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# Rain evaporation, snow melt and entrainment at the heart of water vapor isotopic variations in the tropical troposphere, according to large-eddy simulations and a two-column model

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1 Rain evaporation, snow melt and entrainment at the heart of water  
2 vapor isotopic variations in the tropical troposphere, according to  
3 large-eddy simulations and a two-column model

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5 September 15, 2020

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8  
9 (<=140 characters)

- 10 1. Rain evaporation enriches the tropospheric water vapor more if more snow sublimates before melting and  
11 more rain evaporates.  
12 2. Entrainment of dry air reduces the vertical isotopic gradient and limits the tropospheric depletion of water  
13 vapor.  
14 3. These mechanisms explain the increased depletion of tropospheric water vapor as tropospheric relative hu-  
15 midity increases.

16 **Abstract**

17 (<=250 words)

18  
19 The goal of this study is twofold. First, we aim at developing a simple model as an interpretative framework  
20 for the water vapor isotopic variations in the tropical troposphere over the ocean. We use large-eddy simulations  
21 to justify the underlying assumptions of this simple model, to constrain its input parameters and to evaluate its  
22 results. Second, we aim at interpreting the depletion of the water vapor isotopic composition in the lower and mid-  
23 troposphere as precipitation increases, which is salient features in tropical oceanic observations. This constitutes  
24 a stringent test on the relevance of our interpretative framework. Previous studies, based on observations or  
25 models with parameterized convection, have highlighted the roles of deep convective and meso-scale downdrafts,  
26 rain evaporation, rain-vapor diffusive exchanges and mixing processes.

27 The interpretative framework that we develop is a two-column model representing the net ascent in clouds and  
28 the subsiding environment. We show that the mechanisms for depleting the troposphere when precipitation rate  
29 is larger all stem from the higher tropospheric relative humidity. First, when the relative humidity is larger, less  
30 snow sublimates before melting and a smaller fraction of rain evaporates. Both effects lead to more depleted rain  
31 evaporation and eventually more depleted water vapor. This mechanism dominates in regimes of large-scale ascent.  
32 Second, the entrainment of dry air into clouds reduces the vertical isotopic gradient and limits the tropospheric  
33 depletion of water vapor. This mechanism dominates in regimes of large-scale descent.

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## 64 1 Introduction

### 65 1.1 Looking for an interpretative framework for water vapor isotopic profiles

66 The isotopic composition of water vapor (e.g. its Deuterium content, commonly expressed as  $\delta D = (R/R_{SMOW} - 1) \times$   
67 1000 in ‰, where  $R$  is the ratio of Deuterium over Hydrogen atoms in the water, and SMOW is the Standard Mean  
68 Ocean Water reference) evolves along the water cycle as phase changes are associated with isotopic fractiona-  
69 tion. Consequently the isotopic composition of precipitation recorded in paleoclimate archives has significantly  
70 contributed to the reconstruction of past hydrological changes ([Wang et al., 2001]). It has also been suggested  
71 that observed isotopic composition of water vapor could help better understand atmospheric processes and evaluate  
72 their representation in climate models, in particular convective processes ([Schmidt et al., 2005, Bony et al., 2008,  
73 Lee et al., 2009, Field et al., 2014]). Yet, water isotopes remain rarely used beyond the isotopic community to an-  
74 swer today’s pressing climate questions. A prerequisite to better assess the strengths and weaknesses of the isotopic  
75 tool is to better understand what controls spatio-temporal variations in water vapor isotopic composition ( $\delta D_v$ )  
76 through the tropical troposphere, in particular how convective processes drive these variations.

77 While there are interpretative frameworks for the controls of free tropospheric humidity ([Sherwood, 1996,  
78 Romps, 2014]), no such interpretative framework exist for water isotopes beyond the simple Rayleigh distillation or  
79 mixing lines ([Worden et al., 2007, Bailey et al., 2017]). We aim at filling this gap here. The first goal of this paper  
80 is thus to design an interpretative framework that could be useful in the future to interpret water vapor isotopic  
81 variations in the tropical troposphere in a wide range of contexts. Analogous to that for relative humidity, this  
82 framework will also allow us to compare the processes controlling relative humidity and isotopic composition.

83 Frameworks do exist to interpret the  $\delta D_v$  in the sub-cloud layer (SCL), such as the [Merlivat and Jouzel, 1979]  
84 closure assumption, later extended to account for updrafts and downdrafts ([Risi et al., 2020]). This latter frame-  
85 work highlighted the need to know the steepness of the relationship between  $\delta D_v$  and humidity  $q$  as they evolve  
86 with altitude. This motivates us to develop a framework that allows us to predict the  $\delta D_v$  evolution with altitude  
87 in the troposphere.

## 88 1.2 Large-eddy simulation analysis as a guide to design the interpretative framework

89 Many previous studies investigating the processes controlling tropospheric  $\delta D_v$  have relied on general circulation  
90 models that include convective parameterization ([Lee et al., 2007, Bony et al., 2008, Risi et al., 2008, Field et al., 2010]).  
91 However, parameterizations include numerous simplifications or assumptions that are responsible for a significant  
92 part of biases in the present climate simulated by GCMs and of inter-model spread in climate change projections  
93 ([Randall et al., 2003, Stevens and Bony, 2013, Webb et al., 2015]). Here, we thus use large-eddy simulations (LES)  
94 as a guide to design the interpretative framework. These high-resolution simulations allows us to explicitly resolve  
95 convective motions. These simulations will also provide the input parameters for our interpretative framework, and  
96 a benchmark to evaluate its results.

## 97 1.3 Interpreting the amount effect

98 In the tropics, it has long been observed that in average over a month or longer, the isotopic composition of  
99 the rain is more depleted when the precipitation rate is stronger ([Dansgaard, 1964, Rozanski et al., 1993]). This  
100 phenomenon is called the “amount effect”. Since most of the precipitation in the tropics is associated with deep  
101 convection, understanding the amount effect is a stringent test on our understanding of how convective processes  
102 affect the water vapor isotopic composition through the tropical troposphere. The capacity of our interpretative  
103 framework to predict the amount effect will thus be a stringent test on its relevance. The second goal of this study  
104 is thus to better understand the processes underlying the amount effect, using the interpretative framework.

105 [Dansgaard, 1964] hypothesized that the amount effect could be due to the progressive depletion by convective  
106 storms of the vapor from which the rain forms, and to rain evaporation and diffusive exchanges between the  
107 rain and the vapor. If the case, the amount effect crucially depends on the isotopic composition of the vapor.  
108 From a column-integrated water budget perspective, the isotopic composition of precipitation depends on the  
109 relative proportion of the precipitation that originates from horizontal advection and from surface evaporation  
110 ([Lee et al., 2007, Moore et al., 2014]). More precipitation is generally associated with more large-scale ascent and  
111 thus more large-scale convergence. Since vapor from horizontal advection is more depleted than water from surface  
112 evaporation because it has already been processed in clouds, the precipitation is more depleted. In this view as  
113 well, the amount effect crucially depends on the isotopic composition of the vapor.

114 Water isotopic measurements in the vapor phase, by satellite or in-situ, have confirmed that increased precipi-  
115 tation was associated with more depleted water vapor ([Worden et al., 2007, Kurita, 2013, Lacour et al., 2017]).  
116 Hereafter we will call this the “vapor amount effect”. Actually, the precipitation and water vapor isotopic composi-  
117 tion often vary in concert ([Kurita, 2013, Tremoy et al., 2014]). In this paper, we will thus focus on understanding  
118 the processes underlying the “vapor amount effect”.

119 From previous studies, four hypotheses have emerged to explain the “vapor amount effect”:

- 120 1. Hypothesis 1: As precipitation rate increases, convective or meso-scale downdrafts bring more depleted vapor  
121 from above into the sub-cloud layer (SCL) ([Risi et al., 2008, Kurita et al., 2011, Kurita, 2013]). This is  
122 because the water vapor  $\delta D$  ( $\delta D_v$ ) generally decreases with altitude, because as water vapor is lost through  
123 condensation and specific humidity  $q$  decreases, heavy isotopes are preferentially lost in the condensed phase.  
124 This phenomenon is called Rayleigh distillation and is plotted in a  $q-\delta D_v$  diagram in Figure 1 (blue). However,  
125 downdrafts would both decrease  $\delta D_v$  and  $q$ . This hypothesis is thus inconsistent with the observation that  $q$   
126 generally increases while  $\delta D_v$  decreases as precipitation rate increases. By itself, this hypothesis cannot be  
127 sufficient.
- 128 2. Hypothesis 2: As precipitation rate increases, the moistening effect by rain evaporation increases. If rain  
129 evaporation is more depleted than the vapor, then it depletes the vapor ([Worden et al., 2007]). The effect  
130 of rain evaporation is represented in purple in Figure 1. If the evaporated fraction of the rain is small, rain  
131 evaporation acts to deplete the vapor because light isotopes preferentially evaporate.
- 132 3. Hypothesis 3: As precipitation rate increases, the rain evaporation is more depleted. For example, if precipi-  
133 tation rate increases, the fraction of rain that evaporates is smaller, which leads the evaporation to be more  
134 depleted ([Risi et al., 2008, Risi et al., 2010, Tremoy et al., 2014, Risi et al., 2020], Figure 1, purple). Alter-  
135 natively, larger precipitation rates typically occur in moister environments, which favors rain-vapor diffusive  
136 exchanges rather than pure evaporation [Lawrence et al., 2004, Lee and Fung, 2008]. Since rain comes from  
137 higher altitudes, it is more depleted than if in equilibrium with the local vapor, and thus rain-vapor diffusive  
138 exchanges favor more depleted evaporation.

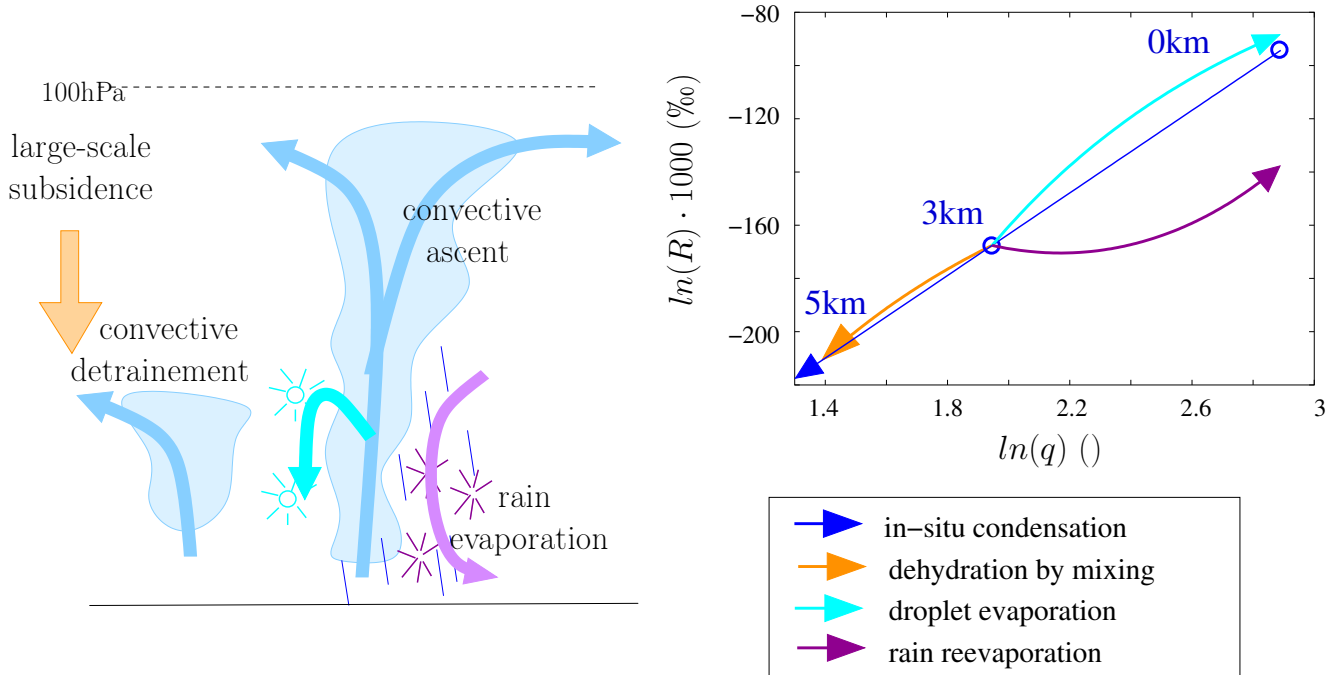


Figure 1: Schematic showing the influence of different processes on  $q$  and  $\delta D_v$ . Condensation and immediate loss of condensate in convective updrafts leads to drying and depleting the water vapor following Rayleigh distillation (blue). During evaporation of cloud droplets, each droplet evaporates totally. Since cloud droplets are enriched in heavy isotopes, this moistens the air and enriches the vapor (cyan). In contrast, during evaporation of rain drops, each drop evaporates progressively. Whereas it moistens the air, it depletes the vapor for small evaporation fractions and enriches the vapor for large evaporation fraction (purple). Finally, mixing of subsiding air with air detrained from convective updrafts dehydrates the air and depletes the vapor following a hyperbolic curve, leading to higher  $\delta D_v$  for a given  $q$  compared to Rayleigh (orange). The curves are plotted following simple Rayleigh and mixing lines with approximate values taken from the control LES described later in the article.

139 4. Hypothesis 4: As precipitation rate increases, dehydration by mixing dominates relatively to dehydration  
 140 by condensation. Due to the hyperbolic shape of the mixing lines in a  $q - \delta D$  diagram, dehydration by  
 141 mixing with a dry source is associated with a smaller depletion than predicted by Rayleigh distillation  
 142 [Dessler and Sherwood, 2003, Galewsky and Hurley, 2010, Galewsky and Rabanus, 2016] (Figure 1 orange).  
 143 [Bailey et al., 2017] argues that in more subsiding regions, mid-tropospheric vapor is more enriched for a  
 144 given specific humidity because air masses result from the mixing between air subsiding from a higher altitude  
 145 and shallow convective detrainment.

146 We notice that hypothesis 2-4 are all associated with an increased steepness as precipitation rate increases (Figure  
 147 1), consistent with the key role of the steepness of the  $q - \delta D_v$  relationship in depleting the SCL water vapor  
 148 highlighted by [Risi et al., 2020]. The mechanisms underlying these hypotheses will thus have to be key ingredients  
 149 of our interpretative framework.

150 The LES will be described and analyzed in section 2. The interpretative framework will be designed and used  
 151 to interpret the “vapor amount effect” in section 3. Finally, section 4 will offer a summary, some discussion and  
 152 perspectives.

## 153 2 Large-eddy simulations

### 154 2.1 Model and simulations

155 We use the same LES model as in [Risi et al., 2020], namely the System for Atmospheric Modeling (SAM) non-  
 156 hydrostatic model ([Khairoutdinov and Randall, 2003]), version 6.10.9, which is enabled with water isotopes ([Blossey et al., 2010])  
 157 This model solves anelastic conservation equations for momentum, mass, energy and water, which is present in the

158 model under six phases: water vapor, cloud liquid, cloud ice, precipitating liquid, precipitating snow, and precipitat-  
 159 ing graupel. We use the bulk, mixed-phase microphysical parameterization from [Thompson et al., 2008] in which  
 160 water isotopes were implemented ([Moore et al., 2016]).

161 The control simulation (“ctrl”) is three-dimensional, with a doubly-periodic domain of 96 km×96 km. The  
 162 horizontal resolution is 750 m. There are 96 vertical levels. The simulation is run in radiative-convective equilibrium  
 163 over an ocean surface. The sea surface temperature (SST) is 30°C. There is no rotation and no diurnal cycle. In  
 164 this simulation, there is no large-scale circulation.

165 The amount effect can be seen only if the precipitation increase is associated with a change in the large-scale  
 166 circulation ([Bony et al., 2008, Dee et al., 2018, Risi et al., 2020]). To compare ctrl to simulations with larger and  
 167 smaller precipitation rate, we thus run simulations with a large-scale vertical velocity profile,  $\omega_{LS}$ . This profile is  
 168 used to compute large-scale tendencies in temperature, humidity and water vapor isotopic composition. We com-  
 169 pute large-scale vertical advection by a simple upstream scheme [Godunov, 1959]. In the computation, large-scale  
 170 horizontal gradients in temperature, humidity and isotopic composition are neglected, i.e. there are no large-scale  
 171 horizontal advective forcing terms. The large-scale vertical velocity  $\omega_{LS}$  has a cubic shape so as to reach its maxi-  
 172 mum  $\omega_{LSmax}$  at a pressure  $p_{max}=500$  hPa and to smoothly reach 0 at the surface and at 100 hPa [Bony et al., 2008].  
 173 We analyze here simulations with  $\omega_{LSmax}=-60$  hPa/d (“ $\omega_{LS}-60$ ”), corresponding to typical deep convective con-  
 174 ditions in the inter-tropical convergence zone, and  $\omega_{LSmax}=+20$  hPa/d (“ $\omega_{LS}+20$ ”), corresponding to subsiding  
 175 trade-wind conditions.

176 The simulations are run for 50 days and the last 10 days are analyzed. We use instantaneous outputs that are  
 177 generated at the end of each simulation day.

## 178 2.2 Simulated amount effect and basic features

179 Figure 2a shows that the ctrl,  $\omega_{LS}-60$  and  $\omega_{LS}+20$  simulations allow us to capture the amount effect both in  
 180 the vapor and in the precipitation, which vary in concert. In case of large-scale ascent, the domain-mean relative  
 181 humidity is larger than in ctrl by more than 10% (Figure 2b), while  $\delta D_v$  is more depleted by more than 50%, in  
 182 most of the troposphere (Figure 2c). We can see that the  $\delta D_v$  difference at all altitudes is similar to that in the SCL.  
 183 This confirms that understanding what controls the SCL  $\delta D_v$  is key to understand what controls  $\delta D_v$  at all altitudes  
 184 ([Risi et al., 2020]). This also explains why models that assume constant SCL  $\delta D_v$  show very little sensitivity to  
 185 all kinds of convective and microphysical processes ([Duan et al., 2018]). We can also see that Rayleigh distillation  
 186 alone (dashed line) is a poor predictor of  $\delta D_v$  profiles and of their sensitivity to large-scale ascent.

## 187 2.3 Steepness of the $q - \delta D_v$ relationship

188 With the goal of understanding the amount effect, as a first step [Risi et al., 2020] focused on understanding what  
 189 controls the  $\delta D_v$  in the SCL, because the SCL ultimately feeds the water vapor at all altitudes in the troposphere.  
 190 They identified the key role of the steepness of the  $q - \delta D_v$  relationship of vertical profiles in the lower troposphere.  
 191 This steepness determines the efficiency with which updrafts and downdrafts near the SCL top deplete the SCL.  
 192 To understand what controls  $\delta D_v$  in the SCL and thus everywhere in the troposphere, we thus need to understand  
 193 what controls the steepness of the  $q - \delta D_v$  relationship.

194 The vertical profiles of  $\ln(R)$  as a function of  $\ln(q)$  for each simulation show a nearly linear relationship (Figure  
 195 2d), consistent with a Rayleigh-like distillation process (Figure 1). If the vertical profiles were dominated by mixing  
 196 processes, as in hypothesis 4, the relationship would look convex ([Bailey et al., 2017], Figure 1 orange). Rather,  
 197 in  $\omega_{LS}-60$ , the curve looks concave near the melting level, consistent with an effect of rain evaporation (Figure 1  
 198 purple).

199 To better quantify the steepness of the  $q - \delta D_v$  relationship, we define the  $q - \delta D_v$  steepness  $\alpha_z$ , as the  
 200 effective fractionation coefficient that would be needed in a distillation to fit the simulated joint  $q - \delta D$  evolution  
 201 ([Risi et al., 2020]):

$$\alpha_z = 1 + \frac{\ln(R(z)/R(z-dz))}{\ln(q(z)/d(z-dz))} \quad (1)$$

202 The steepness  $\alpha_z$  in the ctrl simulation is smaller than that predicted by Rayleigh distillation, i.e.  $\alpha_z <$   
 203  $\alpha_{eq}$ , especially at higher altitudes (Figure 2e) (section 3.2.2 will demonstrate that it is due to entrainment). In  
 204 case of large-scale ascent, just above the SCL top,  $\alpha_z - 1$  is more than three times larger in  $\omega_{LS}-60$  than in  
 205 ctrl. The increased steepness leads the updrafts and downdrafts to deplete more efficiently the SCL water vapor  
 206 ([Risi et al., 2020]), and eventually the full tropospheric profile. Conversely, in  $\omega_{LS}+20$ , the steepness is smaller and

207 responsible for more enriched SCL. Our interpretative framework will allow us to interpret these features (section  
208 3).

## 209 2.4 Effect of de-activating rain-vapor exchanges

210 According to hypotheses 2 and 3, the isotopic composition of the rain plays a key role in the “vapor amount effect”.  
211 At a given instant and for a small increment of rain evaporation fraction, the isotopic composition of the evaporation  
212 flux  $R_{ev}$  is simulated following [Craig and Gordon, 1965]:

$$R_{ev} = \frac{R_r/\alpha_{eq} - h_{ev} \cdot R_v}{\alpha_K \cdot (1 - h_{ev})}$$

213 where  $R_r$  and  $R_v$  are the isotopic ratios in the liquid water and water vapor,  $\alpha_{eq}$  and  $\alpha_K$  are the equilibrium  
214 and kinetic fractionation coefficient and  $h_{ev}$  is the relative humidity. In order to test hypotheses 2 and 3, we run  
215 additional simulations similar to ctrl and  $\omega_{LS} - 60$  but without any fractionation during rain evaporation, named  
216 “nofrac”, where  $R_{ev} = R_r$ . We also run additional simulations with fractionation during evaporation, but with  
217 rain-vapor diffusive exchanges de-activated, named “nodiff”, where  $R_{ev} = R_r/\alpha_{eq}/\alpha_K$ .

218 When fractionation during rain evaporation is de-activated,  $\delta D_v$  is more enriched, consistent with a more  
219 enriched composition of rain evaporation (Figure 3a). In addition, the  $\delta D_v$  difference between  $\omega_{LS} - 60$  and ctrl  
220 is reduced by about 70% compared to when all isotopic exchanges are considered (Figure 3c, red). This confirms  
221 that fractionation during rain evaporation plays a key role in the “vapor amount effect”. When rain-vapor diffusive  
222 exchanges are de-activated, the  $\delta D_v$  difference between  $\omega_{LS} - 60$  and ctrl is reduced by about 30% compared to  
223 when all isotopic exchanges are considered (Figure 3c, green). Rain-vapor vapor diffusive exchanges thus play an  
224 important role as well.

225 We note that the  $\delta D_v$  difference between the simulations is remarkably constant with altitude (Figure 3a,c),  
226 although we expect strong vertical variations in rain evaporation. This is consistent with the important role of the  
227 SCL  $\delta D_v$  as an initial condition for the full  $\delta D_v$  profile. We also note that more enriched  $\delta D_v$  profiles are associated  
228 with a reduced lower-tropospheric steepness  $\alpha_z$  just above the SCL, and larger  $\delta D_v$  differences between simulations  
229 are associated with larger differences in lower-tropospheric  $\alpha_z$ . This is consistent with the SCL  $\delta D_v$  being mainly  
230 driven by the steepness  $\alpha_z$  just above the SCL ([Risi et al., 2020]). Finally, the reduced “vapor amount effect” in  
231 “nofrac” leads to a reduced amount effect in the precipitation  $\delta D$  as well (Figure 3c, circles). This shows that the  
232 column-integrated water budget ([Lee et al., 2007, Moore et al., 2014]) cannot by itself predict the amount effect,  
233 since it depends on the isotopic composition of the advected vapor, which can greatly vary depending on the detailed  
234 representation of rain evaporation processes.

235 To summarize, in the total  $\delta D_v$  difference between  $\omega_{LS} - 60$  and ctrl, there is about one third due to fractionation  
236 during evaporation, one third due to rain-vapor diffusive exchanges, and one third that would remain even in absence  
237 of any fractionation during evaporation. These tests suggest that hypotheses 2 and/or 3 play a key role in the “vapor  
238 amount effect”. In the next sections, we aim at better understanding how rain evaporation impacts  $\delta D_v$  profiles.

## 239 2.5 Vertical profiles binned by moist static energy

240 Previous studies have shown that analyzing variables in isentropic coordinates was a powerful tool to categorize  
241 the different convective structures: undiluted updrafts, diluted updrafts, saturated and unsaturated downdrafts,  
242 and the environment ([Kuang and Bretherton, 2006, Pauluis and Mrowiec, 2013]). This method also has the ad-  
243 vantage of filtering out gravity waves. It has been applied to the analysis of a wide range of convective systems  
244 ([Mrowiec et al., 2015, Mrowiec et al., 2016, Dauhut et al., 2017, Chen et al., 2018]).

245 Here we use the frozen moist static energy  $m$  as a conserved variable because it is conserved during condensation  
246 and evaporation of both liquid and ice water ([Muller and Romps, 2018, Hohenegger and Bretherton, 2011]).

$$m = c_{pd} \cdot T + g \cdot z + L_v \cdot q_v - L_f \cdot q_i$$

247 where  $c_{pd}$  is the specific heat of dry air,  $T$  is temperature,  $g$  is gravity,  $z$  is altitude,  $L_v$  and  $L_f$  are the latent  
248 heat of vaporization and fusion, and  $q_i$  is the total ice water content (cloud ice, graupel and snow). At each level,  
249 we categorize all grid points into bins of  $m$  with a width of 0.4 kJ/kg.

250 The domain-mean frozen moist static energy  $m$  decreases from the upper troposphere down to about 5 km, due  
251 to the loss of energy by radiative cooling, and then increases down to the surface due to the input of energy by  
252 surface fluxes (Figure 4, solid black line). Based on this diagram, we can identify four kinds of air parcels:

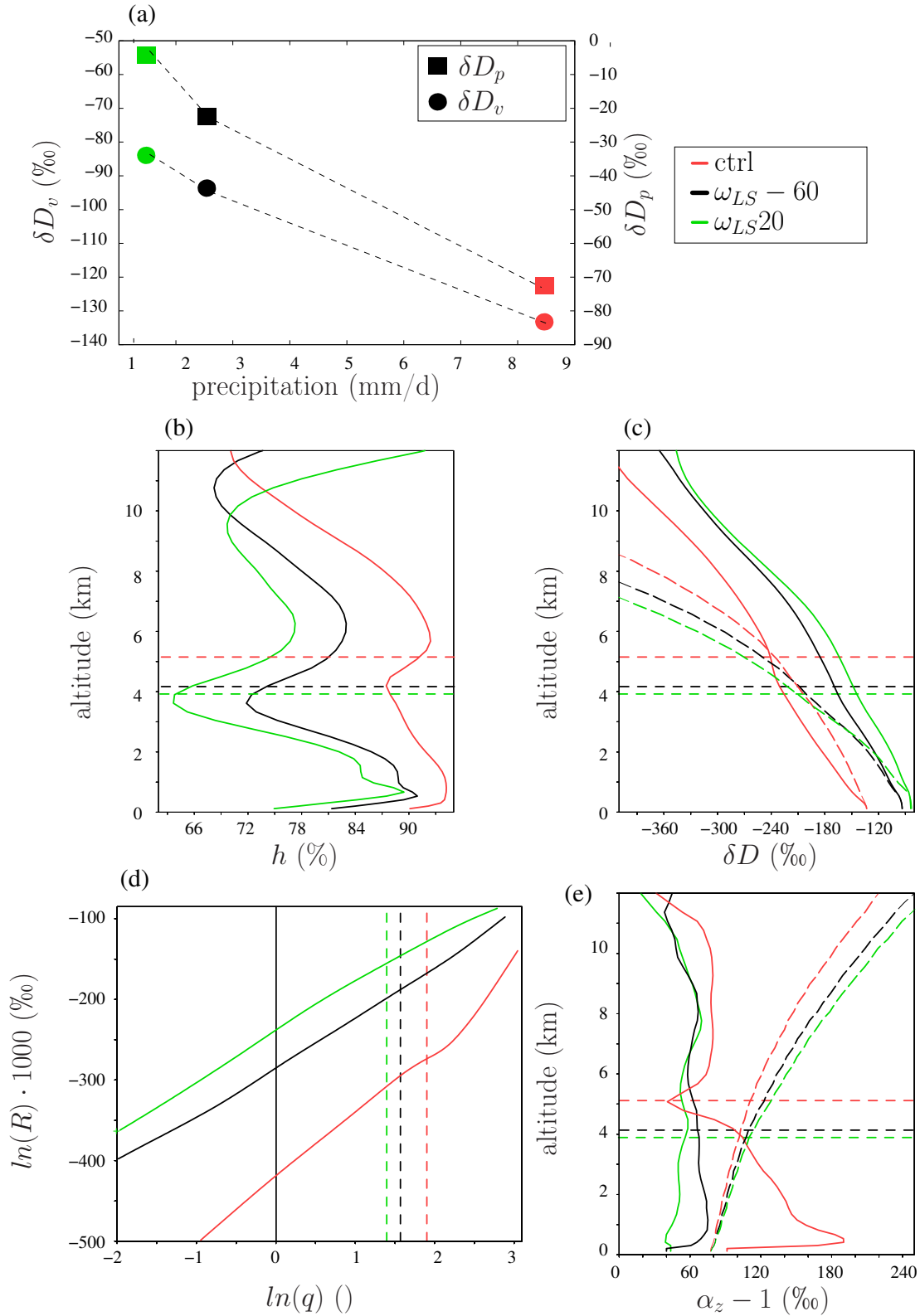


Figure 2: (a) Water vapor (circles) and precipitation (squares)  $\delta D_v$  as a function of precipitation. Vertical distribution of relative humidity (b), water vapor  $\delta D$  (c) and  $\alpha_z$  (e) in ctrl (black),  $\omega_{LS} - 60$  (red) and  $\omega_{LS} + 20$  (green). (c)  $\ln(R(z)) \cdot 1000$  as a function of  $\ln(q(z))$  for different altitudes. In c and e, dashed lines indicate the prediction by Rayleigh distillation. The horizontal lines show the altitude of the melting level.



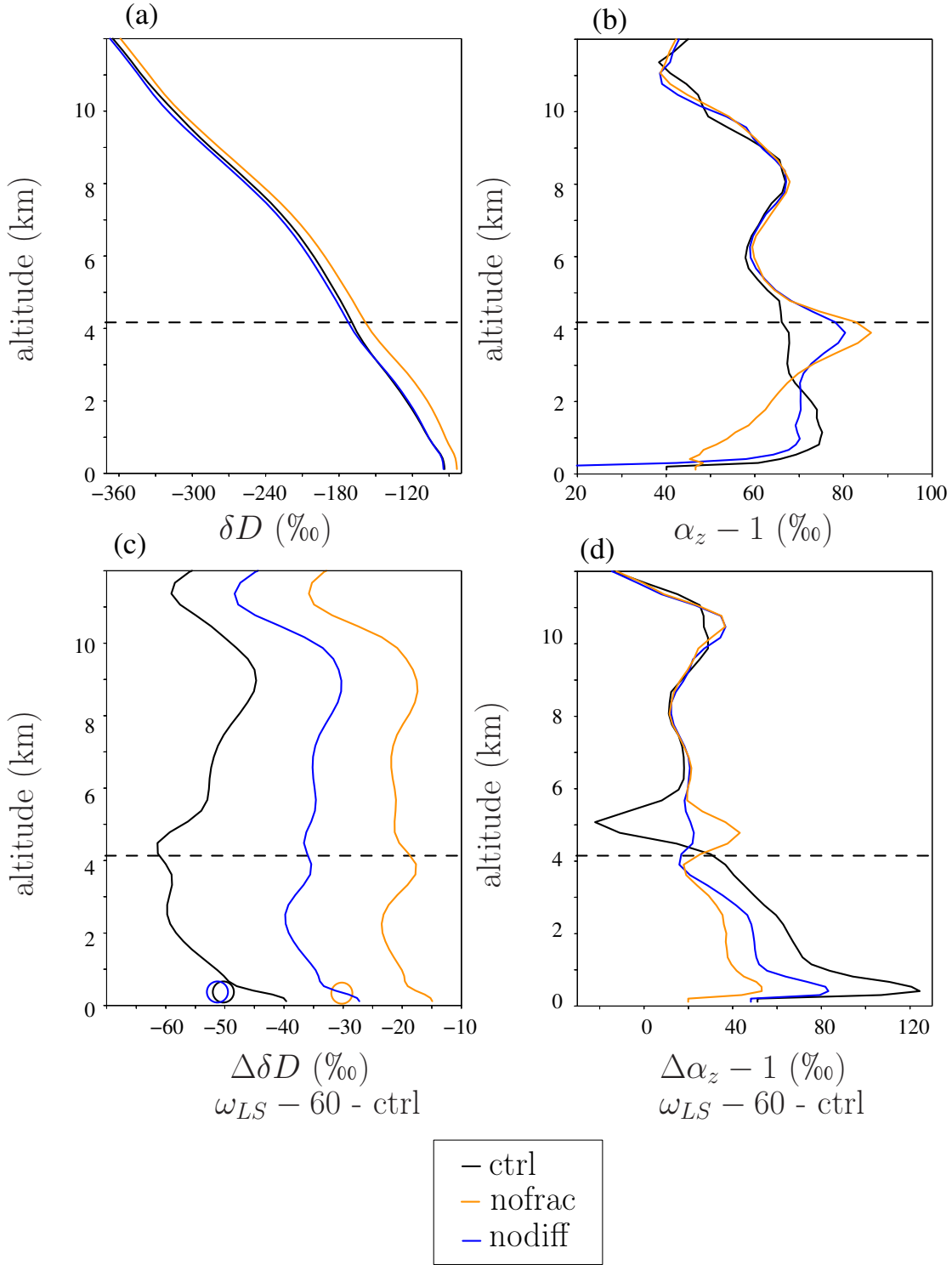


Figure 3: (a) Vertical distribution of  $\delta D_v$  for ctrl, when fractionation during liquid evaporation is turned on (black) or off (red) and when liquid-vapor equilibration is turned off (green). (b) Same as (a) for the vertical profiles of  $\alpha_z$ . (c)  $\delta D_v$  difference between the  $\omega_{LS} - 60$  and ctrl, with (black) and without (red) fractionation during evaporation and when liquid-vapor equilibration is turned off (green). The circles illustrate the difference in the precipitation  $\delta D$ . (d) Same as (c) but for  $\alpha_z$ .

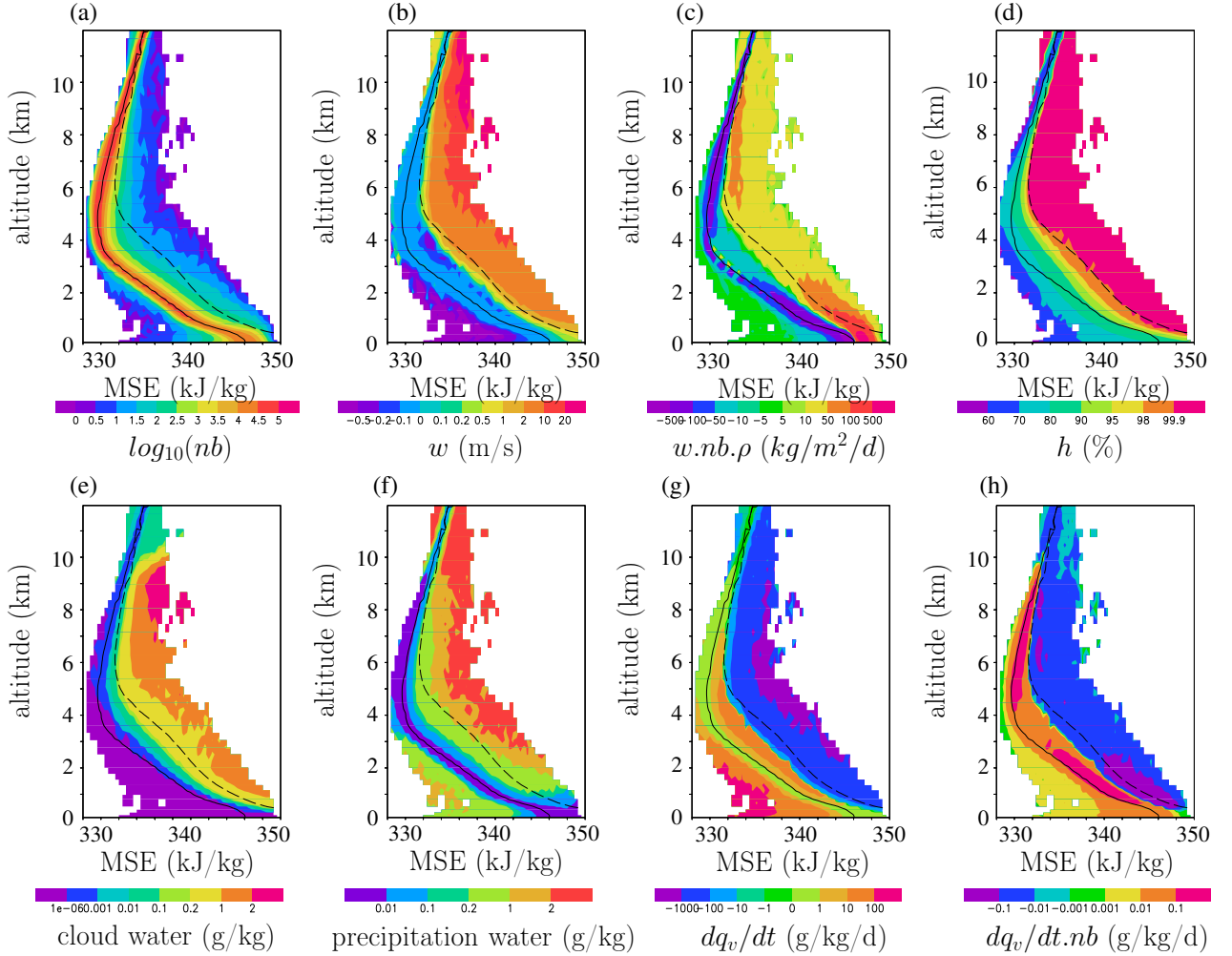


Figure 4: Variables binned as a function of frozen moist static energy  $m$  and of altitude, for the ctrl simulation: (a) number of samples, (b) vertical velocity anomaly, (c) vertical mass flux (vertical velocity multiplied by the proportion of samples and density), (d) relative humidity, (e) cloud water content mixing ratio (liquid and ice), (f) precipitating water mixing ratio (rain, graupel and snow), (g) evaporation and condensation tendency (positive in case of evaporation, negative in case of condensation), (h)  $\delta D_v$  anomaly, (i)  $(\phi - 1) \cdot 1000$ , where  $\phi = R_{ev}/R_v$ ; it is expressed in ‰. The solid black line show the domain-mean frozen moist static energy, while the dashed black line shows the frozen moist static energy at saturation.

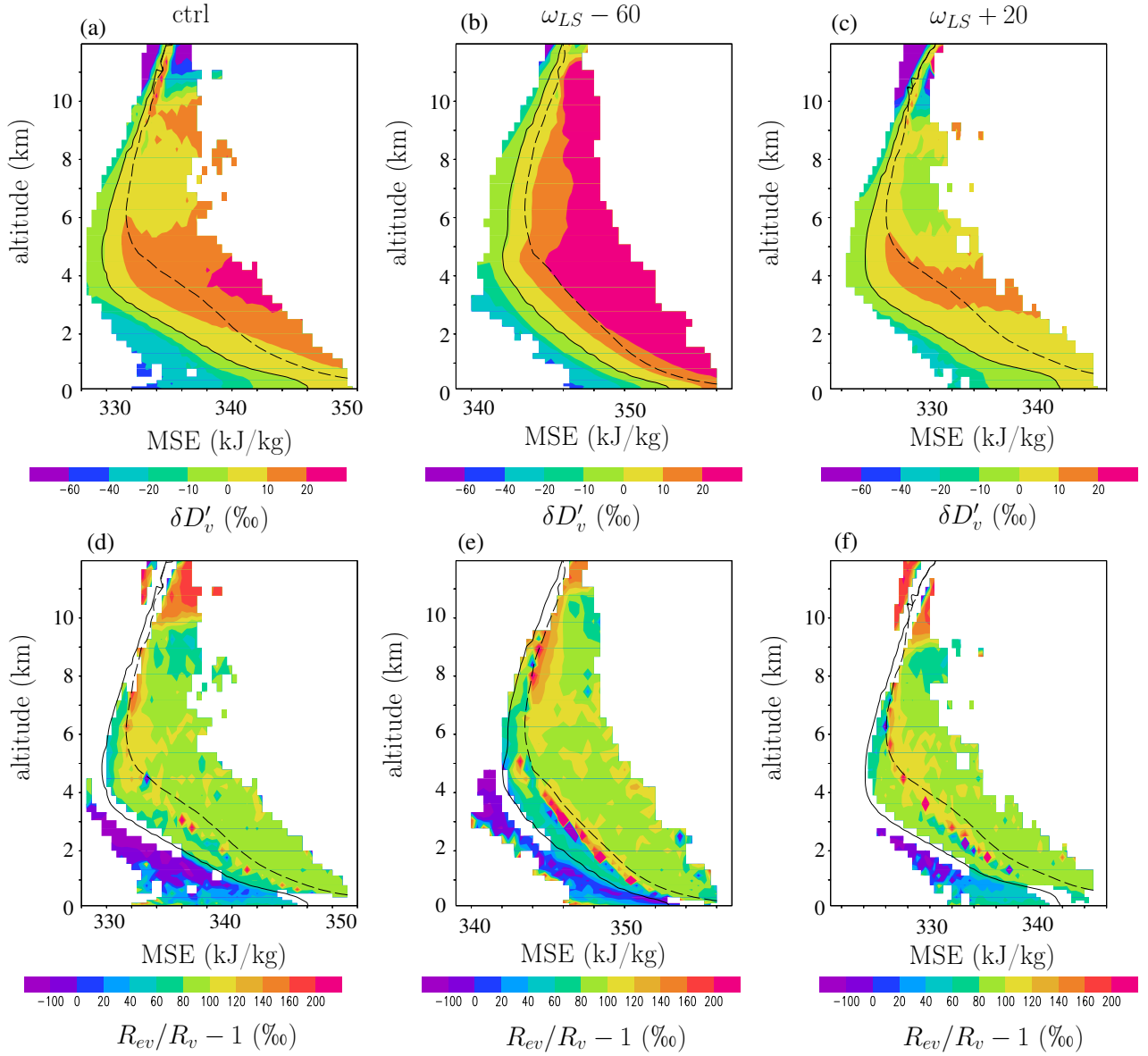


Figure 5: (a,d) As for Figure 4 but for (a)  $\delta D_v$  anomaly, (d)  $(\phi - 1) \cdot 1000$ , where  $\phi = R_{ev}/R_v$ ; it is expressed in ‰. (b,e) As for (a,d) but for  $\omega_{LS} - 60$ . (c,f) As for (a,d) but for  $\omega_{LS} + 20$ .

- 253 1. Environment. They correspond to air parcels whose  $m$  is close to the domain-mean (solid black). They are the  
 254 most numerous (Figure 4a). Their vertical velocity is slightly descending (Figure 4b), but because they are  
 255 very numerous, they account for most of the downward mass flux (Figure 4c). Their relative humidity is close  
 256 to the domain-mean (Figure 4d), they contain only a small cloud water and rain content and phase changes  
 257 are very slow (Figure 4e-g). However, because they cover most of the domain, they contribute significantly to  
 258 the evaporation in the domain-mean (Figure 4h).
- 259 2. Cloudy updrafts. They correspond to air parcels on the right of the domain-mean  $m$  and whose bin-mean  
 260 vertical velocity is ascending (Figure 4b). If air rose adiabatically from the SCL, they would conserve their  
 261  $m$  and they would be located completely on the right of the diagram. In practice,  $m$  decrease because the  
 262 environment air is progressively entrained into ascending parcels. In the diagrams, parcels are more diluted  
 263 when they are closer to the domain-mean, and less diluted when they are more to the right. In spite of their  
 264 dilution with the environment, their humidity is at saturation (Figure 4d). They contain a lot of cloud and  
 265 precipitating water, and vapor undergoes condensation (Figure 4e-g).
- 266 3. Cloudy downdrafts. They correspond to air parcels on the right of the domain-mean  $m$  and whose bin-mean  
 267 vertical velocity is descending (Figure 4b). They are more diluted than cloudy updrafts. Their humidity is  
 268 below saturation (Figure 4d). They contain cloud and precipitating water, but these hydrometeor undergo  
 269 evaporation (Figure 4e-g). Located around the cloudy updrafts in the real space, they mainly correspond to  
 270 subsiding shells ([Glenn and Krueger, 2014]).
- 271 4. Precipitating downdrafts. They correspond air parcels on the bottom-left of the diagrams, with lower  $m$   
 272 relative to the domain-mean. They are among the most strongly descending air parcels (Figure 4b), but  
 273 since they are scarce (Figure 4b), they contribute little to the total descending mass flux (Figure 4c). They  
 274 are very dry, with no cloud water, but with precipitating water (Figure 4d-f). We interpret these parcels  
 275 as unsaturated, precipitating downdrafts. Strong evaporation of rain occur in these downdrafts (Figure 4g),  
 276 but because they cover only a small fraction of the domain, they contribute little to the evaporation in the  
 277 domain-mean (Figure 4h).

278 The isotopic composition of water vapor is the strongest in the least diluted updrafts, and the most depleted in the  
 279 precipitating downdrafts (Figure 5a). To assess the effect of phase changes, we plot  $\phi = R_{ev}/R_v$ , where  $R_{ev}$  is the  
 280 ratio of the water vapor tendency associated with phase changes (evaporation in downdrafts and in the environment,  
 281 or condensation in cloudy updrafts) and  $R_v$  is the isotopic ratio of the water vapor. In cloudy updrafts,  $\phi - 1$  is  
 282 about 100% in the lower troposphere and increases with height (Figure 5b). This corresponds roughly to equilibrium  
 283 fractionation during condensation. In cloudy downdrafts,  $\phi - 1$  is also about 100%. This means that cloud droplets  
 284 evaporate totally without fractionation. In contrast, in precipitating downdrafts,  $\phi - 1$  is much lower. It is around  
 285 30% below 1 km. The fact that  $\phi - 1$  is positive is consistent with the fact that rain evaporation in the SCL acts  
 286 to slightly enrich the water vapor ([Risi et al., 2020]). In contrast, between 2 and 3 km,  $\phi - 1$  is around -100%: at  
 287 these levels, rain evaporation acts to deplete the water vapor, consistent with [Worden et al., 2007].

288 These diagrams look qualitatively similar for the other simulations. One noticeable difference is that in  $\omega_{LS} - 60$ ,  
 289 the  $\delta D_v$  contrast between the environment and the cloudy regions is larger (Figure 5b). This may be associated  
 290 with the more depleted evaporation of the rain in precipitating downdrafts, and even of cloud droplets in cloudy  
 291 downdrafts (Figure 5e). Conversely, in  $\omega_{LS} + 20$ , the  $\delta D_v$  contrast between the environment and the cloudy regions  
 292 is larger (Figure 5c). To quantitatively compare the different simulations, now we plot vertical profiles of variables  
 293 in average over cloudy regions and over the environment.

## 294 2.6 Vertical profiles for cloudy regions and for the environment

295 Here we chose to define cloudy regions as all parcels with a cloud (liquid or ice) water content greater than  $10^{-6}$  g/kg  
 296 (e.g. [Thayer-Calder and Randall, 2015]). In this loose definition, “cloudy regions” correspond to both cloudy  
 297 updrafts and downdrafts, while the “environment” includes both the environment and precipitating downdrafts.  
 298 Including the cloudy downdrafts into the cloudy regions is justified by the fact that a significant portion of the  
 299 water condensed in cloudy updrafts subsequently evaporate in these cloudy downdrafts, without directly affecting  
 300 the environment. Our results below are not crucially sensitive to the definition of the distinction between cloudy  
 301 regions and the environment, provided that the definition of cloudy regions is not too restrictive (Text S1).

302 Cloudy regions cover only a few percent of the domain (Figure 6a). The fraction of water condensed in cloudy  
 303 regions that evaporates into the environment, estimated as  $f_{ev} = -(dq/dt)_{env}/(dq/dt)_{cloud}$ , where  $(dq/dt)_{env}$  and  
 304  $(dq/dt)_{cloud}$  are the humidity tendencies associated with phase changes in average in the environment and in the

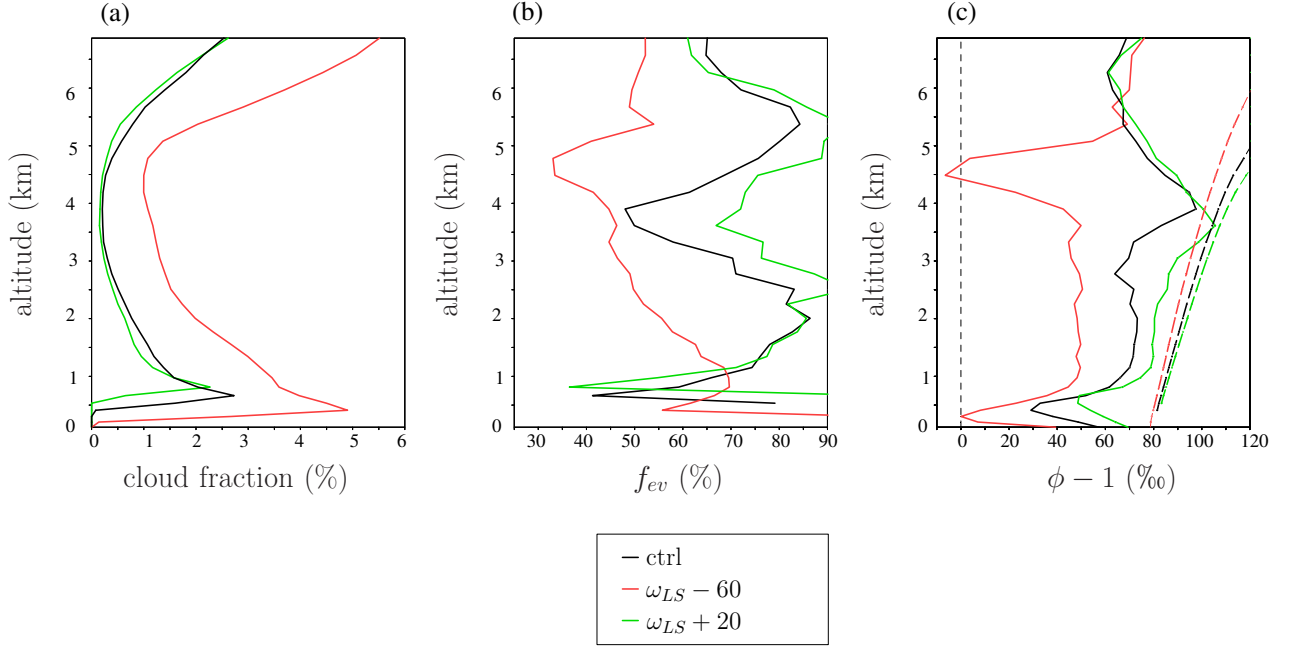


Figure 6: (a) fraction of the domain-area covered by cloudy regions. (b) Fraction of the water condensed in cloudy regions that evaporates into the environment,  $f_{ev}$ . (c)  $(\phi - 1) \cdot 1000$  (solid) and  $(\alpha_{eq} - 1) \cdot 1000$  (dashed), where  $\phi = R_{ev}/R_e$  and  $\alpha_{eq}$  is the equilibrium fractionation coefficient. Both are expressed in ‰. The black, red and green lines are for ctrl,  $\omega_{LS} - 60$  and  $\omega_{LS} + 20$  respectively.

cloudy region respectively, varies between 30% and 90%, depending on altitude (Figure 6b). It is smaller in  $\omega_{LS} - 60$  and than in ctrl, because the environment is moister.

Figure 6c plots  $\phi = R_{ev}/R_e$ , where  $R_{ev} = (dq_{HDO}/dt)_{env}/(dq/dt)_{env}$ ,  $(dq_{HDO}/dt)_{env}$  is the HDO tendency associated with phase changes in the environment and  $R_e$  is the isotopic ratio in the environment. In ctrl,  $\phi > 1$ : the evaporation has an enriching effect on the environment. Yet,  $\phi < \alpha_{eq}$ : the evaporation is not as enriching as if there was total evaporation of condensate. In  $\omega_{LS} - 60$ ,  $\phi$  is smaller than in ctrl: rain evaporation has a weaker enriching effect. At 4.5 km, near the melting level, there is a small layer where  $\phi < 1$ : at this level, the rain evaporation has a depleting effect on the water vapor. The overall enriching effect of evaporation contradicts hypothesis 2, but the weaker enriching effect in  $\omega_{LS} - 60$  than in ctrl supports hypothesis 3. Conversely,  $\phi$  is larger in  $\omega_{LS} + 20$ .

## 2.7 What controls the isotopic composition of the rain evaporation flux?

Why is  $\phi$  smaller in  $\omega_{LS} - 60$  and higher in  $\omega_{LS} + 20$  than in ctrl? It could be because rain-vapor exchanges in a moister environment leads the evaporation to have a more depleting effect ([Lawrence et al., 2004, Risi et al., 2008]), or because rain evaporation is more depleted when the evaporated fraction is small ([Risi et al., 2008, Tremoy et al., 2014]), or because the rain itself is more depleted. We aim here at quantifying these different effects.

Figure 7a plots the vertical profiles of rain  $\delta D$  (solid). Below the melting level, the rain is very close to isotopic equilibrium with the vapor (dashed). Above the melting level, the rain is more enriched than if in equilibrium due to rain lofting. Near the melting level for simulation  $\omega_{LS} - 60$ , the rain is anomalously depleted. This is due to snow melt. Since the snow forms higher in altitude, it is more depleted than the rain. It thus imprints its depleted signature on the rain when melting. In  $\omega_{LS} - 60$ , the moist middle troposphere prevents most of the snow from sublimating: 24% of the precipitation is made of snow at the melting level. The rain is thus strongly depleted by snow melt. In contrast, in ctrl and  $\omega_{LS} + 20$ , the drier middle troposphere favors snow sublimation: only 8% and 3% of the precipitation is made of snow at the melting level respectively.

The quick equilibration between the rain and vapor motivates us to use a simple equation in which some mass  $q_{l0}$  of rain, with isotopic ratio  $R_{l0}$ , partially evaporates and isotopically equilibrates with some mass  $q_{v0}$  of vapor, with isotopic ratio  $R_{v0}$ . As explained in text S2, if  $q_{l0} \gg q_{v0}$ , we get:

$$\phi = \frac{\lambda}{\alpha_{eq} - f_{ev} \cdot (\alpha_{eq} - 1)} \quad (2)$$

331 where  $\phi = R_{ev}/R_{v0}$ ,  $\lambda = R_{l0}/R_{v0}$ ,  $R_{ev}$  is the isotopic ratio of the rain evaporation flux,  $\alpha_{eq}$  is the equilibrium  
 332 fractionation coefficient and  $f_{ev}$  is the fraction of the rain that evaporates. Equation 2 tells us that the rain  
 333 evaporation is more depleted as the rain is more depleted relative to the vapor (quantified by  $\lambda$ ) and as the  
 334 evaporated fraction  $f_{ev}$  is smaller. This simple equation (Figure 7b, red) is able to approximate the simulated  
 335 values of  $\phi$  (black) for the ctrl simulation and is able to capture the smaller and larger values of  $\phi$  for  $\omega_{LS} - 60$  and  
 336  $\omega_{LS} + 20$  respectively (Figure 7c-d).

337 We find that below the melting level,  $\phi$  is smaller in  $\omega_{LS} - 60$  than in ctrl mainly because  $f_{ev}$  is smaller (Figure  
 338 7c, green). Near the melting level,  $\phi$  is smaller in  $\omega_{LS} - 60$  than in ctrl both because  $f_{ev}$  is smaller and because  $\lambda$   
 339 is smaller, i.e. the rain is more depleted due to snow melt. In  $\omega_{LS} + 20$ , the effect of  $f_{ev}$  dominates at all levels.

## 340 2.8 Summary

341 To summarize, the previous sections suggest that rain evaporation in the lower troposphere is a key ingredient of  
 342 the vapor amount effect. The isotopic composition of the rain evaporation flux mainly depends on the evaporated  
 343 fraction of the rain (consistent with [Risi et al., 2008, Tremoy et al., 2014]). Near the melting level in regimes of  
 344 large-scale ascent, it is also impacted by snow melt. We hypothesize that the isotopic effect of rain evaporation  
 345 propagates downward down to the SCL. To test this hypothesis and to understand the underlying mechanisms, in  
 346 the next section we develop a simple two-column model.

## 347 3 A simple two-column model to quantify the relative contributions of 348 different processes

349 The previous section and previous studies provide a guide for developing our simple interpretative framework. First,  
 350 the model needs to represent the effect of rain evaporation, highlighted as a key process in the previous section.  
 351 Second, alternative hypotheses for the ‘‘vapor amount effect’’ have involve mixing between the subsident environment  
 352 and detrained water ([Bailey et al., 2017], hypothesis 4). These processes also need to be represented in our model.  
 353 Third, the steepness of the  $q - \delta D_v$  relationship must be a key ingredient, since it drives  $\delta D_v$  in the SCL and thus  
 354  $\delta D_v$  everywhere. Finally, the previous section has relied on the distinction between the environment and cloudy  
 355 regions. Keeping this distinction, we develop a two-column model.

### 356 3.1 Model equations and numerical application to LES outputs

#### 357 3.1.1 Balance equations

358 This model is inspired by the two-column model used to predict tropospheric relative humidity in [Romps, 2014]  
 359 and  $\delta D$  profiles in [Duan et al., 2018]. The first column represents the cloudy regions, including cloudy updrafts and  
 360 downdrafts, as a bulk entraining plume. The second column represents the subsident environment and precipitating  
 361 downdrafts (Figure 8).

362 The mass balance for the air in the cloudy regions writes:

$$\frac{dM}{dz} = M \cdot (\epsilon - \delta) \quad (3)$$

363 where  $M$  is the bulk mass flux in the cloudy regions (positive upward),  $\epsilon$  and  $\delta$  are the fractional entrainment  
 364 and detrainment rates.

365 We assume that the specific humidity in the cloudy regions is at saturation, and call it  $q_s$ . The water balance  
 366 in the cloudy regions writes:

$$\frac{d(Mq_s)}{dz} = \epsilon \cdot M \cdot q_e - \delta \cdot M \cdot q_s - c \quad (4)$$

367 where  $c$  is the condensation rate and  $q_e$  is the specific humidity in the environment. The terms on the right  
 368 hand side represent the water input by entrainment of environment air, the water loss by detrainment of cloudy air,  
 369 and the water loss by condensation respectively. We assume that all the condensed water is immediately lost by the

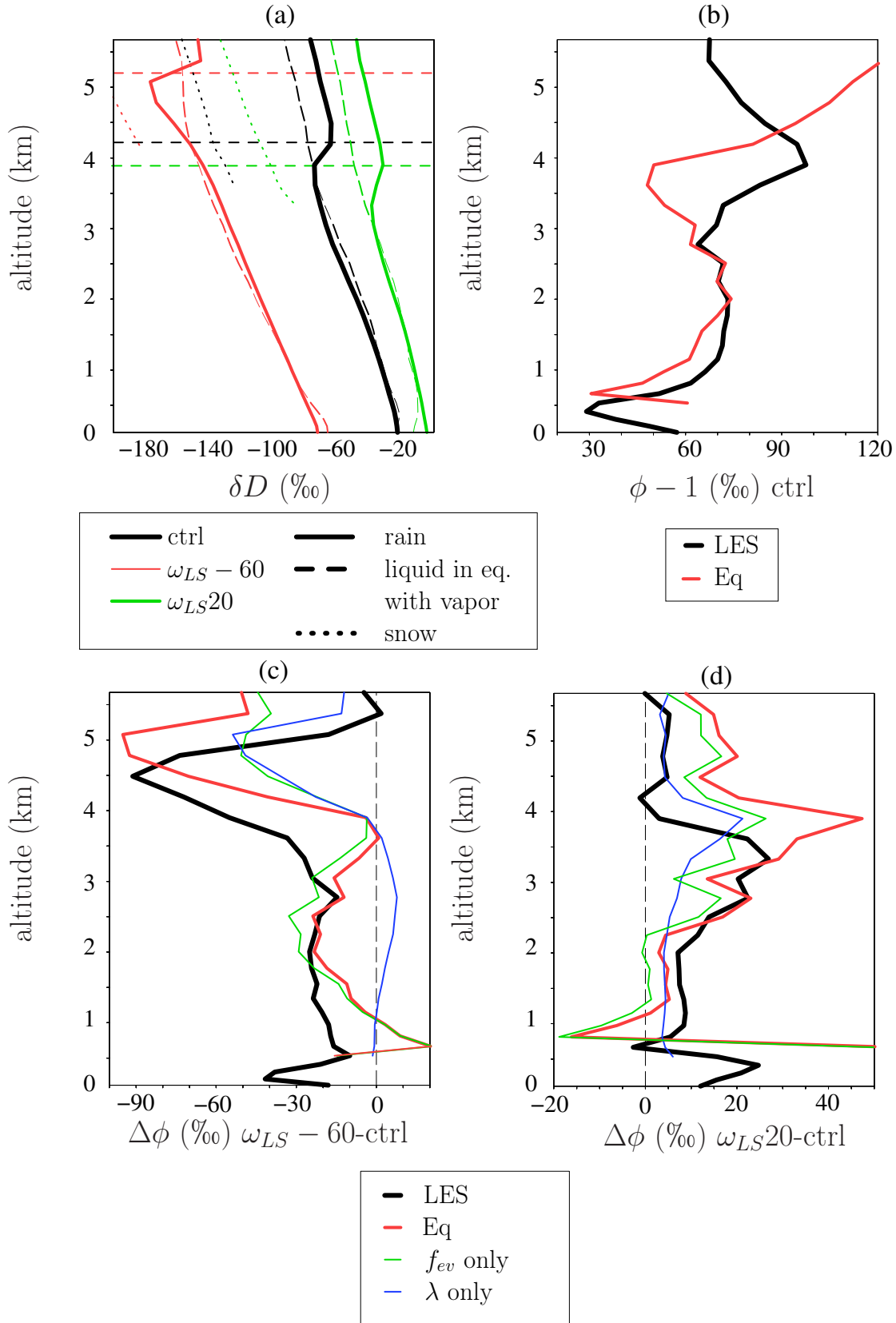


Figure 7: (a)  $\delta D$  profile for rain water (solid) and snow (dotted) falling in the environment. The liquid that would be in equilibrium with the vapor in the environment is shown in dashed. (b) Profile of  $\phi = R_{ev}/R_e$  simulated by the ctrl simulation (black, same as in Figure 6c black) and predicted by equation 2 (red). (c) Difference of  $\phi$  between  $\omega_{LS} - 60$  and ctrl simulated by the LES (black), predicted by the equation 2 (red), predicted by equation 2 if only  $f_{ev}$  varies (green) and if only  $\lambda$  varies (blue). (d) Same as (c) but for the difference between  $\omega_{LS}20$  and ctrl.

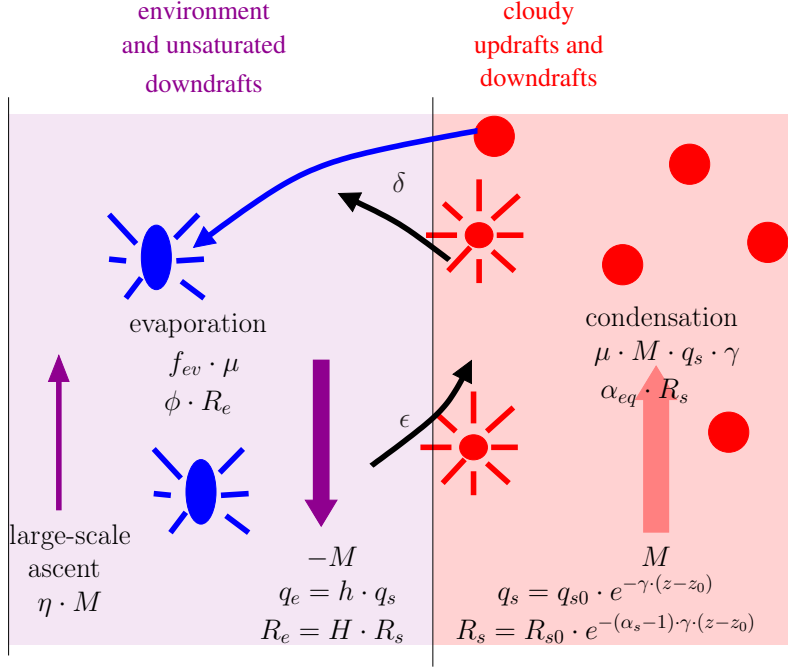


Figure 8: Schematic view of the simple two-column model, and definition of the main variables.

cloudy regions to the environment, and evaporation of this lost water can occur in the sub-saturated environment only (as in [Romps, 2014]).

We assume that mass is conserved within the domain, so that the flux in the environment is  $-M$ . The large-scale ascent, when present, is taken into account through a humidity tendency, consistent with the LES set-up. We assume that the large-scale humidity tendency applies to the environment only, which is a first-order approximation justified by the small fraction of the domain that is covered by cloudy updrafts (less than 10%). The water balance in the environment writes:

$$\frac{d(-Mq_e)}{dz} = -\epsilon \cdot M \cdot q_e + \delta \cdot M \cdot q_s + f_{ev} \cdot c - \eta \cdot M \cdot \frac{\partial q_e}{\partial z} \quad (5)$$

where  $f_{ev}$  is the fraction of the cloud or precipitating water that evaporates in the environment,  $\eta = M_{LS}/M$  and  $M_{LS}$  is the domain-mean large-scale mass flux. The terms on the right hand side represents the water loss by entrainment into cloudy regions, water input by the detrainment of cloudy air, partial evaporation of condensed water and water input by large-scale vertical advection.

Regarding water isotopes, we assume that the cloud water removed by condensation is in isotopic equilibrium with the cloudy region water vapor. The isotopic balance in the cloudy regions thus writes:

$$\frac{d(Mq_s \cdot R_s)}{dz} = \epsilon \cdot M \cdot q_e \cdot R_e - \delta \cdot M \cdot q_s \cdot R_s - c \cdot \alpha_{eq} \cdot R_s \quad (6)$$

where  $\alpha_{eq}$  is the equilibrium fractionation coefficient,  $R_s$  is the isotopic ratio in the cloudy regions and  $R_e$  is the isotopic ratio in the environment.

The isotopic balance in the environment writes:

$$\frac{d(-Mq_e \cdot R_e)}{dz} = -\epsilon \cdot M \cdot q_e \cdot R_e + \delta \cdot M \cdot q_s \cdot R_s + f_{ev} \cdot c \cdot \phi \cdot R_e - \eta \cdot M \cdot \frac{\partial (q_e R_e)}{\partial z} \quad (7)$$

where recall that  $\phi = R_{ev}/R_e$  and  $R_{ev}$  is the ratio of the precipitation evaporation flux.

### 3.1.2 Other simplifying assumptions and differential equations

To simplify the equations, as in [Romps, 2014] we assume that  $q_s$  is an exponential function of altitude:

$$q_s = q_s(z_0) \cdot e^{-\gamma \cdot (z-z_0)} \quad (8)$$



389 where  $\gamma$  is a lapse rate in  $m^{-1}$  calculated as  $d \ln(q_s)/dz$ .  
 390 For isotopes, we assume that the  $R_s$  is a power function of  $q_s$ , consistent with a Rayleigh distillation:

$$R_s = R_s(z_0) (q_s/q_{s0})^{\alpha_s - 1}$$

391 Coefficient  $\alpha_s$  represents the steepness of the  $q - \delta D_v$  gradient in cloudy regions and remains to be estimated.  
 392 As in [Duan et al., 2018],  $R_s$  is thus an exponential function of altitude:

$$R_s = R_s(z_0) \cdot e^{-(\alpha_s - 1) \cdot \gamma \cdot (z - z_0)} \quad (9)$$

393 We set:

$$q_e = h \cdot q_s$$

$$R_e = H \cdot R_s$$

394 Combining equation 5 with equations 3 and 8, we get the following differential equation for  $h$ :

$$\frac{\partial h}{\partial z} = h \cdot \gamma - \frac{\delta}{1 - \eta} (1 - h) - \frac{f_{ev} \cdot \mu \cdot \gamma}{1 - \eta} \quad (10)$$

395 where  $\mu = c/(M \cdot q_s \cdot \gamma)$  represents the ratio of actual condensation ( $c$ ) relative to the condensation if the ascent  
 396 was adiabatic ( $M \cdot q_s \cdot \gamma$ ). Similarly, combining equations 7 with equations 5 and 9, we get the following differential  
 397 equation for  $H$ :

$$\frac{\partial H}{\partial z} = H \cdot \gamma \cdot (\alpha_s - 1) - \frac{\delta}{h \cdot (1 - \eta)} \cdot (1 - H) - \frac{f_{ev} \cdot \mu \cdot \gamma}{h \cdot (1 - \eta)} \cdot H \cdot (\phi - 1) \quad (11)$$

398 We now have two equations with four unknowns:  $h$ ,  $H$ ,  $\mu$  and  $\alpha_s$ . The condensation efficiency  $\mu$  can be deduced  
 399 from equations 4:

$$\mu = 1 - \frac{\epsilon}{\gamma} \cdot (1 - h) \quad (12)$$

400 This equation, similar to one in [Romps, 2014], reflects the fact that condensation efficiency decreases when  
 401 entrainment  $\epsilon$  increases and when the entrained air is drier. If  $\epsilon = 0$  or  $h = 1$ , then  $\mu = 1$ .

402 Similarly, the  $q - \delta D_v$  steepness  $\alpha_s$  in cloudy air can be deduced from equation 6:

$$\alpha_s - 1 = \mu \cdot (\alpha_{eq} - 1) + \frac{\epsilon}{\gamma} \cdot h \cdot (1 - H) \quad (13)$$

403 This equation tells us that two effects control the steepness of the  $q - \delta D_v$  gradient. First, there is a ‘‘dilution  
 404 effect’’: if dry air is entrained, then the condensation efficiency  $\mu$  decreases. This reduces  $\alpha_s$  compared to  $\alpha_{eq}$ ,  
 405 i.e. compared to what we would expect from Rayleigh distillation. Second, there is an ‘‘isotopic contrast effect’’: if  
 406 depleted water vapor is entrained ( $H < 1$ ) is entrained, then  $\alpha_s$  becomes steeper. This is how a depleting effect of  
 407 rain evaporation can translate into a larger steepness, and eventually more depleted SCL.

### 408 3.1.3 Numerical solutions

409 To get analytical solutions for  $h$  and  $H$ , [Romps, 2014] and [Duan et al., 2018] assume that  $h \cdot \frac{\partial q_s}{\partial z} \gg q_s \cdot \frac{\partial h}{\partial z}$  and  
 410 that  $H \cdot \frac{\partial R_s}{\partial z} \gg R_s \cdot \frac{\partial H}{\partial z}$ . This allows them to calculate  $h$  and  $H$  as the solution of a simple linear equation and of  
 411 a second order polynomial respectively. However, there are two issues with these solutions. First, although these  
 412 solutions behave reasonably for  $h$  ([Romps, 2014]), they become very noisy, unstable or unrealistic for values of  
 413  $\epsilon$ ,  $\delta$  and  $f_{ev}$  that are diagnosed from LES outputs. This is because there is a powerful positive feedback between  
 414  $\alpha_s$  and  $H$ : as  $H$  decreases, more depleted vapor is entrained in updrafts which increases the steepness  $\alpha_s$ ; in turn,  
 415 the stronger steepness  $\alpha_s$  makes the subsidence more efficient at depleting the environment, further decreasing  $H$ .  
 416 [Duan et al., 2018] circumvented this problem by assuming  $\epsilon$  and  $\delta$  that are constant with altitude and equal to  
 417 each other, but it is at the cost of artificially reducing freedom for the solutions. Second, our hypothesis is that rain  
 418 evaporation near the melting level affects the isotopic profiles down to the SCL. We thus want each altitude to feel  
 419 the memory of processes at higher altitudes. The term with  $\frac{\partial H}{\partial z}$  is thus a key ingredient in our framework.

420 Therefore, we choose to numerically solve the differential equations 10 and 11. We start from an altitude of  
 421 5 km with  $h = 0.8$  and  $H - 1 = -10\%$ . We do not start above 5 km because entrainment is more difficult to

422 diagnosed above the melting level (section 3.1.4). We integrate equations 10 and 11 down to the SCL top around  
 423 500 m. The resulting  $h$  profile is a function of the profiles of 5 input parameters:  $\gamma$ ,  $\epsilon$ ,  $\delta$ ,  $f_{ev}$  and  $\eta$ . The  $H$  profile  
 424 is a function of 7 input parameters:  $\gamma$ ,  $\epsilon$ ,  $\delta$ ,  $f_{ev}$ ,  $\eta$ ,  $\alpha_{eq}$  and  $\phi$ . These input parameters are all diagnosed from the  
 425 LES simulations as detailed below. In each LES level, the input parameters are assumed constant and equations  
 426 10 and 11 are integrated within each layer over 50 sub-layers.

### 427 3.1.4 Diagnosed input parameters

428 Parameters  $f_{ev}$ ,  $\alpha_{eq}$  and  $\phi$  were already plotted in Figure 6 and discussed in section 2.6. Parameter  $\gamma$  is calculated  
 429 from domain-mean profiles. It is steeper in ctrl than in  $\omega - 60$  because of the steeper temperature gradient resulting  
 430 from the drier air (Figure 9a). Parameter  $\eta = M_{LS}/M$  is calculated from the net upward mass flux in cloudy  
 431 regions  $M$  (Figure 9b), which is calculated as the average vertical velocity in cloudy regions multiplied by the area  
 432 fraction of the cloudy region. Entrainment  $\epsilon$  is diagnosed by using the conservation of the frozen moist static energy  
 433 (e.g. [Hohenegger and Bretherton, 2011, Del Genio and Wu, 2010]):

$$\frac{\partial m_s}{dz} = \epsilon \cdot (m_e - m_s)$$

434 where  $m_s$  and  $m_e$  are the frozen moist static energy in the convective region and the environment respectively.  
 435 The application of this equation is limited to the lower troposphere. Above the melting level, we would need to  
 436 account for the precipitation of ice ([Pauluis and Mrowiec, 2013]) and for the lofting of rain. Therefore, we arbitrarily  
 437 set  $\epsilon = 0.5 \text{ km}^{-1}$  above the melting level. Entrainment is maximal in the sub-cloud layer, and decreases exponentially  
 438 with height (Figure 9c), consistent with previous studies ([Del Genio and Wu, 2010, De Rooy et al., 2013]).

439 Finally, detrainment  $\delta$  is deduced from  $\epsilon$  and  $M$  using equation 3. Detrainment shows the typical trimodal  
 440 distribution ([Johnson et al., 1999]) (Figure 9d), with a first maximum just above the SCL top corresponding to  
 441 the detrainment of shallow convection, a second maximum near the melting level near 4.5 km corresponding to the  
 442 detrainment of congestus convection, and a third maximum in the upper troposphere corresponding to the deep  
 443 convection (not shown in 9d).

### 444 3.1.5 Closure in the sub-cloud layer

445 To calculate the full  $\delta D$  profiles, we need as initial condition the isotopic ratio in the SCL. With this aim, we use a  
 446 simplified version of the sub-cloud layer model of [Risi et al., 2020]. We assume that water enters the SCL through  
 447 surface evaporation and through downdrafts at the SCL top, and exits the SCL through updrafts at the SCL top.  
 448 We neglect large-scale forcing and rain evaporation, since they have a small impact in the SCL ([Risi et al., 2020]).  
 449 The air flux of updrafts equals that of downdrafts. We define  $r_u = q_u/q_1$  and  $r_d = q_d/q_1$ , where  $q_1$  is the mixing  
 450 ratio in the SCL and  $q_u$  and  $q_d$  are the mixing ratios in updrafts and downdrafts at the SCL top. We assume  
 451 that the water vapor is more enriched as the air is moister, following a logarithmic function:  $R_u = R_1 \cdot r_u^{\alpha_u - 1}$   
 452 and  $R_d = R_1 \cdot r_d^{\alpha_d - 1}$  where  $R_u$  and  $R_d$  are isotopic ratios in updrafts and downdrafts, and  $\alpha_u$  and  $\alpha_d$  are  $q - \delta D$   
 453 steepness for updrafts and downdrafts. Water and isotopic budgets yield:

$$R_1 = \frac{R_{oce}/\alpha_{eq}(SST)}{h_1 + \alpha_K \cdot (1 - h_1) \cdot \frac{r_u^{\alpha_u} - r_d^{\alpha_d}}{r_u - r_d}} \quad (14)$$

454 where  $R_{oce}$  is the isotopic ratio at the ocean surface,  $\alpha_{eq}(SST)$  is the equilibrium fractionation coefficient  
 455 at the sea surface temperature,  $\alpha_K$  is kinetic fractionation coefficient ([Merlivat and Jouzel, 1979]) and  $h_1$  is the  
 456 relative humidity normalized at the SST and accounting for ocean salinity:  $h_1 = q_1/q_{sat}^{surf}(SST)$ ,  $q_{sat}^{surf}(SST) =$   
 457  $0.98 \cdot q_{sat}(SST)$  and  $q_{sat}$  is the humidity saturation as a function of temperature at the sea level pressure. We  
 458 assume  $\delta D_{oce} = 0\%$  and  $h_1$  is diagnosed from the LES.

459 For  $r_u$  and  $r_d$ , we use values for the ctrl simulation, because small changes in  $r_u$  and  $r_d$  across simulations  
 460 have only a marginal impact on  $R_1$  ([Risi et al., 2020]). Following [Risi et al., 2020], we set  $r_u - 1 = 1.44\%$  and  
 461  $r_d - 1 = -0.38\%$ . For  $\alpha_u$  and  $\alpha_d$ , [Risi et al., 2020] had shown that they scale with  $\alpha_z$  values just above the SCL  
 462 top, but with larger values especially for simulations with large-scale ascent. We use an empirically-fitting function:  
 463  $\alpha_u = \alpha_d = 1 + 100 \cdot (\widetilde{\alpha}_z - 1)^3$ , where  $\widetilde{\alpha}_z = 1 + \frac{\ln(R(z_{SCT})/R(z_{SCT}+1 \text{ km}))}{\ln(q(z_{SCT})/d(z_{SCT}+1 \text{ km}))}$  where  $z_{SCT}$  is the SCL top.

464 Finally, since the updraft region covers only a very small fraction of the domain, we assume that  $R_e(z_{SCT}) \simeq R_1$ .  
 465 The procedure to calculate the full  $\delta D_v$  profiles is as follows:

- 466 1. vertical profiles for  $h$ ,  $H$  and  $\alpha_s$  are calculated through a downward integration of equations 10-13 following  
 467 3.1.3.

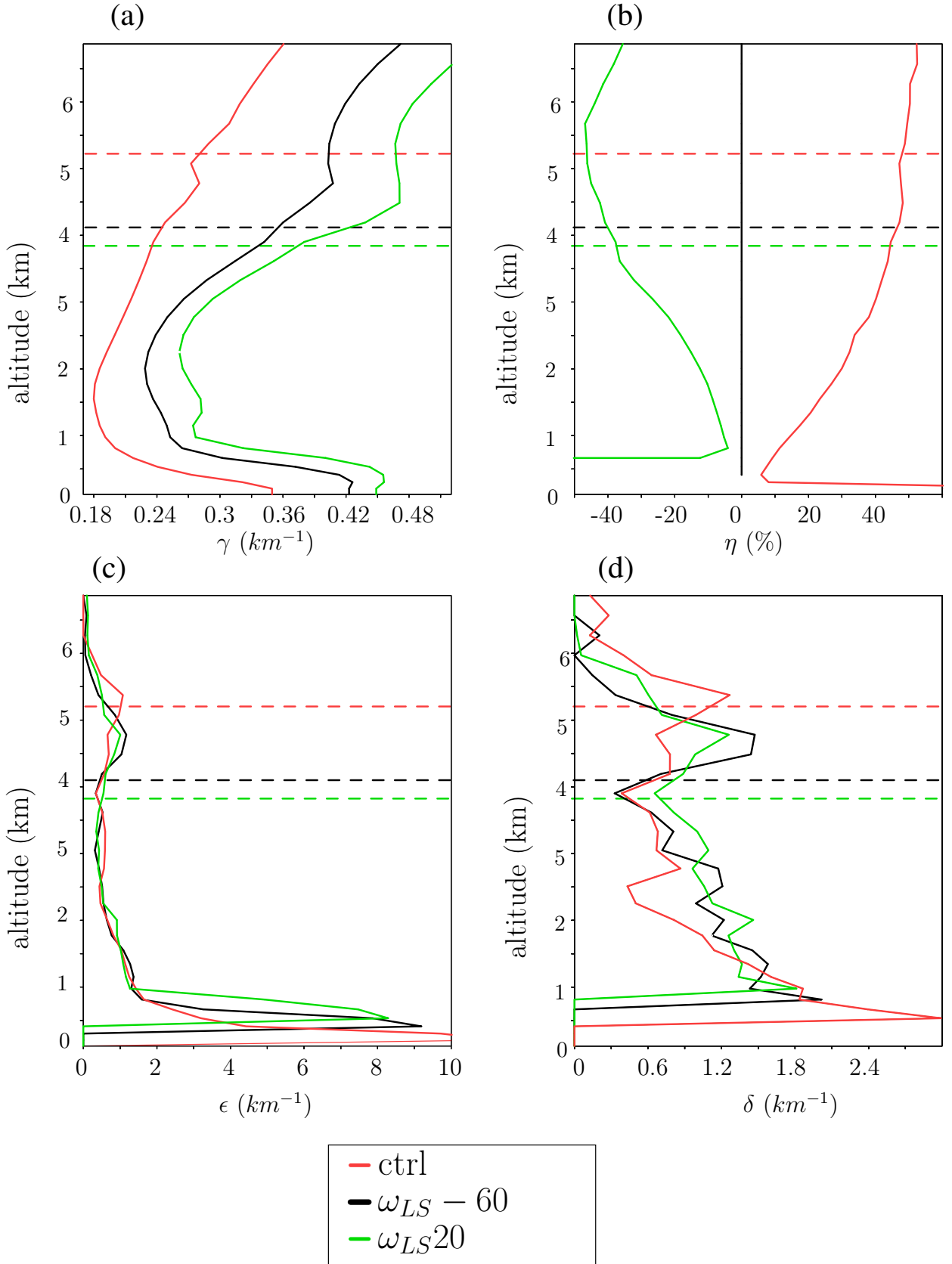


Figure 9: Input parameters for the simple model, for ctrl (black),  $\omega_{LS} - 60$  (red) and  $\omega_{LS} + 20$  (green). (a) saturation specific humidity lapse rate  $\gamma$ ; (b) ratio of large-scale vertical mass flux over the convective cloud mass flux; (c) entrainment rate; (d) detrainment rate.

- 468 2. The vertical profile for a normalized version of  $R_s$ ,  $R_{s,norm}$  that satisfies  $R_{s,norm}(z_{SCT}) = 1$ , is calculated  
469 based on the  $\alpha_s$  profile through an upward integration.
- 470 3. The vertical profile for a normalized version of  $R_e$ ,  $R_{e,norm}$ , is calculated as  $R_{e,norm} = R_{s,norm} \cdot H$ .
- 471 4. From the  $R_{e,norm}$  profile,  $\widetilde{\alpha}_z$  is estimated.
- 472 5. From  $h_1$  and  $\widetilde{\alpha}_z$ ,  $R_1$  is estimated.
- 473 6. The full  $R_e$  profile can finally be calculated so that  $R_e(z_{SCT}) \simeq R_1$ :  $R_e = R_{e,norm} \cdot R_1/H(z_{SCT})$ .

### 474 3.1.6 Evaluation of the two-column model

475 The two-column model successfully captures the order of magnitude and the shape of the vertical profile of relative  
476 humidity for the ctrl simulation (Figure 10a), as well as the moister troposphere in  $\omega_{LS} - 60$  and the drier troposphere  
477 in  $\omega_{LS} + 20$  (Figure 10b-c).

478 It successfully captures the vertical profile of  $\delta D_v$  (Figure 10b) and the more depleted troposphere in  $\omega_{LS} - 60$   
479 but underestimate the  $\delta D_v$  difference by about half (Figure 10e). It also captures the more enriched troposphere in  
480  $\omega_{LS} + 20$  but again underestimate the  $\delta D_v$  difference especially in the middle troposphere (Figure 10f). Similarly, it  
481 approximately captures the steepness  $\alpha_z$  and the sign of the  $\alpha_z$  differences across simulations, but underestimates  
482 the  $\alpha_z$  differences (Figure 10g-i).

483 These mismatches are caused by mismatches in the estimate of the relative enrichment of the environment  
484 relative to the updrafts  $H$ . Although it is reasonably well predicted for the ctrl simulation (Figure 10c), the model  
485 fails to simulate the smaller  $H$  for  $\omega_{LS} - 60$  in the middle troposphere and the larger  $H$  for  $\omega_{LS} + 20$  almost  
486 everywhere. The two-column model overestimates the impact of  $\eta$  and predicts a behavior for  $H$  that is too similar  
487 to that of  $h$ . We could not find the exact reason for this shortcoming, but we have to acknowledge that the  
488 two-column model hides many horizontal heterogeneities. We will have to keep this shortcoming in mind when  
489 interpreting the results.

## 490 3.2 Decomposition of relative humidity and $\delta D$ variations

491 To estimate the impact of the different input parameters on the relative humidity profile  $h$ , we modify them one by  
492 one from the ctrl simulation to the  $\omega_{LS} - 60$  and from the ctrl simulation to  $\omega_{LS} + 20$  simulations.

### 493 3.2.1 Decomposition of relative humidity

494 The moister troposphere in  $\omega_{LS} - 60$  is mainly due to the larger  $\eta$ , i.e. the direct moistening effect of large-scale  
495 ascent (Figure 11a). The thermodynamic structure, entrainment, detrainment and rain evaporation have a much  
496 smaller effect. Similarly, The drier troposphere in  $\omega_{LS} + 20$  is mainly due to the more negative  $\eta$ , i.e. the direct  
497 drying effect of large-scale descent (Figure 11b).

498 Note that the direct effect of  $\eta$  on environment relative humidity may be an artifact of the way the large-  
499 scale forcing is prescribed in our LES. The large-scale circulation is accounted for through tendencies in humidity,  
500 temperature and isotopes that are horizontally homogeneous. Therefore, since the environment covers most of the  
501 domain, most of the large-scale ascent or descent is felt in the environment. In reality, in case of large-scale ascent  
502 for example, we expect the environment to keep the same subsidence rate, corresponding to radiative subsidence.  
503 Rather, we expect the large-scale ascent to be absorbed by the larger cloud fraction ([Emanuel et al., 1994]). In our  
504 interpretative framework, this would be equivalent to adding the large-scale in the cloudy column rather than in  
505 the environment. This would increase the condensation efficiency  $\mu$ , and would eventually moisten the environment  
506 through rain evaporation, but less than if the large-scale ascent was homogeneous (Text S3). We have to keep this  
507 in mind if trying to connect the sensitivity of  $h$  to large-scale circulation in our LES and in reality.

### 508 3.2.2 Dilution effect on $\delta D$

509 A first effect impacting  $\delta D_v$  profiles is the dilution by entrainment (section 3.1.2). In the absence of entrainment  
510 ( $\epsilon = 0$ ), the steepness in the updraft column would be  $\alpha_s = \alpha_{eq}$  (Figure 12a, black). Because dry air is entrained,  
511 the condensation rate is reduced by the factor  $\mu$  following equation 12. According to equation 13, this reduces the  
512 steepness (Figure 12a, green). This effect of entrainment can be understood as a mixing process: as the air rises and  
513 condensation proceeds, the remaining air is mixed with dry air from entrainment and with droplets that evaporate.  
514 Consistent with the convex shape of the mixing lines, this leads to a reduction of the  $q - \delta D$  steepness (Figure 1).

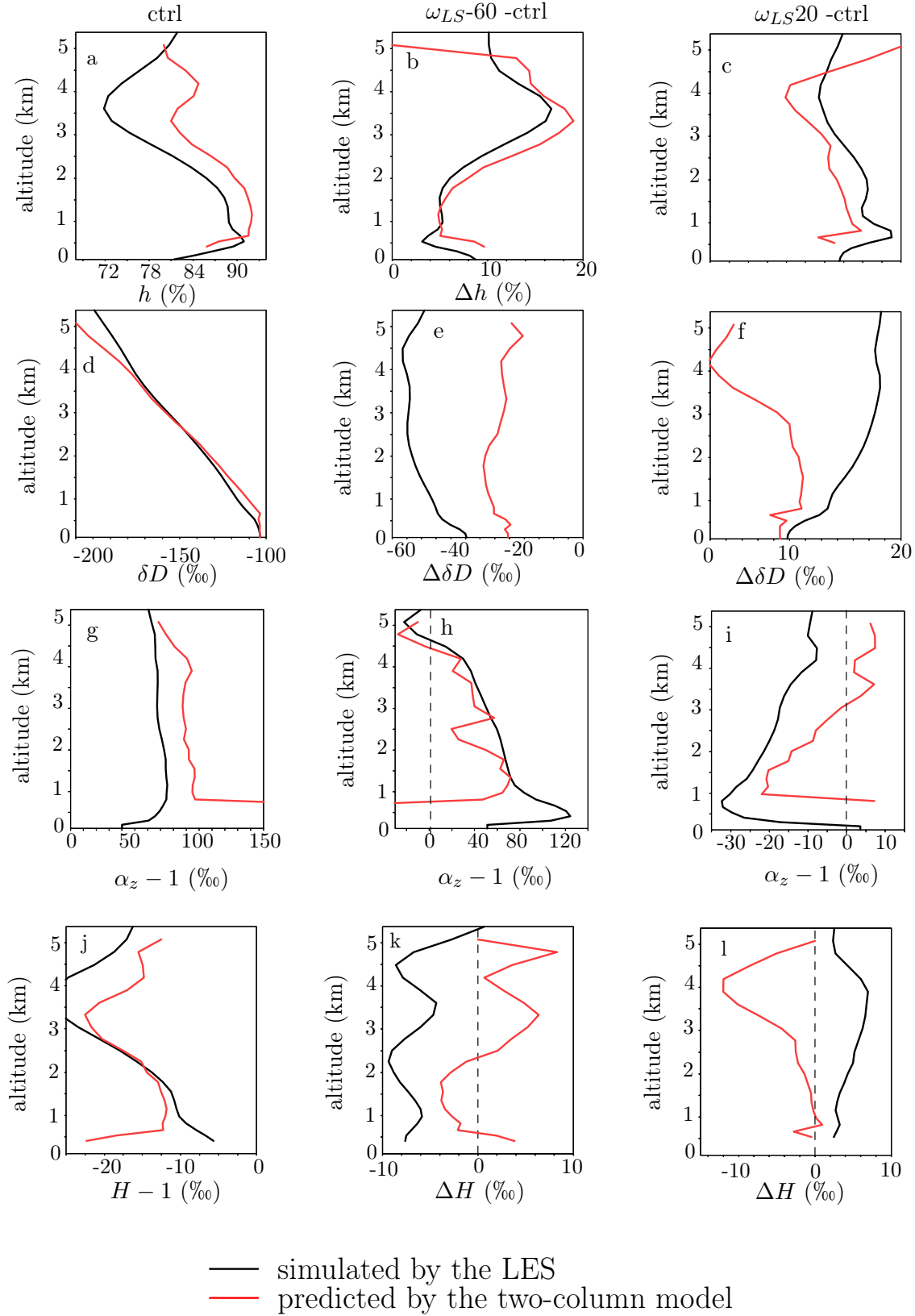


Figure 10: (a) Relative humidity  $h$  simulated by the LES (black) and predicted by the two-column model (red) for the ctrl simulation. (b) Same as (a) but for the difference between  $\omega_{LS} - 60$  and ctrl. (c) Same as (b) but for the difference between  $\omega_{LS} + 20$  and ctrl. (d-f) Same as (a-c) but for the water vapor  $\delta D$ . (g-i) Same as (a-c) but for the steepness  $\alpha_z$ . (j-l) Same as (a-c) but for the relative enrichment of the environment relative to the updrafts  $H$ .

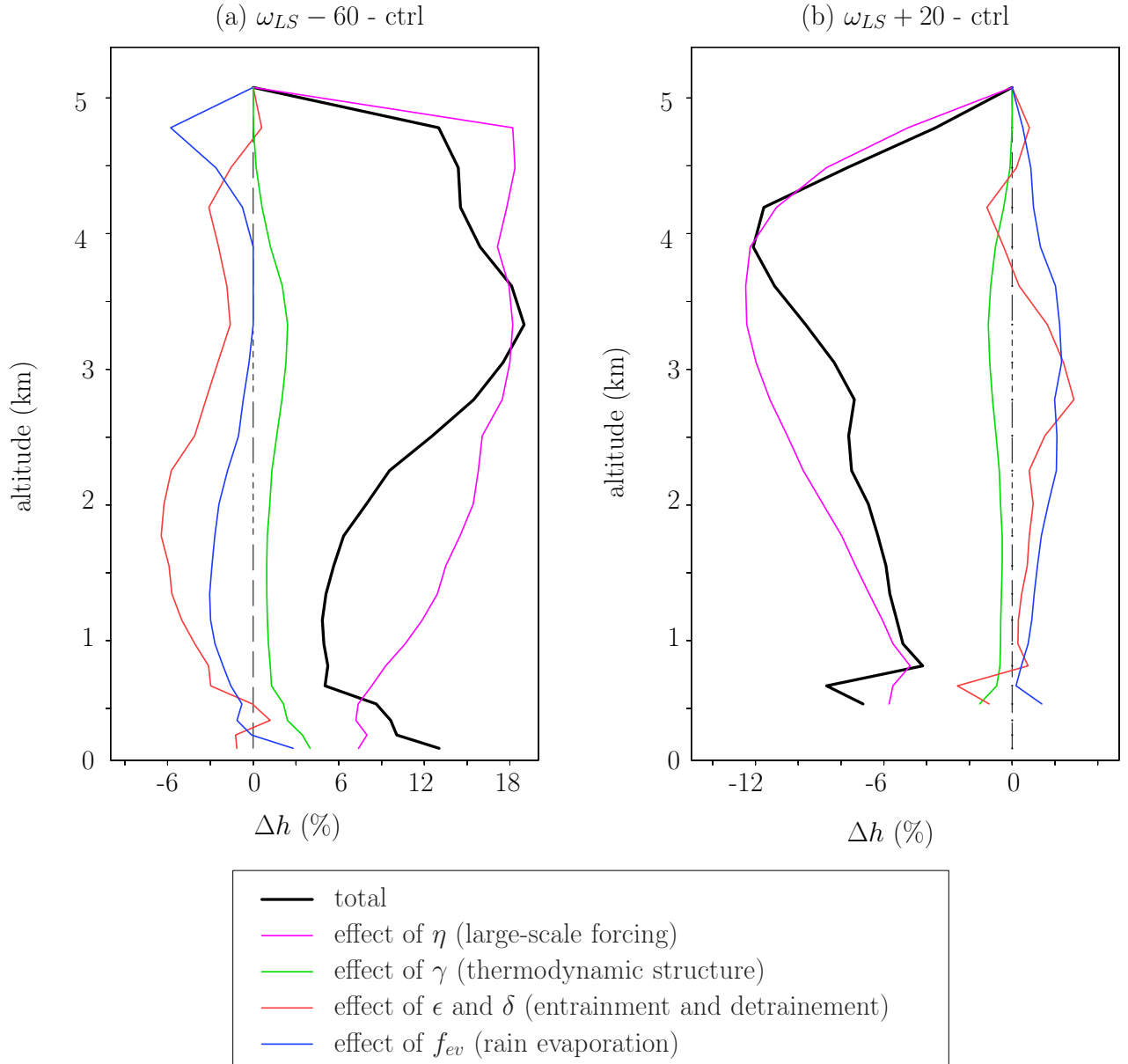


Figure 11: (a) Relative humidity difference between  $\omega_{LS} - 60$  and ctrl predicted by the two-column model (black) and its contributions from variations of input parameters one by one:  $\eta$  (pink),  $\gamma$  (green),  $\epsilon$  and  $\delta$  (red) and  $f_{ev}$  (blue). (b) Same as (a) but for the difference between  $\omega_{LS} + 20$  and ctrl.

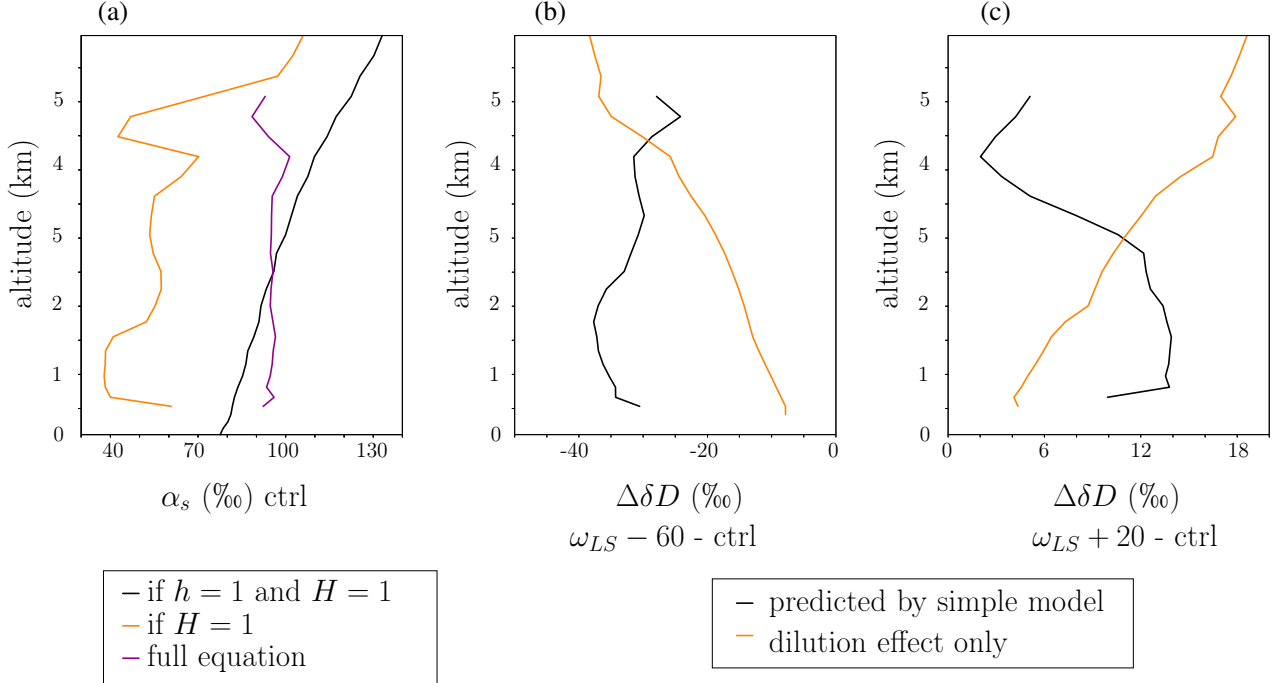


Figure 12: (a) Fractionation coefficient  $\alpha_{eq}$  (black), corresponding to the steepness in the updraft column  $\alpha_s$  if  $h = 1$  and  $H = 0$ ; steepness  $\alpha_s$  predicted if  $h < 1$  and  $H = 0$  ( $\alpha_s = 1 + \mu \cdot (\alpha_{eq} - 1)$ ) (green); steepness  $\alpha_s$  from the full equation 13 (red). (b) Difference in  $\delta D_v$  from ctrl to  $\omega_{LS} - 60$  predicted by the two-column model (red) and predicted if accounting only for the dilution effect (green). (c) Same as (b) but for  $\omega_{LS} + 20$ .

Difference from ctrl	$\omega_{LS} - 60$	$\omega_{LS} + 20$
Total simulated by the LES (‰)	-40	10
Total predicted by the two-column model (‰)	-30	11
Dilution effect (‰, %)	-9 (29%)	5 (47%)

Table 1: Difference of  $\delta D_v$  between  $\omega_{LS} - 60$  and ctrl and between  $\omega_{LS} + 20$  and ctrl simulated by the LES and predicted by the two-column model, and the contribution of the dilution effect.

As a consequence of this “dilution effect”, tropospheric  $\delta D_v$  is less depleted than predicted by Rayleigh distillation. Since the troposphere is moister in  $\omega_{LS} - 60$ , this “dilution effect” contributes to the more depleted  $\delta D_v$  in  $\omega_{LS} - 60$  (Figure 12b, green). Reciprocally, since the troposphere is drier in  $\omega_{LS} + 20$ , this “dilution effect” contributes to the more enriched  $\delta D_v$  in  $\omega_{LS} + 20$  (Figure 12c, green). Quantitatively, the contribution of this dilution effect on the SCL  $\delta D_v$  difference is 29% for  $\omega_{LS} - 60$  and 47% for  $\omega_{LS} + 20$  (table 2). The contribution increases with altitude.

Note that the two-column model likely overestimates this contribution, because of the shortcoming mentioned in section 3.1.6. The fact that only one third of the  $\delta D_v$  difference remains when post-condensation effects are turned off confirms that these contributions are overestimated.

### 3.2.3 Decomposition of $\delta D$

In  $\omega_{LS} - 60$ , the more depleted troposphere is driven primarily by the effect of the smaller  $\phi$ , i.e. the more depleted rain evaporation (Figure 13a, cyan). It explains 147% of the  $\delta D_v$  difference in the SCL (Table 2). The smaller rain evaporated fraction (smaller  $f_{ev}$ ) is the second main contributor (Figure 13a, blue, 43% in the SCL). This positive contribution is explained by the fact that evaporation has an overall enriching effect. The third main contributor is the larger  $\eta$  (i.e. large-scale ascent), contributing to 26% of the  $\delta D_v$  difference. This contribution corresponds mainly to the “dilution effect” explained in section 3.2.2. The sum of these contributions exceeds 100%, because there are some dampening effects, especially  $h_1$ : the moister surface relative humidity reduces the kinetic fractionation during surface evaporation.

In  $\omega_{LS} + 20$ ,  $\eta$  becomes the main contribution to the  $\delta D_v$  difference in the SCL (126%), associated with the

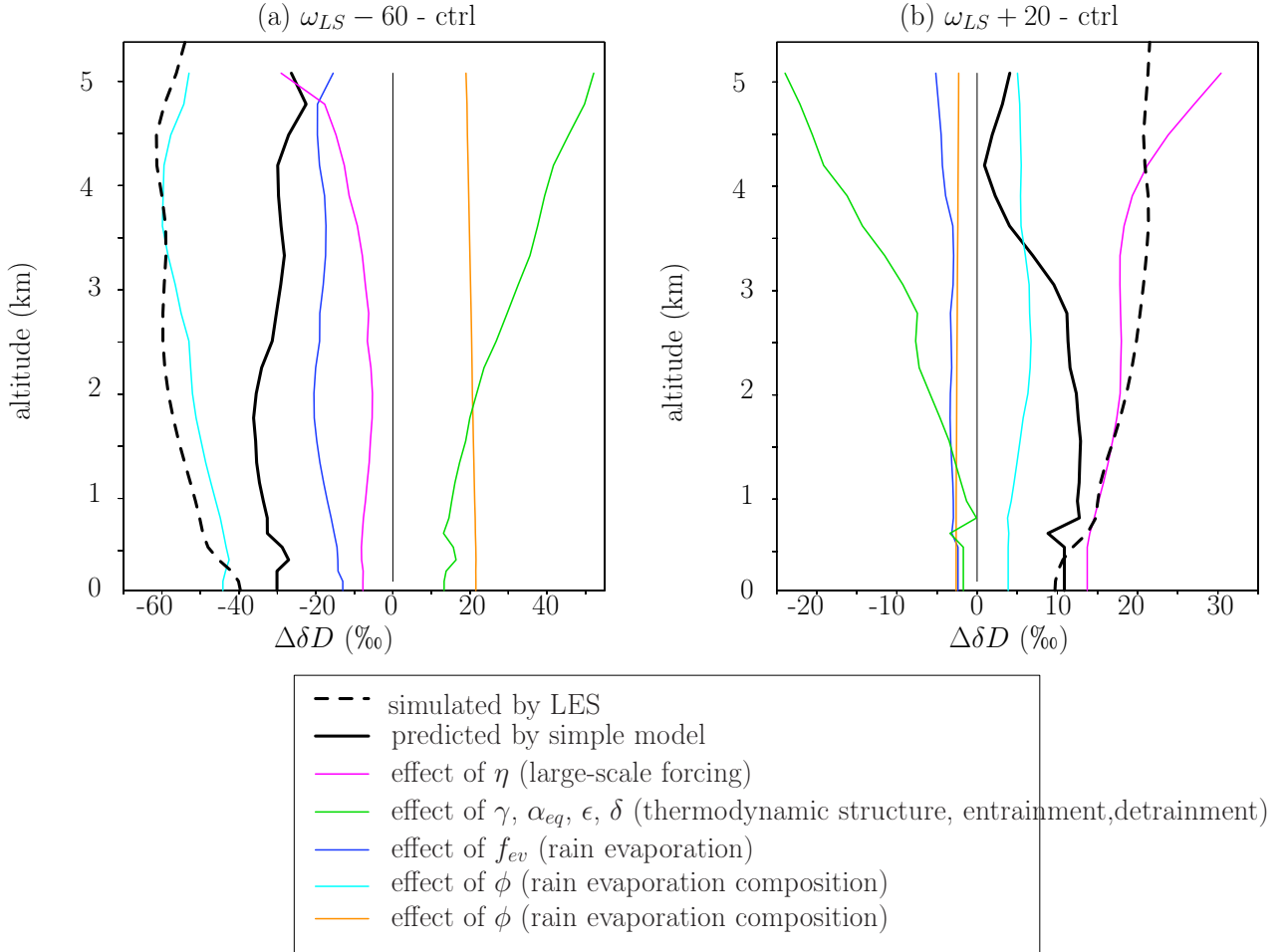


Figure 13: (a)  $\delta D_v$  difference between  $\omega_{LS} - 60$  and ctrl predicted by the two-column model (black) and its contributions from variations of input parameters one by one:  $\eta$  (pink),  $\gamma$  and  $\alpha_{eq}$  (green),  $\epsilon$  and  $\delta$  (red),  $f_{ev}$  (blue),  $\phi$  (cyan) and  $h_1$  (orange). (b) Same as (a) but for the difference between  $\omega_{LS} + 20$  and ctrl.

dilution effect. The effect of the larger  $\phi$ , i.e. the more enriched rain evaporation, contributes to 36% to the  $\delta D_v$  difference in the SCL.

This decomposition can be reconciled with the result that about one third of the  $\delta D_v$  difference from ctrl to  $\omega_{LS} - 60$  remains when the fractionation during condensate evaporation is de-activated. This remaining difference is associated with (1) the dilution effect, and (2) the portion of the  $\phi$  contribution that is due to the more depleted rain due to more snow melt. The fact that the sum of this two contributions exceeds one third suggests that the underestimate of  $\delta D_v$  variations by the simple model is due to the underestimate of the effect of rain evaporation.

We note that the relative contributions of the different processes are very homogeneous in the vertical. For example, in the SCL, half of the contribution of  $\phi$  comes from  $\phi$  above 3 km. This shows the strong “memory” of water vapor  $\delta D$ , which integrates processes downwards in the environment column, and then upward in the updraft column.

## 4 Conclusion

### 4.1 Summary

This study gives a comprehensive and quantitative understanding of the processes underlying the vapor amount effect, at least in our LES simulations. This understanding is illustrated in Figure 14:

1. When the troposphere is moister (in terms of relative humidity), less snow sublimates and thus more snow is available for melting. Snow melt results in rain that is more depleted relative to a liquid in equilibrium with



SCL $\delta D_v$ difference from ctrl	$\omega_{LS} - 60$	$\omega_{LS} + 20$
Total simulated by the LES ( $\%$ )	-40	10
Total predicted by the two-column model ( $\%$ )	-30	11
Effect of $\gamma$ and $\alpha_{eq}$ ( $\%$ , $\%$ )	8 (-25%)	-3 (-30%)
Effect of $\epsilon$ and $\delta$ ( $\%$ , $\%$ )	6 (-19%)	2 (14%)
Effect of $\eta$ ( $\%$ , $\%$ )	-8 (26%)	14 (126%)
Effect of $f_{ev}$ ( $\%$ , $\%$ )	-13 (43%)	-2 (-22%)
Effect of $\phi$ ( $\%$ , $\%$ )	-44 (147%)	4 (36%)
Including $\phi$ above 3 km ( $\%$ , $\%$ )	-23 (76%)	2 (23%)
Effect of $h_1$ ( $\%$ , $\%$ )	22 (-72%)	-3 (-24%)

Table 2: Difference of  $\delta D_v$  in the SCL between  $\omega_{LS} - 60$  and ctrl and between  $\omega_{LS} + 20$  and ctrl simulated by the LES and predicted by the two-column model, and the contribution of different effects. The sum of all the different effects, except the line “Including  $\phi$  above 3 km”, is 100% of the predicted  $\delta D_v$  difference. The line “Including  $\phi$  above 3 km” is a part of “Effect of  $\phi$ ”

the vapor, which leads to more depleted rain evaporation flux. When the troposphere is moister, the rain evaporated fraction is also smaller, making the rain evaporation flux even more depleted.

2. The more depleted evaporation depletes the environment more efficiently relative to updrafts. When this more depleted environment is entrained into the updrafts, it makes the vertical  $q - \delta D$  gradient steeper. In turn, the steeper  $q - \delta D$  vertical gradient makes the subsidence more efficient at depleting the environment, in a positive feedback that makes the vertical  $q - \delta D$  gradient even steeper. Overall, this mechanisms allows to propagate the isotopic anomalies associated with rain evaporation downwards.
3. When the troposphere is moister, the updrafts get less diluted as they rise. They condense water vapor more efficiently, which also contributes to the steeper vertical  $q - \delta D$  gradient.
4. The steeper gradient in the lower troposphere makes updrafts and downdrafts at the SCL top more efficient in depleting the SCL water vapor ([Risi et al., 2020]).
5. Finally, since the SCL serves as the initial condition for the full  $\delta D$  vertical profiles, the water vapor is more depleted at all altitudes in the troposphere.

Coming back to our initial hypotheses, the dominant role of rain evaporation and rain-vapor diffusive exchanges confirm hypothesis 3 ([Lawrence et al., 2004, Risi et al., 2008, Lee and Fung, 2008]). The role of entrainment in diluting updrafts and reducing their condensation efficiency is reminiscent of hypothesis 4.

We notice that the root of the amount effect in the water vapor is higher relative humidity, with a triple effect on reducing sublimation, reducing the rain evaporated fraction and reducing the dilution of updrafts by entrainment. This explains why the amount effect can be observed only when the precipitation increase is associated with a change in the large-scale circulation ([Bony et al., 2008, Moore et al., 2014, Bailey et al., 2017, Risi et al., 2020]). While the tropospheric relative humidity is very sensitive to the large-scale circulation, it is almost invariant with sea surface temperature ([Romps, 2014]). For example, if precipitation increases because sea surface temperature increases without any change in large-scale circulation, then the tropospheric humidity would remain almost constant ([Romps, 2014]), the above-mentioned mechanism cannot take place and there is no amount effect.

## 4.2 Discussion

This study has investigated processes controlling isotopic profiles in idealized conditions. In particular, horizontal gradients in humidity and  $\delta D$  were neglected. In reality, these gradients are expected to dampen the humidity and  $\delta D$  variations as a function of large-scale vertical velocity ([Risi et al., 2019]).

The large-scale circulation is accounted for through tendencies in humidity, temperature and isotopes that are horizontally homogeneous. This is equivalent to adding an horizontally-constant vertical velocity in each grid point, thus shifting the probability distribution of grid-point vertical velocities  $w$ . In reality, in case of large-scale ascent, we expect the  $w$  distribution to be modified, with a thicker tail corresponding to a larger cloud fraction, while keeping its maximum corresponding to slow radiative subsidence ([Emanuel et al., 1994]). To check this, we would need to analyze LES on large-domains in which a large-scale circulation is allowed to develop, without needing to artificially prescribe large-scale tendencies ([Bretherton and Khairoutdinov, 2015, Khairoutdinov and Emanuel, 2018,

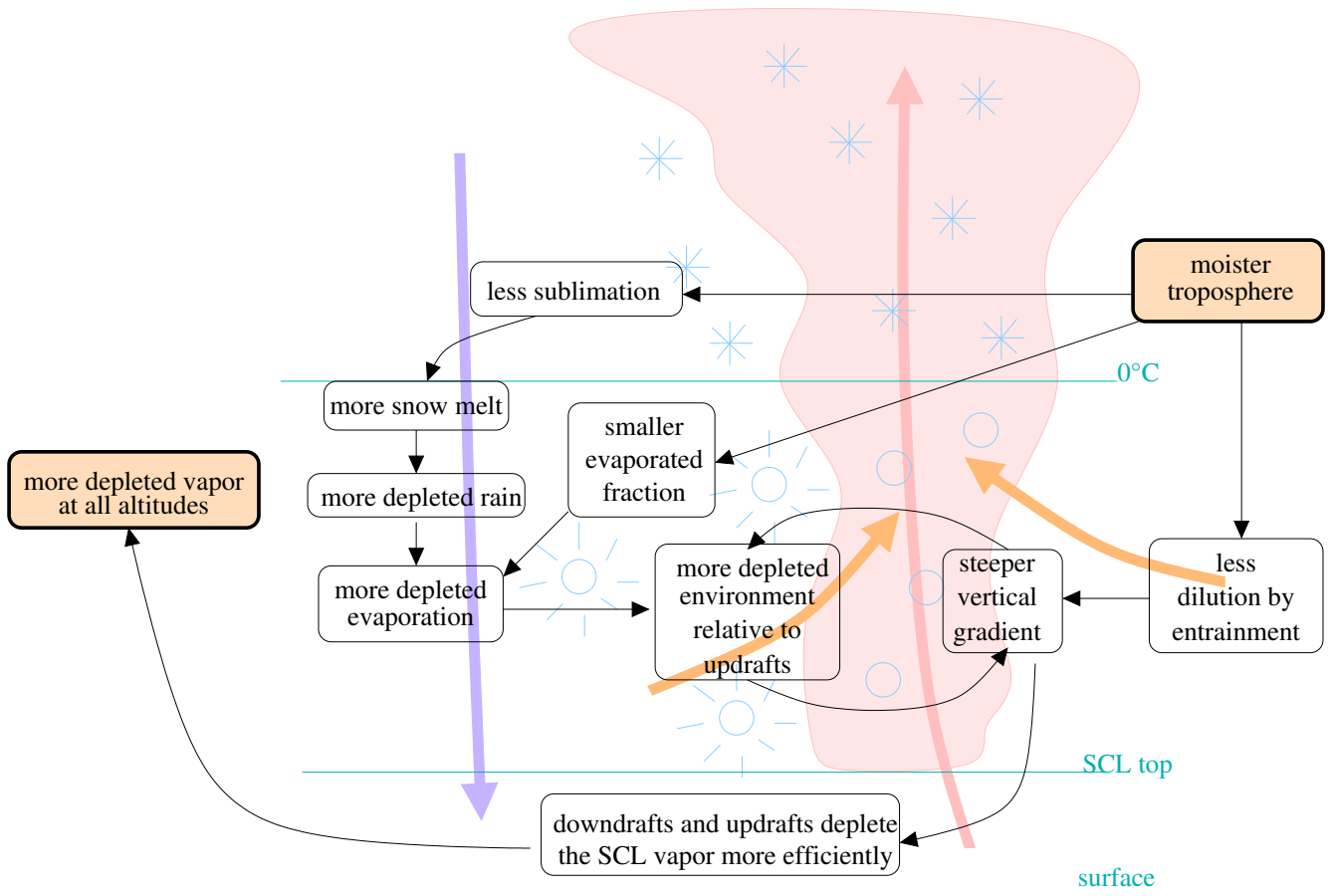


Figure 14: Schematic summarizing how a moister troposphere leads to more depleted vapor in the troposphere.

585 Stevens et al., 2019]). According to our analytical model, restricting the large-scale forcing to the cloudy regions  
586 would dampen the humidity variations (Text S3) and amplify the  $\delta D$  variations as a function of large-scale vertical  
587 velocity.

588 To assess to what extent our idealized simulations in radiative-convective equilibrium over the ocean are  
589 relevant for interpreting observations, it would be useful to compare our LES simulations with different large-scale  
590 velocities to in-situ and remote-sensing observations. This raises the question of the spatial scales at which the  
591 amount effect can be observed and of the spatial representativeness of both observations and LES simulations. This  
592 will also be investigated in a future study.

### 593 4.3 Perspectives

594 This paper highlights the important role of snow melt and rain evaporation in depleting the water vapor in case  
595 of large-scale ascent. These processes are expected to be even stronger in stratiform regions of meso-scale systems,  
596 where all the rain arises from the brutal melting of snow near the melting level, and where the rain evaporation  
597 is boosted by the meso-scale downdraft drying the lower troposphere ([Houze, 1977]). This may explain  
598 why observations show that stratiform regions are often more depleted than convective regions in squall lines  
599 ([Risi et al., 2010, Tremoy et al., 2014]), and why the amount effect is stronger where the fraction of stratiform  
600 clouds is larger ([Kurita, 2013, Aggarwal et al., 2016, Sengupta et al., 2020]. To check this hypothesis, we plan to  
601 analyze in a future study the dependence of water vapor isotopic profiles to large-scale circulation in LES with dif-  
602 ferent convective organizations, such as squall lines ([Robe and Emanuel, 2001, Muller, 2013]) or tropical cyclones  
603 ([Khairoutdinov and Emanuel, 2013, Muller and Romps, 2018]).

604 Finally, this study highlights the key role of both microphysical processes (evaporation, snow melt) and macro-  
605 physical processes (entrainment) in the amount effect. These processes are typically parameterized in LES through  
606 the microphysical and sub-grid-scale turbulence schemes in LES. What is the sensitivity of the amount effect to these  
607 parameterizations? These processes are even more crudely parameterized in general circulation models (GCMs).  
608 How do GCMs represent these processes? More generally, what would be the added value of adding isotopic diag-  
609 nostics when routinely comparing single-column versions of GCMs to LES simulations? This is yet another question  
610 that we plan to address in the future.

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617 Information on SAM can be found on this web page: <http://rossby.msrc.sunysb.edu/~marat/SAM.html> All simu-  
618 lation outputs used in this article will be submitted to the PANGAEA data repository.

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