text{precip}}$. However, the sensitivity depends on
both the mountain height and the region above the mountain surface in which precipitation forms and grows. The model could be beneficial to those planning observational campaigns in terms of choosing locations to collect samples. For example, those interested in studying the influence of aerosols on snowfall could identify the regions where precipitation is likely to be most sensitive or least sensitive to aerosol loading.

In this idealized work, we were able to distinguish isotopic signatures of the microphysical growth processes. As the climatology will vary between locations or even seasonally at one location, the model can be used to identify the isotopic signatures of microphysical processes in specific locations, which would help to determine growth pathways of measured precipitation. The next steps are to use the isotope-enabled microphysics scheme in a realistic setting to study snowfall events at Storm Peak Lab in Colorado observed during the Isotopic Fractionation in Snow (IFRACS) campaign. In this future work, we hope to expand upon our current research by studying the isotopic signatures of the microphysical growth processes that produce the observed precipitation.
Microphysical Controls on the Isotopic Composition of Wintertime Orographic Snowfall During IFRACS

This chapter is a summary of the modeling component in a collaboration with the Desert Research Institution (DRI) for the observational campaign that took place at Storm Peak Laboratory (SPL) near Steamboat Springs, CO. For the Isotopic Fractionation in Snow (IFRACS) campaign, an isotope-enabled version of the Weather Research and Forecasting (WRF) model is used to simulate two events during the 2014 campaign to further investigate the role of the microphysical processes as they contribute to snow growth in orographic mixed-phase clouds. The model is able to adequately reproduce observations and thus two sampling periods with snow of similar isotopic composition are analyzed. In both cases, riming and vapor deposition onto ice are the two main growth processes of snow. When the snow and cloud liquid $\delta^{18}O$ are small, the orographic cloud is deep and both riming and vapor deposition onto ice occur near the surface and up to several kilometers above the mountain surface.

4.1 Introduction

In Chapter 3, the motivation for studying orographic precipitation has been well-established; as a major water resource, understanding how different atmospheric regimes will influence the
growth and distribution of orographic precipitation is essential. The first step to addressing this question involved the idealized simulations of Chapter 3, while here, a more detailed, and perhaps meaningful, analysis is applied to a specific location using observations and realistic model simulations. The location used in this analysis is the Storm Peak Laboratory (SPL), a facility operated by the Desert Research Institute, which is situated 3210 m above sea level on the summit of Mt. Werner in the Park Range near Steamboat Springs, CO (Borys and Wetzel (8)). Observations were collected during the Isotope Fractionation in Snow (IFRACS) campaign that ran from January 20th - February 27th, 2014.

SPL has been used in previous work to study orographic clouds and precipitation, usually with the emphasis on the impact from aerosols (e.g. Hallar et al. (28), Lowenthal et al. (55), Saleeby and Cotton (90), Saleeby et al. (91, 93)). As noted by Borys and Wetzel (8), the orientation of the Park Range combined with the lifting of near-surface air masses (often) from the west, leads to mixed-phase clouds both at and near SPL in the winter. Often, in mixed-phase orographic clouds, precipitation will grow by the seeder-feeder mechanism (Reinking et al. (81)), whereby sedimentation of ice and snow formed by vapor deposition in an ice cloud aloft collide with and collect supercooled droplets; a process known as riming. The degree to which ice/snow particles are rimed, may determine how efficiently they are removed from the cloud (Borys et al. (10), Mitchell et al. (65)), and may also have a significant impact on the isotopic composition of the precipitation (Anderson et al. (3), Demoz et al. (22), Lowenthal et al. (55)). While in the case of Lowenthal et al. (55), observations determined that heavily rimed snowfall had an isotopic signature that was similar to cloud droplets near SPL and was more enriched than collected unrimed snow. Demoz et al. (22) however, emphasize that the isotopic signal of the rimed snowfall will depend upon where the supercooled liquid is collected within the cloud, and therefore will not necessarily resemble the isotopic content of cloud liquid near cloud base.

The variation in the isotopic content of the precipitation due to riming leads into the main objectives of the IFRACS campaign. One of the first questions this campaign hopes to address is if there is a significant difference in the isotopic signatures associated with heavily rimed and unrimed precipitation. As noted above, the work of Demoz et al. (22) introduces the additional factor to consider: if riming occurs, where does it occur in the cloud and how does this impact the observed isotopic signature of the snowfall. This last point factors into the second objective of the IFRACS campaign, which is to study the relative contributions
to snow growth from riming and vapor deposition onto ice. The goal is to understand how
important each process is, the altitude at which they occur and how distinctive the isotopic
signatures of these growth pathways are.

To some extent, Chapter 3 has addressed these objectives by determining in an idealized
setting that riming and vapor deposition onto ice contribute most to the growth of orographic
precipitation. This work was also able to determine distinct ranges of the isotopic signatures
for each of the precipitation source terms, such that despite the mountain height or initial
temperature profile used, the isotopic signature of riming was always 3-8‰ more enriched
than vapor deposition.

Building upon this idealized work, this chapter investigates the microphysical processes
of orographic snowfall in a realistic setting. Detailed analysis of two specific events during
the IFRACS campaign are the focus of this work. First, the model simulations are compared
to the observations available from SPL and for the region (more discussion in section 4.3.1),
and then the model microphysics are analyzed to determine how each microphysical process
contributes to snow formation at SPL and also how each helps to determine the observed
isotopic signature of the collected snow.

4.2 Model Setup

4.2.1 WRF Specifications

The Weather Research and Forecasting (WRF) model version 3.5.1 (Skamarock and Klemp
(94)) provided by the Mesoscale and Microscale Meteorology Division of the National Center
for Atmospheric Research is used for the two numerical simulations. The model is initiated
with three nested domains (using a one-way nesting process) as shown in Fig. 4.1. The
parent domain covers most of the United States and uses 30 km grid spacing over 100 × 56
grid points. The first nest is focused more on the western half of the country and uses 10 km
spacing throughout the 100 × 94 gridded domain, while the smallest nest is centered over
SPL (labeled by the star in Fig. 4.1), and uses 3.33 km spacing for the 91 × 136 grid points.
Within each nest there are 41 vertical sigma levels, with the top layer at 100 mb. Note that
no convection scheme was used for either the parent domain or any of the subdomains.
Figure 4.1: WRFiso Nested Domains - Nested domain setup for WRF. Location of SPL is noted by yellow star.
4.2 Model Setup

As in Chapter 3, the setup uses the Thompson microphysics scheme within WRF (Thompson et al. (100)), but with stable water isotopologue physics included (Moore et al. (68)). A thorough description of the microphysics scheme is provided in Chapter 3, and Appendix B, but to summarize here briefly, the isotopic composition of vapor and each hydrometeor species (cloud liquid, cloud ice, rain, snow and graupel) is tracked. The exchanges of the heavy isotopologues are accounted for during all of the microphysical processes present in the Thompson scheme. Fractionation is set to occur during processes of deposition of vapor onto liquid or ice hydrometeors and evaporation of liquid phase hydrometeors (rain or cloud liquid). Other phase changes are assumed to be non-fractionating. Going forward, the isotope-enabled version of the model will be referred to as WRFiso.

In addition to the Thompson microphysics scheme, the Noah Land Surface model (Chen and Dudhia (12)) is also applied in these simulations. Both the microphysics scheme and the land surface model are chosen as they have been used successfully in previous studies modeling areas similar to SPL in terms of terrain and snow cover (e.g. Ikeda et al. (34), Liu et al. (53), Rasmussen et al. (79)). The land surface model does not account for the isotopic composition of surface fluxes (e.g. evaporation and sublimation of snow), and instead the fluxes are fixed at \(-63^\circ\text{h}(\delta D)\) and \(-9^\circ\text{h}(\delta^{18}O)\). These values are the mean composition of precipitation during February 2014 in the global model that provides the initial conditions for isotopes, which is discussed further in the next section.

4.2.2 Initial and Boundary Conditions

Initial and boundary conditions for each domain are created from data from the 32-km resolution North American Regional Reanalysis (NARR) (Mesinger et al. (63)). Using realistic data allows for a reasonable comparison between the model simulations and observations, which is one of the main goals of this work. NARR data is available at 3-hour increments and is used to set up the lateral boundary conditions and initial conditions for both of the 2014 events simulated with WRFiso.

To generate the initial and boundary conditions for the stable water isotopes, output from an isotope-enabled general circulation model was used to interpolate the isotopic composition of water vapor onto the initial and boundary conditions of WRFiso. This was accomplished
by using output from the LMDZ4 general circulation model, which is the atmospheric component of IPSL-CM4 coupled model used in the 4th IPCC assessment report (Hourdin et al. (31)). The output used to generate the initial and boundary conditions were run by Camille Risi using an isotope-enabled version of LMDZ4 (LMDZ-iso). In this model, the heavy isotopologues (H$_{18}^2$O and HDO) are treated just like the standard water molecule (H$_2$O) except that these molecules are subject to fractionation during phase changes, such as evaporation and vapor deposition (Bony et al. (7)). As in the isotope-enabled Thompson scheme used in this work and in Chapter 3, fractionation does not occur during freezing, melting, riming or sublimation of ice hydrometeors. The LMDZiso model has been successfully used and validated in many different research studies (e.g. Lee et al. (50), Risi et al. (82, 84), Vimeux et al. (101)), and therefore provides confidence in using the output to initialize WRFiso. 

In order to create the isotope initial and boundary conditions, LMDZiso was run for the entire month of February with hourly time steps and a subset of this output corresponding to the region of the parent domain of WRFiso was defined. Using the dates that correspond to the time periods analyzed in sections 4.3 and 4.4, $\delta$-values of vapor are calculated from the LMDZiso output. The WRF Processing System (WPS) then interpolates the wind, temperature, vapor and $\delta$-values on to the WRFiso x-y grid. Next, the WRF program real.exe interpolates these values in the vertical and then the mixing ratios of the heavy isotopologues in vapor are calculated from the delta-values, such that the mixing ratio of HDO vapor is calculated as:

$$HDO_{vapor} = H_2O_{vapor} \times (1 + 1 e^{-3} \times \delta D).$$ (4.1)

Thus with some adjustments, the boundary and initial conditions are filled in for the isotopes, just as they are for all of the other values in WRFiso during pre-processing.

### 4.3 February 17 00Z – 19 00Z: Clear Day

This particular time period was chosen to examine first as it represents a particularly clear time during the campaign. During this 48-hour period, skies were relatively cloud-free (see Fig. 4.2) and SPL was never in cloud, nor was any precipitation recorded at the site. Assessment of the model by comparing with observations in the region and at SPL are detailed in
the following sections.

**Figure 4.2: Infrared Composite for 2/17 2014 12Z** - Infrared image from a GOES East and West composite at 12Z on February 17th, 2014. Image courtesy of Unisys Weather.

### 4.3.1 Comparison with Observations

Before analyzing the model output at SPL, information gathered from the snow telemetry (SNOTEL) automated network is used to assess the output for the smallest nest. The SNOTEL system is operated by the Natural Resource Conservation Service (NRCS) and has broad coverage over the western United States. There are approximately 116 automated stations in Colorado alone, 60 of which fall within the region covered by the smallest nest. An additional 10 sites in southern Wyoming are also located within the simulated domain and therefore are also used for analysis. The location and elevation of the SNOTEL sites are shown in Fig. 4.3 (left). The standard SNOTEL station measures air temperature, precipitation, snow water content and snow depth. Therefore, while the SNOTEL sites provide no isotopic measurements, useful meteorological information can be obtained to compare with the model output.
4.3 February 17 00Z – 19 00Z: Clear Day

Figure 4.3: SNOTEL and WRFiso Average Surface Temperature: 2/17–2/19 - (Left) SNOTEL sites and elevation [m] with SPL denoted by black triangle and (Right) Comparison of average surface temperature [°C] for SNOTEL locations (circles) and WRFiso between 2/17 00Z and 2/19 00Z.
Figure 4.4: SNOTEL and WRFiso Surface Temperature Difference: 2/17–2/19 - WRFiso – SNOTEL surface temperature as a function of longitude averaged over 2/17 00Z and 2/19 00Z.
The right panel of Fig. 4.3 shows the surface temperatures as simulated by WRFiso (contours) and as measured by SNOTEL (circles) averaged over the period of 00Z on February 17th to 00Z on February 19th. The temperatures simulated by WRFiso tend to follow the terrain, in that the coldest temperatures occur in the highest elevations. In these regions in particular, the model tends to underestimate the temperature and on average, the model is approximately 1.6°C colder than temperatures recorded by SNOTEL stations, as noted in Fig. 4.4. However, this value is biased by some very warm temperatures and most SNOTEL stations are about 3-5°C warmer than what is simulated by the model.

Though the model does not capture the temperatures perfectly well throughout the domain, there is generally good agreement between the model and observations near SPL. There are four SNOTEL towers by the lab (SPL is marked by black triangle in Fig. 4.3), and all of these stations report average temperatures similar to what is simulated by WRFiso. Further discussion of the model’s simulation of SPL conditions during this clear event are left to section 4.3.2.

While this time period was chosen due to the lack of precipitation and relatively clear conditions, there was still some precipitation within the modeled domain. The center panel of Fig. 4.5 depicts the accumulated precipitation for WRFiso and the SNOTEL stations. During the 48-hour time period, very little precipitation fell over the domain, with the model indicating accumulations of less than 1 mm. The SNOTEL sites also indicate trace amounts of precipitation, though there are some larger discrepancies near SPL. The two stations in question, one to the northeast of SPL and the other to the southeast, indicate much more accumulated precipitation than any other SNOTEL station. Investigating this further, it appears that precipitation amounts at these two stations are suspect as the daily averages (those listed at 00Z) do not match up with the hourly observations listed in the archives. This discrepancy appears to only occur for the accumulated precipitation and only at these two SNOTEL locations.

The other two stations closest to SPL indicate that at least some precipitation was recorded over the 2-day period, though in both cases the accumulation is less than 1 mm. As both of these stations are at lower elevations than SPL, it is possible that a precipitating cloud encompassed these locations but did not extend up to the lab. The dynamics of this scenario are discussed in Chapter 3, such that an air mass may go around the mountain peak rather than up and over (Galewsky (26), Lechler and Galewsky (47), Smith (95)), and thus
result in precipitation below and around, but not over, the mountain peak. This scenario is supported by the fact that the flow at SPL is likely to be subcritical as determined by the Froude number calculated with model output at the location of the lab. Using the methods of Muccilli (69) and Chen et al. (13), the average Froude number during the observation period is $\ll 1$ ($\approx 0.16$), indicating blocked flow near SPL.

As mentioned previously, the SNOTEL stations do not make isotopic measurements, however, the average $\delta^{18}O$ of precipitation ($\delta^{18}O_{\text{precip}}$) is presented in the right panel of Fig. 4.5 for reference. The isotopic composition of the precipitation ranges between $-32\%_o$ and $-12\%_o$, but on average is around $-18\%_o$. It should be noted that the degree of depletion does not appear to correspond with elevation and thus temperature (comparing the left and right panels of Fig. 4.5), which is an indication that the local microphysical processes are likely determining the isotopic signal of the precipitation in these regions. However, as the focus of this work is on SPL, analysis of precipitation elsewhere in the domain is left to future work.

![Figure 4.5: SNOTEL and WRFiso Average Precipitation: 2/17–2/19](image)

- (Left) SNOTEL sites and elevation [m] with SPL denoted by black triangle; (Center) Accumulated precipitation [mm] over the 48-hour period for both WRFiso (contours) and SNOTEL (circles) and (Right) average $\delta^{18}O_{\text{precip}}$ from WRFiso between 2/17 00Z and 2/19 00Z.
4.3 February 17 00Z – 19 00Z: Clear Day

4.3.2 Storm Peak Lab

The model has proven capable of simulating conditions similar to observations, though typically biased toward colder temperatures and less precipitation. This sections considers direct comparisons between WRFiso and observations at SPL. In Fig. 4.6, temperature measured every 5 minutes by a rooftop sensor at SPL is plotted in red, while temperature from the lowest model level of WRFiso is presented in blue. During the day of February 17th, WRFiso does not match up very well with observations and appears to lag slightly; however by the end of the day on the 17th and throughout the 18th, temperatures are relatively similar and differ by at most 2°C. The initial difference between WRFiso and SPL is likely related to the initial conditions provided by NARR (black line in Fig. 4.6), which has a much coarser topography than the model and therefore places the location of SPL at a different elevation.

At SPL, a Picarro L-2130i analyzer was used to continuously measure water vapor and its isotopic content, the methods and calibrations for which are explained in Lowenthal et al. (56). According to Fig. 4.7, the model underestimates the vapor at SPL between February 17th and 18th by ≤1 kg kg\(^{-1}\). This difference is largest at 00Z on the 18th, but is closer to the measured values before and after this time point. The same relationship is seen in the \(\delta^{18}O_{\text{vapor}}\) and \(\delta D_{\text{vapor}}\) in the middle and bottom panels respectively. However, for the isotopic measurements, the model underestimates the values on the 17th, but slightly overestimates on the 18th. During this period, wind direction at SPL ranged between westerly and southwesterly (Lowenthal et al. (56)), but the WRFiso-produced winds range between westerly and northwesterly (not shown). The more northerly winds are likely bringing drier air into the region and could help to explain the lower vapor values compared to observations.

By looking at model transects at the latitude of SPL, the evolution of the vapor can be investigated to help understand why the model values are lower than what is measured at SPL. The output for February 17th is shown in Fig. 4.8. When the model simulation starts, the vapor mixing ratio is relatively high and only slightly less than what is observed at SPL. As time progresses, it seems that the air mass with higher water vapor becomes trapped in the valley area upstream of SPL, such that drier air is flowing over this moist air mass and bringing drier air downstream, as indicated by the colder colors at SPL between 09Z – 15Z. After this time, the moist air has dissipated and the vapor mixing ratio has decreased throughout the whole transect.
Figure 4.6: SPL, WRFiso and NARR Surface Temperature: 2/17–2/19 - Comparison of the surface temperature [°C] recorded at SPL (red), the lowest model level of WRFiso (blue), and NARR between 00Z on 2/17 and 00Z on 2/19, 2014.
4.3 February 17 00Z – 19 00Z: Clear Day

Figure 4.7: SPL, WRFiso and LMDZiso Vapor, $\delta^{18}O_{vapor}$ and $\delta D_{vapor}$ - Comparison of the (top) in situ water vapor [kg kg$^{-1}$] recorded at SPL (black), output from WRFiso (blue) and LMDZiso (red); (center) same as the top panel for the $\delta^{18}O$ of vapor; (bottom) same as middle but for $\delta D$ of vapor. Values are provided for the time period of 00Z on 2/11 to 00Z on 2/20, 2014.
By 00Z on the 18th, Fig. 4.9 shows that over SPL the vapor mixing ratio is low, but some higher mixing ratios are progressing into the transect from the western boundary. Recall that 00Z on the 18th is when the vapor deviates most from the observations (Fig 4.7), but as the day progresses, the model vapor increases and becomes more like the vapor measurements at SPL. This increase in the vapor mixing ratio can be seen in the transects for 09Z – 21Z on the 18th.

![Figure 4.8: WRFiso Vapor and δ^{18}O_v Transects at SPL 2/17](image)

While the δD_{vapor} is provided in Fig. 4.7, the analysis in this chapter will focus on H_2^{18}O and additional information that can gained by considering HDO is left to future work. The evolution of the δ^{18}O_{vapor} for both the 17th and 18th are provided in the right columns of Figs. 4.8 and 4.9, respectively. Patterns similar to the vapor mixing ratio discussed above are also depicted in the WRFiso output for the δ^{18}O_{vapor}. 

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Figure 4.9: WRFiso Vapor and $\delta^{18}O_v$ Transects at SPL 2/18 - WRFiso output for vapor (left column) [kg kg$^{-1}$] and $\delta^{18}O$ (right column) of vapor at 40.5° latitude (latitude of SPL). Each plot shows the vapor transect for model output at 03Z, 09Z, 15Z and 21Z on February 18th, 2014. The vertical black line represents the location of SPL.
The model appears to simulate the region well, and while the comparisons with SPL observations are not perfect matches, the patterns are similar and values are within the same order of magnitude. Given this performance, the following sections investigate a snow event at SPL and provide a deeper investigation of the microphysical processes.

### 4.4 February 12 00Z-14 00Z: Snow Event

During this 48-hour term, the lab was in cloud for periods of 30 minutes or more and 10 samples were collected during the intensive observation periods (IOPs). Of these 10 samples, isotopic analysis of cloud liquid, vapor and snow is available for 4 IOPs. The analysis in this section will focus on 2 sample periods that both reported snow of similar isotopic content, but varied in the offset from the concurrently sampled cloud liquid $\delta^{18}O$. Using the information collected during the sample periods and output from WRFiso, in-depth analysis of the microphysical processes during these events are conducted.

#### 4.4.1 Comparison with Observations

As with the clear period, the output from WRFiso for February 12th – 14th is first compared to SNOTEL sites that are within the model domain. Fig. 4.10 (right) indicates that similar patterns, as discussed in Section 4.3.1, are also present here: WRFiso temperatures are strongly correlated with the elevation, such that colder temperatures are simulated at higher elevations. Overall, WRFiso simulates temperatures throughout the domain that are colder than the SNOTEL observations by about 3.2°C (Fig. 4.11. At SPL, the model is in good agreement with the average temperatures recorded at nearby SNOTEL stations, though more detailed temperature comparisons are left to Section 4.4.2.

The accumulated precipitation during this time period is shown in Fig. 4.12 (middle panel) and the difference as a function of longitude is shown in Fig. 4.13. Higher amounts of precipitation are simulated by the model, as indicated by the warmer colored contours. On average, the model overestimate accumulated precipitation by about 7.5mm, though there are some regions of severe underestimation (Fig. 4.13), though these data points are associated with the suspect SNOTEL stations to the east of SPL, and thus the comparison
4.4 February 12 00Z-14 00Z: Snow Event

Figure 4.10: SNOTEL and WRFiso Average Surface Temperature: 2/12-2/14
- (Left) SNOTEL sites and elevation [m] with SPL denoted by black triangle and (Right) Comparison of average surface temperature [°C] for SNOTEL locations (circles) and WRFiso between 2/12 00Z and 2/14 00Z.
Figure 4.11: SNOTEL and WRFiso Surface Temperature Difference: 2/12–2/14
- WRFiso – SNOTEL surface temperature as a function of longitude averaged over 2/12 00Z and 2/14 00Z.
of these sites in particular is not meaningful. However, focusing on the area surrounding SPL, the two nearby SNOTEL stations to the west match up well with WRFiso output.

The average isotopic content of the simulated precipitation is provided in the right panel of Fig. 4.12. There is no obvious relationship between the amount of precipitation and the $\delta^{18}O_{\text{precip}}$, such that regions of high accumulations do not also simulate more depleted or enriched precipitation. The lack of a correlation implies that factors other than elevation and temperature are important for setting the isotopic signature of the precipitation.

The domain precipitation on average has a $\delta^{18}O$ of approximately -18‰. There are regions in the east and north that are slightly more depleted, but the overall gradients are not large. Additionally, there is a low elevation area at 39.5° latitude with relatively enriched $\delta^{18}O_{\text{precip}}$; however, the accumulated precipitation in this region is near-zero and thus these higher $\delta^{18}O$-values are likely not significant.

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**Figure 4.12: SNOTEL and WRFiso Average Precipitation: 2/12–2/14**

- (Left) SNOTEL sites and elevation [m] with SPL denoted by black triangle;
- (Center) Accumulated precipitation [mm] over the 48-hour period for both WRFiso (contours) and SNOTEL (circles) and
- (Right) average $\delta^{18}O_{\text{precip}}$ from WRFiso between 2/12 00Z and 2/14 00Z.
Figure 4.13: SNOTEL and WRFiso Precipitation Difference: 2/12–2/14 - WRFiso – SNOTEL accumulated precipitation as a function of longitude between 2/12 00Z and 2/14 00Z.
4.4 February 12 00Z-14 00Z: Snow Event

4.4.2 Storm Peak Lab

While the comparisons with SNOTEL observations indicate that WRFiso does better in some regions of the domain than others, the main focus here is how well the model simulates conditions as SPL. In Fig. 4.14, the air temperature at SPL is compared to output from the lowest model level at the coordinates of SPL. The trends in the temperature throughout the time period are similar, though the model tends to run colder than observations. Similarly, the simulated vapor at SPL is slightly lower than what is measured at the lab (Fig. 4.7), though the δ-values are very similar, and the agreement is much better than during the clear period. As on the 17th and 18th, winds measured at SPL range between westerly and southwesterly (Lowenthal et al. (56)), while the WRFiso winds are westerly and northwesterly, which could explain the lower vapor mixing ratios over this period.

The precipitation during this 48-hour period occurs later on the 12th and into the 13th. Trace amounts of precipitation also occur earlier on the 12th (not shown), though not all conditions for performing an IOP were met and thus samples were not collected. In Fig. 4.15, the SPL samples and simulated values of δ18O for vapor, cloud liquid and snow are compared. The points on the figure (either triangles or asterisks) indicate the value of the observation or model output at the time it was recorded, while the lines help to indicate the trends over the time period. Overall, there is good agreement between the model and the observations in the δ18O values of the hydrometeors and vapor. The output from LMDZiso (solid red line) is also included to illustrate how the WRF implementation has helped to improve the simulation of vapor δ18O to be closer to the observations at SPL.

Immediately, one can see that the vapor is more depleted than either the snow or cloud liquid, while simultaneously the snow and cloud liquid δ18O values are relatively similar. The degree to which the δ18O of cloud liquid and snow at SPL differ can be an indication of the degree of riming (Lowenthal et al. (55)). The general idea behind this relationship is that as snow crystals rime supercooled cloud droplets, the isotopic signature of the snow becomes more similar to the cloud liquid. In comparison, snow that grows mostly through vapor deposition will have isotopic signatures that are different from the cloud liquid and typically more depleted.

While there are several samples available during this period, there are two that stand out as particularly interesting to investigate further. On February 12th, a sampling period
Figure 4.14: SPL, WRFiso and NARR Surface Temperature: 2/12–2/14 - Comparison of the surface temperature [°C] recorded at SPL (red), the lowest model level of WRFiso (blue), and NARR (black) between 00Z on 2/12 and 00Z on 2/14, 2014.
starting at 20:05Z collected snow with an isotopic value measuring -17.1‰ and was more depleted than the cloud liquid at SPL by approximately 0.4‰. Similarly, snow collected during a sampling period on February 13th beginning at 9:51Z had a similar isotopic value of -17.3‰, but was more depleted than cloud liquid at SPL by 2.6‰. To understand what may be the underlying driver for the differences in the relationship with cloud liquid δ¹⁸O for snow of similar isotopic content, the remainder of this chapter will investigate the microphysical processes contributing to the snow growth during these two time periods.

![Diagram](image-url)

**Figure 4.15:** SPL, WRFiso and LMDZiso δ¹⁸O of Vapor, Cloud and Snow - δ¹⁸O values of cloud liquid (blue), snow (black) and vapor (red) for SPL samples (solid line with triangles), WRFiso (dashed lines with asterisks), and LMDZiso (solid line).
4.4.3 Microphysics

4.4.3.1 February 12th, 2014: 21Z Analysis

To analyze the sampling period on February 12th, WRFiso output from 21Z is used. Fig. 4.16 shows transects at the latitude of SPL for cloud liquid, vapor, cloud ice and snow to illustrate the structure of the orographic cloud that produces the snowfall. Note that transects of graupel and rain were not included as the model did not produce either of these hydrometeors at SPL, nor were they observed during the IOPs. As discussed in Chapter 3, the Thompson microphysics scheme quickly leads to the conversion of cloud ice to snow (Thompson et al. (100)), which is why the mixing ratio of ice is so low compared to snow, and snow that is above the surface can be considered as part of the glaciated/mixed-phase cloud.

At SPL, the figure shows that there is a mixed-phased orographic cloud as indicated by the cloud liquid and snow transects. Above 650 mb, the cloud is mostly composed of snow, though cloud liquid is present between at the surface and as high as 450 mb. This setup is favorable for riming as the snow crystals in the glaciated cloud could continue to grow by riming of supercooled cloud liquid (as indicated by the temperature contours in the top left panel of Fig 4.16), which will be discussed shortly.

The corresponding isotopic signatures for each of the transects are shown in Fig. 4.17. As would be expected in all of the fields, the isotopic values are more enriched near the surface and decrease with altitude. Near SPL, snow and cloud liquid are both approximately -17‰ (-16.9‰ and -16.6‰, respectively), while vapor in the same location is closer to -29‰. Snow and ice aloft are much more depleted (approximately -55‰ between 300–200 mb). The cloud liquid present at 500 mb has a very similar isotopic content as snow at this same altitude, and just as in the observations, the difference between snow and cloud liquid at SPL is small (0.3‰).

While the exact budgets for precipitation are difficult to close at specific locations (unlike for the entire domain as was done in Chapter 3), transects of microphysical processes occurring near SPL are examined to get a sense of the precipitation growth processes occurring in this region. In Fig. 4.16, only two microphysical processes are presented as the other source terms were at least 10 times smaller. As was also the case in the idealized simulations of Chapter 3 (and the references cited therein), riming and vapor deposition onto ice are the main growth processes, with small contributions from autoconversion and accretion of cloud
Figure 4.16: WRFiso Cloud, Vapor, Ice and Snow Mixing Ratios: 2/12 21Z - Simulated mixing ratios in kg kg\(^{-1}\) of (top right) cloud liquid, (top left) vapor, (bottom right) ice, and (bottom left) snow for February 12\(^{th}\) 21Z.
Figure 4.17: WRFiso $\delta^{18}O$ of Cloud, Vapor, Ice and Snow: 2/12 21Z - Simulated $\delta^{18}O$ values in per mil for (top right) cloud liquid, (top left) vapor, (bottom right) ice, and (bottom left) snow for February 12th 21Z.
4.4 February 12 00Z-14 00Z: Snow Event

liquid (not shown). The transects show that riming is a slightly larger source than vapor deposition, though both sources are of similar magnitudes. Both processes occur relatively close to the mountain surface, though there appears to be a small amount of riming occurring as high as 450 mb, corresponding with the cloud liquid present in this region (top right of Fig. 4.16), but is strongest near the surface.

The isotopic content of the precipitation is going to depend on the source terms and their respective isotopic signatures. In this case, the isotopic signature of the snowfall is predominately determined by the degree of riming and vapor deposition, with an apparent bias toward riming, as this source is slightly larger (at least according to the transects). Note that the source terms are accumulated through time and it is not necessarily the case that they occur simultaneously. However, while an exact budget cannot be closed for the precipitation at SPL, the source terms are still suggestive of the contributions to snow growth and thus are analyzed as such.

The isotopic signatures for each of the microphysical processes are presented in Fig. 4.18, which shows that riming is slightly more depleted than vapor deposition. Near SPL, the simulated $\delta^{18}O$ of riming is around -18‰, while vapor deposition is closer to -12‰. The isotopic signature of the riming is similar to that of cloud liquid near SPL, and thus supports the small difference between the $\delta^{18}O$ of snow and cloud liquid (0.3‰). The altitude at which these processes occur is also an important consideration as the source vapor below 600 mb is relatively enriched (compared to higher altitudes), and thus contributes to the high isotopic signatures of both riming and vapor deposition onto ice.

While the low-level growth and relatively similar $\delta^{18}O$ values makes it difficult to differentiate between the riming and deposition sources, output does indicate that snowfall produced by WRFiso at 21Z on February 12th grows predominantly through a combination of riming and vapor deposition, with riming being a slightly larger source. The isotopic signature of the snowfall is thus likely weighted toward the isotopic signature of riming, and as riming implies growth by collection of supercooled cloud droplets and occurs near the surface, there is a small difference in the isotopic signatures of the snowfall and cloud liquid.

4.4.3.2 February 13th, 2014: 12Z Analysis

Samples collected on February 13th at 9:51Z at SPL indicated that snow had an isotopic content of -17.3‰, which was similar to snow collected during the sampling period on the
4.4 February 12 00Z-14 00Z: Snow Event

Figure 4.18: Transects of Riming and Vapor Deposition: 2/12 21Z - Simulated mixing ratios in kg kg$^{-1}$ of (top left) riming of cloud liquid and (bottom left) vapor deposition onto ice as well as the corresponding $\delta^{18}O$ of (top right) riming and (bottom right) vapor deposition for February 12$^{th}$ 21Z.
12th. However, the snow collected on the 13th was more depleted than cloud liquid by 2.6%. Based on this difference, it would seem that the microphysical conditions were different on the 13th, and thus snow growth resulted by slightly different pathways.

WRFiso output from 12Z on February 13th is used to evaluate the microphysics during sampling period at SPL. Unlike the earlier IOP, the model produces snow at SPL that is more enriched than what is actually measured (-15.4‰ compared to -17.3‰), but the simulated cloud liquid $\delta^{18}O$ is very close to observations: -14.4‰ and -14.8‰, respectively. Despite the difference in the snow $\delta^{18}O$ values, the model accurately simulates snow that is more depleted than the cloud liquid. Therefore, the microphysical analysis is still meaningful for this time period.

Transects of orographic cloud, vapor and snow at SPL latitude are presented in Fig. 4.19. (Note: no graupel or rain were simulated or observed at SPL.) The liquid cloud contains supercooled liquid, as indicated by the temperature contours, and extends from the surface near SPL to 500 mb. The glaciated cloud, which is composed mostly of snow with traces of ice above 600 mb, extends from the surface to 300 mb. The glaciated cloud during this period is much deeper than in the earlier sampling period, but similarly has a setup that is promising for the seeder-feeder mechanism to occur. Another important difference on the 13th is the higher vapor mixing ratio near SPL; the gradient is much steeper as the mixing ratio drops off rapidly with height.

To further analyze the microphysical conditions on the 13th, isotopic signatures of the vapor and orographic cloud are shown in Fig. 4.20. Near SPL, the cloud isotopic signature is approximately -15‰, which matches up with the $\delta^{18}O$ of cloud liquid mentioned above (-14.4‰). Snow at SPL is also approximately -15‰, and decreases with height to -40‰ at 300 mb, which is also the case for the cloud ice. The $\delta^{18}O$ of vapor during this period is about -26‰ near the surface and decreases to -55‰ at 300 mb.

Similar to the sampling period on the 12th, the main source terms of precipitation are examined in Fig. 4.19 (left column). Four microphysical processes are non-zero near SPL, but only riming and vapor deposition onto ice are presented as autoconversion and accretion of cloud liquid are much smaller than these source terms. One major difference during this sampling period (compared to the previous day), is that the model output indicates the microphysical processes occur at greater atmospheric depths. While both riming and vapor deposition are highest near the surface, both occur as high as 450 mb. Near the surface,
4.4 February 12 00Z-14 00Z: Snow Event

Figure 4.19: WRFiso Cloud, Vapor, Ice and Snow Mixing Ratios: 2/13 12Z - Simulated mixing ratios in kg kg$^{-1}$ of (top right) cloud liquid, (top left) vapor, (bottom right) ice, and (bottom left) snow for February 13th 12Z.
Figure 4.20: WRFiso δ¹⁸O of Cloud, Vapor, Ice and Snow: 2/13 12Z - Simulated δ¹⁸O values in per mil for (top right) cloud liquid, (top left) vapor, (bottom right) ice, and (bottom left) snow for February 13th 12Z.
riming is a larger source term than vapor deposition; however there is more vapor deposition occurring in the region aloft as well as immediately upstream of the lab, which also has the potential to contribute to snowfall measured at SPL.

The isotopic composition of the microphysical processes indicate that vapor deposition is the most enriched of the source terms presented in the right column Fig. 4.20. While both processes occur at high altitudes and near the surface, vapor deposition onto ice seems to be the larger source term in this region, though riming is larger near the surface. Given the deeper orographic cloud and the higher extents at which riming and vapor deposition occur, snow is forming at higher altitudes from more depleted source vapor. However, as both source terms are much larger near the surface where the isotopic compositions are more enriched (as is the source vapor), it is likely that the more depleted signal of snow formed aloft is lost by the enhanced low-level growth processes. The longer growth pathway (i.e. greater cloud depth over which snow growth occurs) on the 13th, is one possible explanation for the larger difference between the $\delta^{18}O$ of snow and cloud liquid compared to measurements on the 12th

4.5 Conclusions

The work presented in this chapter evaluates how capable the WRFiso model is at simulating realistic events. This evaluation was conducted by comparing output during the same period as the IFRACS observational campaign at Storm Peak Laboratory (SPL) in Colorado. Overall the model proved sufficient in simulating realistic conditions that were similar to observations provided by SNOTEL stations located throughout Colorado in addition to measurements taken at SPL. The model was biased toward colder temperatures and lower amounts of atmospheric vapor over the domain as a whole; however, direct comparisons with observations at SPL indicate that the model is capable of producing conditions of similar magnitude and trends.

One of the main objectives of the IFRACS campaign is to determine the relative importance of snow growth by vapor deposition and riming. To address this objective, two sampling periods were analyzed with the model to investigate the microphysical tendencies and the sources of snow growth. The sampling periods were chosen based on the fact that snow of similar isotopic content was collected during both events, though snow and
4.5 Conclusions

Figure 4.21: Transects of Riming and Vapor Deposition: 2/13 12Z - Simulated mixing ratios in kg kg\(^{-1}\) of (top left) riming of cloud liquid and (bottom left) vapor deposition onto ice as well as the corresponding \(\delta^{18}O\) of (top right) riming and (bottom right) vapor deposition for February 13\(^{th}\) 12Z.
cloud $\delta^{18}O$ differences were quite different, indicating likely differences in the microphysical processes of snow growth.

Analysis of the microphysical tendencies indicate that riming and vapor deposition onto ice are the two main growth pathways of snowfall during both sampling periods. In the event where the snow and cloud $\delta^{18}O$ difference is small, riming is found to contribute more to snow growth, while deposition is more significant in the event with the larger snow and cloud $\delta^{18}O$ difference. This finding supports the correlation between the small snow and cloud $\delta^{18}O$ differences and snow growth by riming as suggested by Lowenthal et al. (55). However, the altitude at which these processes occur and the source vapor are also important to consider. On the day where cloud and snow $\delta^{18}O$ differences are small (February 12th), microphysical processes are contained at low altitudes with relatively enriched source vapor. Conversely, the period with the larger snow and cloud $\delta^{18}O$ difference is associated with growth processes occurring at higher altitudes and more depleted source vapor. However, enhanced-low level growth appears to compensate for the depleted isotopic signal associated with snow formed at higher altitudes resulting in precipitation that is relatively enriched.

Such results are promising as the WRFiso model has demonstrated to be a useful tool for analyzing the microphysics of orographic snow. While this work only considers two sampling periods, future work will investigate the relative roles of riming and vapor deposition further. Additionally, this work does not address the impact of increased aerosols on riming and snowfall, which is another objective of the IFRACS campaign. For the simulations presented here, only one cloud droplet number concentration (100 cm$^{-3}$) was used, which as discussed in Chapter 3, can serve as a proxy for aerosol loading. This cloud droplet number concentration was chosen as it represents the average value generally observed at SPL. In future work, additional realistic simulations with varying cloud droplet number concentrations could help to address the sensitivity of riming to changes in aerosol content.
5

Conclusions

In this thesis, isotope-enabled models are used to study moisture budgets and microphysical processes of convection in the tropics and orographic precipitation in the midlatitudes. Using stable water isotopes as tracers of water and precipitation sources, additional information has been obtained about the relationship between the observed isotopic signals of precipitation in different global regions. The analysis of model output as well as the comparisons with measurements provides important information about the factors that seem to have the strongest control on determining the isotopic composition of precipitation collected at the surface.

Looking at the large scale moisture budgets of precipitation in the tropics, idealized model analysis indicates that during strong convective events capable of producing high amounts of precipitation, the increasing contribution of more depleted converged water vapor to the precipitation budget results in the decreasing $\delta^{18}O$ of precipitation. Large-scale convergence of vapor is found to be more important than local evaporation and also explains the amount effect observed in tropical precipitation on monthly or longer timescales. Isotopic depletion of precipitation is more sensitive to the source vapor than to depletion caused by post-condensational processes. However, such processes likely contribute more to amount effect relationships occasionally observed on shorter (less than monthly) timescales than what is considered in this modeling framework.

In idealized model simulations of 2D flow over a mountain, the sensitivity of orographic precipitation to different atmospheric regimes is investigated by changing initial temperature profiles, mountain heights, and cloud droplet number concentrations. Using stable water
isotopes as tracers of precipitation and microphysical processes, the results from the model simulations find that changes in the domain average $\delta^{18}O$ of precipitation is more sensitive to changes in temperature and mountain height than varying levels of cloud droplet number concentration. Varying the mountain peak height and the temperature profile alters the microphysical growth processes in terms of where these processes occur and how much they contribute to the total orographic precipitation. Simulated decreases in the $\delta^{18}O$ of precipitation is a reflection of the decreasing contribution of riming in the precipitation budget and increasing growth by vapor deposition onto ice. These two processes are associated with distinct isotopic signatures that are insensitive to the initial temperature profile. Overall, this idealized isotope-enabled model provides a useful framework for analyzing the microphysics of orographic precipitation and determining how the source terms of precipitation contribute to the accumulated precipitation at the mountain surface.

The isotope-enabled model is a useful resource for examining observations of orographic snow. Our model is capable of reproducing conditions similar to observations at a specific location over a given time period during a wintertime campaign. In conjunction with the measurements collected, our model is able to assess the microphysical conditions during sampling periods and determine the general pathways of snowfall growth, which helps to explain the isotopic signature of the collected snow. In the realistic setting, riming and deposition of vapor onto ice are still the most important source terms in the precipitation budget, and while their isotopic signatures are not as distinct as in the idealized simulations, this information, as well as output detailing other atmospheric conditions that are not easy to measure, make this model a valuable tool in understanding the physics of orographic precipitation.

The work presented in this thesis indicates that isotope-enabled models are useful tools in the study of meteorology and climatology. The addition of the isotopic tracers provides the opportunity to determine the isotopic signatures characteristic of various atmospheric processes. In this work, isotopes provided additional information in precipitation budgets to assess large-scale properties, like moisture source regions, but also prove useful in small scale cloud physics by tracking the microphysical processes leading to orographic precipitation. The successful implementation of this model makes it a useful tool to continue with further analysis of the topics presented in this thesis, but also in much wider applications, such as other types of synoptic events (monsoons, hurricanes, etc.) as well as paleoclimate studies.
Appendices
Appendix A

Supplemental Material for Chapter 2

This supplement contains further explanation of the method for calculating the large-scale tendencies, additional information about the vertical velocity profiles and the relationship between the isotopic content of boundary layer vapor and precipitation.

A.1 Large-scale tendencies

Fig. A.1 depicts the different terms used to calculate the tendencies of moisture and temperature. The first term represents the vertical advection. The second term represents the horizontal convergence, where the value of $\chi^*$ depends on whether there is large-scale convergence or divergence. In cases of large-scale divergence ($\frac{d(\rho w)}{dz} < 0$), the upwind value of $\chi^* = \bar{\chi}$, which represents the domain average value. However, when there is large-scale convergence ($\frac{d(\rho w)}{dz} > 0$), $\chi^* = \chi^{\text{ref}}$. Here $\chi^{\text{ref}}$ refers to the reference value that is calculated when the model reaches equilibrium (before WTG initiation) and represents the environment surrounding the convecting column. The final term, horizontal advection of upstream air, is neglected in our calculations because we do not have information concerning the horizontal gradient of isotopes in this region. We thus assume that the isotopologue distribution is horizontally uniform and therefore the advection term is 0.
A.2 Vertical Velocity Profile

Back and Bretherton (4) found distinct vertical velocity profiles associated with East and West Pacific locations and we noted in our paper that the structure of the vertical velocity profile could influence the strength of the amount effect as well as the potential role of boundary layer (BL) vapor. In Fig. A.2, we provide the domain average vertical velocity profiles, averaged over each SST regime. Excluding the coldest SST, we find that $w$ peaks in the upper troposphere. While we have not investigated this relationship further, it is possible that the top heavy profile could result in a lower $\delta D_v$ of converged vapor, as convergence will be strongest in the mid-troposphere where vapor is more depleted.

A.3 Boundary Layer Vapor

A secondary result of our simulations was to find that the isotope ratios of BL vapor and precipitation rates change together. Fig. A.3 depicts the linear relationship between the $\delta D_p$
Figure A.2: Vertical Velocity Profiles - Domain averaged vertical velocity profiles for each SST regime.
and $\delta D_{BL}$ values as well as the isotopic ratios of precipitation and BL vapor ($R_p$ and $R_{BL}$). For the $\delta D_p$ and $\delta D_{BL}$ plot, we include the best fit line, which compares well with that in Fig. 8 of Kurita (43) who found a linear relationship of $\delta D_{vap} = 0.72\delta D_p - 78.55$. We note that here we used the BL vapor for our comparison, which is the average of vapor in the lower 50mb, while Kurita (43) used near-surface vapor. The plot of $R_p$ and $R_{BL}$ is included to demonstrate that we expect to find an $\alpha_{eff} \sim 1.05 - 1.08$ since the $R_{BL}$ values are $\sim 0.9$

![Figure A.3: Isotopic Signatures and Ratios](image)

- Showing the linear relationship between (a) $\delta D_p$ and $\delta D_{BL}$ and (b) $R_p$ and $R_{BL}$, where $R$ is the ratio of HDO to H$_2$O. $R^2$ values are included in both cases as well as the best fit line in (a) for comparison with Kurita (43).
Appendix B

Supplemental Material for Chapter 3

This supplement contains further explanation of the methods used and assumptions made in order to add stable water isotopologue physics to the Thompson microphysics scheme.

B.1 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. (6, 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. (6), 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
B.1 Incorporating Isotopologues Into the Microphysics

For example, the freezing of cloud liquid droplets (wfz) to form cloud ice transfers heavy isotopologues to cloud ice as follows:

\[
\frac{dr'_i}{dt}_{wfz} = \frac{dr_i}{dt}_{wfz} \mathcal{R}_c
\]  

(B.1)

where \( \mathcal{R}_c = r'_c/r_c \) 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. (6, 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:

\[
\mathcal{R}_c = \alpha_l \mathcal{R}_v
\]  

(B.2)

where \( \mathcal{R}_v \) is the isotopic ratio of water vapor and \( \alpha_l \) is the equilibrium fractionation coefficient for liquid (Majoube (60)). As the equilibration time for isotopic composition of small liquid
B.1 Incorporating Isotopologues Into the Microphysics

water droplets is on the order of a few seconds (Ciais and Jouzel (14)), 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

$$\frac{dr_i}{dt}_{dep} = \alpha_s \alpha_k R_v \frac{dr_i}{dt}_{dep}$$

(Ciais and Jouzel (14)), where $\alpha_s$ is the equilibrium fractionation coefficient for ice (Majoube (59), Merlivat and Nief (62) T and the kinetic fractionation coefficient, $\alpha_k$, represents the effects of the relative diffusion of the heavy and light isotopologues (Jouzel and Merlivat (37)).

The effects of these two processes on the isotopic composition of liquid and ice in mixed-phase clouds is depicted in Fig. B.1. 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 B.2 and B.3 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. (55),

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

(B.4)

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 (85, 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.

**Figure B.1: Saturation Ratios and $\delta^{18}O$ with Temperature** - [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.
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