



**HAL**  
open science

**What controls the mesoscale variations in water isotopic composition within tropical cyclones and squall lines?  
Cloud resolving model simulations in radiative-convective equilibrium**

Camille Risi, Caroline Muller, Françoise Vimeux, Peter Blossey, Grégoire Védeau, Clarisse Dufaux, Sophie Abramian

► **To cite this version:**

Camille Risi, Caroline Muller, Françoise Vimeux, Peter Blossey, Grégoire Védeau, et al.. What controls the mesoscale variations in water isotopic composition within tropical cyclones and squall lines? Cloud resolving model simulations in radiative-convective equilibrium. *Journal of Advances in Modeling Earth Systems*, 2023, 15 (4), pp.e2022MS003331. 10.1029/2022MS003331 . hal-04080289

**HAL Id: hal-04080289**

**<https://hal.science/hal-04080289>**

Submitted on 24 Apr 2023

**HAL** is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1                   **What controls the mesoscale variations in water**  
2                   **isotopic composition within tropical cyclones and squall**  
3                   **lines? Cloud resolving model simulations in**  
4                   **radiative-convective equilibrium**

5                   **Camille Risi <sup>1</sup>, Caroline Muller <sup>1</sup>, Françoise Vimeux <sup>2,3</sup>, Peter Blossey <sup>4</sup>,**  
6                   **Grégoire Vèdeau <sup>1,3</sup>, Clarisse Dufaux <sup>1</sup>, Sophie Abramian <sup>1</sup>**

7                   <sup>1</sup>Laboratoire de Météorologie Dynamique, IPSL, CNRS, École Normale Supérieure, Sorbonne Université,  
8   PSL Research University, Paris, France

9                   <sup>2</sup> HydroSciences Montpellier, Université de Montpellier, CNRS, IMT, IRD, Montpellier, France

10                   <sup>3</sup>Laboratoire des Sciences du Climat et de l'Environnement (CEA, CNRS, UVSQ), IPSL, Université Paris  
11   Saclay, Gif-sur-Yvette, France

12                   <sup>4</sup>Department of Atmospheric Sciences, University of Washington, Seattle, USA

13                   **Key Points:**

- 14                   • Cloud resolving model simulations capture robust aspects of mesoscale isotopic  
15                   variations observed in tropical cyclones and squall lines  
16                   • Rain evaporation and rain-vapor diffusive exchanges are the main drivers of iso-  
17                   topic depletion within tropical cyclones and squall lines  
18                   • Horizontal advection spreads isotopic anomalies, thus reshaping the stable isotope  
19                   patterns at the mesoscale.

---

Corresponding author: Camille RISI, [Camille.Risi@lmd.ipsl.fr](mailto:Camille.Risi@lmd.ipsl.fr)

## Abstract

Water isotopes are tracers of convective processes and are often used as proxies for past precipitation. These applications require a better understanding of the impact of convective processes on the isotopic composition of water vapor and precipitation. One way to advance this understanding is to analyze the isotopic mesoscale variations during organized convective systems such as tropical cyclones or squall lines. The goal of this study is to understand these isotopic mesoscale variations with particular attention to isotopic signals in near-surface vapor and precipitation that may be present in observations and in paleoclimate proxies. With this aim, we run cloud resolving model simulations in radiative-convective equilibrium in which rotation or wind shear is added, allowing us to simulate tropical cyclones or squall lines. The simulations capture the robust aspects of mesoscale isotopic variations in observed tropical cyclones and squall lines. We interpret these variations using a simple water budget model for the sub-cloud layer of different parts of the domain. We find that rain evaporation and rain-vapor diffusive exchanges are the main drivers of isotopic depletion within tropical cyclones and squall lines. Horizontal advection spreads isotopic anomalies, thus reshaping the mesoscale isotopic pattern. Variations in near-surface relative humidity and wind speed have a significant impact on d-excess variations within tropical cyclones. This study contributes to our understanding of mesoscale isotopic variability and provides physical arguments supporting the interpretation of paleoclimate isotopic archives in tropical regions in terms of past cyclonic activity.

## Plain Language Summary

Naturally-available, stable water molecules can be light (one oxygen atom and two hydrogen atoms) or heavy (one hydrogen atom is replaced by a deuterium atom). These different molecules are called water isotopes. In large, long-lived, severe storms such as tropical cyclones or squall lines (thunderstorms that organize into lines), the rain is observed to be more depleted in heavy isotopes. Several studies have exploited this property to reconstruct the past variations in the frequency of occurrence of tropical cyclones or severe thunderstorms based on isotope variations observed in speleothems. The aim of this study is to understand what controls the depletion in heavy isotopes of the rain in tropical cyclones and squall lines. With this aim, for the first time we use high-resolution simulations (2-4 km in horizontal) to simulate the internal dynamics of tropical cyclones and squall lines and their isotope composition. We design a simple model to interpret the results. We show that the rain evaporation and rain-vapor exchanges deplete the water vapor, and the subsequent rainfall, in heavy isotopes.

## 1 Introduction

The isotopic composition of water vapor ( $HDO$  or  $H_2^{18}O$ ) evolves along the water cycle as phase changes are associated with isotopic fractionation. The isotopic composition of precipitation recorded in paleoclimate archives has significantly contributed to the reconstruction of past hydrological changes across the tropics (Wang et al., 2001; Cruz et al., 2009). Indeed, over tropical oceans, the precipitation is usually more depleted in heavy isotopes as precipitation rate increases, an observation called the amount effect (Dansgaard, 1964). In concert with the precipitation, the water vapor over tropical oceans is also more depleted as precipitation rate increases according to satellite and in-situ observations (Worden et al., 2007; Kurita, 2013). Over tropical land, both the precipitation and water vapor are generally observed to be more depleted as the precipitation rate increases on average over the previous days before the observation of isotopic depletion and on average over some large-scale domain upstream of the region of depletion, for example in Western Africa (Risi et al., 2008; Tremoy et al., 2012), Southeast Tibetan Plateau (Gao et al., 2013), Southern India (Lekshmy et al., 2014; Sinha & Chakraborty,

2020), Southern tropical America (Vimeux et al., 2005; Vimeux et al., 2011) or the Maritime Continent (Moerman et al., 2013; Conroy et al., 2016). In the tropics, the importance of precipitation rate, either at the local or at the regional scale, in controlling the water isotopic composition of water vapor and precipitation is thus well established. However, the relationship between the water isotopic composition and precipitation rate can vary temporally and spatially. For example, it may depend on the proportion of stratiform versus convective rain (Aggarwal et al., 2016), on the organization of convection (Lawrence et al., 2004; Risi et al., 2008; Chakraborty et al., 2016) or on the shape of vertical velocity profiles (Moore et al., 2014; Torri et al., 2017; Lacour et al., 2017). For a more robust and quantitative interpretation of water isotopic archives in terms of past hydrological changes or cyclonic activity, a better understanding of how the precipitation rate impacts the isotopic composition of water vapor and precipitation is thus necessary.

In the tropics, most of the precipitation falls under deep convective systems (Houze, 2004). It is associated with processes which deplete the water vapor in heavy isotopes. In particular, observational studies have highlighted the role of rain evaporation (Worden et al., 2007), diffusive liquid-vapor exchanges (Lawrence et al., 2004), mesoscale downdrafts (Risi et al., 2010; Kurita, 2013) and microphysical processes in stratiform regions of convective systems (Aggarwal et al., 2016). Modeling studies with high resolution simulations have confirmed the key role of rain evaporation and diffusive liquid-vapor (Torri, 2021, 2022), and of microphysical processes in stratiform regions of convective systems, especially melting of depleted snow that subsequently evaporates (Risi et al., 2021).

One way to test the importance of these processes is to investigate the observed evolution of the isotopic composition of precipitation or near-surface water vapor within “organized” convective systems (Risi et al., 2010). By organized, we mean that the convective system has different parts, characterized by different convective or microphysical processes, and connected through some mesoscale circulation. For example, tropical cyclones are a spectacular manifestation of convective organization, usually with an eye at the center, surrounded by convective walls with intense rainfall and spiral rain bands reaching several hundreds of kilometers (Houze, 2010) (Figure 1a). The precipitation and near-surface water vapor collected in the vicinity of tropical cyclones often show stronger depletion towards the cyclone center, more depleted water vapor in spiral bands than in between bands (Gedzelman et al., 2003; Xu et al., 2019; Bhattacharya et al., 2022), and more enriched water vapor in the eye (Fudeyasu et al., 2008). The depletion has been interpreted in terms of progressive rain out towards the center and rain-vapor diffusive exchanges (Lawrence et al., 2002). The enrichment in the eye has been interpreted in terms of sea spray evaporation (Fudeyasu et al., 2008). As another example, squall lines are elongated, propagative convective systems with a gust front, followed by a convective region of intense rainfall, a transition region with a paused rainfall, and a trailing stratiform region of light rainfall (Houze, 2004) (Figure 1b). The precipitation collected during squall lines often features a W shape with more depleted rain in convective and stratiform regions (Taupin & Gallaire, 1998; Risi et al., 2010). In the near-surface water vapor, many squall lines show isotopic depletion in the convective and stratiform regions (Tremoy et al., 2014). This pattern has been interpreted in terms of rain evaporation and mesoscale downdrafts.

Better understanding the evolution of the isotopic composition of precipitation or near-surface water vapor within “organized” convective systems is also useful to better understand how convective organization is recorded in paleoclimate archives. In particular, more organized convective systems, such as squall lines (Risi et al., 2008; Tremoy et al., 2014; Maupin et al., 2021) or tropical cyclones (Lawrence & Gedzelman, 1996; Lawrence et al., 2004; Price et al., 2008; Chakraborty et al., 2016; Jackisch et al., 2022), have been observed to be associated with water vapor and precipitation that are more depleted in heavy isotopes than unorganized systems. In particular, the depleted rain of tropical cy-

123 clones leaves a depleted imprint in surface waters (Welsh & Sánchez-Murillo, 2020) and  
 124 can significantly affect long-term averages of the isotopic composition of precipitation  
 125 or surface waters (Lawrence, 1998; Baldini et al., 2016). This suggests that the annual-  
 126 mean or multi-annual-mean isotopic composition of precipitation recorded in speleothems  
 127 could be used to reconstruct past cyclonic activity (Lawrence & Gedzelman, 2003; Miller  
 128 et al., 2006; Frappier et al., 2007; Chen et al., 2021), though this can be challenging (Jackisch  
 129 et al., 2022). In the past few years, several studies have interpreted speleothems in terms  
 130 of cyclonic frequency (Nott et al., 2007; Medina-Elizalde & Rohling, 2012; Baldini et al.,  
 131 2016). Similarly, the depletion observed in Texan speleothems has been interpreted as  
 132 enhanced activity of large, long-lived, organized convective systems (Maupin et al., 2021).  
 133 In addition to paleoclimate applications, better understanding how convective processes  
 134 impact the water isotopic composition is also relevant to assess the added value of wa-  
 135 ter isotopic measurements to better quantify convective processes and better evaluate  
 136 their representation in models (Bony et al., 2008; Field et al., 2014; Diekmann et al., 2021;  
 137 Ramos et al., 2022).

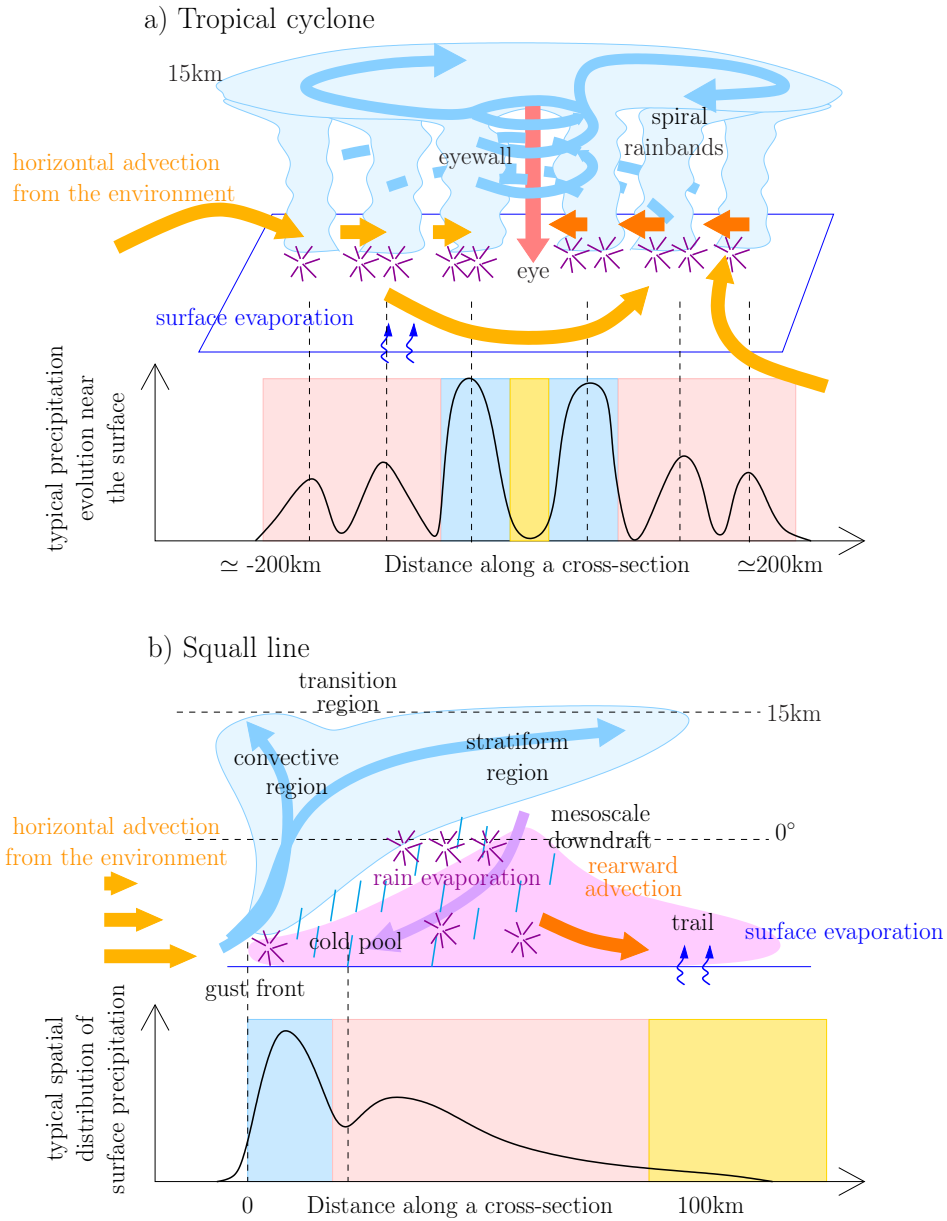
138 The goal of this paper is to investigate the processes controlling the evolution of  
 139 near-surface water vapor and precipitation within squall lines and tropical cyclones. So  
 140 far, this question has often been addressed using observational studies or simple box mod-  
 141 els (Gedzelman et al., 2003; Lawrence et al., 2002; Fudeyasu et al., 2008; Tremoy et al.,  
 142 2014; Xu et al., 2019; Bhattacharya et al., 2022). Here for the first time, we used three-  
 143 dimensional high-resolution simulations, in which convective motions are explicitly rep-  
 144 resented, to simulate the isotopic composition of water vapor and precipitation in tropi-  
 145 cal cyclones and squall lines. Using these simulations and a simple box model of the sub-  
 146 cloud layer, we quantified the relative importance of the different processes that have been  
 147 suggested by previous studies, e.g. progressive rain out, rain-vapor exchange, downdrafts  
 148 (Lawrence et al., 2002; Risi et al., 2010; Tremoy et al., 2014; Xu et al., 2019; Bhattacharya  
 149 et al., 2022). Our simulations are run in a radiative-convective equilibrium configura-  
 150 tion. Because of this idealized configuration, no one-to-one comparison can be made with  
 151 real observed system. Therefore, we focus on robust features that have been observed  
 152 in most squall lines and tropical cyclones in previous studies.

## 153 2 Methods

### 154 2.1 Isotopic variables

155 The water content in heavy isotopes ( $HDO$  or  $H_2^{18}O$ ) is expressed in ‰ as  $\delta D =$   
 156  $(R_D/R_{D,SMOW} - 1) \times 1000$  and  $\delta^{18}O = (R_{18O}/R_{18O,SMOW} - 1) \times 1000$ , where  $R_D$   
 157 and  $R_{18O}$  respectively are the ratio of Deuterium over Hydrogen atoms and of  $^{18}O$  over  
 158  $^{16}O$  atoms in the water, and SMOW is the Standard Mean Ocean Water reference. Here-  
 159 after, when there is a relatively large ratio of heavy over light atoms in some water, we  
 160 will refer to this water as relatively enriched in heavy atoms, or simply relatively enriched.  
 161 Conversely, when there is a relatively small ratio of heavy over light atoms in some wa-  
 162 ter, we will refer to this water as relatively depleted in heavy atoms, or simply relatively  
 163 depleted.

164 To first order,  $\delta D$  variations are 8 times those in  $\delta^{18}O$  (Craig, 1961), so we will fo-  
 165 cus on  $\delta D$  here. However, slight deviations in the  $\delta D - \delta^{18}O$  relationship can be quan-  
 166 tified by the second-order parameter d-excess:  $d = \delta D - 8 \cdot \delta^{18}O$ . It reflects kinetic ef-  
 167 fects associated with diffusivity differences between the different water isotopologues. We  
 168 will also show some results for d-excess as it can reflect kinetic effects in rain evapora-  
 169 tion or surface evaporation.



**Figure 1.** A schematic of the main structural elements of tropical cyclones (a) and squall lines (b). The observed typical spatial distribution of the surface precipitation is also indicated. The vertical and horizontal scales are approximate and may vary depending on events. The yellow, blue and pink rectangles in (a) respectively represent the eye, eyewall and rain band regions of the cyclone, as will be defined later in the article; the blue, pink and yellow rectangles in (b) respectively represent the convective, stratiform and trail regions of the squall line, as will also be defined later in the article.

170 **2.2 Model and simulations**

171 **2.2.1 Cloud Resolving Model**

172 We used the same Cloud Resolving Model (CRM) as that used in Risi et al. (2020),  
 173 namely the System for Atmospheric Modeling (SAM) non-hydrostatic model (Khairoutdinov  
 174 & Randall, 2003), version 6.10.9, which is enabled with water isotopes (Blossey et al.,  
 175 2010). This model solves anelastic conservation equations for momentum, mass, energy  
 176 and water, which is present in the model under six phases: water vapor, cloud liquid, cloud  
 177 ice, precipitating liquid, precipitating snow, and precipitating graupel. We used the bulk,  
 178 mixed-phase microphysical parameterization from (Thompson et al., 2008) in which wa-  
 179 ter isotopes were implemented (Moore et al., 2016). At the ocean surface, there is no rep-  
 180 resentation of sea spray. Therefore, we did not simulate the possible impact of sea spray  
 181 on the isotopic composition in the eye (Fudeyasu et al., 2008).

182 **2.2.2 Radiative-convective equilibrium with large-scale forcing**

183 Simulations are three-dimensional, with a doubly-periodic domain. They are run  
 184 in radiative-convective equilibrium over an ocean surface. The sea surface temperature  
 185 (SST) was set at 30°C. We did not prescribe any diurnal cycle.

186 Organized convection is typically observed in regions of large-scale ascent (Tan et  
 187 al., 2013; Jakob et al., 2019). Therefore, we impose a large-scale vertical ascent with a  
 188 cubic shape, reaching -40 hPa/d at 5 km and 0 hPa/d at the surface and above 100 hPa  
 189 (Risi et al., 2020). Simulations were also run without vertical ascent, and gave similar  
 190 results except that the convective systems were smaller and with a less well-defined in-  
 191 ternal structure. For example, the tropical cyclone without ascent does not show any eye  
 192 at the center. We thus focus on the simulations with large-scale ascent in the following.

193 The simulations were run for 50 days. The last 10 days of simulation are analyzed  
 194 with one three-dimensional output file every day. In radiative-convective equilibrium,  
 195 all snapshots exhibit similar behavior (videos in SI).

196 **2.2.3 Set-up for the cyclone simulation**

197 We used a domain of 1024 km×1024 km with a horizontal resolution of 4 km and  
 198 96 vertical levels. This horizontal resolution is sufficient to properly simulate the inter-  
 199 nal structure of a cyclone (Gentry & Lackmann, 2010). Cyclones spontaneously develop  
 200 in radiative-convective equilibrium simulations when some rotation is added (Khairoutdinov  
 201 & Emanuel, 2013; Muller & Romps, 2018). Here the effect of rotation is added through  
 202 a Coriolis parameter that corresponds to a latitude of 40°. Although no tropical cyclones  
 203 are expected to form at such latitudes, a strong rotation allows us to simulate a small  
 204 cyclone (Chavas & Emanuel, 2014) that can fit our small domain. This allows the sim-  
 205 ulation to remain computationally reasonable.

206 The initial conditions are spatially homogeneous and an unique cyclone develops  
 207 spontaneously through self-aggregation mechanisms after a few days. This is consistent  
 208 with the time scale for cyclogenesis in other self-aggregation studies (Muller & Romps,  
 209 2018).

210 **2.2.4 Set-up for the squall line simulation**

211 We used a domain of 256 km×256 km with a horizontal resolution of 2 km and 96  
 212 vertical levels. Squall lines spontaneously develop in radiative-convective equilibrium sim-  
 213 ulations when horizontal wind shear is added (Robe & Emanuel, 2001; Muller, 2013; Abramian  
 214 et al., 2022). We added a horizontally uniform wind in the x direction that reaches 10 m/s  
 215 at the surface and linearly decrease to 0 m/s at 1 km. According to Rotunno et al. (1988),

216 the shear with our settings leads to the formation of a strong and long-lived squall line,  
 217 perpendicular to the background wind. The uniform surface wind is subtracted when  
 218 calculating surface fluxes, to avoid this simulation from having significantly higher sur-  
 219 face fluxes. The radiative fluxes are imposed, because interactive radiation leads to some  
 220 radiative feedbacks that disfavors the organization into squall lines. The convection quickly  
 221 organizes into a line, after about one day of simulation.

## 222 **2.3 Simple box model of the sub-cloud layer**

223 To quantify the relative importance of processes in determining the isotopic com-  
 224 position in the different parts of the domain, we design a simple box model for the sub-  
 225 cloud layer (SCL) (Figure 2a). The model is the same as that in Risi et al. (2020), ex-  
 226 cept that here we account for horizontal advection and non-stationary effects. Account-  
 227 ing for these effects is necessary because the simple model will be applied in the differ-  
 228 ent sub-domains highlighted in Figure 1. Whereas the SCL is in quasi-equilibrium in the  
 229 domain-mean, the SCL is not in quasi-equilibrium in sub-domains. For example, the eye  
 230 of the cyclone wanders across the domain and is thus never in quasi-equilibrium.

231 Previous studies have used model tendencies along backward trajectories to quan-  
 232 tify the relative importance of processes on local conditions (Dütsch et al., 2018; Attinger  
 233 et al., 2019). This approach is most relevant when large-scale horizontal transport plays  
 234 a key role. In radiative-convective equilibrium however, local processes and vertical trans-  
 235 port dominate. A simple box model was thus judged more relevant.

### 236 **2.3.1 Water vapor budget**

237 The SCL is defined as the atmospheric levels from the surface to the SCL top, where  
 238 the SCL top is the level just below the lowest level where the domain-mean massic cloud  
 239 water content exceeds 10% of its maximum value. With this definition, we find that the  
 240 SCL is approximately well-mixed in our simulations. Therefore, in the box model, we  
 241 can assume that the specific humidity is vertically uniform throughout the SCL, consis-  
 242 tent with previous mixed-layer models (Stevens, 2006; Neggers et al., 2006; Albright et  
 243 al., 2022). We define  $W$  as the water mass in the SCL per area unit (in  $kg.m^{-2}$ ):

$$W = \int_{z=0}^{z_t} \rho \cdot q \cdot dz \quad (1)$$

244 where  $q$  is the specific humidity near the surface,  $\rho$  is the air density,  $z$  is the al-  
 245 titude above sea level and  $z_t$  is the altitude of the SCL top. The water budget of the SCL  
 246 in a given sub-domain writes (Figure 2a):

$$\frac{dW}{dt} = E_{sfc} + F_d \cdot q \cdot (r_d - 1) - F_u \cdot q \cdot (r_u - 1) + E_{horiz} + E_{ev} - E_c \quad (2)$$

247 where  $E_{sfc}$  is surface evaporation,  $F_d$  and  $F_u$  are the downward and upward mass  
 248 fluxes at SCL top,  $E_{horiz}$  is the flux of water through horizontal advection,  $E_{ev}$  is the  
 249 rain evaporation,  $E_c$  is some condensation that may occur if the SCL top is not horizon-  
 250 tally uniform,  $r_u = q_u/q$ ,  $r_d = q_d/q$ ,  $q_u$  and  $q_d$  are the specific humidity in updrafts  
 251 and downdrafts. All these variables are directly diagnosed from the simulations for each  
 252 sub-domains as detailed in (Risi et al., 2021).

### 253 **2.3.2 Isotopic budget**

254 As for the water vapor budget, the budget for heavy water vapor isotopes in the  
 255 SCL is:



$$\frac{d(R \cdot W)}{dt} = R_{sfc} \cdot E_{sfc} + F_d(R_d \cdot q_d - R \cdot q) - F_u(R_u \cdot q_u - R \cdot q) + R_{horiz} \cdot E_{horiz} + R_{ev} \cdot E_{ev} - R_c \cdot E_c \quad (3)$$

where  $R$  is the isotopic ratio of the near-surface vapor,  $R_u$  and  $R_d$  are the isotopic ratios in updrafts and downdrafts,  $R_{sfc}$ ,  $R_{horiz}$ ,  $R_{ev}$  and  $R_c$  are the isotopic compositions of the surface evaporation, horizontal advection, rain evaporation and condensation fluxes.

We define:

$$E_{res} = E_{horiz} - \frac{dW}{dt} \quad (4)$$

$E_{res}$  is the flux of water through both horizontal advection and non-stationary effects, and is calculated as a residual. For example, in the cyclone's eyewall where the air is very moist (the relative humidity is greater than 95%), we expect that horizontal advection will have a drying effect, i.e.  $E_{horiz} < 0$ . In addition, since the cyclone wanders across the domain, the eyewall often arrives in dry parts of the domain, i.e.  $\frac{dW}{dt} > 0$ . Therefore, both horizontal advection and non-stationary effects contribute to drying the eyewall, i.e.  $E_{res} < 0$ .

Similarly, we define the isotopic ratio of the flux  $R_{res}$ :

$$R_{res} = \frac{R_{horiz} \cdot E_{horiz} - \frac{d(R \cdot W)}{dt}}{E_{horiz} - \frac{dW}{dt}} \quad (5)$$

To solve the isotopic budget equation for  $R$ , the isotopic ratios  $R_{sfc}$ ,  $R_d$ ,  $R_u$ ,  $R_{res}$ ,  $R_{ev}$  and  $R_c$  are all expressed as a function of  $R$ . The isotopic ratio of surface evaporation is given by (Craig & Gordon, 1965):

$$R_{sfc} = \frac{R_{oce}/\alpha_{eq}(SST) - h \cdot R}{\alpha_K \cdot (1 - h)} \quad (6)$$

where  $R_{oce}$  is the isotopic ratio at the ocean surface,  $\alpha_{eq}(SST)$  is the equilibrium fractionation coefficient at the sea surface temperature,  $\alpha_K$  is kinetic fractionation coefficient (Merlivat & Jouzel, 1979) and  $h$  is the relative humidity normalized at the SST and accounting for ocean salinity calculated as follows:

$$h = q/q_{sat}^{surf}(SST) \quad (7)$$

where  $q_{sat}^{surf}(SST) = 0.98 \cdot q_{sat}(SST)$  and  $q_{sat}$  is the humidity saturation as a function of temperature at the sea level pressure. Variables  $h$  and the kinetic fractionation as a function of surface wind speed are diagnosed from the CRM. We assume that  $\delta D_{oce} = 0\text{‰}$ .

The isotopic ratios in updrafts and downdrafts are assumed to follow logarithmic functions:  $R_u = R \cdot (r_u)^{\alpha_u - 1}$  and  $R_d = R \cdot (r_d)^{\alpha_d - 1}$  where  $R_u$  and  $R_d$  are isotopic ratios in updrafts and downdrafts, and  $\alpha_u$  and  $\alpha_d$  are the  $q - \delta D_v$  steepness coefficients for updrafts and downdrafts (Risi et al., 2020). We set  $R_{res} = \phi_{res} \cdot R$ ,  $R_{ev} = \phi_{ev} \cdot R$  and  $R_c = \phi_c \cdot R$ . All parameters  $\alpha_u$ ,  $\alpha_d$ ,  $\phi_{res}$ ,  $\phi_{ev}$  and  $\phi_c$  can be diagnosed from the simulation for each sub-domain.

$$R = \frac{R_{oce}/\alpha_{eq}(SST)}{h + \alpha_K \cdot (1 - h) \cdot A} \quad (8)$$

286

where

$$A = \frac{((r_u)^{\alpha_u} - 1) + \frac{F_d}{F_u} \cdot (1 - (r_d)^{\alpha_d}) - \frac{E_{ev}}{qF_u} \cdot \phi_{ev} + \frac{E_c}{qF_u} \cdot \phi_c - \frac{E_{res}}{qF_u} \cdot \phi_{res}}{(r_u - 1) + \frac{F_d}{F_u} \cdot (1 - r_d) - \frac{E_{ev}}{qF_u} + \frac{E_c}{qF_u} - \frac{E_{res}}{qF_u}} \quad (9)$$

287

288

289

290

In absence of vertical mixing ( $r_u = r_d = 1$ ), rain evaporation, condensation, horizontal advection and non-stationary effects ( $E_{ev} = E_c = E_{res} = 0$ ), then  $A = 1$  and equation 8 reduced to the classical closure (Merlivat & Jouzel, 1979) (hereafter MJ79 closure).

291

292

293

294

295

296

297

298

299

300

301

302

303

304

Note that the diagnostic of  $E_{res}$  and  $\phi_{res}$  as residuals guarantees that the water and isotopic budgets of the SCL are closed. However, it does not guarantee that equation 8 with input parameters diagnosed from the CRM simulations yields exactly the same isotopic ratios as those directly simulated by the CRM. This is because many simplifying assumptions underlie the simple model. In particular, we neglect spatial and temporal co-variations between all the different parameters within each sub-domain. In addition, we diagnose the parameters related to vertical mixing ( $q_u$ ,  $q_d$ ,  $\alpha_u$ ,  $\alpha_d$ ) assuming a simple upstream advection scheme (Godunov, 1959), whereas the advection scheme in the SAM model is more sophisticated (Smolarkiewicz & Grabowski, 1990). As will be illustrated in sections 4.3 and 4.4, the simple box model of this study systematically overestimates the  $\delta D_v$  simulated by the CRM by a few ‰, and systematically underestimates  $d_v$  by a few ‰. While we keep in mind these mismatches, we will show that the simple box model is nevertheless able to simulate the isotopic variations between different sub-domains of the cyclone and the squall line, justifying its relevance.

305

### 2.3.3 Method to decompose anomalies in isotopic ratios

306

307

308

To decompose the anomaly in the isotopic ratio in each sub-domain relative to that in the environment, we categorize the parameters controlling  $R$  in our simple box model into 6 processes:

309

310

311

312

313

314

315

316

317

318

319

320

321

322

323

324

325

326

327

328

329

330

331

332

1. Parameter  $h$ : near-surface relative humidity, which impacts the kinetic processes during ocean surface evaporation (Merlivat & Jouzel, 1979) (Figure 2b, red).
2. Parameter  $\alpha_K$ : kinetic fractionation coefficient, which depends on the surface wind speed (Merlivat & Jouzel, 1979) (Figure 2b, orange).
3. Parameters  $F_d/F_u$ ,  $r_u$ ,  $r_d$ ,  $\alpha_u$  and  $\alpha_d$ : vertical mixing through the SCL top (green in Figure 2b). This includes the effect of the horizontal humidity contrasts ( $r_u$  and  $r_d$ ). When horizontal humidity contrasts between dry and moist zones of a sub-domain are larger, dry subsiding regions import drier air and more depleted water vapor into the SCL, while ascending regions export moister air and more enriched water vapor from the SCL. This has thus a depleting effect. This also includes the effect of variations in the steepness of the relationship between  $q$  and  $\delta D_v$  for updrafts and downdrafts ( $\alpha_u$  and  $\alpha_d$ ). When the  $q - \delta D_v$  steepness is larger, downdrafts import more depleted vapor into the SCL and updrafts export more enriched vapor out of the SCL (Risi et al., 2021). The  $q - \delta D_v$  steepness depends on the enriching or depleting processes that occur above the SCL. Typically, the dominant effect is rain evaporation above the SCL, which depletes the water vapor, especially near the melting level (Risi et al., 2021). While the effects of horizontal humidity contrasts and  $q - \delta D_v$  steepness were separated in (Risi et al., 2021), here we combine these two effects to reduce non-linear effects during the decomposition.
4. Parameters  $E_{ev}/(qF_u)$  and  $\phi_{ev}$ : rain evaporation in the SCL (purple in Figure 2b).
5. Parameters  $E_c/(qF_u)$  and  $\phi_c$ : condensation in the SCL (cyan in Figure 2b).
6. Parameters  $E_{res}/(qF_u)$  and  $\phi_{res}$ : horizontal advection and non-stationary effects (dark green in Figure 2b).

333 We can thus write the isotopic ratio predicted by the simple box model in each sub-domain  
 334  $s$  as

335  $R_s = R(p_{s,1}, p_{s,2}, p_{s,3}, p_{s,4}, p_{s,5}, p_{s,6}) = R\left((p_{s,i})_{i=1,6}\right)$ , where  $p_{s,i}$  is the vector  
 336 of parameters corresponding to the process  $i$  and the sub-domain  $s$ , and  $R$  is the func-  
 337 tion defined in equation 8. The difference between the isotopic ratio in the sub-domain  
 338 and that in the environment thus writes:

$$\Delta R_s = R\left((p_{s,i})_{i=1,6}\right) - R\left((p_{e,i})_{i=1,6}\right) \quad (10)$$

339 where  $p_{e,i}$  is the vector of parameters corresponding to the process  $i$  and the environ-  
 340 ment. For each sub-domain  $s$  and each process  $j$ , we estimate the contribution of pro-  
 341 cess  $j$  to  $\Delta R_s$  as:

$$c_{s,j} = R\left(\left(\frac{p_{s,i} + p_{e,i}}{2}\right)_{i=1,6,i \neq j}, p_{s,j}\right) - R\left(\left(\frac{p_{s,i} + p_{e,i}}{2}\right)_{i=1,6,i \neq j}, p_{e,j}\right) \quad (11)$$

342 Because  $R$  is a non-linear function of its parameters,  $\sum_{j=1}^6 c_{s,j}$  does not necessar-  
 343 ily equal  $\Delta R_s$ . We estimate the impact of non-linear effects as a residual:

$$c_{s,non\ linear} = \Delta R_s - \sum_{j=1}^6 c_{s,j} \quad (12)$$

### 344 **3 Simulated patterns and qualitative comparison with observations**

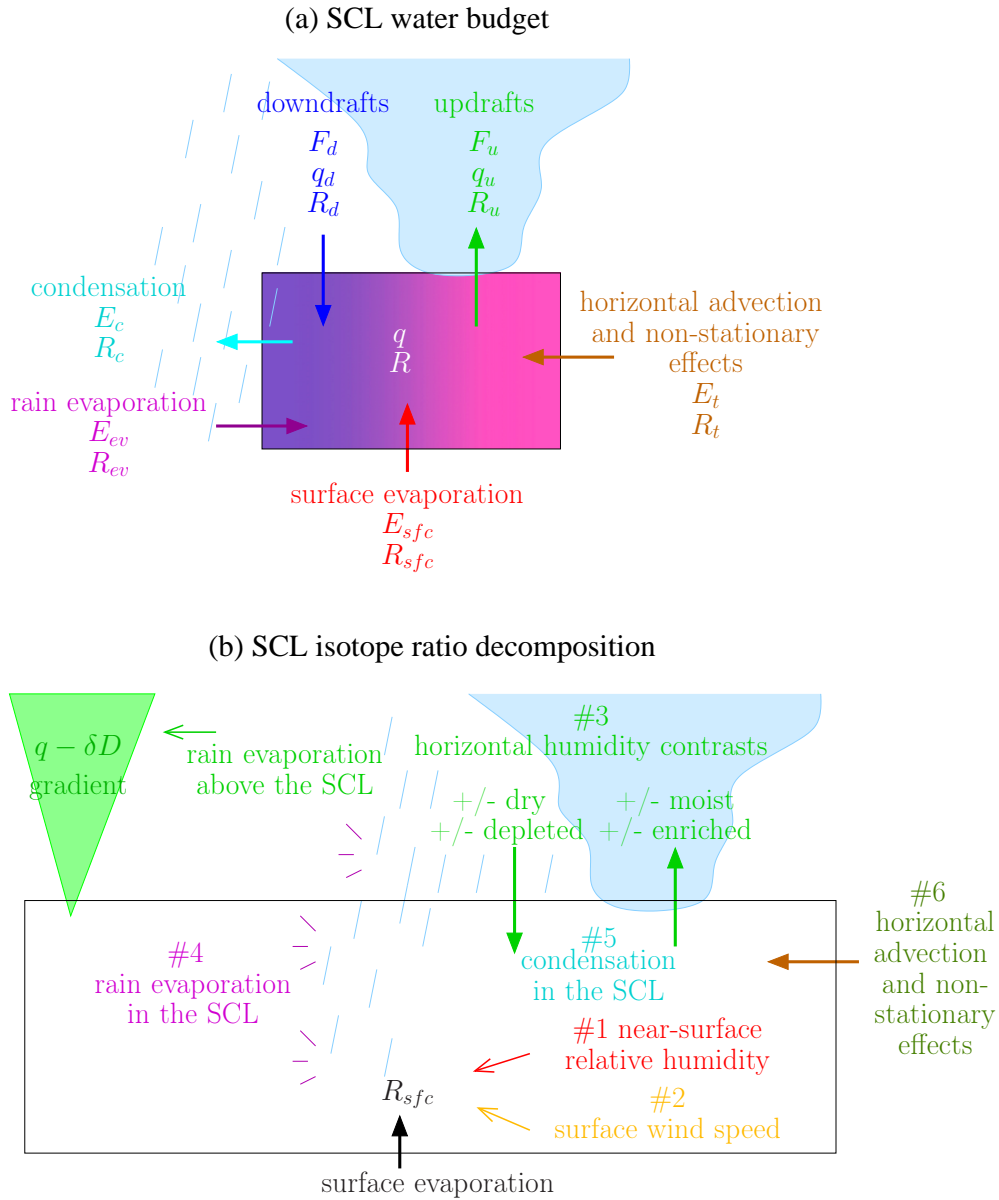
#### 345 **3.1 Tropical cyclone**

##### 346 **3.1.1 mesoscale structure**

347 To visualize the mesoscale structure of the tropical cyclone, in Figure 3 we plot maps  
 348 of the precipitation rate, near-surface air temperature, surface pressure anomaly, near-  
 349 surface relative humidity, near-surface water vapor  $\delta D$  and surface rain  $\delta D$  for an ar-  
 350 bitrary snapshot. The simulated cyclone exhibits features that are typical of observed  
 351 cyclones (Houze, 2010). It exhibits a small eye with weak precipitation (Figure 3a) and  
 352 warm air (Figure 3b), consistent with the subsidence in the eye. The eye is surrounded  
 353 by an eyewall and spiraling rain bands with intense precipitation and strong cyclonic winds.  
 354 Around the cyclone, strong compensating subsidence develops, leading to a dry environ-  
 355 ment and some scattered, isolated cumulus and cumulonimbus clouds and their cold pools  
 356 (Figure 3a-b,d).

357 To better document the different parts of the tropical cyclone, we plot composites  
 358 of meteorological and isotopic variables as a function of the distance  $r$  to the storm  
 359 center (Figure 4 and 5). All 10 snapshots were used to compute the composites. The storm  
 360 center is defined as the minimum surface pressure over the domain for each snapshot.  
 361 The typical structure of a tropical cyclone is well captured:

- 362 • The eye is associated with minimum pressure (around 50hPa lower than in the en-  
 363 vironment, typical of category 4 cyclones), a local minimum in precipitation, max-  
 364 imum near-surface air temperature and relative humidity and weak winds (Fig-  
 365 ure 4a-c). The eye is, however, too small to see the expected subsidence in Fig 4.
- 366 • The eyewall is associated with maximum precipitation and horizontal winds. The  
 367 air is strongly ascending, almost saturated throughout the full troposphere (Fig-  
 368 ure 5a), and condensation is intense except in the shallow sub-cloud layer (Fig-  
 369 ure 5c).



**Figure 2.** A description of the simple box model and of the decomposition of the SCL water vapor isotopic ratio into different contributions. (a) Simple model to predict the SCL water vapor composition. It accounts for surface evaporation, rain evaporation, cloud condensation, updrafts and downdrafts at the SCL top, and horizontal advection and non-stationary effects quantified as a water budget residual. (b) The 6 contributions in the decomposition of the isotopic ratio: near-surface relative humidity (#1, red) and surface wind speed (#2, orange), which both contribute to control the isotope composition of the surface evaporation; horizontal humidity contrasts and rain evaporation above the SCL (#3, green), which both contribute to the SCL depletion by vertical mixing; rain evaporation (#4, purple); condensation (#5, cyan) within the SCL; and horizontal and non-stationary effects (#6, dark green).

- 370 • Rainbands beyond the eyewall are associated with significant but weaker precip-  
 371 itation and winds. In rain band clouds, the condensation rate is large (orange shades  
 372 in Figure 5c). Below the rain band clouds, the air is dry (Figure 5a), allowing large  
 373 rates of snow sublimation and rain evaporation (blue shades in Figure 5c).

### 374 **3.1.2 Definition of sub-domains**

375 Based on the previous description of mesoscale structure, we divide all grid points  
 376 into 5 sub-domains. These sub-domains are defined automatically based on some arbi-  
 377 trary thresholds, to which results are not crucially sensitive. We define:

- 378 • the “eye” as grid points with  $r \leq r_{wall}$ , where  $r_{wall}$  is the first  $r$  value for which  
 379 the precipitation exceeds 20 times the domain-average precipitation (yellow rect-  
 380 angles in Figure 4).  
 381 • the “eyewall” as grid points with  $r_{wall} < r \leq r_{band}$ , where  $r_{band}$  is the first  $r$   
 382 value greater than  $r_{wall}$  and with the precipitation lower than 20 times the domain-  
 383 average precipitation (blue rectangles in Figure 4).  
 384 • the “environment” as grid points with  $r > r_{env}$ , where  $r_{env}$  is the first  $r$  value  
 385 greater than  $r_{band}$  and with the precipitation lower than 0.8 times the domain-  
 386 average precipitation (white in Figure 4).

387 The precipitation thresholds have been adjusted so as to match the simulated cyclone  
 388 structure to that described in previous studies (Houze, 2010). In between the eyewall  
 389 and the environment (pink rectangles in Figure 4), rain bands are not radially symmet-  
 390 ric. Therefore, we define “rain bands” as grid points with  $r_{band} < r \leq r_{env}$  and pre-  
 391 cipitation greater than 4 times the domain-average precipitation, and “in between rain  
 392 bands” as the other points.

### 393 **3.1.3 Simulated isotopic evolution**

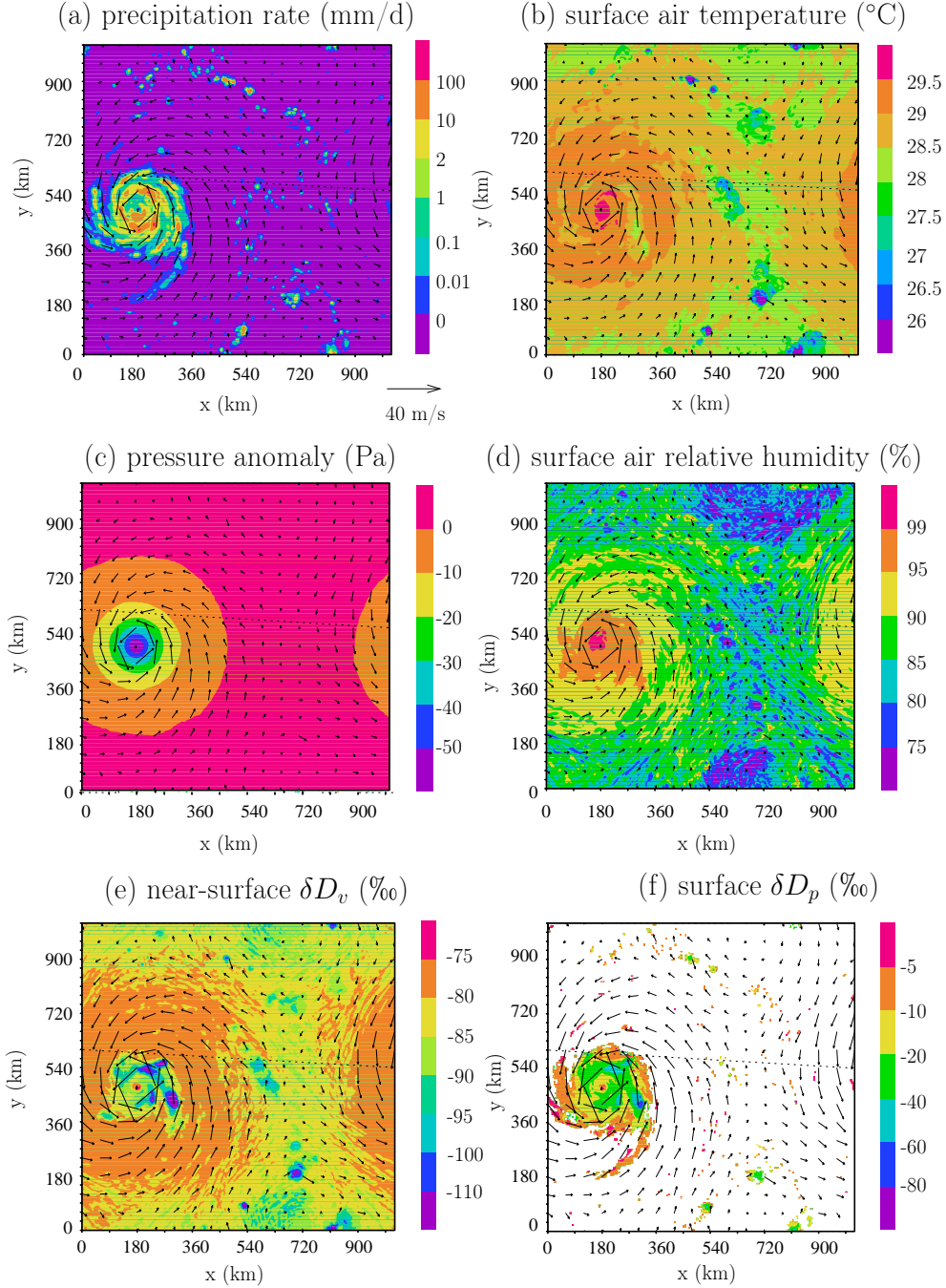
394 The water vapor is most enriched in the eye and in the dry environment (Figure  
 395 3e-f, Figure 4e), and most depleted in heavy isotopes in the eyewall and spiraling rain  
 396 bands. The water vapor d-excess is lower in the eye, and higher in the eyewall and rain  
 397 bands (4e-f). The water vapor is more depleted and has a higher d-excess within the rain  
 398 bands than between the rain bands (Figure 4d-e, dashed black).

399 The precipitation  $\delta D$  ( $\delta D_p$ ) varies in concert with the water vapor  $\delta D$  ( $\delta D_v$ ) where  
 400 the precipitation is highest (Figure 4e, dashed black). The precipitation is slightly more  
 401 depleted than if in equilibrium with the vapor. This suggests that the rain originates from  
 402 condensation at a higher altitude and undergoes little evaporative enrichment as it falls,  
 403 consistent with the high relative humidity. In addition, the rain quickly falls to the ground,  
 404 leaving less time for the rain to isotopically equilibrate with the vapor.

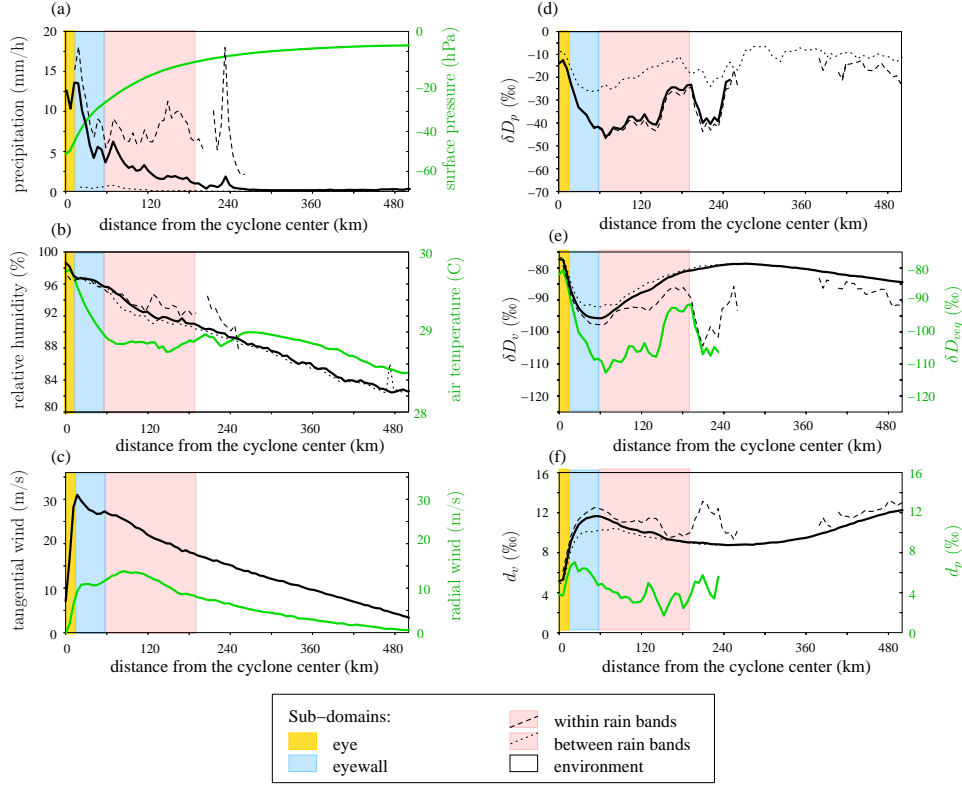
### 405 **3.1.4 Comparison with isotopic observations**

406 Observed isotopic patterns in tropical cyclones can be diverse (Guilpart, 2018). How-  
 407 ever, some robust features emerge from the literature.

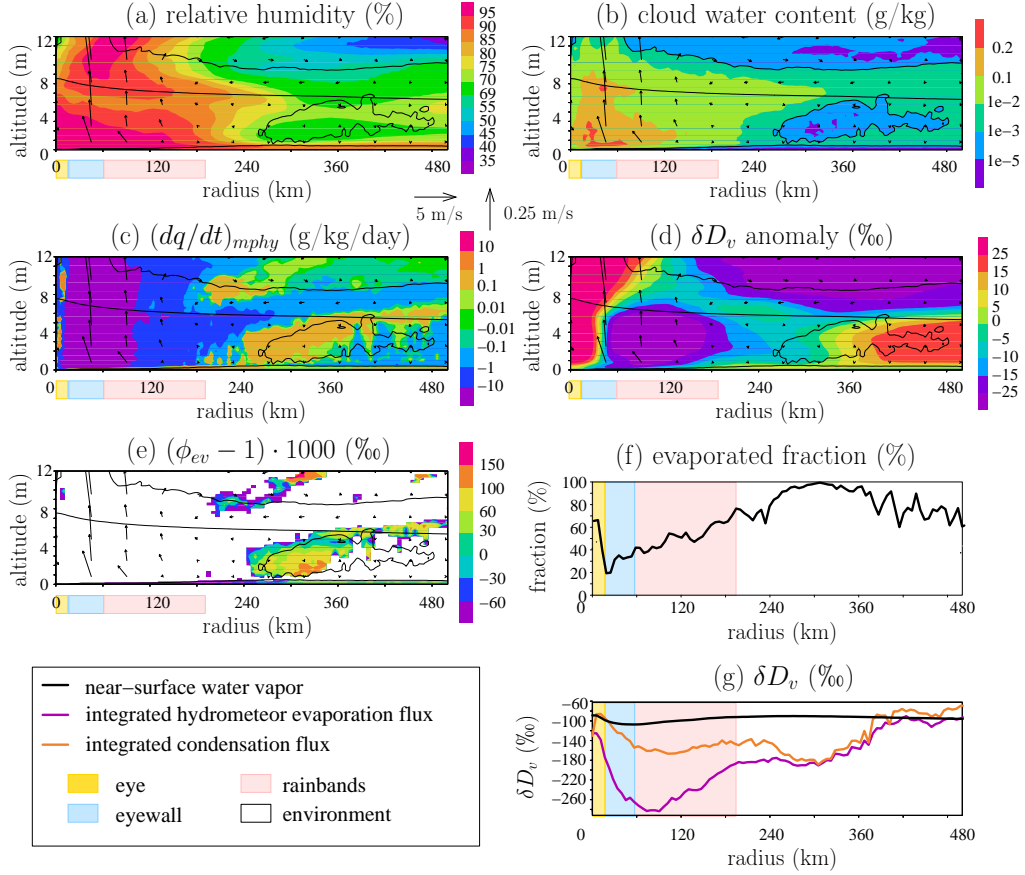
- 408 • There is a local maximum in  $\delta D_v$  and  $\delta D_p$  in the eye (Gedzelman et al., 2003; Fudeyasu  
 409 et al., 2008). This is what the CRM simulates, with  $\delta D_v$  and  $\delta D_p$  slightly greater  
 410 in the eye than in the environment (Figure 4e)  
 411 • Outside the eye, many studies have observed that the water vapor and precipi-  
 412 tation are more depleted towards the storm center (Gedzelman et al., 2003; Fudeyasu  
 413 et al., 2008; Munksgaard et al., 2015; Skrzypek et al., 2019; Xu et al., 2019; Sanchez-



**Figure 3.** Snapshots of the cyclone simulation: (a) precipitation rate, (b) near-surface air temperature, (c) surface pressure anomaly relative to the domain-mean, (d) near-surface relative humidity, (e) near-surface  $\delta D_v$  and (f)  $\delta D_p$ . The near surface winds are shown as arrows. Note that due to the doubly-periodic domain, the missing part of the cyclone on the left edge of the domain appears on the right edge of the domain. The snapshot was chosen as the one where the cyclone is the closest to the center of the domain, for easier visualization.



**Figure 4.** Evolution of surface variables as a function of distance to the cyclone center: precipitation rate (a, black), surface pressure (a, green), near-surface air temperature (b, black), near-surface relative humidity (b, green), tangential (c, black) and radial (c, green) wind, surface precipitation  $\delta D$  (d), near-surface water vapor  $\delta D$  (e, black), water vapor  $\delta D$  that would be in equilibrium with the precipitation (e, green), near-surface water vapor d-excess (f, black) and precipitation d-excess (f, green). The colored rectangles indicate the different sub-domains: the eye (yellow), the eyewall (blue), within and between rain bands (pink). The environment is in white. To distinguish between within rain bands and between rain bands, dashed and dotted black lines indicate the same as black lines but for grid points where the precipitation rate is respectively higher and lower than 4 times the domain-mean precipitation, representing respectively within rain bands and between rain bands.



**Figure 5.** Variables as a function of altitude and of the distance  $r$  to the cyclone center: (a) Relative humidity; (b) cloud water content (cloud condensate and cloud ice); (c) specific humidity tendency due to phase changes (negative and positive values represent condensation/deposition and evaporation/sublimation respectively); (d) water vapor  $\delta D$  anomaly relative to the domain-mean  $\delta D_v$  at each level; (e)  $(\phi_{ev} - 1) \cdot 1000$ , where  $\phi_{ev} = R_{ev}/R_v$  is the relative enrichment of the isotopic ratio of the hydrometeor evaporation/sublimation relative to the water vapor isotopic ratio.  $\phi_{ev}$  is shown only when significant hydrometeor evaporation/sublimation occurs; (f) fraction of the vertically-integrated condensation flux that evaporates before reaching the ground; (g)  $\delta D$  in the near-surface vapor (black), in the vertically-integrated condensation flux (orange) and in the vertically-integrated evaporation flux (purple). The vertically-integrated condensation and evaporation fluxes are calculated as the volumetric-mass weighted integrals of the specific humidity tendency due to phase changes (c) where the tendency is respectively negative and positive. In (a-e), the vectors show the radial and vertical components of the wind, with the vertical wind multiplied by 20 for better readability. The nearly-horizontal black line shows the  $0^\circ\text{C}$  isotherm. The black contours highlights the  $10^{-3}$  g/kg contour for cloud water content. The yellow, blue and pink boxes indicate the eye, eyewall and rainbands defined in section 3.2.2.



414 Murillo et al., 2019; Jackisch et al., 2022). This is what the CRM simulates (black  
415 line in the pink rectangle in Figure 4e).

- 416 • At a given distance from the storm center, the water vapor or precipitation is of-  
417 ten more depleted within rain bands than in between (Munksgaard et al., 2015;  
418 Guilpart, 2018; Sun et al., 2022). This is also what the CRM simulates (Figure  
419 4e, dashed black line relative to dotted black line).
- 420 • The observed d-excess in water vapor or precipitation is lower in the eye (Fudeyasu  
421 et al., 2008), higher in the environment and higher in the rain bands than in be-  
422 tween (Munksgaard et al., 2015). These observations are consistent with our sim-  
423 ulation (Figure 4f).

## 424 3.2 Squall line

### 425 3.2.1 mesoscale structure

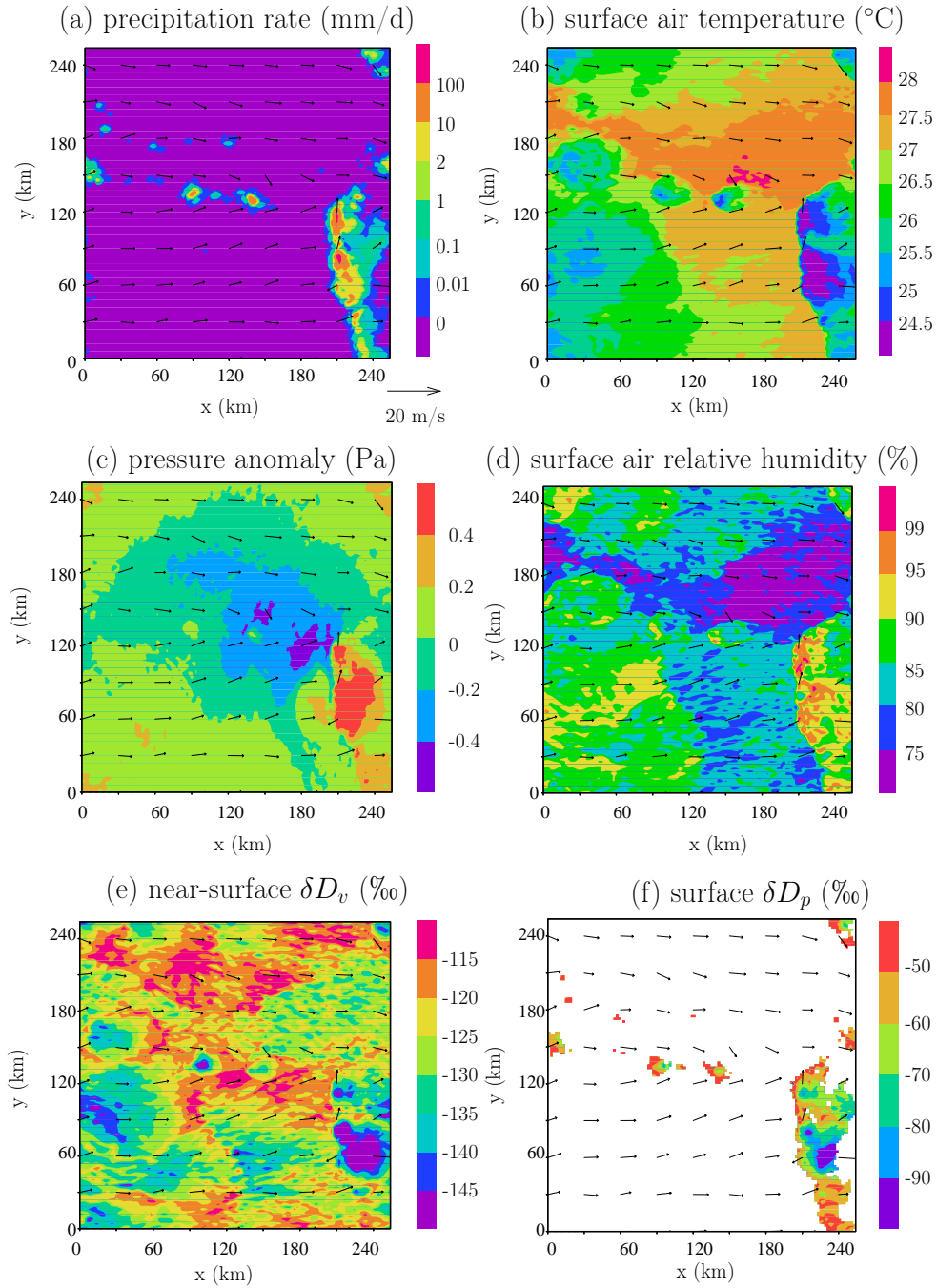
426 In presence of wind shear, the convection organizes into lines of intense precipita-  
427 tion perpendicular to the imposed surface winds (Figure 6a). The environment is dry,  
428 with only a few isolated cumulonimbi. Under the squall line, a cold pool is driven by mesoscale  
429 downdrafts (Zipser, 1977; Gamache & Houze, 1981). The cold pool has a sharp edge at  
430 the front of the line, corresponding to the gust front, and a long trail due to the imposed  
431 rearward horizontal winds near the surface (Figure 6b).

432 To better document the different parts of the squall line, we plot composites of me-  
433 teorological and isotopic variables as a function of the along-x distance to the gust front  
434 (Figures 7 and 8). For each snapshot and each value of  $y$  in the  $x$ - $y$  domain, we select  
435 the value of  $x$  where the along- $x$  gradient in surface pressure is at maximum. If the  $x$ -  
436 mean precipitation rate exceeds the  $x$ - $y$ -mean value and if the gradient in surface pres-  
437 sure exceeds 1.7 Pa/km, we assume that it is a gust front. This procedure was visually  
438 defined to optimally detect gust fronts. We define a new  $x$ -axis and translate all rows  
439 so that all gust fronts of the different rows are aligned at  $x_{gust}=30$  km. We arbitrary set  
440  $x_{gust}=30$  km so that the squall lines stand in the middle of the composite plots. Rows  
441 of the domain where the precipitation is lower than the domain mean, or where a gust  
442 front could not be identified, are considered “environment” and are not taken into ac-  
443 count in the composite.

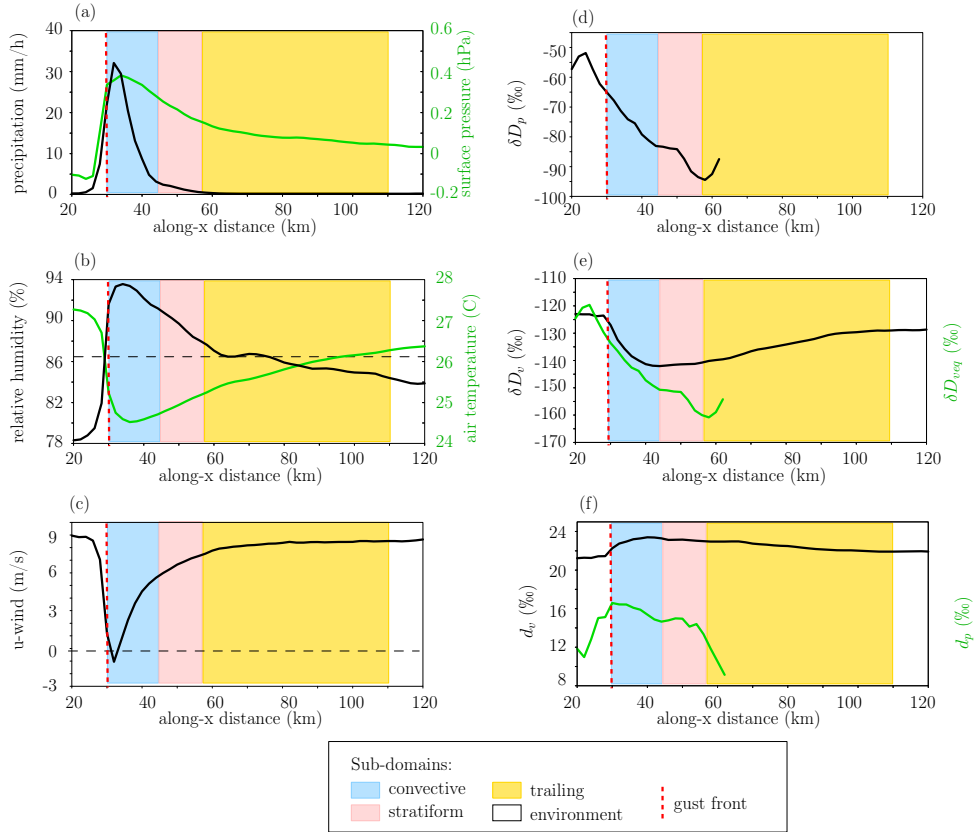
444 The precipitation rate is maximum just after the gust front (Figure 7a), consistent  
445 with observations (Chong, 2010). The precipitation maximum is located where the along-  
446  $x$  near-surface surface wind becomes null (Figure 7c), favoring the maintenance of strong  
447 updrafts (Rotunno et al., 1988). Elsewhere, the surface wind blows rearward. Near the  
448 gust front, the temperature drops and the relative humidity rises (Figure 7b). The re-  
449 covery to their environment value is slow due to the rearward advection.

450 Our simulated squall line shows only one precipitation peak. This is at odds with  
451 observations that often show two peaks, one for the convective region and one for the  
452 stratiform region, separated by a transition region (Biggerstaff & Houze Jr, 1991; Chong,  
453 2010). In our simulation, the convective region transitions continuously to the stratiform  
454 region. Increasing the horizontal resolution to 1 km did not help to simulate a transi-  
455 tion region.

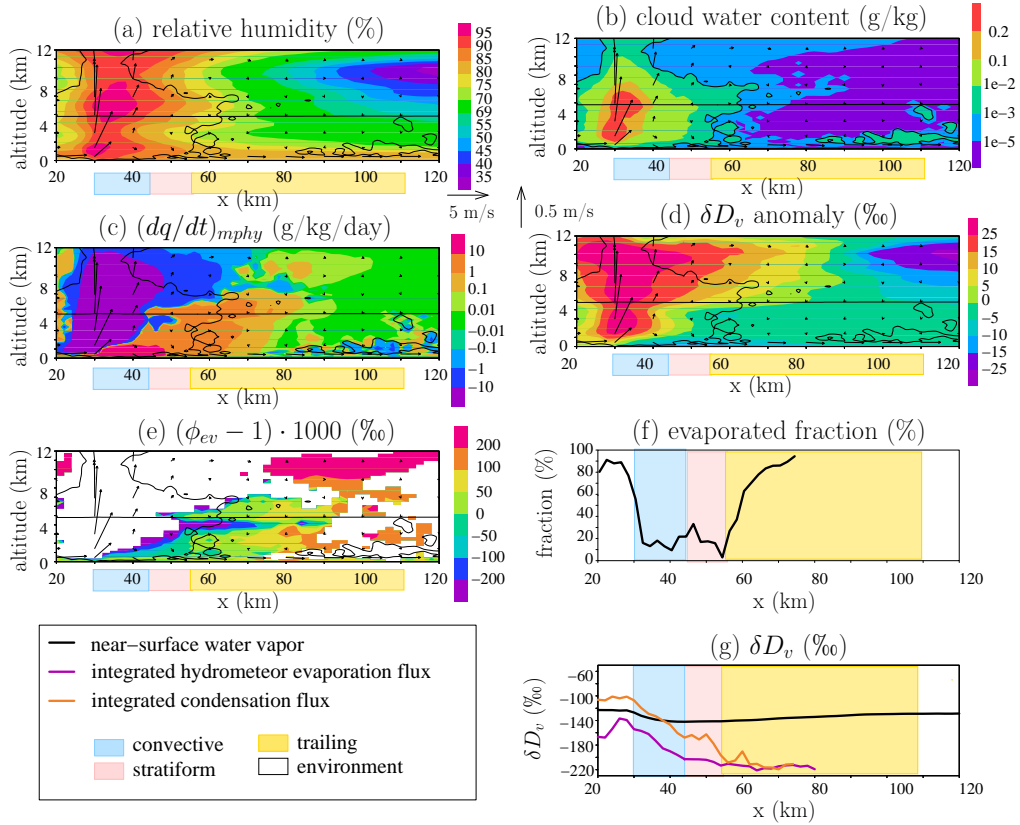
456 In spite of this shortcoming, the convective and stratiform regions of the squall line  
457 can be identified from water vapor tendencies (Figure 8c). The convective region can be  
458 identified by its intense condensation throughout the entire troposphere (Figure 8c, around  
459 50-60 km). The stratiform region can be identified by the condensation restricted to the  
460 upper troposphere (the anvil) and evaporation below (mesoscale downdraft) (Figure 8c,  
461 around 60-80 km). This pattern of condensation and evaporation is consistent with what  
462 we know from the squall line water budgets (Gamache & Houze, 1983; Chong & Hauser,  
463 1990).



**Figure 6.** Same as Figure 3, except that a snapshot of the squall line simulation is shown. Note that due to the doubly-periodic domain, the trailing region on the right edge of the domain continues on the left edge.



**Figure 7.** Same as Figure 4, except that variables for the squall line simulation are shown as a function of distance along the x-axis. The colored rectangles indicate the convective (blue), stratiform (pink) and trailing (yellow) sub-domains defined in section 3.2.2. The location of the gust front is indicated by the vertical red dashed line.



**Figure 8.** Same as Figure 5, except that variables for the squall line simulation are shown as a function of the distance along the x-axis. The blue, pink and yellow boxes indicate the convective, stratiform and trailing regions defined in section 3.2.2.

464

### 3.2.2 Definition of the sub-domains

465

466

467

468

469

Based on the above description of the mesoscale structure, we divide the grid points into 4 sub-domains: convective, stratiform, trailing and environment, as detailed below. Given the continuous transition in our simulations, we define the convective and stratiform sub-domains based on a precipitation threshold. For rows where  $x_{gust}$  is defined, we define:

470

471

472

473

474

475

476

477

478

479

480

481

- the convective region with  $x$  between  $x_{gust}$  and  $x_{conv}$ , where  $x_{conv}$  is the lowest  $x$  value greater than  $x_{gust}$  and where the precipitation is lower than 8 times the domain-average precipitation (yellow rectangle in Figure 7). This precipitation threshold was adjusted to coincide with the transition from condensation through the entire troposphere to condensation above the melting level and rain evaporation below (Figure 8c), consistent with the transition from the convective to the stratiform region (Gamache & Houze, 1983; Chong & Hauser, 1990).
- the stratiform region with  $x$  between  $x_{conv}$  and  $x_{strati}$ , where  $x_{strati}$  is the lowest  $x$  value greater than  $x_{conv}$  and where the precipitation is below the domain-average precipitation (blue rectangle in Figure 7). This precipitation threshold was adjusted to coincide with the maximum extension of the anvil clouds, consistent with the typical structure of a squall line (Houze, 1977).

482

483

484

485

486

487

The horizontal winds near the surface spread the cold pool rearward beyond the precipitating region. Therefore, we also define a sub-domain called “trailing”, with  $x$  between  $x_{strati}$  and  $x_{trail}$ , where  $x_{trail}$  is the  $x$  value greater than  $x_{strati}$  and  $T(x) < T(x_{gust}) - 1$ , where  $T$  is the near-surface temperature in K (pink rectangle in Figure 7). All grid points that are not categorized as “convective”, “stratiform” or “trailing” are called “environment” (white in Figure 7).

488

### 3.2.3 Simulated isotopic evolution

489

490

491

492

Simulated squall lines show a progressive depletion of the vapor in the convective region, maximum depletion at the end of the convective region, and a long recovery in the stratiform and trailing regions (Figure 7e). The  $\delta D_v$  reaches its environment value about 100 km after the convective peak.

493

494

495

496

497

The  $\delta D_p$  varies in concert with  $\delta D_v$  (Figure 7d). In the convective and stratiform regions,  $\delta D_p$  is lower than in equilibrium with the vapor (green in Figure 7e), consistent with a quick fall and little evaporative enrichment. The weak precipitation that falls upwind of the convective region, where the air is dry, has a  $\delta D_p$  higher than that in equilibrium with vapor, indicating evaporative enrichment during rain evaporation.

498

499

500

501

The d-excess in the vapor is higher in the convective and stratiform regions, and to a lesser extent in the trailing region, than in the environment (Figure 7f). The low d-excess in the precipitation reflect the effect of evaporative enrichment, especially before the gust front and in the trailing region.

502

### 3.2.4 Comparison with isotopic observations

503

504

505

506

507

Isotopic observations during squall lines often show a “W” shape with minimum  $\delta D_p$  in the convective and stratiform regions and a local maximum in the transition region (Taupin & Gallaire, 1998; Risi et al., 2010). Our simulation is consistent with this observation, except that since our simulation does not exhibit any transition region, it shows a “V” shape instead of a “W” shape.

508

509

Investigation of the vapor isotopes of the squall lines in the Sahel showed that the isotopic evolution can be diverse, but some robust features emerge (Tremoy et al., 2014).

- 510 • In 80% of 74 observed squall lines (Tremoy et al., 2014), there is a depletion in  
511 the convective region compared to the environment before the squall line. This  
512 is consistent with our simulation.
- 513 • More than half of the observed squall lines show additional depletion in the strat-  
514 iform region (Tremoy et al., 2014). This is also consistent with our simulation.
- 515 • For squall lines showing an isotopic depletion in the convective or stratiform re-  
516 gion, the recovery from this depletion takes several hours after the end of the rain  
517 (Tremoy et al., 2014). Considering a propagation speed of about 20 m/s, this is  
518 consistent with the recovery distance of about 100 km in our simulation.
- 519 • In 78% of observed squall lines, the “W” shape often observed in the precipita-  
520 tion is not observed in the vapor (Tremoy et al., 2014). Our simulations are thus  
521 consistent with this majority of squall lines

522 Our simulated isotopic evolution during the squall line thus captures the features that  
523 are most commonly observed in squall lines. Some squall lines may feature different vari-  
524 ations, and even enrichment in the convective and stratiform regions (Tremoy et al., 2014).  
525 To check whether our simulations could capture such a diversity of isotopic variations,  
526 we performed many sensitivity tests, including simulations without large-scale ascent,  
527 with large-scale ascent peaking in the upper troposphere to favor stratiform development  
528 (Su et al., 2000), increased horizontal resolution, interactive radiation, reduced sublima-  
529 tion or reduced rain evaporation to favor the maintenance of the stratiform region (Yang  
530 & Houze Jr, 1995; Bryan & Morrison, 2012), bowling alley domain, or prescribed hor-  
531 izontal wind in the upper troposphere to favor the development of the stratiform region  
532 (Caniaux et al., 1994). Depending on the simulations, the stratiform region is more or  
533 less extended and the squall lines are more or less organized, but the meteorological and  
534 isotopic evolution is always similar. We thus keep in mind that our simulations match  
535 the majority of squall lines, but not all of them. In addition, some observed feature in  
536 squall lines over land, such as the enrichment of the water vapor in some stratiform re-  
537 gions due to nearly total evaporation of the rain in dry air (Tremoy et al., 2014), can-  
538 not be captured by our simulations with an oceanic setting that favors a relatively moist  
539 environment.

## 540 4 Understanding mesoscale isotopic variations

541 Since  $\delta D$  in the precipitation varies in concert with  $\delta D$  in the SCL water vapor,  
542 we will focus most of this section on understanding mesoscale isotopic variations in the  
543 water vapor.

### 544 4.1 Impact of rain evaporation

545 Observational and modeling studies highlighted the key role of rain evaporation and  
546 rain-vapor exchanges in depleting the water vapor within organized systems (Lawrence  
547 et al., 2004; Tremoy et al., 2014; Xu et al., 2019). Around tropical cyclones and down-  
548 wind of squall lines, dry air (Figures 5a 8a) favors thick layer of rain evaporation in the  
549 middle and lower troposphere (orange shade in Figures 5c, 8c). Maximum evaporation  
550 occurs in the cold pool under the convective and stratiform regions of the squall line (red  
551 shade in Figure 8c). The fraction of the formed condensate that evaporates before reach-  
552 ing the ground is in the range of 0-30% in the convective and stratiform regions of the  
553 squall line and 20-40% in the eyewall of the cyclone. It increases to 40-80% in the rain  
554 bands of the cyclone and reaches 100% in the trailing region of the squall line and in the  
555 environment of both systems (Figures 5f, 8f).

556 To analyze the isotopic effect of rain evaporation in more detail, we calculate  $\phi_{ev} =$   
557  $R_{ev}/R_v$ , where  $R_v$  and  $R_{ev}$  are the isotopic ratios in water vapor and in hydrometeor  
558 evaporation/sublimation. Hydrometeor evaporation/sublimation includes the evapora-

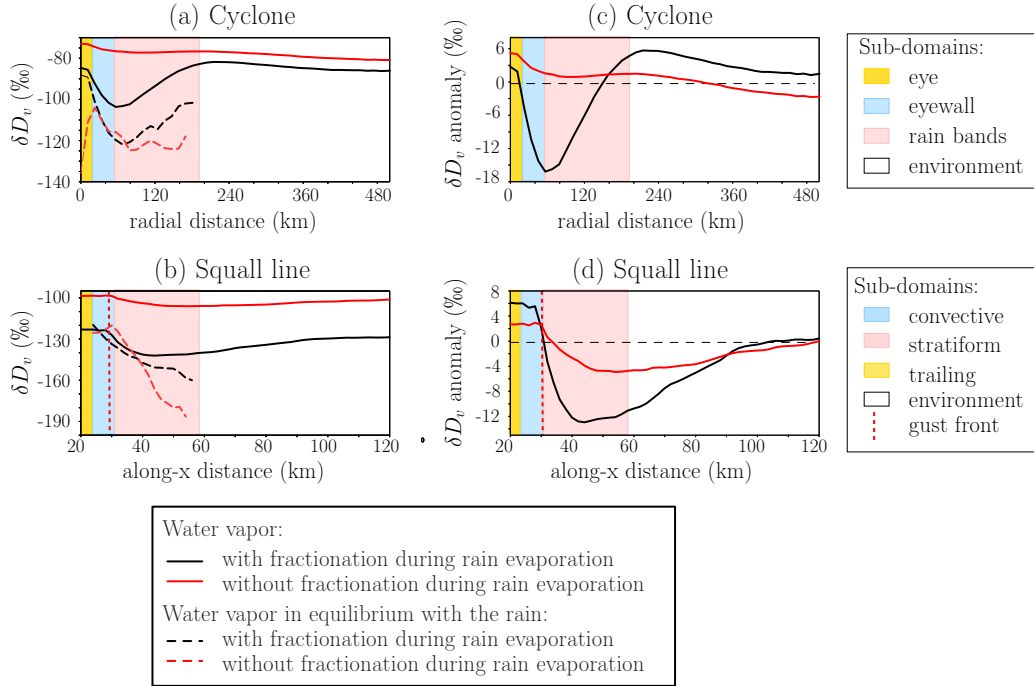
559 tion of rain and cloud droplets (below the 0°C isotherm) and the sublimation of ice, snow  
 560 or graupel (above the 0°C isotherm).  $\phi_{ev}$  represents the isotope enrichment of hydrometeor  
 561 evaporation relative to water vapor: if  $(\phi_{ev} - 1) \cdot 1000 > 0\text{‰}$ , hydrometeor evap-  
 562 oration/sublimation enriches the water vapor; if  $(\phi_{ev} - 1) \cdot 1000 < 0\text{‰}$ , hydrometeor  
 563 evaporation/sublimation depletes the water vapor.  $R_{ev}$  is calculated as  $(dq_{HDO}/dt)_{mphy}/(dq/dt)_{mphy}$ ,  
 564 where  $q_{HDO}$  is the mixing ratio for  $HDO$  (i.e. the ratio of the mass of  $HDO$  molecules  
 565 over the mass of dry air),  $q$  is the water vapor mixing ratio, and  $(dq/dt)_{mphy}$  and  $(dq_{HDO}/dt)_{mphy}$   
 566 are the water vapor and  $HDO$  tendencies associated with phase changes. If hydrometeor  
 567 evaporation/sublimation occurs, then the  $(dq/dt)_{mphy}$  tendency is positive. In con-  
 568 trast, if cloud condensation or deposition onto the snow, cloud ice or graupel occurs, then  
 569 the  $(dq/dt)_{mphy}$  tendency is negative.  $R_{ev}$  is calculated only where  $(dq/dt)_{mphy} > 0$ .  
 570 We can see that near the rain bands of the cyclone and in the stratiform region of the  
 571 squall line, there is a strongly depleting effect of rain evaporation just below the melt-  
 572 ing level (purple shade in Figures 5e and 8e, where the nearly horizontal black line in-  
 573 dicates the 0°C isotherm). This is because just below the melting level, most of the rain  
 574 originates from snow melt, which is more depleted than locally formed droplets because  
 575 the snow has formed at higher altitudes (Risi et al., 2021). There is also a depleting ef-  
 576 fect in the sub-cloud layer of the cyclone. The depleting effect of rain evaporation is as-  
 577 sociated with negative anomalies in water vapor  $\delta D$  in the rain bands and under the eye-  
 578 wall of the cyclone, and under the stratiform region of the squall line (purple shade in  
 579 Figures 5d and 8d).

580 Figures 5g and 8g compare the  $\delta D$  in the near-surface vapor, the vertically-integrated  
 581 condensation flux and the vertically-integrated hydrometeor evaporation flux. The  $\delta D$   
 582 in the vertically-integrated condensation flux is most negative in the eyewall and rain  
 583 bands of the cyclone and in the convective and stratiform regions of the squall line (or-  
 584 ange in Figures 5f and 8f), consistent with the condensation occurring higher in altitude  
 585 in anvil clouds (Figures 5b-c and 8b-c). Where the formed condensate evaporates totally,  
 586 or nearly totally, as is the case in the environment, then the  $\delta D$  in the vertically-integrated  
 587 hydrometeor evaporation flux equals that in the vertically-integrated condensation flux  
 588 (orange and purple lines are almost identical in Figures 5f and 8f). There, the rain evap-  
 589 oration has an enriching effect (yellow and orange shades in Figures 5e and 8e), consis-  
 590 tent with Tremoy et al. (2014). In contrast, where the evaporated fraction is smaller and  
 591 the air is moister, the  $\delta D$  in the vertically-integrated hydrometeor evaporation flux is  
 592 more depleted than in the vertically-integrated condensation flux (orange in Figures 5f  
 593 and 8f). This contributes to the depleting effect of the rain evaporation in the eyewall  
 594 and rain bands of the cyclone and in the convective and stratiform regions of the squall  
 595 line (purple shade in Figures 5e and 8e), and thus to the depleted water vapor in these  
 596 regions (purple shade in Figures 5d and 8d, black line in Figures 5f and 8f).

597 To summarize, rain evaporation and diffusive exchanges deplete the water vapor  
 598 under the convective and stratiform regions of the cyclone and of the squall line. This  
 599 depletion is due to the higher condensation altitude, the lower evaporated fraction of the  
 600 rain and the moister air in which the evaporation occurs. The depleting effect of rain  
 601 evaporation and diffusive exchanges in stratiform regions of convective systems has al-  
 602 ready been highlighted in previous studies (Kurita, 2013; Aggarwal et al., 2016), includ-  
 603 ing in cyclones (Lawrence et al., 2004; Munksgaard et al., 2015).

## 604 4.2 Simulations with de-activated fractionation during rain evaporation

605 To quantify the effect of rain evaporation and rain-vapor diffusive exchanges, ad-  
 606 ditional simulations in which these effects are de-activated were run, as in Field et al.  
 607 (2010); Risi et al. (2021). The fractionation coefficients were set to unity during all post-  
 608 condensational processes. This impacts the simulation of isotopic variables, but not the  
 609 simulation of meteorological variables.



**Figure 9.** Results of the simulations with de-activated fractionation during rain evaporation. (a-b) Evolution of near-surface  $\delta D_v$  (solid) and of the water vapor  $\delta D$  in equilibrium with the near-surface precipitation (dashed) as a function of  $r$  for tropical cyclones (a) and as a function of  $x$  for squall lines (b), in simulations where rain evaporation and rain-vapor diffusive exchanges are activated (black) and de-activated (red). (c-d) Evolution of the near surface  $\delta D_v$  anomaly relative to the environment as a function of  $r$  for tropical cyclones (c) and as a function of  $x$  for squall lines (d).

610 Both in the cyclone and squall line simulations,  $\delta D_v$  would be higher without frac-  
 611 tionation during rain evaporation than with fractionation (solid lines in Figure 9a-b),  
 612 and the rain would be more depleted than if in equilibrium with the vapor (dashed lines  
 613 in Figure 9a-b). The  $\delta D_p$  would be minimum under the rain bands in the cyclone sim-  
 614 ulation, and under the stratiform region in the squall line simulation, reflecting the higher  
 615 condensation altitude in these regions (section 4.1). With fractionation during rain evap-  
 616 oration, the rain-vapor diffusive exchanges make the vapor more isotopically enriched  
 617 and the rain more isotopically depleted, thus bringing the rain and the vapor close to  
 618 isotopic equilibrium (black lines in Figure 9a-b). This explains why  $\delta D_v$  and  $\delta D_p$  vary  
 619 in concert. The  $\delta D_v$  is most negative where the condensation forms high in altitude and  
 620 where the rain rate is strong enough to isotopically impact the water vapor. This explains  
 621 why the  $\delta D_v$  is most negative near the transition between the eyewall and rainbands for  
 622 the cyclone, and between the convective and stratiform region for the squall line (black  
 623 solid lines in Figure 9a-b).

624 Without fractionation during rain evaporation, the mesoscale  $\delta D_v$  variations would  
 625 be strongly reduced (Figure 9c-d). Peak-to-peak variations are respectively about 22‰  
 626 and 8‰ with and without fractionation in the cyclone case, and respectively 18‰ and  
 627 8‰ in the squall line case. This confirms the key role of rain evaporation and rain-vapor  
 628 diffusive exchanges to deplete the low-level water vapor at the mesoscale scale.



629

### 4.3 Insights of the simple box model for the tropical cyclone

630

631

632

633

634

635

To quantify the relative importance of different processes in different regions of the domain in the case of the cyclone, a simple box model of the SCL was designed (section 2.3). We quantify the different terms of the water vapor budget in the different regions (section 4.3.1), before evaluating the predictions by the simple box model and decomposing the isotopic differences for water vapor  $\delta D$  (section 4.3.2) and d-excess (section 4.3.3).

636

#### 4.3.1 SCL water vapor budget

637

638

639

640

641

642

643

644

645

646

647

648

649

650

651

652

653

654

The cyclone in itself (sub-domains eye, eyewall, rain bands and between rain bands) covers less than 10% of the domain (Figure 10a). In all sub-domains, the main source of water vapor is surface evaporation (red in Figure 10b). Surface evaporation is more than twice larger in the eyewall, in rain bands and in-between rain bands than in the environment, consistent with the maximum winds. Surface evaporation is half smaller in the eye than in the environment, due to weak winds and moist near-surface air. Rain evaporation is also a significant moistening term in the eyewall and the rain bands (purple in Figure 10b). Condensation is insignificant (cyan in Figure 10b). Everywhere except in the eye, updrafts and downdrafts have a drying effect on the SCL (green and blue), because updrafts are preferentially moister and downdrafts are preferentially drier. In the eye, updrafts and downdrafts slightly moisten the SCL because the core of the eye is descending and almost saturated whereas air parcels near the eyewall may be drier and ascending. Horizontal advection and non-stationary effects dry the cyclone and slightly moisten the environment (dark green in Figure 10b). This is because dry air from the environment converge towards to cyclone center (horizontal advection effect). In addition, the cyclone wanders across the domain and thus mixes with air that was previously in the dry portions of the domain (non-stationary effect). In turn, in the wake of the cyclone, the environment is left moistened.

655

#### 4.3.2 Evaluating and decomposing the water vapor $\delta D_v$ in each sub-domain

656

657

658

659

660

661

662

In spite of its overestimate of the  $\delta D_v$  simulated by the CRM (green compared to red in Figure 11c), the simple box model is able to simulate the more depleted water vapor in all sub-domains relative to the  $\delta D_v$  simulated by the MJ79 closure (blue in Figure 10c) and to capture the main  $\delta D_v$  differences between the sub-domains: the slightly more enriched vapor in the eye relative to environment, the more depleted vapor in the eyewall, rain bands and between bands relative to the environment, and the more depleted vapor in the bands relative to in-between bands (Figure 10c).

663

664

665

666

667

668

669

670

671

672

673

The slightly more enriched vapor in the eye relative to the environment is mainly explained by the higher relative humidity (red in Figure 10d). When the relative humidity is close to 1, the vapor is nearly in equilibrium with the ocean. In this case, the influence of  $A$  in equation 8 becomes small. As a consequence, the prediction by the simple box model is close to that of the MJ79 closure in the eye (blue in in Figure 10c). In addition, in the eye, the rain evaporation is weak because there is little rain, and the horizontal advection into the eye is weak because the winds are mainly tangential, further limiting the influence of  $A$  in equation 8. We note that a relative enrichment inside the eye is simulated in spite of the neglect of sea spray in our simulations. This does not exclude the possibility for a role of sea spray in nature (Fudeyasu et al., 2008), but this role is not necessary to explain the enrichment in the eye.

674

675

676

677

678

Figure 10d shows the decomposition of the  $\delta D_v$  anomaly for the different sub-domains relative to the environment. The more depleted vapor in the eyewall relative to the environment is mainly due to the enhanced depletion by vertical mixing trough the SCL top (green in Figure 10d). The steeper isotopic gradient above the SCL, driven by the

678 stronger rain evaporation above the SCL (section 4.1), likely contributes to this term.  
 679 The rain evaporation in the SCL also contributes to the more depleted vapor in the eye-  
 680 wall (purple in Figure 10d).

681 The more depleted vapor in the rain bands relative to the environment (and rela-  
 682 tive to between rain bands) is mainly due to the enhanced depletion by rain evapora-  
 683 tion within the SCL (purple in Figure 10d). Alone, rain evaporation in the SCL would  
 684 deplete the vapor in the rain bands by nearly 50‰ relative to the environment. As in  
 685 the eyewall, the enhanced depletion by vertical mixing through the SCL top also contributes  
 686 to the depletion (green in Figure 10d).

687 The more depleted vapor between rain bands relative to the environment is mainly  
 688 due to the horizontal advection and non-stationary effects (dark green in Figure 10d).  
 689 Horizontal winds bring depleted water vapor from the rain bands (Figure 10d). Consis-  
 690 tently, horizontal advection has an enriching effect in rain bands, where the isotopic gra-  
 691 dients are reversed: horizontal winds bring enriched water vapor from in-between rain  
 692 bands.

693 The contributions of the kinetic fractionation during surface evaporation and of con-  
 694 densation in the SCL are marginal. The marginal impact of condensation is consistent  
 695 with the absence of clouds in the SCL, and confirms that rain-out does not directly im-  
 696 pact the SCL isotopic composition.

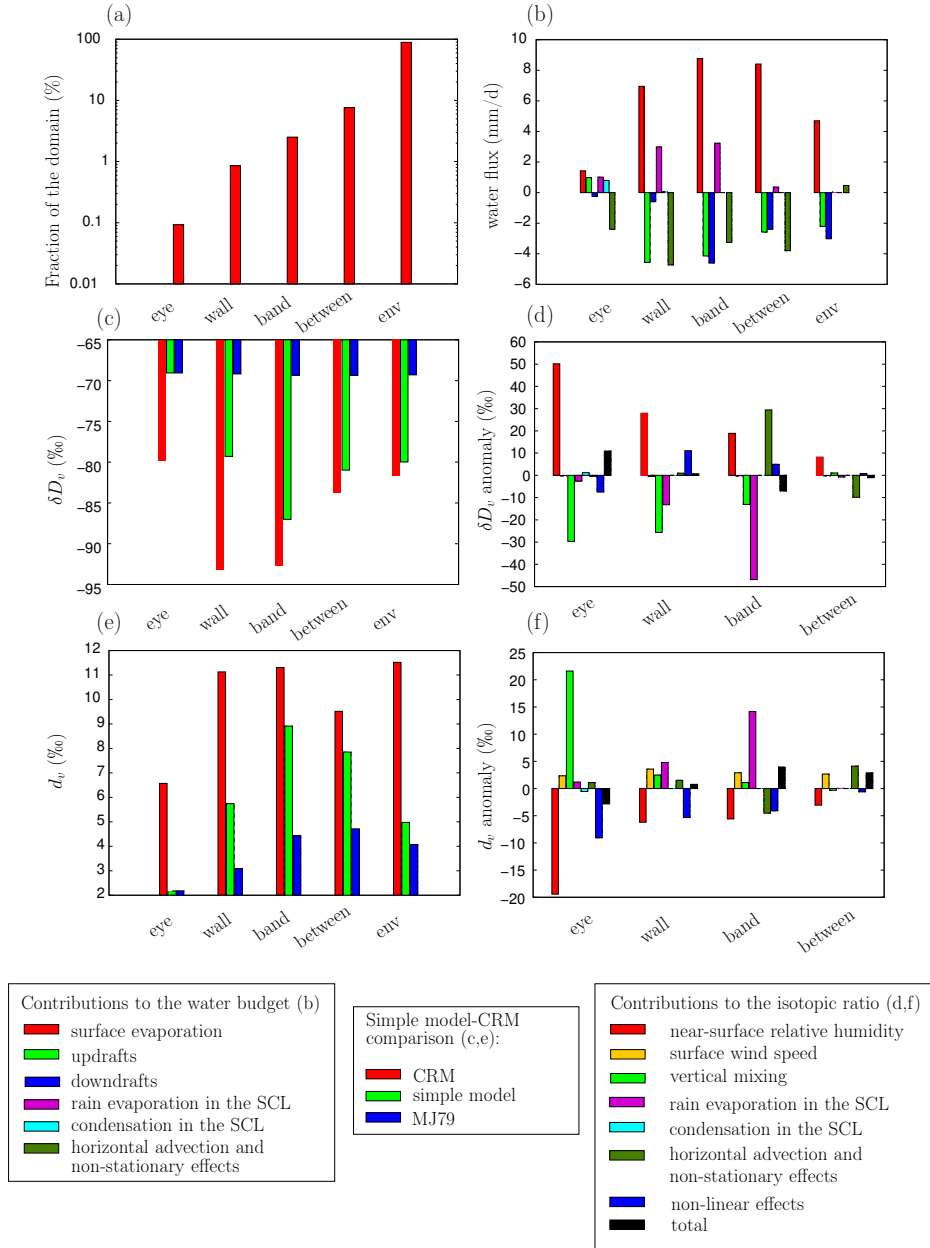
697 To summarize, the  $\delta D_v$  differences between the sub-domains are mainly explained  
 698 by rain evaporation and rain-vapor diffusive exchanges inside the SCL, which deplete the  
 699 eyewall and rain bands, consistent with previous studies (Gedzelman et al., 2003). Hor-  
 700 izontal advection plays a key role to smooth the isotopic patterns, contributing to the  
 701 progressive depletion as the air converges towards the center of cyclones suggested in pre-  
 702 vious studies (Gedzelman et al., 2003; Xu et al., 2019).

### 703 ***4.3.3 Evaluating and decomposing the water vapor d-excess in each sub-*** 704 ***domain***

705 In spite of its underestimate of the  $d_v$  simulated by the CRM (green compared to  
 706 red in Figure 10e), the model is able to capture the higher  $d_v$  than predicted by MJ79  
 707 closure (blue in Figure 10c) and several features simulated by the CRM: (1) the mini-  
 708 mum  $d_v$  in the eye, (2) the higher  $d_v$  in the rain bands relative to between bands (Fig-  
 709 ure 10e). We will thus focus on understanding these two features:

- 710 1. The minimum  $d_v$  in the eye is due to the moist conditions that reduce the kinetic  
 711 effects during surface evaporation (red in Figure 10f).
- 712 2. The higher  $d_v$  in the rain bands relative to the environment or between bands is  
 713 mainly due to the rain evaporation in the SCL (purple in Figure 10f), which yields  
 714 water vapor with high d-excess because of the relatively larger diffusivity of  $HDO$   
 715 relative to that of  $H_2^{18}O$ . Alone, it would contribute to an increase in d-excess by  
 716 more than 15‰ relative to the environment.

717 As for  $\delta D_v$ , horizontal advection acts to smooth the  $d_v$  patterns, decreasing  $d_v$  in the  
 718 rain bands and increasing it between rain bands (dark green in Figure 10f). The stronger  
 719 the winds, the larger the kinetic fractionation during surface evaporation (yellow in Fig-  
 720 ure 10f). The contribution of this effect is however relatively small: alone, it would in-  
 721 crease the d-excess by less than 5‰ in the eyewall, bands and in-between bands, rela-  
 722 tive to the environment (yellow in Figure 10f).



**Figure 10.** Results for the simple model of the SCL water vapor budget applied to the cyclone simulation. (a) Fraction of the domain, in log scale, covered by the five sub-domains of the cyclone: eye, eyewall (“wall”), rain bands (“bands”), in-between bands (“between”) and the environment (“env”). (b) Water fluxes contributing to the water budgets of the different sub-domains of the cyclone: Surface evaporation (red), updrafts (green), downdrafts (blue), rain evaporation (pink), condensation (cyan) and residual term associated with horizontal advection and non-stationary effects (dark green). (c)  $\delta D_v$  simulated by SAM (red), by the simple model (green) and by the MJ79 closure (blue bars). (d) Decomposition of the  $\delta D_v$  anomaly in different sub-domains relative to the environment into 6 contributions: near-surface relative-humidity (red), surface wind speed (yellow), vertical mixing through the SCL top (green), rain evaporation within the SCL (purple), condensation within the SCL (cyan) and horizontal advection and non-stationary effects (dark green). The total is shown in black and non-linear effects are shown in blue. (e-f) Same as (c-d) but for  $d_v$ .

723 **4.4 Insights of the simple box model for the squall line**

724 **4.4.1 SCL water vapor budget**

725 The convective, stratiform and trailing regions of the squall line cover about 15%  
 726 of the domain (Figure 11a). Surface evaporation is the main source of water in the SCL  
 727 and is approximately uniform in all sub-domains (red in Figure 11b). The rain evapo-  
 728 ration is a significant source in both the convective and stratiform parts (purple). In the  
 729 convective part, the main sink of water is the export of moist air through updrafts (green  
 730 in Figure 11b), consistent with the vigorous updrafts. Horizontal advection and non-stationary  
 731 effects moisten the convective zone by advecting air from the stratiform region moist-  
 732 ened by rain evaporation, and dries the trailing region by advecting drier air from the  
 733 environment (dark green in Figure 11b).

734 **4.4.2 Evaluating and decomposing the water vapor  $\delta D_v$  in each sub-domain**

735 The box model performs better for the squall line than for the cyclone simulation.  
 736 It captures well the relative variations across the sub-domains, in particular the max-  
 737 imum depletion in the stratiform region and the depletion in the convective and trail-  
 738 ing regions relative to the environment (Figure 11c). We will thus focus on understand-  
 739 ing these differences.

740 In both the convective and stratiform parts of the squall line, the depletion rela-  
 741 tive to the environment is mainly due to the rain evaporation and rain-vapor diffusive  
 742 exchanges in the SCL that deplete the water vapor (purple in Figure 11d). Alone, it would  
 743 deplete the convective and stratiform regions by about 30 and 45 % respectively, far ex-  
 744 ceeding the total difference relative to the environment. In the stratiform region, ver-  
 745 tical mixing, which likely includes the effect of rain evaporation above the SCL, also sig-  
 746 nificantly contributes to the depletion (green in Figure 11d). This is consistent with the  
 747 depleting effect of rain evaporation both in the SCL and near the melting level (Figure  
 748 8d), and with the major effect of fractionation during rain evaporation on the isotopic  
 749 evolution in squall lines (section 4.2).

750 In the trailing region, where the precipitation is weak, the depletion in heavy iso-  
 751 topes of the vapor is explained by horizontal advection and non-stationary effects (dark  
 752 green in Figure 11d). Alone, it would contribute to a 55% depletion relative to the en-  
 753 vironment. This reflects the effect of horizontal advection spreading the depleted water  
 754 vapor from the stratiform region. In contrast, the horizontal advection explains why the  
 755 convective region is less depleted than the stratiform region: horizontal advection brings  
 756 enriched water vapor from the environment towards the squall line front.

757 To summarize, the  $\delta D_v$  differences between the sub-domains are mainly explained  
 758 by rain evaporation and rain-vapor diffusive exchanges inside the SCL, and also prob-  
 759 ably above the SCL. This is consistent with previous studies (Risi et al., 2010; Tremoy  
 760 et al., 2014). Horizontal advection then plays a key role in spreading the isotopic anomaly  
 761 rearward.

762 **4.4.3 Evaluating and decomposing the water vapor  $d_v$ -excess in each sub-**  
 763 **domain**

764 The simple box model is able to capture the fact that  $d_v$  is much larger than would  
 765 be predicted by the MJ79 closure. It also captures the higher  $d_v$  in the stratiform region  
 766 relative to the convective and trailing region.

767 The higher  $d_v$  in the stratiform region is mainly explained by the vertical mixing  
 768 (green in Figure 11f). Vertical mixing probably brings water vapor with high  $d_v$  due to  
 769 rain evaporation above the SCL. Rain evaporation in the SCL strongly increases  $d_v$  in

770 the convective region (purple in Figure 11f), but this is compensated by the advection  
 771 of vapor (dark green in Figure 11d). Since the environment has a lower  $d_v$  than the con-  
 772 vective region (red in Figure 11c), the advection of air from the environment to the con-  
 773 vective region acts to deplete the convective region. Variations in near-surface relative  
 774 humidity and winds are smaller than in the cyclone simulations, so the contributions of  
 775 these effects are marginal.

#### 776 4.5 Discussion

777 We find many common aspects for the mesoscale isotopic variability between cy-  
 778 clone and squall line simulations.

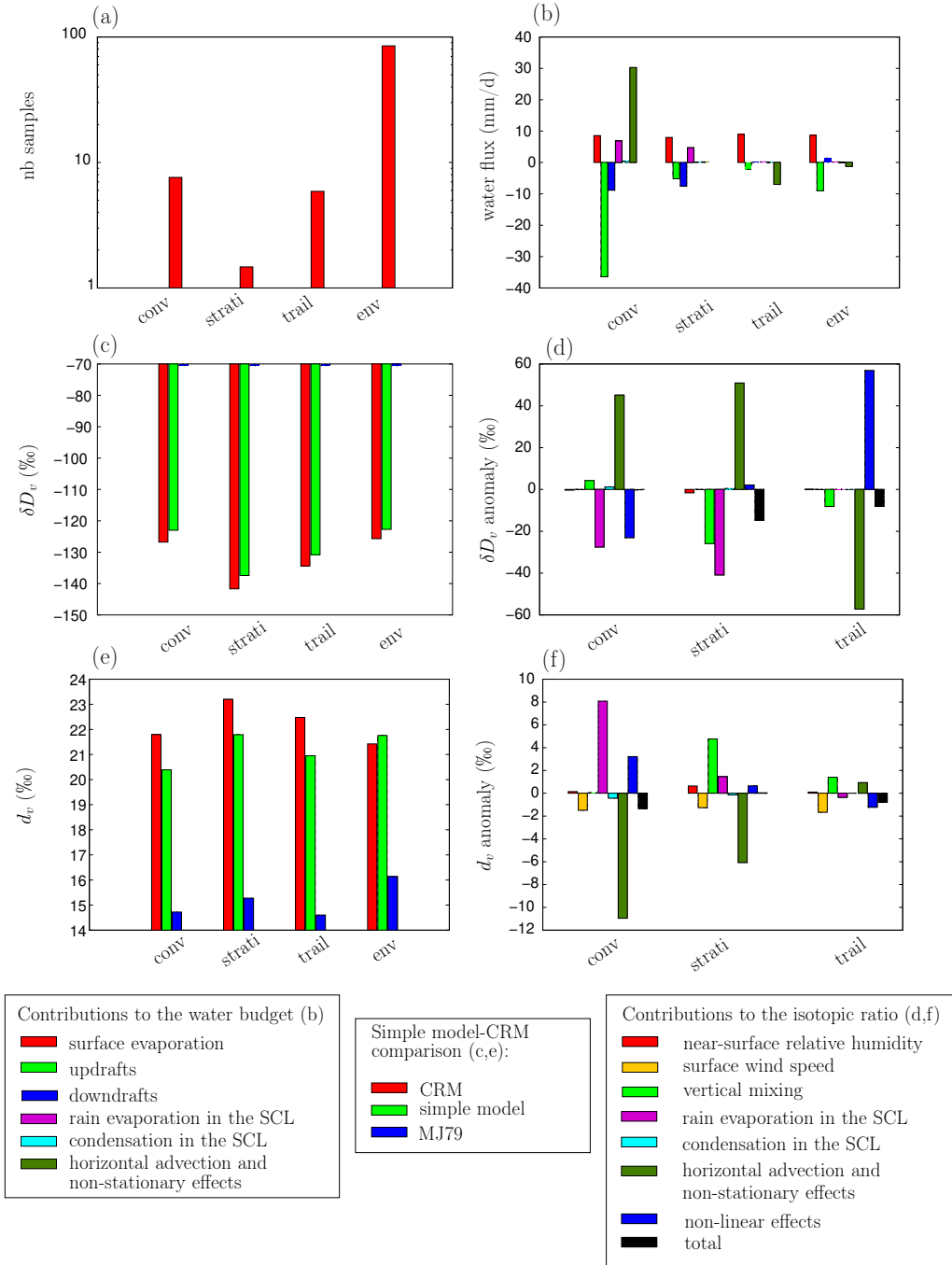
- 779 • In both convective systems, rain evaporation and rain-vapor diffusive exchanges  
 780 both within and above the SCL (especially near the melting level) are the main  
 781 drivers of the depletion in heavy isotopes of the vapor in the eyewall and rain bands  
 782 for the cyclone, and in the convective and stratiform parts of the squall line. This  
 783 is consistent with previous studies (Gedzelman et al., 2003; Tremoy et al., 2014;  
 784 Xu et al., 2019). These processes have also a crucial impact on d-excess.
- 785 • In both cases, horizontal advection and non-stationary effects act to smooth the  
 786 isotopic patterns. It leads to the gradual depletion towards the eyewall observed  
 787 in tropical cyclones, and it spreads the isotopic anomalies rearward in the case of  
 788 the squall line. This is consistent with previous studies (Xu et al., 2019).
- 789 • In both cases, the effect of condensation on the SCL water vapor is indirect. Con-  
 790 densation in the free troposphere maintains the vertical gradient in  $\delta D_v$  that al-  
 791 lows the rain evaporation to have a depleting effect on the SCL water vapor. Through  
 792 this indirect effect, condensation and precipitation are necessary for the depletion  
 793 of the water vapor SCL (Yoshimura et al., 2003).

794 The main difference between the two convective systems is in the effect of kinetic effects  
 795 during surface evaporation. Strong winds and moist conditions in tropical cyclones sig-  
 796 nificantly impact the d-excess, whereas they have a marginal effect in squall lines.

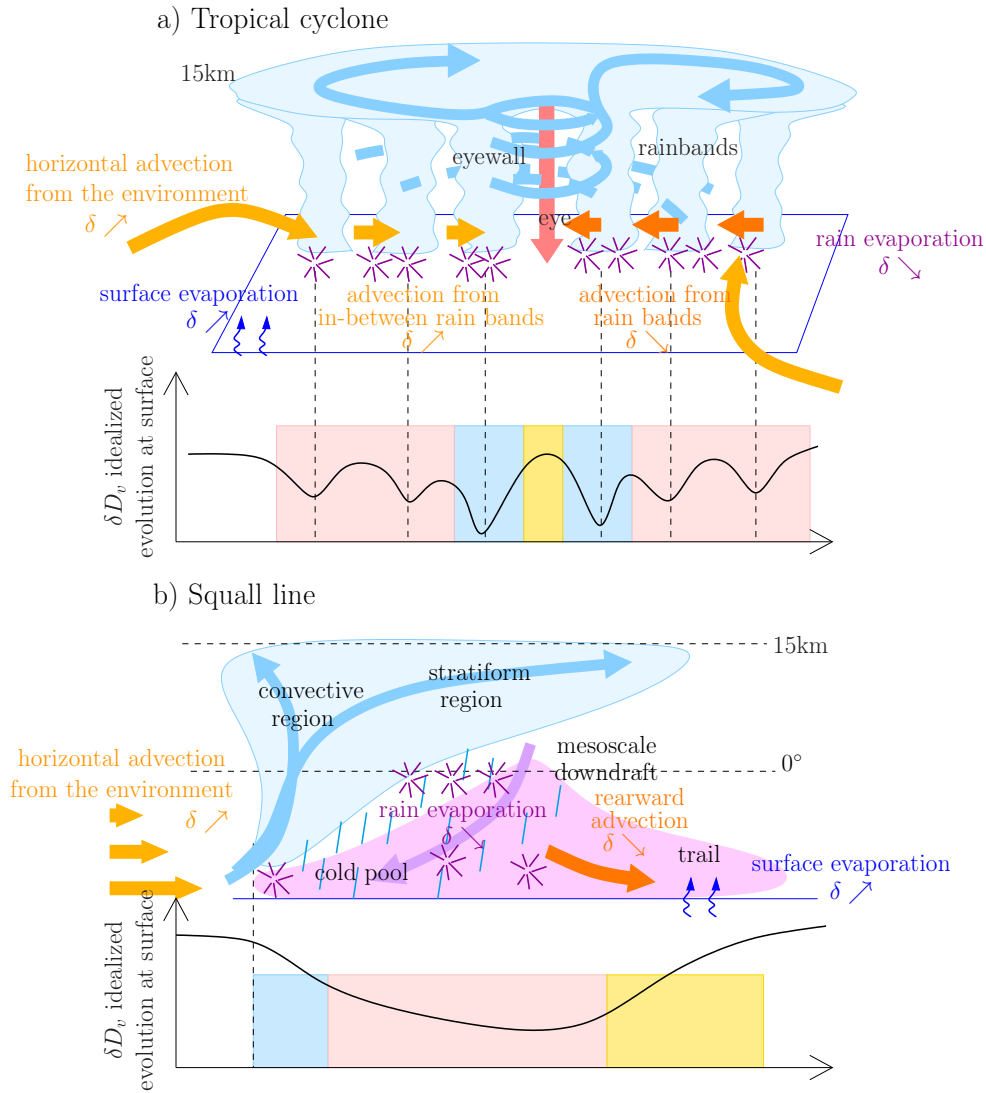
797 Some similarities can also be found between processes controlling the isotopic com-  
 798 position in tropical cyclones and those in extra-tropical cyclones and cold fronts, as doc-  
 799 umented by previous studies (Pfahl et al., 2012; Aemisegger et al., 2015; Thurnherr &  
 800 Aemisegger, 2022). Isotopically depleted water vapor and rain were observed in extra-  
 801 tropical cyclones. This was explained by the interplay between air mass advection and  
 802 depletion by rain-vapor interaction, consistent with our study, although in extratropi-  
 803 cal cyclones the large-scale advection and large-scale gradients in temperature also con-  
 804 tributed to depleting the vapor (Pfahl et al., 2012; Aemisegger et al., 2015). Low  $d_v$  was  
 805 also observed in the warm sector of extratropical cyclones. As in tropical cyclones, this  
 806 was explained by ocean evaporation in a moist environment, although in extratropical  
 807 cyclones dew formation also contributes to lowering  $d_v$  (Thurnherr & Aemisegger, 2022).

## 808 5 Conclusion

809 Using cloud resolving model simulations of cyclones and squall lines, and a sim-  
 810 ple model for the SCL water vapor budget, we investigate how convective processes im-  
 811 pact the isotopic composition of water vapor and precipitation at the mesoscale. We show  
 812 that the main factors depleting heavy isotopes of the water vapor at the mesoscale is rain  
 813 evaporation, especially in the sub-cloud layer of rain bands and of the eyewall in tropi-  
 814 cal cyclones, and in the mesoscale downdraft of the stratiform region in squall lines (pur-  
 815 ple in Figure 12). The mesoscale  $\delta D_v$  patterns are subsequently reshaped by horizon-  
 816 tal advection (orange in Figure 12). These mechanisms are overall consistent with those  
 817 suggested in previous studies (Gedzelman et al., 2003; Tremoy et al., 2014; Xu et al., 2019).



**Figure 11.** Results for the simple model of the SCL of the SCL water vapor budget applied to the squall line simulation. Same as Figure 10 but for the four sub-domains of the squall line: convective (“conv”), stratiform (“strati”) and trailing (“trail”) regions, and the environment (“env”). Note that in (c), the prediction by the MJ79 closure (blue) is nearly -70‰ in all sub-domains, so it is hard to see.



**Figure 12.** A schematic of the processes controlling the water vapor composition inside tropical cyclones (a) and squall lines (b). The key driver is rain evaporation, indicated by purple stars. Rain evaporation depletes water vapor in the rain bands and eyewall of tropical cyclones and in the convective and stratiform regions of squall lines. Horizontal advection then reshapes this pattern. Dark orange arrows indicate horizontal advection from depleted regions to less depleted regions, contributing to the spread of the depleted anomalies inward in cyclones and rearward in squall lines. Light orange arrows indicate horizontal advection from less depleted regions to more isotope depleted regions, partially compensating the depletion in most depleted regions.

818 In contrast to these previous studies however, we highlight that condensation has no di-  
819 rect impact and that the evaporation of sea spray is not necessary to explain the rela-  
820 tive enrichment in the cyclone eye.

821 This study contributes to our understanding of mesoscale isotopic variability. It  
822 provides physical arguments for the more depleted rain observed in tropical cyclones and  
823 squall lines relative to the rain in small-scale convection. Therefore, this study supports  
824 the interpretation of paleoclimate isotopic archives in tropical regions in terms of past  
825 cyclonic activity (Nott et al., 2007; Medina-Elizalde & Rohling, 2012; Baldini et al., 2016)  
826 or past frequency of large, long-lived, organized convective systems such as squall lines  
827 (Maupin et al., 2021).

828 However, when considering paleoclimate records at the annual scale or larger, the  
829 isotopic composition reflects an average over many convective systems of different organ-  
830 ization types. In our simulations, this is equivalent to the domain-mean  $\delta D$  in simula-  
831 tions of tropical cyclones or squall lines relative to the domain-mean  $\delta D$  in simulations  
832 of isolated cumulonimbi, rather than the  $\delta D$  in tropical cyclone or squall lines relative  
833 to their environment in a given simulation. This paper focuses on mesoscale isotopic vari-  
834 ations and does not discuss domain-mean values, because the realism of simulated mesoscale  
835 variations could be more easily assessed than the realism of domain-mean values. In par-  
836 ticular, we realized that the domain-mean  $\delta D$  in the precipitation or water vapor of our  
837 tropical cyclone simulation was more enriched in heavy isotopes than that in simulations  
838 of squall lines or even of isolated cumulonimbi (Risi et al., 2020). This is at odds with  
839 observations of depleted cyclonic rains in the tropics (Lawrence et al., 2004). This dis-  
840 crepancy may be due to limitations in the radiative-convective equilibrium configura-  
841 tion. In radiative-convective equilibrium, the cyclone maintains a strong subsidence in  
842 its environment, which favors unrealistically dry conditions that allows enriched water  
843 vapor in the SCL to accumulate. In reality, tropical cyclones propagate and are thus not  
844 in equilibrium with their environment. In addition, tropical cyclones may occur in spe-  
845 cific large-scale conditions. To account for the propagation of tropical cyclones and for  
846 the interplay with the large-scale circulation, more realistic simulations are necessary,  
847 e.g. using regional cloud-resolving models (de Vries et al., 2022). In parallel, the inter-  
848 play between large-scale circulation and convective organization deserves to be better  
849 documented in observations. It is possible that part of the observed depletion associated  
850 with tropical cyclones is mediated by large-scale conditions and domain-mean precip-  
851 itation. To rigorously assess the role of convective organization, we would need to com-  
852 pare isotopic observations for different kinds of convective organization but for the same  
853 precipitation rate and large-scale context, as is now done for humidity (Tobin et al., 2012).  
854 This will allow us to rigorously assess the realism of the domain-mean isotopic compo-  
855 sition in our simulations and will be the subject of a future study.

856 Finally, many other processes need to be investigated before drawing any paleo-  
857 climate conclusions from this study, including large-scale horizontal advection (Chen et  
858 al., 2021), land-atmosphere interactions along the air mass trajectories, infiltration pro-  
859 cesses, processes in the karstic systems and during calcite formation (Lases-Hernández  
860 et al., 2020). Our study is a first step towards a more comprehensive understanding of  
861 water isotopic variations.

## 862 Open Research Section

863 Information on SAM can be found on this web page:

864 <http://rossby.msrc.sunysb.edu/~marat/SAM.html>. All simulation outputs used  
865 in this article are archived in the PANGAEA data repository: [https://doi.pangaea.de/  
866 10.1594/PANGAEA.937534](https://doi.pangaea.de/10.1594/PANGAEA.937534).



867 **Acknowledgments**

868 This work was granted access to the HPC resources of TGCC under the allocation  
 869 2092 made by GENCI. CR, CM, and FV, acknowledge the national french program INSU-  
 870 LEFE for the funding of the SAM-iso proposal at the AO LEFE 2021. C.M. gratefully  
 871 acknowledges funding from the European Research Council (ERC) under the European  
 872 Union’s Horizon 2020 research and innovation programme (Project CLUSTER, grant  
 873 agreement No 805041). P.B. was supported by the National Science Foundation under  
 874 Grant No. AGS-1938108. G.V. contributed as part of his internship as a Master student  
 875 at Sorbonne Université, with a funding from Paris Sciences & Lettres “ANR-10-IDEX-  
 876 0001-02”. C. D. contributed as part of her internship as a Bachelor student at Sorbonne  
 877 Université. S.A. contributed as part of her PhD, with a funding from Ecole Normale Su-  
 878 périeure de Paris-Saclay. We thank all reviewers for their constructive comments.

879 **References**

- 880 Abramian, S., Muller, C., & Risi, C. (2022). Shear-convection interactions and ori-  
 881 entation of tropical squall lines. *Geophys. Res. Lett.*, *49*(1), e2021GL095184,  
 882 DOI: doi.org/10.1029/2021GL095184.
- 883 Aemisegger, F., Spiegel, J., Pfahl, S., Sodemann, H., Eugster, W., & Wernli, H.  
 884 (2015). Isotope meteorology of cold front passages: A case study combining  
 885 observations and modeling. *Geophys. Res. Lett.*, *42*(13), 5652–5660.
- 886 Aggarwal, P. K., Romatschke, U., Araguas-Araguas, L., Belachew, D., Longstaffe,  
 887 F. J., Berg, P., ... Funk, A. (2016). Proportions of convective and stratiform  
 888 precipitation revealed in water isotope ratios. *Nat. Geosci.*, *9*(8), 624-629,  
 889 <https://doi.org/10.1038/ngeo2739>.
- 890 Albright, A. L., Bony, S., Stevens, B., & Vogel, R. (2022). Observed subcloud layer  
 891 moisture and heat budgets in the trades. *J. Atmos. Sci.*
- 892 Attinger, R., Spreitzer, E., Boettcher, M., Forbes, R., Wernli, H., & Joos, H. (2019).  
 893 Quantifying the role of individual diabatic processes for the formation of pv  
 894 anomalies in a north pacific cyclone. *Quart. J. R. Meteor. soc.*, *145*(723),  
 895 2454–2476.
- 896 Baldini, L. M., Baldini, J. U., McElwaine, J. N., Frappier, A. B., Asmerom, Y., Liu,  
 897 K.-b., ... Breitenbach, S. F. M. (2016). Persistent northward north atlantic  
 898 tropical cyclone track migration over the past five centuries. *Sci. Rep.*, *6*,  
 899 37522, <https://doi.org/10.1038/srep37522>.
- 900 Bhattacharya, S. K., Sarkar, A., & Liang, M.-C. (2022). Vapor isotope prob-  
 901 ing of typhoons invading the taiwan region in 2016. *J. Geophys. Res.*,  
 902 e2022JD036578.
- 903 Biggerstaff, M. I., & Houze Jr, R. (1991). Kinematic and precipitation structure  
 904 of the 10–11 june 1985 squall line. *Mon. Wea. Rev.*, *119*(12), 3034–3065,  
 905 <https://doi.org/10.1175/1520-0493>.
- 906 Blossey, P. N., Kuang, Z., & Romps, D. M. (2010). Isotopic composition  
 907 of water in the tropical tropopause layer in cloud-resolving simulations  
 908 of an idealized tropical circulation. *J. Geophys. Res.*, *115*, D24309,  
 909 doi:10.1029/2010JD014554.
- 910 Bony, S., Risi, C., & Vimeux, F. (2008). Influence of convective processes on the  
 911 isotopic composition ( $\delta^{18}\text{O}$  and  $\delta^2\text{H}$ ) of precipitation and water vapor  
 912 in the Tropics. Part 1: Radiative-convective equilibrium and TOGA-COARE  
 913 simulations. *J. Geophys. Res.*, *113*, D19305, doi:10.1029/2008JD009942.
- 914 Bryan, G. H., & Morrison, H. (2012). Sensitivity of a simulated squall line to hor-  
 915 izontal resolution and parameterization of microphysics. *Mon. Weather Rev.*,  
 916 *140*(1), 202–225, <https://doi.org/10.1175/MWR-D-11-00046.1>.
- 917 Caniaux, G., Redelsperger, J.-L., & Lafore, J.-P. (1994). A numerical study  
 918 of the stratiform region of a fast moving squall line. part 1: general

- 919 description and water and heat budgets. *J. Atm. Sci.*, *51*, 2046-2074,  
920 <https://doi.org/10.1175/1520-0469>.
- 921 Chakraborty, S., Sinha, N., Chattopadhyay, R., Sengupta, S., Mohan, P., &  
922 Datye, A. (2016). Atmospheric controls on the precipitation iso-  
923 topes over the Andaman Islands, Bay of Bengal. *Sci. Rep.*, *6*(1), 1–11,  
924 <https://doi.org/10.1038/srep19555>.
- 925 Chavas, D. R., & Emanuel, K. (2014). Equilibrium tropical cyclone size in an ide-  
926 alized state of axisymmetric radiative–convective equilibrium. *J. Atmos. Sci.*,  
927 *71*(5), 1663–1680, <https://doi.org/10.1175/JAS-D-13-0155.1>.
- 928 Chen, F., Huang, C., Lao, Q., Zhang, S., Chen, C., Zhou, X., . . . Zhu, Q. (2021).  
929 Typhoon control of precipitation dual isotopes in southern china and its  
930 palaeoenvironmental implications. *J. Geophys. Res.: Atmospheres*, *126*(14),  
931 2020JD034336, <https://doi.org/10.1029/2020JD034336>.
- 932 Chong, M. (2010). The 11 August 2006 squall line system as observed from MIT  
933 Doppler radar during the AMMA SOP. *Quart. J. R. Meteor. soc.*, *136*(S1),  
934 <https://doi.org/10.1002/qj.466>.
- 935 Chong, M., & Hauser, D. (1990). A tropical squall line observed during the COPT81  
936 experiment in West Africa: Part III: heat and moisture budgets. *Mon. Wea.*  
937 *Rev.*, *118*, 1696-1706, <https://doi.org/10.1175/1520-0493>.
- 938 Conroy, J. L., Noone, D., Cobb, K. M., Moerman, J. W., & Konecky, B. L. (2016).  
939 Paired stable isotopologues in precipitation and vapor: A case study of the  
940 amount effect within western tropical pacific storms. *J. Geophys. Res.*, *121*(7),  
941 3290–3303, <https://doi.org/10.1002/2015JD023844>.
- 942 Craig, H. (1961). Isotopic variations in meteoric waters. *Science*, *133*, 1702-1703,  
943 DOI: 10.1126/science.133.3465.170.
- 944 Craig, H., & Gordon, L. I. (1965). Deuterium and oxygen-18 variations in the  
945 ocean and marine atmosphere. *Stable Isotope in Oceanographic Studies and*  
946 *Paleotemperatures, Laboratorio di Geologia Nucleate, Pisa, Italy*, 9-130.
- 947 Cruz, F. W., Vuille, M., Burns, S. J., Wang, X., Cheng, H., Werner, M., . . . Nguyen,  
948 H. (2009). Orbitally driven east-west antiphasing of South American precipita-  
949 tion. *Nat. Geosci.*, *2*, 210-214, <https://doi.org/10.1038/ngeo444>.
- 950 Dansgaard. (1964). Stable isotopes in precipitation. *Tellus*, *16*, 436-468,  
951 <https://doi.org/10.3402/tellusa.v16i4.8993>.
- 952 de Vries, A. J., Aemisegger, F., Pfahl, S., & Wernli, H. (2022). Stable water isotope  
953 signals in tropical ice clouds in the west african monsoon simulated with a re-  
954 gional convection-permitting model. *Atm. Chem. Phys.*, *22*, 8863-8895, DOI:  
955 <https://doi.org/10.5194/acp-22-8863-2022>.
- 956 Diekmann, C. J., Schneider, M., Knippertz, P., de Vries, A. J., Pfahl, S., Aemiseg-  
957 ger, F., . . . others (2021). A lagrangian perspective on stable water isotopes  
958 during the west african monsoon. *J. Geophys. Res.: Atmospheres*, *126*(19),  
959 e2021JD034895.
- 960 Dütsch, M., Pfahl, S., Meyer, M., & Wernli, H. (2018). Lagrangian process attri-  
961 bution of isotopic variations in near-surface water vapour in a 30-year regional  
962 climate simulation over europe. *Atm. Chem. Phys.*, *18*(3), 1653–1669.
- 963 Field, R. D., Jones, D. B. A., & Brown, D. P. (2010). The effects of post-  
964 condensation exchange on the isotopic composition of water in the atmosphere.  
965 *J. Geophys. Res.*, *115*, D24305, doi:10.1029/2010JD014334.
- 966 Field, R. D., Kim, D., LeGrande, A. N., Worden, J., Kelley, M., & Schmidt, G. A.  
967 (2014). Evaluating climate model performance in the tropics with retrievals  
968 of water isotopic composition from Aura TES. *Geophys. Res. Lett.*, DOI:  
969 10.1002/2014GL060572.
- 970 Frappier, A. B., Sahagian, D., Carpenter, S. J., González, L. A., & Frappier, B. R.  
971 (2007). Stalagmite stable isotope record of recent tropical cyclone events.  
972 *Geology*, *35*(2), 111–114, <https://doi.org/10.1130/G23145A.1>.
- 973 Fudeyasu, H., Ichiyanagi, K., Sugimoto, A., Yoshimura, K., Ueta, A., Yamanaka,

- 974 M. D., . . . Ozawa, K. (2008). Isotope ratios of precipitation and water  
 975 vapor observed in Typhoon Shanshan. *J. Geophys. Res.*, *113*, D12113,  
 976 doi:10.1029/2007JD009313.
- 977 Gamache, J. F., & Houze, R. A. (1981). Mesoscale air motions associ-  
 978 ated with a tropical squall line. *Mon. Weather Rev.*, *110*, 118-135,  
 979 https://doi.org/10.1175/1520-0493.
- 980 Gamache, J. F., & Houze, R. A. (1983). Water budget of a mesoscale  
 981 convective system in the tropics. *J. Atmos. Sci.*, *40*, 1835-1850, oi:  
 982 http://dx.doi.org/10.1175/1520-0469(1983)040.
- 983 Gao, J., Masson-Delmotte, V., Risi, C., He, Y., & Yao, T. (2013). What con-  
 984 trols southern Tibetan Plateau precipitation  $\delta^{18}O$  at seasonal and  
 985 intra-seasonal scales? A case study at Lhasa and Nyalam. *Tellus*, *65(1)*,  
 986 https://doi.org/10.3402/tellusb.v65i0.21043.
- 987 Gedzelman, S., Lawrence, J., Gamache, J., Black, M., Hindman, E., Black, R., . . .  
 988 Zhang, X. (2003). Probing hurricanes with stable isotopes of rain and water  
 989 vapor. *Mon. Wea. Rev.*, *131*, 112-1127, DOI: https://doi.org/10.1175/1520-  
 990 0493.
- 991 Gentry, M. S., & Lackmann, G. M. (2010). Sensitivity of simulated tropical cyclone  
 992 structure and intensity to horizontal resolution. *Mon. Wea. Rev.*, *138(3)*, 688–  
 993 704, DOI: https://doi.org/10.1175/2009MWR2976.1.
- 994 Godunov, S. K. (1959). Finite-difference methods for the numerical computations of  
 995 equations of gas dynamics. *Math. Sb.*, *7*, 271-290.
- 996 Guilpart, E. (2018). *Etude de la composition isotopique (deutérium et oxygène*  
 997 *18) de la vapeur d'eau dans l'atmosphère sur l'île de la réunion: apport à la*  
 998 *compréhension des processus humides atmosphériques en région tropicale.*  
 999 Unpublished doctoral dissertation, Université Paris-Saclay (ComUE).
- 1000 Houze, R. A. (1977). Structure and dynamics of a tropical squall line system. *Mon.*  
 1001 *Wea. Rev.*, *105*, 1540-1567.
- 1002 Houze, R. A. (2004). Mesoscale convective systems. *Rev. Geophys.*, *42(4)*, DOI:  
 1003 10.1029/2004RG000150.
- 1004 Houze, R. A. (2010). Clouds in tropical cyclones. *Mon. Wea. Rev.*, *138(2)*, 293–344,  
 1005 DOI: https://doi.org/10.1175/2009MWR2989.1.
- 1006 Jackisch, D., Yeo, B. X., Switzer, A. D., He, S., Cantarero, D. L. M., Siringan, F. P.,  
 1007 & Goodkin, N. F. (2022). Precipitation stable isotopic signatures of trop-  
 1008 ical cyclones in metropolitan manila, philippines, show significant negative  
 1009 isotopic excursions. *Nat. Hazards Earth Syst. Sci.*, *22(1)*, 213–226, DOI:  
 1010 https://doi.org/10.5194/nhess-22-213-2022.
- 1011 Jakob, C., Singh, M., & Jungandreas, L. (2019). Radiative convective equilibrium  
 1012 and organized convection: An observational perspective. *J. Geophys. Res.*,  
 1013 *124(10)*, 5418–5430, DOI: https://doi.org/10.1029/2018JD030092.
- 1014 Khairoutdinov, M., & Emanuel, K. (2013). Rotating radiative-convective equi-  
 1015 librium simulated by a cloud-resolving model. *J. Adv. Model. Earth Sci.*, *5(4)*,  
 1016 816–825, DOI: https://doi.org/10.1002/2013MS000253.
- 1017 Khairoutdinov, M., & Randall, D. A. (2003). Cloud resolving modeling of the ARM  
 1018 summer 1997 IOP: Model formulation, results, uncertainties, and sensitivities.  
 1019 *J. Atm. Sci.*, *60(4)*, 607–625, DOI: https://doi.org/10.1175/1520-0469.
- 1020 Kurita, N. (2013). Water isotopic variability in response to mesoscale convec-  
 1021 tive system over the tropical ocean. *J. Geophys. Res.*, *118(18)*, 10-376, DOI:  
 1022 https://doi.org/10.1002/jgrd.50754.
- 1023 Lacour, J.-L., Risi, C., Worden, J., Clerbaux, C., & Coheur, P.-F. (2017). Isotopic  
 1024 signature of convection's depth in water vapor as seen from iasi and tes d  
 1025 observations. *Earth Planet. Sci. Lett.*, *7*, 9645-9663, doi.org/10.5194/acp-17-  
 1026 9645-2017.
- 1027 Lasés-Hernández, F., Medina-Elizalde, M., & Frappier, A. B. (2020). Drip  
 1028 water  $\delta^{18}O$  variability in the northeastern Yucatán Peninsula, Mexico:

- 1029 Implications for tropical cyclone detection and rainfall reconstruction  
 1030 from speleothems. *Geochim. Cosmochim. Acta*, 285, 237–256, DOI:  
 1031 <https://doi.org/10.1016/j.gca.2020.07.008>.
- 1032 Lawrence, J. R. (1998). Isotopic spikes from tropical cyclones in surface waters:  
 1033 Opportunities in hydrology and paleoclimatology. *Chem. Geol.*, 144(1-2), 153–  
 1034 160.
- 1035 Lawrence, J. R., & Gedzelman, S. D. (1996). Low stable isotope ratios  
 1036 of tropical cyclone rains. *Geophys. Res. Lett.*, 23, 527–530, DOI:  
 1037 <https://doi.org/10.1029/96GL00425>. doi: 10.1029/96GL00425
- 1038 Lawrence, J. R., & Gedzelman, S. D. (2003). Tropical ice core isotopes: Do they  
 1039 reflect changes in storm activity? *Geophys. Res. Lett.*, 30(2), 44–1, DOI:  
 1040 <https://doi.org/10.1029/2002GL015906>.
- 1041 Lawrence, J. R., Gedzelman, S. D., Dexheimer, D., Cho, H.-K., Carrie, G. D.,  
 1042 Gasparini, R., ... Biggerstaff, M. I. (2004, March). Stable isotopic com-  
 1043 position of water vapor in the tropics. *J. Geophys. Res.*, 109, D06115,  
 1044 doi:10.1029/2003JD004046. doi: 10.1029/2003JD004046
- 1045 Lawrence, J. R., Gedzelman, S. D., Gamache, J., & Black, M. (2002). Sta-  
 1046 ble isotope ratios: Hurricane Olivia. *J. Atmos. Chem.*, 41, 67–82, DOI:  
 1047 <https://doi.org/10.1023/A:1013808530364>.
- 1048 Lekshmy, P., Midhun, M., Ramesh, R., & Jani, R. (2014).  $\delta^{18}\text{O}$  depletion in mon-  
 1049 soon rain relates to large scale organized convection rather than the amount of  
 1050 rainfall. *Scientific reports*, 4, 5661, DOI: <https://doi.org/10.1038/srep05661>.
- 1051 Maupin, C. R., Roark, E. B., Thirumalai, K., Shen, C.-C., Schumacher, C., Van  
 1052 Kampen-Lewis, S., ... others (2021). Abrupt southern great plains thunder-  
 1053 storm shifts linked to glacial climate variability. *Nat. Geosci.*, 14(6), 396–401,  
 1054 DOI: <https://doi.org/10.1038/s41561-021-00729-w>.
- 1055 Medina-Elizalde, M., & Rohling, E. J. (2012). Collapse of classic maya civiliza-  
 1056 tion related to modest reduction in precipitation. *Science*, 335(6071), 956–959,  
 1057 DOI: 10.1126/science.1216629.
- 1058 Merlivat, L., & Jouzel, J. (1979). Global climatic interpretation of the Deuterium-  
 1059 Oxygen 18 relationship for precipitation. *J. Geophys. Res.*, 84, 5029–5332,  
 1060 DOI: <https://doi.org/10.1029/JC084iC08p05029>.
- 1061 Miller, D. L., Mora, C. I., Grissino-Mayer, H. D., Mock, C. J., Uhle, M. E., & Sharp,  
 1062 Z. (2006). Tree-ring isotope records of tropical cyclone activity. *Proc. Natl.*  
 1063 *Acad. Sci. USA*, 103(39), 14294–14297.
- 1064 Moerman, J. W., Cobb, K. M., Adkins, J. F., Sodemann, H., Clark, B., & Tuen,  
 1065 A. A. (2013). Diurnal to interannual rainfall  $\delta^{18}\text{O}$  variations in northern Bor-  
 1066 neo driven by regional hydrology. *Earth Planet. Sci. Lett.*, 369, 108–119, DOI:  
 1067 <https://doi.org/10.1016/j.epsl.2013.03.014>.
- 1068 Moore, M., Blossey, P., Muhlbauer, A., & Kuang, Z. (2016). Microphysical controls  
 1069 on the isotopic composition of wintertime orographic precipitation. *J. Geo-*  
 1070 *phys. Res.*, 121(12), 7235–7253, DOI: <https://doi.org/10.1002/2015JD023763>.
- 1071 Moore, M., Kuang, Z., & Blossey, P. N. (2014). A moisture budget per-  
 1072 spective of the amount effect. *Geophys. Res. Lett.*, 41, 1329–1335,  
 1073 doi:10.1002/2013GL058302.
- 1074 Muller, C. (2013). Impact of convective organization on the response of tropi-  
 1075 cal precipitation extremes to warming. *J. Clim.*, 26(14), 5028–5043, DOI:  
 1076 <https://doi.org/10.1175/JCLI-D-12-00655.1>.
- 1077 Muller, C., & Romps, D. M. (2018). Acceleration of tropical cyclogenesis by self-  
 1078 aggregation feedbacks. *Proc. Natl. Acad. Sci.*, 115(12), 2930–2935, DOI:  
 1079 <https://doi.org/10.1073/pnas.1719967115>.
- 1080 Munksgaard, N. C., Zwart, C., Kurita, N., Bass, A., Nott, J., & Bird,  
 1081 M. I. (2015). Stable isotope anatomy of tropical cyclone ita,  
 1082 north-eastern Australia, april 2014. *PLoS one*, 10(3), e0119728,  
 1083 <https://doi.org/10.1371/journal.pone.0119728>.

- 1084 Neggers, R., Stevens, B., & Neelin, J. D. (2006). A simple equilibrium model for  
 1085 shallow-cumulus-topped mixed layers. *Theor. Comput. Fluid Dyn.*, *20*(5-6),  
 1086 305–322.
- 1087 Nott, J., Haig, J., Neil, H., & Gillieson, D. (2007). Greater frequency vari-  
 1088 ability of landfalling tropical cyclones at centennial compared to seasonal  
 1089 and decadal scales. *Earth Planet. Sci. Lett.*, *255*(3-4), 367–372, DOI:  
 1090 <https://doi.org/10.1016/j.epsl.2006.12.023>.
- 1091 Pfahl, S., Wernli, H., Yoshimura, K., & Dubey, M. (2012). The isotopic composition  
 1092 of precipitation from a winter storm—a case study with the limited-area model  
 1093 cosmo iso. *Atmos. Chem. Phys.*, *12*(3).
- 1094 Price, R. M., Swart, P. K., & Willoughby, H. E. (2008). Seasonal and  
 1095 spatial variation in the stable isotopic composition ( $\delta^{18}\text{O}$  and  $\delta\text{D}$ ) of  
 1096 precipitation in south Florida. *J. Hydrol.*, *358*(3-4), 193–205, DOI:  
 1097 <https://doi.org/10.1016/j.jhydrol.2008.06.003>.
- 1098 Ramos, R., LeGrande, A., Griffiths, M., Elsaesser, G., Litchmore, D., Tierney, J., ...  
 1099 Nusbaumer, J. (2022). Constraining clouds and convective parameterizations  
 1100 in a climate model using paleoclimate data. *J. Adv. Model. Earth Syst.*, *14*(8),  
 1101 e2021MS002893.
- 1102 Risi, C., Bony, S., Vimeux, F., Chong, M., & Descroix, L. (2010). Evolution  
 1103 of the water stable isotopic composition of the rain sampled along Sahe-  
 1104 lian squall lines. *Quart. J. Roy. Meteor. Soc.*, *136* (S1), 227 - 242, DOI:  
 1105 <https://doi.org/10.1002/qj.485>.
- 1106 Risi, C., Bony, S., Vimeux, F., Descroix, L., Ibrahim, B., Lebreton, E., ... Sultan,  
 1107 B. (2008). What controls the isotopic composition of the African monsoon pre-  
 1108 cipitation? Insights from event-based precipitation collected during the 2006  
 1109 AMMA campaign. *Geophys. Res. Lett.*, *35*, doi:10.1029/2008GL035920. doi:  
 1110 10.1029/2008GL035920
- 1111 Risi, C., Muller, C., & Blossey, P. (2021). Rain evaporation, snow melt,  
 1112 and entrainment at the heart of water vapor isotopic variations in the  
 1113 tropical troposphere, according to large-eddy simulations and a two-  
 1114 column model. *J. Adv. Model. Earth Sci.*, *13*(4), e2020MS002381, DOI:  
 1115 <https://doi.org/10.1029/2020MS002381>.
- 1116 Risi, C., Muller, C., & N, B. P. (2020). What controls the water vapor isotopic com-  
 1117 position near the surface of tropical oceans? Results from an analytical model  
 1118 constrained by large-eddy simulations. *J. Adv. Model. Earth Sci.*, *12*(8), DOI:  
 1119 <https://doi.org/10.1029/2020MS002106>.
- 1120 Robe, F. R., & Emanuel, K. A. (2001). The effect of vertical wind shear on  
 1121 radiative–convective equilibrium states. *J. Atm. Sci.*, *58*(11), 1427–1445,  
 1122 DOI: <https://doi.org/10.1175/1520-0469>.
- 1123 Rotunno, R., Klemp, J. B., & Weisman, M. L. (1988). A theory for  
 1124 strong, long-lived squall lines. *J. Atm. Sci.*, *45*(3), 463–485, DOI:  
 1125 <https://doi.org/10.1175/1520-0469>.
- 1126 Sanchez-Murillo, R., Durán-Quesada, A. M., Esquivel-Hernández, G., Rojas-  
 1127 Cantillano, D., Birkel, C., Welsh, K., ... others (2019). Deciphering key  
 1128 processes controlling rainfall isotopic variability during extreme tropical cy-  
 1129 clones. *Nat. Commun.*, *10*(1), 1–10, DOI: <https://doi.org/10.1038/s41467-019-12062-3>.
- 1130
- 1131 Sinha, N., & Chakraborty, S. (2020). Isotopic interaction and source moisture con-  
 1132 trol on the isotopic composition of rainfall over the Bay of Bengal. *Atmos.*  
 1133 *Res.*, *235*, 104760, DOI: <https://doi.org/10.1016/j.atmosres.2019.104760>.
- 1134 Skrzypek, G., Dogramaci, S., Page, G. F., Rouillard, A., & Grierson, P. F. (2019).  
 1135 Unique stable isotope signatures of large cyclonic events as a tracer of soil  
 1136 moisture dynamics in the semiarid subtropics. *J. Hydrol.*, *578*, 124124, DOI:  
 1137 <https://doi.org/10.1016/j.jhydrol.2019.124124>.
- 1138 Smolarkiewicz, P. K., & Grabowski, W. W. (1990). The multi-dimensional posi-

- 1139 tive definite advection transport algorithm: Non-oscillatory option. *J. Comput.*  
 1140 *Phys.*, *86*, 355-375, DOI: [https://doi.org/10.1016/0021-9991\(90\)90105-A](https://doi.org/10.1016/0021-9991(90)90105-A).
- 1141 Stevens, B. (2006). Bulk boundary-layer concepts for simplified models of tropical  
 1142 dynamics. *Theor. Comput. Fluid Dyn.*, *20*(5-6), 279-304.
- 1143 Su, H., Bretherton, C. S., & Chen, S. S. (2000). Self-aggregation and large-scale control  
 1144 of tropical deep convection: A modeling study. *J. Atm. Sci.*, *57*(11), 1797-  
 1145 1816, DOI: <https://doi.org/10.1175/1520-0469>.
- 1146 Sun, C., Tian, L., Shanahan, T. M., Partin, J. W., Gao, Y., Piatrunia, N., & Ban-  
 1147 ner, J. (2022). Isotopic variability in tropical cyclone precipitation is controlled  
 1148 by Rayleigh distillation and cloud microphysics. *Commun. Earth Environ.*,  
 1149 *3*(1), 1-10, DOI: <https://doi.org/10.1038/s43247-022-00381-1>.
- 1150 Tan, J., Jakob, C., & Lane, T. P. (2013). On the identification of the large-scale  
 1151 properties of tropical convection using cloud regimes. *J. Clim.*, *26*(17), 6618-  
 1152 6632, DOI: <https://doi.org/10.1175/JCLI-D-12-00624.1>.
- 1153 Taupin, J.-D., & Gallaire, R. (1998). Variabilité isotopique à l'échelle infra-  
 1154 événement de quelques épisodes pluvieux dans la région de Niamey, Niger.  
 1155 *Comptes Rendus de l'Académie des Sciences-Series IIA-Earth and Planetary*  
 1156 *Science*, *326*(7), 493-498.
- 1157 Thompson, G., Field, P. R., Rasmussen, R. M., & Hall, W. D. (2008). Explicit  
 1158 forecasts of winter precipitation using an improved bulk microphysics scheme.  
 1159 Part II: Implementation of a new snow parameterization. *Mon. Weather Rev.*,  
 1160 *136*(12), 5095-5115, DOI: <https://doi.org/10.1175/2008MWR2387.1>.
- 1161 Thurnherr, I., & Aemisegger, F. (2022). Disentangling the impact of air-sea inter-  
 1162 action and boundary layer cloud formation on stable water isotope signals in  
 1163 the warm sector of a southern ocean cyclone. *Atmos. Chem. Phys.*, *22*(15),  
 1164 10353-10373.
- 1165 Tobin, I., Bony, S., & Roca, R. (2012). Observational evidence for relation-  
 1166 ships between the degree of aggregation of deep convection, water va-  
 1167 por, surface fluxes and radiation. *J. Clim.*, *25*(20), 6885-6904, DOI:  
 1168 <https://doi.org/10.1175/JCLI-D-11-00258.1>.
- 1169 Torri, G. (2021). On the isotopic composition of cold pools in radiative-convective  
 1170 equilibrium. *J. Geophys. Res.*, *126*(10), e2020JD033139.
- 1171 Torri, G. (2022). Isotopic equilibration in convective downdrafts. *Geophys. Res.*  
 1172 *Lett.*, *49*(15), e2022GL098743.
- 1173 Torri, G., Ma, D., & Kuang, Z. (2017). Stable water isotopes and large-scale  
 1174 vertical motions in the tropics. *J. Geophys. Res.*, *122*(7), 3703-3717, DOI:  
 1175 <https://doi.org/10.1002/2016JD026154>.
- 1176 Tremoy, G., Vimeux, F., Mayaki, S., Souley, I., Cattani, O., Favreau, G., & Oi, M.  
 1177 (2012). A 1-year long delta18O record of water vapor in Niamey (Niger) re-  
 1178 veals insightful atmospheric processes at different timescales. *Geophys. Res.*  
 1179 *Lett.*, *39*(8), DOI: <https://doi.org/10.1029/2012GL051298>.
- 1180 Tremoy, G., Vimeux, F., Soumana, S., Souley, I., Risi, C., Cattani, O., ... Oi, M.  
 1181 (2014). Clustering mesoscale convective systems with laser-based water vapor  
 1182 delta18O monitoring in Niamey (Niger). *J. Geophys. Res.*, *119*(9), 5079-5103,  
 1183 DOI: 10.1002/2013JD020968.
- 1184 Vimeux, F., Gallaire, R., Bony, S., Hoffmann, G., & Chiang, J. C. H. (2005,  
 1185 December). What are the climate controls on deltaD in precipita-  
 1186 tion in the Zongo Valley (Bolivia)? Implications for the Illimani ice  
 1187 core interpretation. *Earth Planet. Sci. Lett.*, *240*, 205-220: DOI:  
 1188 <https://doi.org/10.1016/j.epsl.2005.09.031>. doi: 10.1016/j.epsl.2005.09.031
- 1189 Vimeux, F., Tremoy, G., Risi, C., & Gallaire, R. (2011). A strong control of the  
 1190 South American SeeSaw on the intraseasonal variability of the isotopic com-  
 1191 position of precipitation in the Bolivian Andes. *Earth. Planet. Sci. Lett.*, *307*  
 1192 *(1-2)*, 47-58, DOI: <https://doi.org/10.1016/j.epsl.2011.04.031>.
- 1193 Wang, Y. J., Cheng, H., Edwards, R. L., An, Z. S., Wu, J. Y., Shen, C. C., & Do-

- 1194           rale, J. A. (2001). A high-resolution absolute-dated late Pleistocene Monsoon  
 1195           record from Hulu Cave, China. *Science*, *294*(5550), 2345-8, DOI: 10.1126/sci-  
 1196           ence.106461.
- 1197           Welsh, K., & Sánchez-Murillo, R. (2020). Rainfall, groundwater, and surface  
 1198           water isotope data from extreme tropical cyclones (2016-2019) within the  
 1199           caribbean sea and atlantic ocean basins. *Data in Brief*, *30*, 105633, DOI:  
 1200           https://doi.org/10.1016/j.dib.2020.105633.
- 1201           Worden, J., Noone, D., & Bowman, K. (2007). Importance of rain evaporation and  
 1202           continental convection in the tropical water cycle. *Nature*, *445*, 528-532, DOI;  
 1203           https://doi.org/10.1038/nature05508.
- 1204           Xu, T., Sun, X., Hong, H., Wang, X., Cui, M., Lei, G., . . . Jiang, X. (2019). Stable  
 1205           isotope ratios of typhoon rains in Fuzhou, Southeast China, during 2013–2017.  
 1206           *J. Hydrol.*, *570*, 445–453, DOI: https://doi.org/10.1016/j.jhydrol.2019.01.017.
- 1207           Yang, M.-H., & Houze Jr, R. A. (1995). Sensitivity of squall-line rear inflow to  
 1208           ice microphysics and environmental humidity. *Mon. Wea. Rev.*, *123*(11), 3175–  
 1209           3193, DOI: https://doi.org/10.1175/1520-0493.
- 1210           Yoshimura, K., Oki, T., Ohte, N., & Kanae, S. (2003). A quantitative analysis of  
 1211           short-term O18 variability with a Rayleigh-type isotope circulation model. *J.*  
 1212           *Geophys. Res.*, *108*, D20, 4647, doi:10.1029/2003JD003477.
- 1213           Zipser, E. (1977). Mesoscale and convective scale downdrafts as distinct com-  
 1214           ponents of squall-line structure. *Mon. Wea. Rev.*, *105*, 1568-1589, DOI:  
 1215           https://doi.org/10.1175/1520-0493.