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Les isotopes stables de l'eau : application pour évaluer et améliorer les modèles de climat

Water stable stable isotopes: application to evaluate and improve climate models

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Résumé

Les projections de changement climatique diffèrent selon les modèles de climat utilisés, notamment concernant l'amplitude du réchauffement à venir et les changements de précipitation associés. La représentation des processus convectifs et nuageux est une source majeure d'incertitude. L'enjeu de mes recherches est donc de mieux évaluer la représentation de ces processus dans les modèles de climat et la crédibilité de leurs projections.

Dans ce cadre, j'essaye d'exploiter les mesures de composition isotopique de l'eau. L'eau se présente en effet sous différentes formes isotopiques ($H_2^{16}O$, $H_2^{18}O$, HDO ou $H_2^{17}O$) dont l'abondance est affectée par les changements de phase au cours du cycle de l'eau. Si l'apport des mesures isotopiques en paléo-climatologie est indéniable, leur apport pour mieux comprendre les processus atmosphériques dans le climat présent, et mieux les représenter dans les modèles de climat, est plus difficile à démontrer. Depuis plusieurs années, une révolution technologique a eu lieu permettant de mesurer la composition isotopique de la vapeur d'eau au sol et par satellite avec une fréquence, une couverture spatio-temporelle et une résolution verticale sans précédent. Dans le même temps, la modélisation isotopique s'est développée et diversifiée. Mes travaux reposent sur ces nouvelles opportunités.

Je discute d'abord des processus contrôlant la composition isotopique de la vapeur d'eau dans la troposphère tropicale. Les processus convectifs et nuageux jouent un rôle crucial. En particulier, les descentes convectives et la ré-évaporation de la pluie appauvrissent la vapeur d'eau dans la basse troposphère, tandis que le détrainement convectif enrichit la vapeur dans la haute troposphère. Dans la moyenne troposphère, l'un ou l'autre de ces effets domine selon la profondeur de la convection. Enfin, sur les continents, le recyclage de la pluie par évapo-transpiration joue un rôle significatif, mais il est difficile à démêler du rôle de la convection.

Forte de cette compréhension, je propose ensuite quelques exemples d'utilisation des mesures isotopiques dans la vapeur d'eau pour évaluer et améliorer la représentation des processus atmosphériques dans les modèles de climat. En étudiant la saisonnalité de la composition isotopique de la vapeur d'eau dans la troposphère subtropicale, je montre que dans la plupart des modèles de climat, le biais humide dans la haute troposphère tropicale et subtropicale est causé par une advection verticale trop diffusive. Je montre aussi que les modèles sont incapables de reproduire la distribution latitudinale de la composition isotopique de la vapeur d'eau dans la haute troposphère, à moins de considérer que l'efficacité de précipitation dépend de l'intensité de la convection. Enfin, les mesures isotopiques dans la vapeur d'eau pourraient permettre de mieux contraindre les profils verticaux de mélange convectif.

Enfin, je discute l'utilisation des archives paléo-climatiques d'isotopes de l'eau dans les tropiques pour évaluer la capacité des modèles à simuler les changements de précipitation en réponse aux variations climatiques. Je montre qu'on peut contraindre les changements de précipitation futurs en connaissant les changements de précipitations passé, par exemple en Amérique du Sud. Dans les tropiques, l'interprétation des archives isotopiques est débattue. Mes simulations confirment la difficulté du débat, montrant que le signal isotopique reflète la température ou les précipitations selon les régions, la période paléo-climatique et la version du modèle.

Summary

Climate change projections depend on climate models, in particular regarding the magnitude of future warming and associated precipitation changes. The representation of convective and cloud processes is a major source of uncertainty. The overall goal of my research is to better evaluate climate models and the credibility of their projections.

In this context, I try to exploit water isotopic composition measurements. The water molecule has several isotopic forms ($H_2^{16}O$, $H_2^{18}O$, HDO or $H_2^{17}O$) whose concentrations are affected by phase changes along the water cycle. While the contribution of isotopic measurements to paleoclimatology has been crucial, it is more difficult to demonstrate the added value of water isotopes their to better understand atmospheric processes in the present climate, and better evaluate their representation in climate models. For several years, a technological revolution has led to isotopic measurements in water vapor from the ground and by satellite with unprecedented frequency, spatio-temporal coverage and vertical resolution. In the meanwhile, isotopic modeling has considerably grown. My work is based on these new opportunities.

First, I will discuss the processes controlling the isotopic composition of water vapor in the tropical troposphere. Convective cloud processes and play a crucial role. In particular, convective descents and rain re-evaporation deplete the water vapor in the lower troposphere, while convective detrainment enriches water vapor in the upper troposphere. In the middle troposphere, one of these effects dominate depending on the convective depth. Finally, over land, rain recycling by evapotranspiration plays a significant role, but it is difficult to disentangle it from the role of convection.

With this understanding, I present some examples in which water vapor isotopic measurements are used to evaluate the representation of atmospheric processes in climate models. Studying the seasonality of water vapor isotopic composition in the subtropical troposphere, I show that in most climate models, the wet bias in the tropical and subtropical upper troposphere is caused by an excessively diffusive vertical advection. I also show that models are unable to reproduce the latitudinal distribution of the water vapor isotopic composition in the upper troposphere, unless they consider that precipitation efficiency depends on convective intensity. Finally, isotopic measurements may help better constrain the profiles of convective mixing.

Finally, I discuss using paleo-climatic water isotope records in the tropics to assess the ability of models to simulate rainfall changes in response to climate variations. I show that past precipitation changes could help constrain future precipitation changes, for example in South America. In the tropics, the intepretation isotopic records is debated. My simulations confirm the difficulty of the debate, showing that the isotopic signal reflects temperature or precipitation depending on the region, paleo-climatic period and model version.

Foreword: justification of some choices on the subject of this document

To write a coherent and concise document, I have deliberately chosen to focus on the heart of my current activities. In particular, I leave aside many pieces of work that I have done through collaborations.

First, I chose to focus exclusively **in the tropics**. The cloud response in the tropics is at the heart of the inter-model spread for climate sensitivity ([Bony and Dufresne, 2005]). In addition, the models spread in terms of precipitation change projections is largest in the tropics ([Oueslati et al., 2016]). Besides, I have extensively studied the role of **convective processes** on the water isotopic composition ([Risi et al., 2008a], section 2.1). In the tropics, convective processes, whose representation in models remains a challenge, play a crucial role. Finally, the tropics are where the interpretation of isotopic paleo-climatic archives is the more debated (sections 1.2.2, 4.2).

Second, I chose to focus on **water isotopes in the vapor phase**. I leave aside a significant piece of my PhD work that was devoted to understanding the isotopic composition in the precipitation. I do not want this document to be a repetition of my PhD thesis. Furthermore, in the past few years, the emergence of water vapor isotopic measurements from ground and space (section 1.3.1) have dramatically modified the context of isotopic research. Since my post-doc, I have focused almost exclusively on these new measurements, and I will continue to focus on these in the future. I will discuss the interpretation of precipitation paleo-climatic archives (section 4.2), evidencing the importance of better understanding isotopic controls in the vapor phase.

Third, I decided to focus on the **evaluation and improvement of climate models**, rather than on the understanding of processes in nature. Evaluation and improvement of climate models is an important issue in itself, given the persistent spread in climate change projections in terms of climate sensitivity and precipitation changes (section 1.1). In addition, using water isotopic measurements to understand of processes in nature is very promising but challenging. On the road to achieve this challenging goal, I think that the evaluation and improvement of climate models is a necessary step (section 1.2.3).

Fourth, I chose to write mainly about my **recent work**. I do not feel like repeating once again what I have already written in my thesis or in articles and what I have explained many times in conferences. Rather, I want this document to be a springboard for my future work. This is why several sections of this document are based on unpublished work (sections 2.1.5, 2.1.6, 3.3 and 4.2). In appendix A, I give some ideas of subjects for future students, post-docs or myself.

Contents

Chapter 1

Introduction

1.1 Motivation: inter-model spread in climate change projections

For several decades, there has been a persistent spread in climate change projections simulated by climate models, in particular for the magnitude of future global warming and for the patterns of precipitation changes ([Meehl et al., 2007, Knutti and Sedláček, 2013], figure 1.1). So the overall goal of my research is to **better assess climate models and the credibility of their projections**. In this document, I summarize how I try to contribute to this goal using water stable isotopic measurements.

The representation of atmospheric convection and cloud processes is a major source of uncertainty. In particular, the magnitude of future global warming is strongly impacted by cloud feedbacks in tropical regions ([Bony et al., 2004, Bony and Dufresne, 2005, Sherwood et al., 2014]). Changes in precipitation are very sensitive to the representation of deep convection cloud processes ([Frierson, 2007, Kang et al., 2008, Frierson and Hwang, 2012, Hwang and Frierson, 2013]). Since the isotopic composition of water vapor is very sensitive to convective and cloud processes (chapter 2), I try to **use water stable isotopic measurements to evaluate the representation of convective and cloud processes by climate models** (chapter 3).

The climate response of a model to a given forcing could be evaluated using past climate changes ([Schmidt, 2010, Schmidt et al., 2014, Hargreaves et al., 2013, Hargreaves et al., 2012]). Water isotopic proxies are potentially useful proxies for past precipitation changes ([Pausata et al., 2011]), especially over tropical land regions where the projection spread is the largest ([Knutti and Sedláček, 2013, Oueslati et al., 2016]). Therefore, I try to understand **how precipitation changes are imprinted into isotopic archives and how these records could be used to assess our confidence in precipitation projections** (chapter 4).

1.2 Why use water isotopes?

1.2.1 What are water isotopes?

Water molecules exhibit different isotopic species ($H_2^{16}O$, $H_2^{18}O$, HDO or $H_2^{17}O$) that undergo fractionation during phase changes. In particular, the lightest water molecules ($H_2^{16}O$) preferentially evaporate, while the heaviest molecules preferentially concentrate in the most condensed phase (liquid or solid) ([Dansgaard, 1964]). As a consequence, the isotopic composition of water vapor or precipitation is affected by the history of phase changes throughout the water cycle (figure 1.2).

The relative isotopic abundance in a sample is quantified as an anomaly compared to a reference ocean water using the δ notation and expressed in ‰. For example, the relative abundance in HDO is quantified by δD:

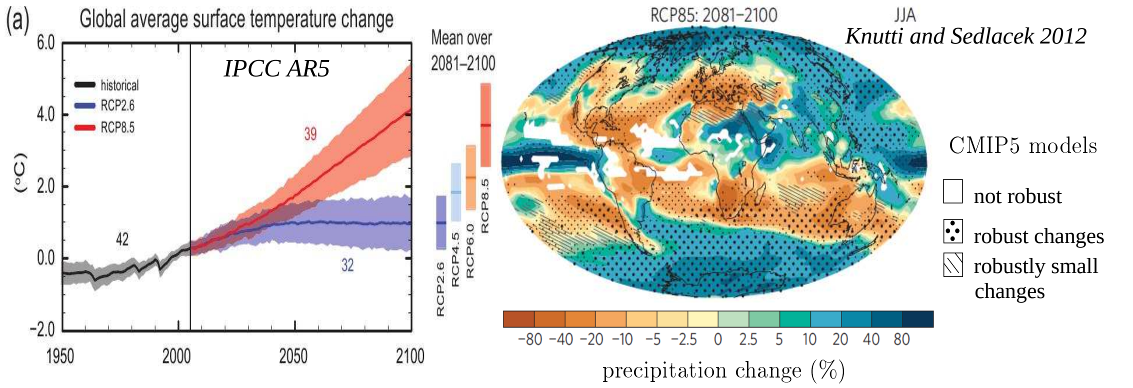

Figure 1.1: a) Global average surface temperature change for two different emission scenarios. The solid line shows the multi-model mean, whereas the envelope shows the range of individual model results. b) Precipitation change from present-day to the last two decade of the 21st century. Colors show the multi-model average, whereas stippling and Hatching illustrates the robustness of model results.

$$\delta D = \left(\left(\frac{HDO}{H_2O} \right)_{sample} / \left(\frac{HDO}{H_2O} \right)_{reference} - 1 \right) \cdot 1000$$

When δD is high, we say that the sample is enriched. Conversely, when δD is low, we say that the sample is depleted.

$\delta^{18}O$ is defined in the same way. To first order, $\delta^{18}O$ exhibit similar variations as δD, but 8 times smaller. The second order parameter d-excess, defined as $d = \delta D - 8 \cdot \delta^{18}O$, reflects the effect of non-equilibrium fractionation.

1.2.2 An important contribution to paleoclimate reconstructions

Water stable isotopes in precipitation can be stored in a variety of paleo-climatic archives. Since the isotopic composition of precipitation reflects the numerous phase change processes that occurred along the air mass trajectory, isotopic archives yield information on past climates. The contribution of water isotopic research to paleo-climatology is impressive. In polar regions, isotopic variations recorded in ice cores can be used as a proxy for temperature ([Jouzel et al., 2000]). They have allowed to document the glacial-interglacial cycles ([Dansgaard et al., 1969, Johnsen et al., 1972, Lorius et al., 1979, Jouzel et al., 1987a]) and to discover abrupt climate variations ([Dansgaard et al., 1989, Dansgaard et al., 1993]). In other regions, isotopic variations observed in speleothems ([Wang et al., 2001, Fleitmann et al., 2007]) or mountain ice cores ([Thompson et al., 1995, Thompson et al., 1998, Thompson et al., 2000]) have evidenced worldwide responses to these modes of variability.

Even though the precision of the isotopic thermometer in polar regions can be discussed ([Jouzel and Alley, 1997, Krinner and Werner, 2003, Sime et al., 2009]) and the interpretation of isotopic records in tropical regions remains debated ([Vuille et al., 2003], section 4.2), the value of water stable isotopes for paleoclimate reconstructions is well established.

1.2.3 More challenging applications to study present-day processes

A wide range of potential applications

More recently, water stable isotopic observations have been used in an attempt to better understand present-day hydrological and atmospheric processes. Since water isotopic composition is affected by so many processes, it has been used for many different applications (figure 1.3). Near the surface, the isotopic enrichment in water vapor has been used to try to estimate continental recycling ([Salati

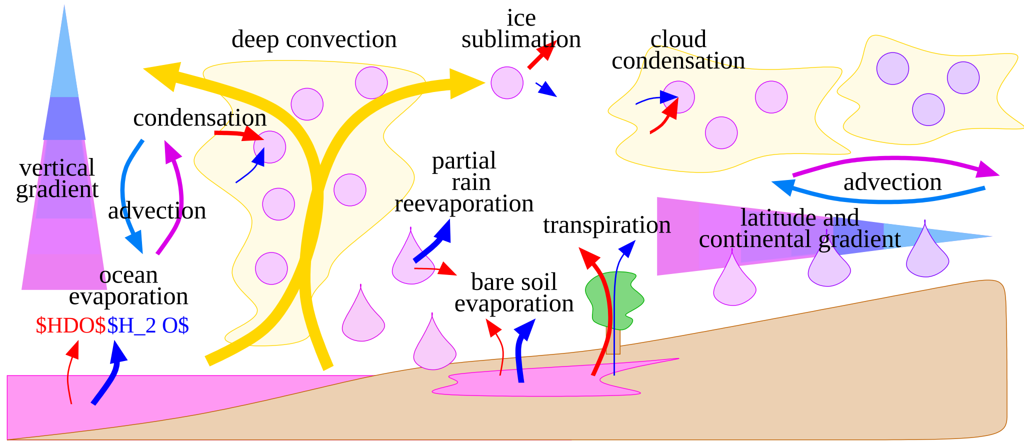

Figure 1.2: Overview of atmospheric processes affecting the isotopic composition of water vapor. Red and blue arrows illustrate the HDO and H_2O fluxes respectively during phase changes. Red-ish and blue-ish colors indicate more enriched and depleted colors respectively. HDO preferentially stays in the liquid phase during evaporation from ocean and bare soil. Transpiration does not fractionate. HDO preferentially condenses in clouds. This results in vertical, latitudinal and continental gradients. Advection downstream these gradients has a depleting effect, while advection against these gradients has an enriching effect. Sublimation of ice crystals does not fractionate and imprints the enriched signature of the condensate.

et al., 1979, Risi et al., 2013], section 2.2). A second order parameter called d-excess had been used to try to estimate the proportion of evapo-transpiration that occurs as evaporation or transpiration ([Gat and Matsui, 1991, Aemisegger et al., 2014]). In addition, the surface water vapor is impacted by convective processes, such as unsaturated downdrafts ([Risi et al., 2008a, Risi et al., 2010a], section 2.1.3), rain re-evaporation ([Worden et al., 2007]), or the degree of organization ([Lawrence et al., 2004]). In the middle troposphere, the water isotopic composition reflects the proportion of precipitation that occurs as convective or large-scale precipitation ([Lee et al., 2009, Kurita, 2013], section 2.1.6). In the sub-tropics, the water isotopic composition reflects vertical mixing ([Risi et al., 2012b], section 3.1.2). In the upper troposphere, the isotopic composition has been used to try to quantify convective detrainment ([Moyer et al., 1996, Webster and Heymsfield, 2003], section 2.1.2) and study ice microphysics ([Bolot et al., 2013]).

Water isotopic composition are affected by many processes allowing for many potential applications, but the flip side is that it is difficult to attribute isotopic variations to only one cause. Isotopic signals are very complex to interpret. In spite of many attempts to make quantitative estimates of the above-mentioned processes, most papers remain at a stage where they argue that water isotopes are *promising* or *have the potential* to solve problems. Going beyond this stage is a challenge.

Two examples of attempts to use water isotope measurements to quantify processes in nature, and their limitations

For example, clouds associated with deep convective detrainment and cirrus clouds associated with in-situ condensation have very distinct isotopic signature ([Webster and Heymsfield, 2003], figure 1.4, physical explanation in section 2.1.2). Based on the enriched signature of the isotopic composition observed in the tropical tropopause layer, [Kuang et al., 2003] concluded that convective detrainment contributes to the water budget at this altitude. But this conclusion was already known from non-isotopic measurements, such as cloud observations ([Gettelman et al., 2002, Gettelman et al., 2002]). The remaining question is: what is the quantitative contribution of convective detrainment to water budget? Water isotope measurements have not been able to provide any quantitative answer so far.

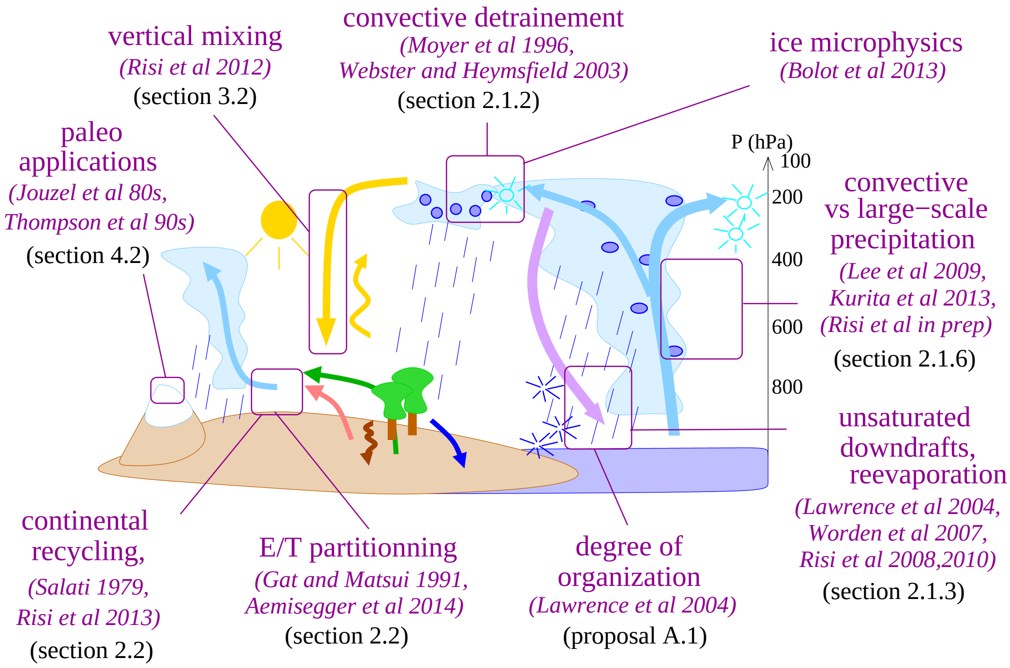

Figure 1.3: Different applications of water isotopic measurements. References and sections of this manuscript where these applications are discussed are indicated.

As another example, [Worden et al., 2007] proposed to use water vapor δD observations to quantify the proportion of tropospheric water vapor that orginates from rain-reevaporation. They developped a simple model for the effect of rain reevaporation on water vapor δD. Fitting their measurements with their model (figure 1.5), they endded up with values ranging from 0 to 50%. But what is the accuracy of these estimates, given the strong approximations in their model for rain reevaporation (proposal A.3), the numerous other atmospheric processes that were ignored (section 1.3.3) and the isotopic measurement uncertainties (section 1.3.1)? It is likely that the uncertainty range of such estimates is much larger than what we can obtain by other methods, such as water and energy budget considerations ([Folkins and Martin, 2005]) or model simulation of intensive field campaign cases ([Sud and Walker, 1993]).

1.2.4 Use water isotopes to evaluate and improve models: a necessary first step and a goal in itself

A major difficulty in trying to use water vapor isotopes to understand and quantify processes in nature is that they are sensitive to so many processes that comprehensive and complex models need to be used, such as climate models. For example, if a relationship can be found between the isotopic composition and some processes of interest among model simulations, then isotopic observations can discriminate the most realistic simulation, and processes can be analyzed and quantified in this simulation (figure 1.6). In this perspective, **evaluating and improving climate models using water isotopic measurements is a necessary step before we can actually use water isotopic measurements to understand and quantify processes in nature**.

In addition, evaluating and improving climate models is an important goal in itself (section 1.1) to which water isotopes can contribute.

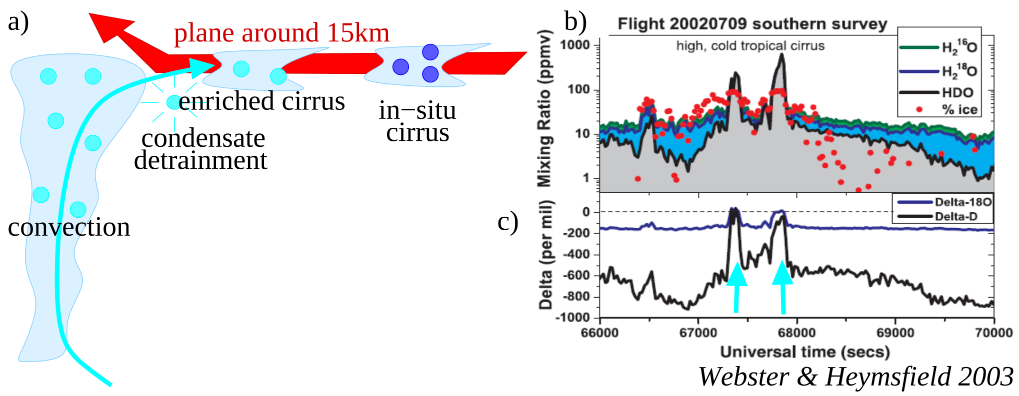

Figure 1.4: a) Measurements by plane in the upper troposphere. The plane crossed different kinds of cirrus clouds, some associated with convective detrainment and some associated with in-situ condensation. b) Humidity mixing ratio (green) along the plave path. c) Total-water δD (black) along the path. The two peaks in humidity and δD correspond to cirrus clouds originating from deep convection (light blue arrows).

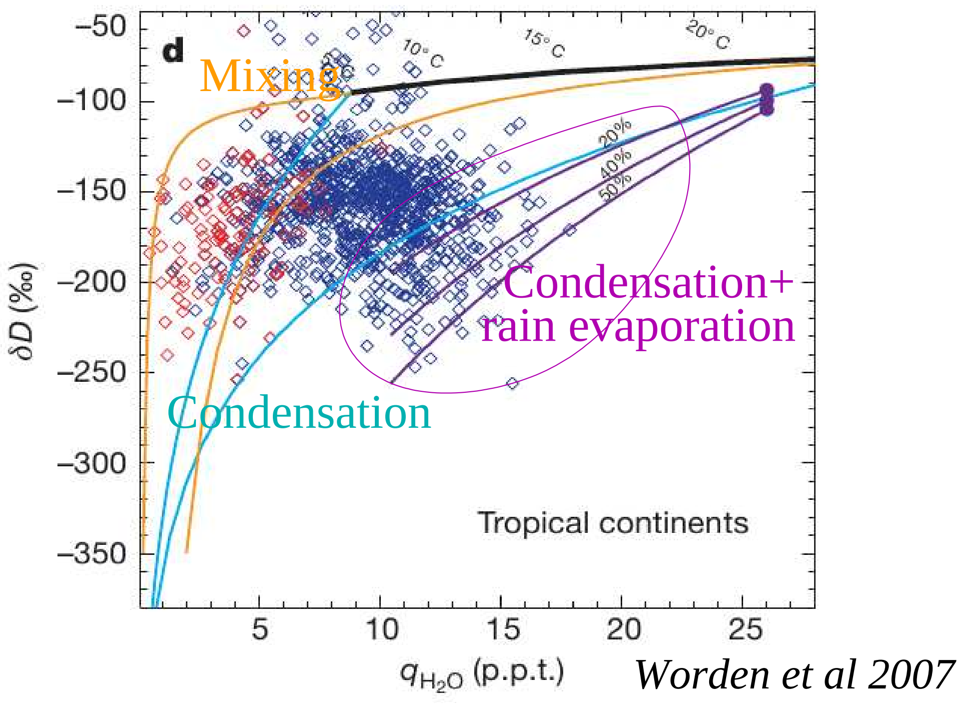

Figure 1.5: q-δD diagram (water vapor δD as a function of specific humidity q), as popularized by [Worden et al., 2007]. According to their simple model, the orange lines show the ensemble of q-δD values that can be explained by mixing processes, the blue lines show the ensemble of q-δD values that can be explained by condensation processes, and the purple lines show the ensemble of q-δD values that can be explained by condensation plus different contributions of rain reevaporation (from 20 to 50%). Markers illustrate the TES satellite observations in the lower middle troposphere over tropical continents for clear-sky (red) and cloudy (blue) scenes. This diagram was used to attempt to estimate the contribution of rain reevaporation to tropospheric humidity. Several cloudy scenes are consistent with the purple lines indicating between 20% and 50% of the vapor originating from rain reevaporation.

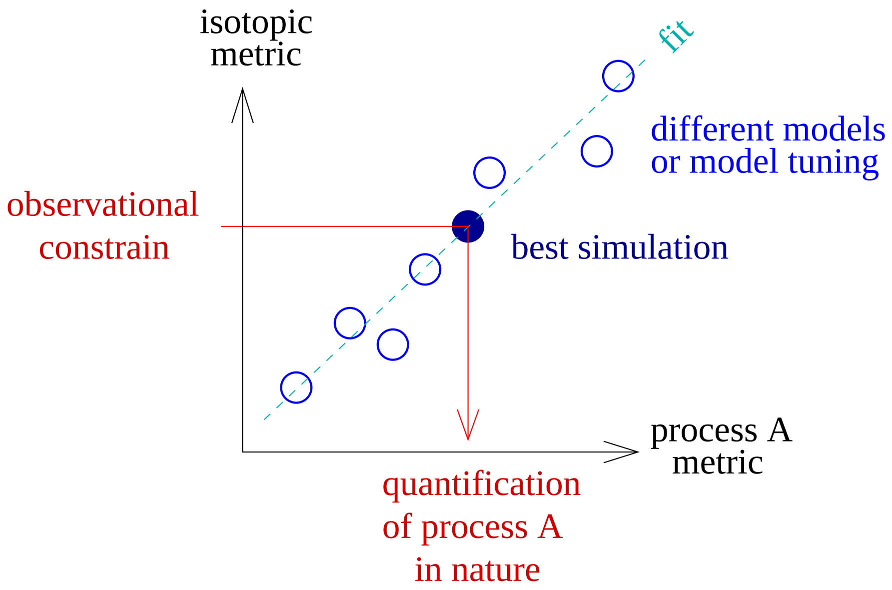

Figure 1.6: Schematic showing how combining isotopic measurements and models could help us quantify processes in nature. First, there should be a relationship among different models (or different model tuning) between a metric for process A and an isotopic metric. Isotopic observations can discriminate the most realistic simulation. Then process A can be estimated in this simulation, and we assume that this estimate is representative of what goes on in nature.

1.2.5 Personal opinion on trying to make isotopes useful

Sometimes I wonder why isotope scientists try so hard to make water isotopes useful. Ideally, research should **start with science questions** (figure 1.7, orange): for example, what is the contribution of different processes to the observed distribution of a meteorological variable? Do models represent it properly? What is the cause of a model bias for a given meteorological variable? Then, scientists wonder which observations are the most adequate to discriminate between different hypotheses. In most cases, non-isotopic observations (e.g. observations of conventional meteorological variables, cloud properties, flux tower measurements...) are the most suited to address the science questions. In some (rare) cases, the isotopic observations, combined with other observations, are the most suited. In this case, this motivates the implementation of field campaigns or satellite missions to perform isotopic measurements.

But in practice, research is often conducted the other way round (figure 1.7, pink). Isotopic measurements are performed first, motivated mainly by curiosity. Then we wonder: what can these measurements be useful for? To address this question, first we need to understand what processes control the isotopic composition displayed by the measurements. Once the main process is identified, we look for science questions associated with this process, and we try to answer it using the isotopic measurements. Although this approach looks like conducted the wrong way, accumulating understanding on processes controlling the isotopic composition is useful. It allows us to know whether a science question can be best answered using water isotopic composition or not.

Therefore, I think that at this stage of isotopic research, pretending to use water isotopes to better "evaluate and improve climate models", which I often do myself, is a bit premature. I think more and more that studies whose goal is **"simply" to understand isotopic controls**, without pretending to make any use of it, should receive more credit. This is why in this document I devote chapter 2 to "simply" understanding isotopic controls, before trying to use isotopic measurements to evaluate and improve models (chapter 3). In appendix I suggest several ideas to deepen our understanding isotopic controls (proposals A.2, A.3, A.4, A.5).

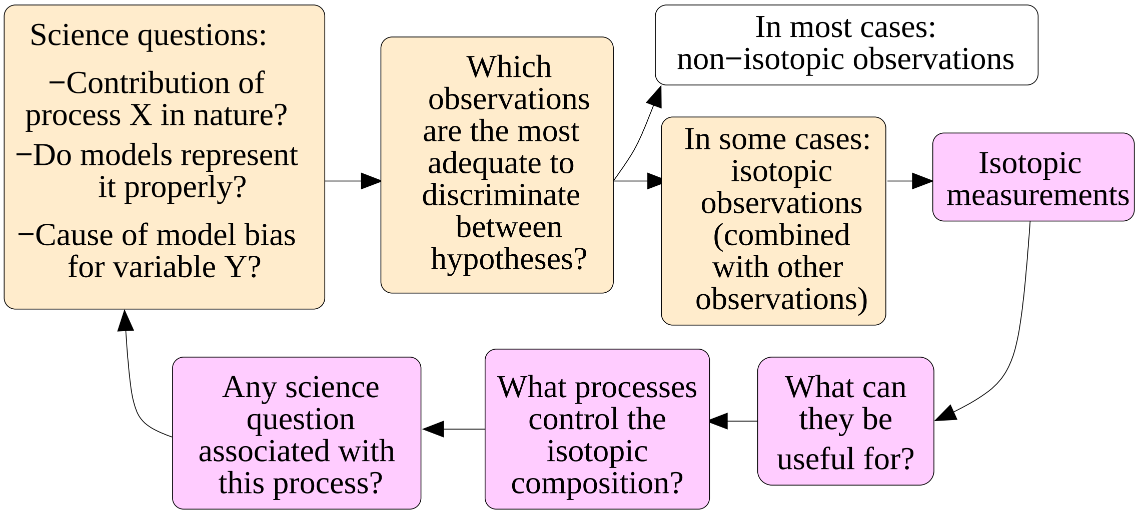

Figure 1.7: Schematics illustrating the scientific approach from science questions to isotopic measurements, and vice versa.

1.3 New opportunities and challenges in current water isotopic research

Nowadays is a good time to study isotopic controls and exploring the use of isotopic measurements to evaluate and improve models. The recent development of new measurement techniques (section 1.3.1) and new modelling tools (section 1.3.2) offers unprecedented opportunities.

1.3.1 Measurements

Why should we measure water isotopes is the water vapor phase?

Water stable isotopes are measured in various water reservoirs and fluxes of the Earth system, such as precipitation ([Dansgaard, 1964]), water vapor (next sections), rivers ([Kendall and Coplen, 2001]), lakes ([Gibson et al., 2005]), underground waters ([Fontes, 1980]), soil water ([Brunel et al., 1997]), stem water and leaf water ([Lai et al., 2006, Twining et al., 2006]), surface evaporation flux ([Griffis et al., 2010]), paleo-climatic proxies (section 4.2)... In this document, we focus mainly on **water vapor measurements**:

- We focus on atmospheric processes, which are most directly imprinted in water vapor Isotopic composition in land water reservoirs are affected by additional fractionation processes.
- The isotopic composition of water vapor is the only atmospheric *prognostic* variable, i.e. the isotopic composition in water vapor and all water reservoirs at time t depends on the isotopic composition of water vapor at time $t - dt$. In contrast, the isotopic composition of precipitation is purely *diagnostic*, i.e. at each time t it can be deduced from the that of water vapor. Understanding isotopic variations in water vapor is thus the first step before understanding those in all water reservoirs, including precipitation and derived paleo-climatic proxies.
- The isotopic composition of water vapor is defined continuously in time and in the three spatial dimensions every where in the atmosphere. In contrast, the isotopic composition of the precipitation, for example, is defined only during precipitation events.
- Studying the isotopic composition of water vapor has bloomed over the past decade, due to an instrumental revolution of water vapor isotopic measurements.

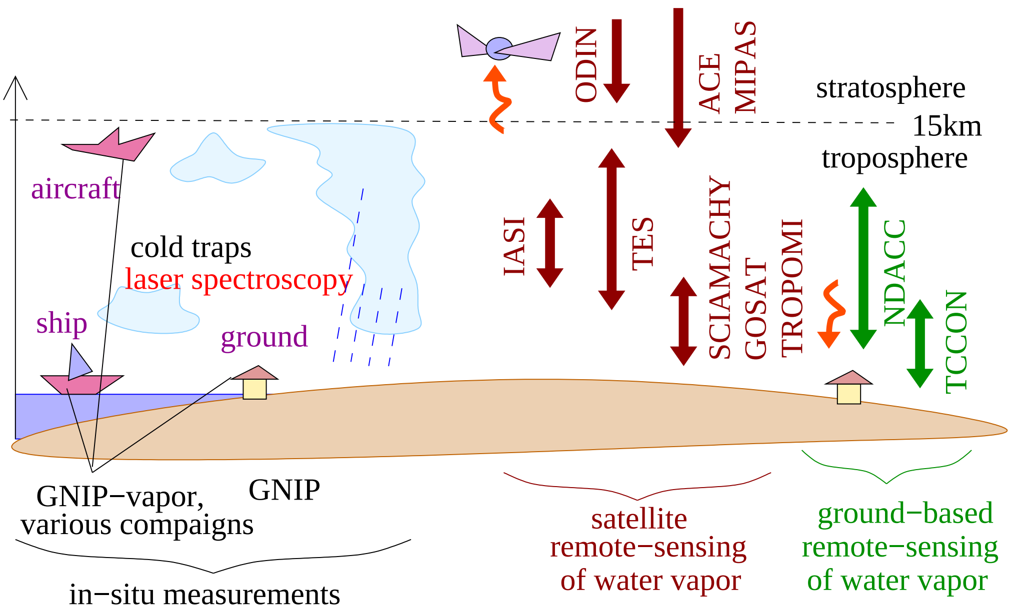

Figure 1.8: Overview of available isotopic measurements in water vapor and the altitude they sample.

Instrumental revolution of water vapor isotopic measurements

First isotopic measurements in water vapor were performed in-situ by cryogenic sampling followed by mass spectrometer analysis ([Ehhalt, 1974, Moreira et al., 1997, Lawrence et al., 2004, Uemura et al., 2008]). However, this method was long and tedious, so that temporal resolution and spatio-temporal coverage was weak. Therefore, most isotopic studies have long been limited to measurements in precipitation ([Dansgaard, 1964, Rozanski et al., 1993]).

In the past decade, an instrumental revolution has led to the development of two kinds of water vapor measurements: by in-situ laser instruments and by remote-sensing instruments.

Laser instruments

Laser instruments have been developed to perform in-situ isotopic measurements continuously and at a high frequency ([Gupta et al., 2009]), possibly during several years ([Tremoy et al., 2014, Steen-Larsen et al., 2014]). This allows to study isotopic variations in the near-surface vapor

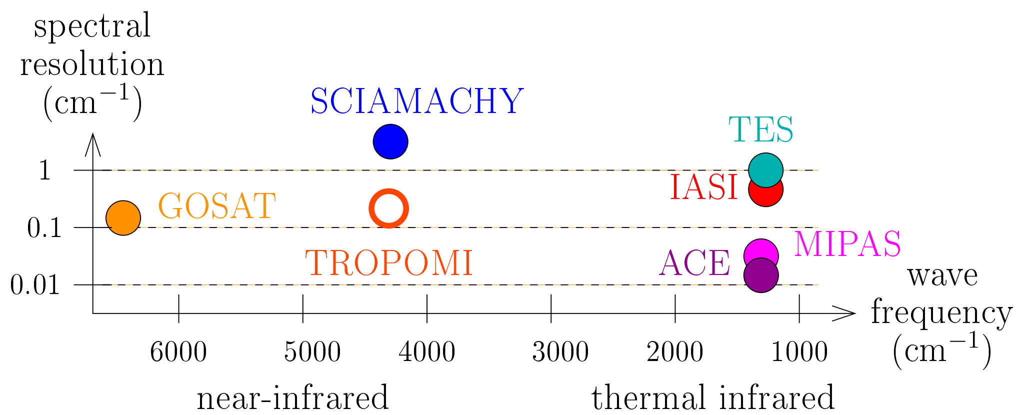

Figure 1.9: Overview of the spectral domain and resolution used for the water vapor isotopic satellite measurements.

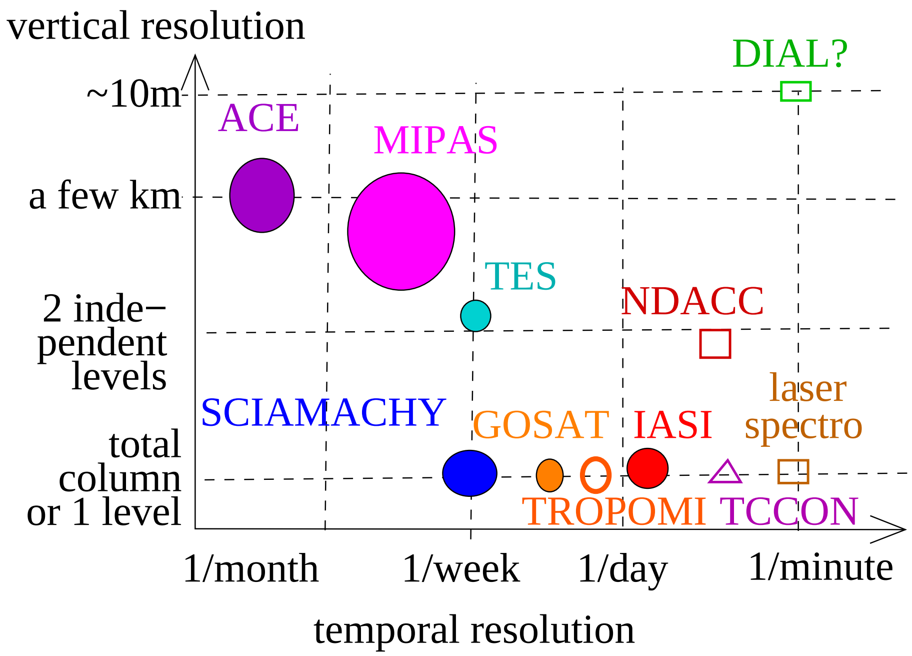

Figure 1.10: Overview of the vertical and temporal resolutions of the different measurements of water vapor δD. The temporal resolution for satellite measurements considers the return time for a measurement in a typical GCM grid box. The size of markers indicate the precision: the bigger the marker, the less precise, for the corresponding temporal resolution.

over a wide range of time scales, from the intra-rain event scale ([Tremoy et al., 2014]) to the diurnal ([Bailey et al., 2013]), synoptic and intra-seasonal ([Bonne et al., 015 , Galewsky and Samuels-Crow, 2014, Tremoy et al., 2012]), seasonal ([Tremoy et al., 2012, Steen-Larsen et al., 2014]) and inter-annual time scales. The measurement precision is as good as with older spectrometer techniques, i.e. ≃0.5 ‰ for δD and ≃5-10 ‰ for d-excess ([Tremoy et al., 2011]).

Remote-sensing instruments

Remote-sensing instruments have been developed for both ground-based and satellited measurements. They are passive sensors that used near-infrared radiance (in this case, they measure the vertically integrated δD) or thermal infrared radiance (in this case, they may have some profiling capacities) (figures 1.8, 1.9).

1. **Ground-based** instruments can measure the isotopic composition in the total-column (e.g. TCCON network, [Wunch et al., 2011]) or at different tropospheric levels (e.g. NDACC network, [Schneider et al., 2010a, Schneider et al., 2015]). Compared to in-situ measurements, they observe higher altitudes, and have the longest data records so far ([Schneider et al., 2010b]). The drawbacks are to record only δD ($\delta^{18}O$ is not precise enough to be useful), to be less precise (≃5 to 50 ‰ for δD depending on conditions and instruments) and to be limited to clear-sky conditions. Averaging measurements over some period of time refine the precision, but decrease the temporal resolution of the record.

2. **Satellite** measurements can measure the isotopic composition in the total-column (e.g. SCIAMACHY: [Frankenberg et al., 2009, Scheepmaker et al., 2012], GOSAT: [Frankenberg et al., 2013], soon TROPOMI: [Scheepmaker et al., 2016]) , at different tropospheric levels (e.g. TES: [Worden et al., 2006, Worden et al., 2007, Worden et al., 2012], IASI: [Schneider, 2011, Lacour et al., 2012]), or at different levels through the upper troposphere, stratosphere and higher (ACE: [Nassar et al., 2007], MIPAS: [Payne et al., 2007, Steinwagner et al., 2010], ODIN: [Urban et al., 2007]). The major advantage is that they allow us to visualize the global δD distribution. All these instruments retrieve only δD and only in clear-sky con-

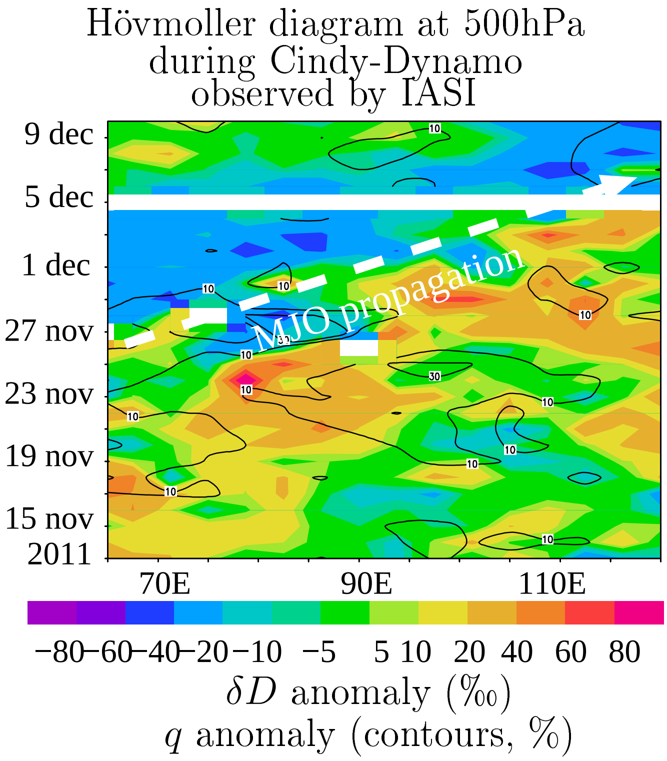

Figure 1.11: Time-longitude diagram for specific humidity (contours) and water vapor δD (shaded) at 500hPa as observed by IASI during the November 2011 Cindy-Dynamo field campaign case. This illustrates the unprecedented spatio-temporal coverage of IASI.

ditions, but they have various spatio-temporal coverage, vertical resolutions and precision properties (figure 1.10).

Figures 1.8, 1.9 and 1.10 compare the vertical resolutions and coverage, temporal resolution, spectral domains and resolutions of these different datasets. Most of the above-mentioned datasets are detailed and compared in [Risi et al., 2012a]. Table 1.1 summarizes the main advantages and drawbacks of the different insrument techniques.

Among satellite measurements, the **IASI** instrument on-board MetOp has the **best spatio-temporal coverage**: it retrieves δD globally twice a day! This allows to study for the first time the isotopic evolution in time and space during individual synoptic or intra-seasonal events ([Bonne et al., 015 , Tuinenburg et al., 2015], for example during an Madden-Julian Oscillation (MJO) event (figure 1.11). The future TROPOMI instrument should have a similar coverage.

The **TES** instrument has the **best vertical resolution**. It retrieves vertical profiles of δD with several independent levels of information through the troposphere. This is a unique opportunity to study the effect of some meteorological conditions or some convective processes as a function of altitude (figure 1.12).

In the future, the **DIAL** (differential absorption lidar) remote-sensing instrument, operated from the ground or from an aircraft, is very promising ([Bruneau et al., 2001]). Contrary to all instrument listed earlier, the DIAL is based on active remote-sensing technology. Preliminary studies suggest a vertical resolution of 10-100m, a temporal resolution of a few minutes and a precision arounf 10‰ (project by Cyrille Flamant).

measurement technique	advantages	drawbacks
cryogenic sampling and mass spectrometry	very precise small horizontal footprint	tedious very scarce
in-situ laser spectrometry	very precise continuous high frequency ($<$ 1 minute) small horizontal footprint	local measurements
TCCON	continuous small horizontal footprint	local measurements clear-sky only
NDACC	continuous long time series information on vertical profiles small horizontal footprint	local measurements clear-sky only
DIAL	continuous precise good vertical resolution (100m) high frequency (a few minutes) small horizontal footprint	not available yet: under development local clear-sky only
TES	spatial distribution information on vertical profiles (3 independant levels)	relatively weak spatio-temporal coverage clear-sky only
IASI	spatial distribution very good spatio-temporal coverage (global twice daily) plans for long time series	clear-sky only very costly retrievals, available only on demand.
SCIAMACHY	spatial distribution	relatively weak spatio-temporal coverage clear-sky only
GOSAT	spatial distribution good spatio-temporal coverage over land	clear-sky only
TROPOMI	spatial distribution very good spatio-temporal coverage (global twice daily)	not available yet clear-sky only
ACE	spatial distribution good vertical resolution in upper-troposphere-stratosphere	weak spatio-temporal coverage clear-sky only
MIPAS	spatial distribution good vertical resolution in upper-troposphere-stratosphere	clear-sky only

Table 1.1: Table summarizing the main advantages and drawbacks of the different instruments measuring water vapor isotopes.

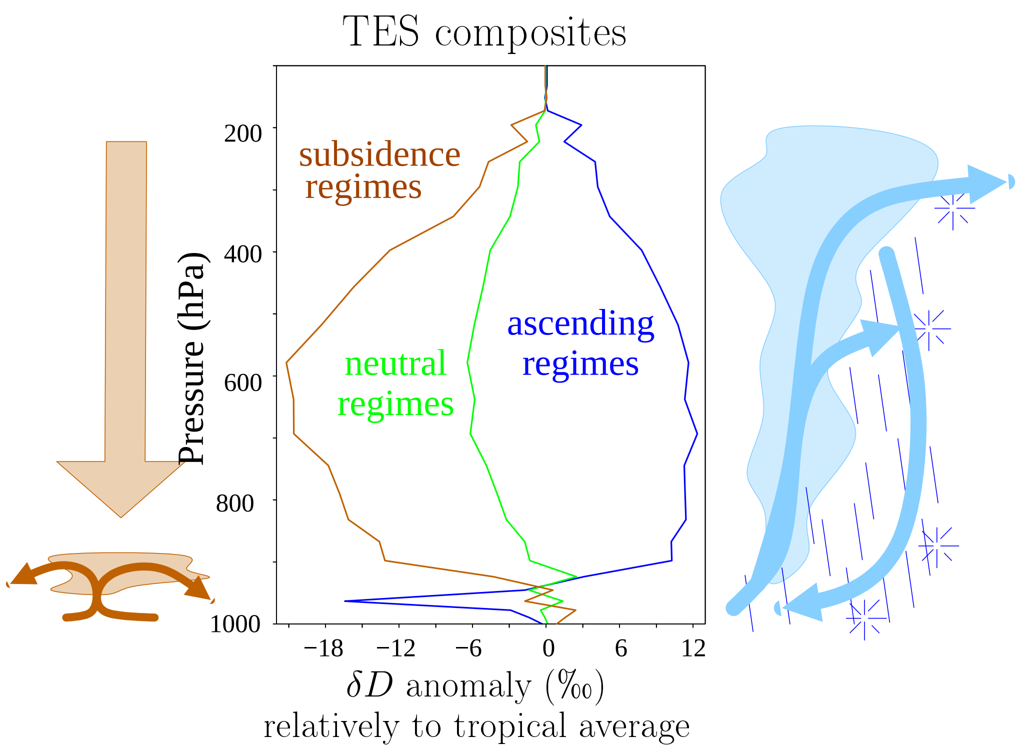

Figure 1.12: Vertical profiles of water vapor δD as measured by TES in the tropics, on average for dynamical regimes based on NCEP monthly large-scale vertical velocity: ascending regimes ($\omega < -20hPa/d$, blue), neutral regimes ($-20hPa/d \leqslant \omega \leqslant 20hPa/d$, green) and subsidence regimes ($\omega > 20hPa/d$, brown). This illustrates the vertical resolution of TES, with up to 3 independent levels of information.

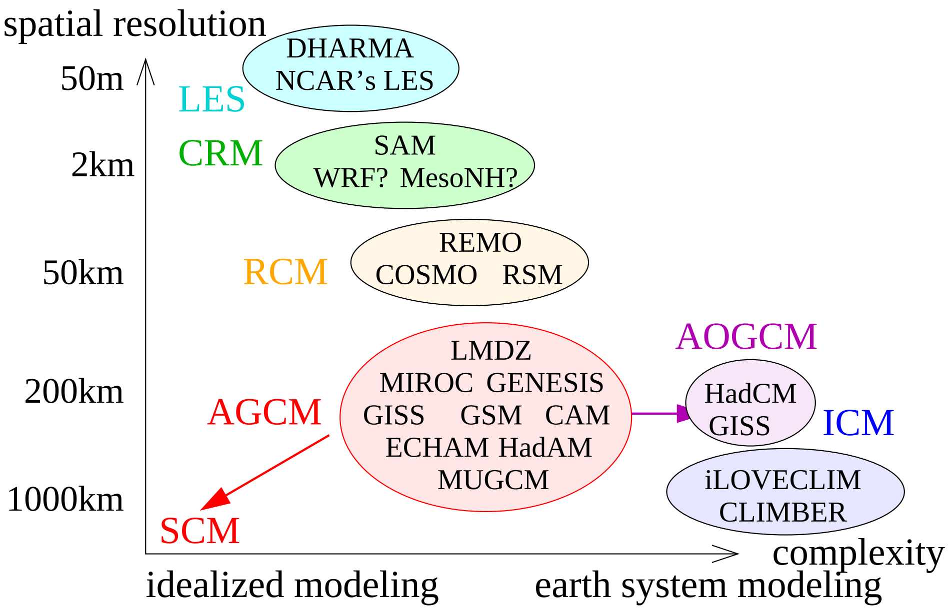

Figure 1.13: Overview of available isotopic models, as a function of spatial resolution and complexity in terms of the diversity in modeled processes.

1.3.2 Modeling

Many modeling tools simulating water isotopes have became available in the past few years (figure 1.13).

General circulation models

General circulation models (GCMs) are the best tool to try and understand the complexity of the controls on the isotopic composition of water vapor and precipitation, because they represent all physical processes involved in these controls (figure 1.2), albeit in a parameterized way. In addition, they are global, so that the only necessary boundary condition for water isotopes is the ocean water composition. Their limitation is their relatively coarse resolution ($\geqslant$100 km).

Since the pioneering work of [Joussaume et al., 1984], water isotopes have been implemented in at least a semi-dozen GCMs: GISS ([Jouzel et al., 1987b, Schmidt et al., 2007]), ECHAM ([Hoffmann et al., 1998]), MUGCM ([Noone and Simmonds, 2002]), GENESIS ([Mathieu et al., 2002]), CAM ([Lee et al., 2007]), GSM ([Yoshimura et al., 2008]), Hadley GCM ([Tindall et al., 2009]), LMDZ ([Risi et al., 2010b]); MIROC ([Kurita et al., 2011]). These models have been used for a wide range of applications, from paleo-climates ([Jouzel et al., 2000, Vuille et al., 2003, Pausata et al., 2011]) to isotopic data assimilation ([Yoshimura et al., 2014]). Some of these models can be coupled to isotope-enabled land surface models, to investigate land-atmosphere interactions ([Yoshimura et al., 2006, Risi et al., 2013]), or to isotope-enabled ocean GCMs, to investigate the feedback of ocean isotopic composition on atmospheric isotopic composition ([LeGrande and Schmidt, 2008]).

The number of isotopic GCMs allows us to perform **model inter-comparison studies**. The SWING (Stable Water Inter-comparison Group) inter-comparison project, phases 1 and 2, provide monthly outputs for AMIP simulations (i.e. with prescribed sea surface conditions) ([Noone and Sturm, 2010, Xi, 2014]). Using the SWING2 archive, inter-comparison studies have been useful to:

- check the robustness between simulated results ([Yao et al., 2013, Conroy et al., 2013, Gryazin et al., 2014]),
- understand the cause of model biases ([Risi et al., 2012b], section 3.2).

In the future possible extensions of SWING2 could include:

- Daily outputs from the existing AMIP simulations, to investigate synoptic and intra-seasonal variability, or even higher-frequency outputs, to investigate the isotopic response to the diurnal cycle.
- Simulations for paleo-climate conditions, i.e. an isotopic version of the Paleo-climate Model Inter-comparison Project (PMIP, [Braconnot et al., 2007, Braconnot et al., 2012]), to investigate how climate variations are imprinted in isotopic archives.
- Simulations using Single Column Model (SCM) versions of GCMs. Outside the isotopic community, studies with SCM are very useful to compare GCMs with data gathered during field campaigns ([Randall et al., 1996, Randall et al., 2003]), to develop and test parameterizations ([Rio and Hourdin, 2007, Hourdin et al., 2012]) or to study processes in an idealized framework ([Sobel and Bretherton, 2000]). This tool could be useful for the isotopic community as well. Vertical isotopic profiles simulated in convective-radiative equilibrium could provide an isotopic "ID" for each GCM.

Higher resolution models

As in GCMs, water isotopes have been implemented in regional models (RCMs): REMO ([Sturm et al., 2005]); isoRSM ([Yoshimura et al., 2010]), COSMO ([Pfahl et al., 2012]). With its zoom functionality, the LMDZ GCM also falls into this category. These models can be use for the same applications as GCMs, but their resolution (≃10-50km) allows to investigate processes at a much finer scale, e.g. during specific weather events ([Yoshimura et al., 2010, Pfahl et al., 2012]). High-resolution simulations are especially necessary in regions of high and complex topography ([Gao et al., 2013]) and can be used to for paleo-climate applications in such regions ([Eagle et al., 2013]).

Like GCMs, RCMs rely on parameterizations to represent sub-grid scale processes such as atmospheric shallow and deep convection. To represent explicitly these processes, Cloud Resolving Models (CRMs) or Large Eddy Simulations (LES) are necessary. Isotopes have been implemented in a few of such models: SAM ([Blossey et al., 2010]), DHARMA ([Smith et al., 2006]), NCAR's LES ([Lee et al., 2012]). There are on-going projects for WRF and Meso-NH. Such simulations are very useful to study the isotopic response to convective processes without relying on approximations and uncertainties inherent to parameterizations ([Smith et al., 2006, Moore et al., 2014], proposal A.4).

CRMs or LES simulations are very promising tools for the isotopic community and remain to be exploited:

- Idealized simulations could be directly compared to SCMs ([Rio et al., 2009, Couvreux et al., 2010, Hourdin et al., 2012]);
- CRMs or LES applied for case studies derived from observational campaigns could help to understand processes and evaluate GCMs. The lack of isotopic measurements that can provide initial and boundary conditions is a challenge for this application.

Earth system modeling

For paleo-climate applications, water isotopes have been implemented in Intermediate Complexity Models (ICM) (CLIMBER: [Roche et al., 2004], iLOVECLIM: [Roche, 2013]). Compared to GCMs, the advantage of these models is that they are more time efficient, allowing them to run several thousands of years, and they represent more components of the Earth system, including ice sheets. The drawback is that the atmospheric component is less detailed, with much coarser horizontal and vertical resolution and much simpler physical parameterizations.

Model-data synergy

Following the recent increase in data and model availability, tools have been developed in the past few years to rigorously compare models with observations.

Observations must be **co-located** in time and space with the available measurements. GCM simulations in which winds are **nudged** towards reanalyses are useful to ensure that the simulated meteorological conditions are realistic on a day-to-day basis ([Yoshimura et al., 2008, Risi et al., 2010b]). In the case of remote-sensing observations, the instrument sensitivity is taken into account by applying **averaging kernels** to the model outputs ([Rodgers and Connor, 2003]). These kernels may come directly from retrieval analysis ([Risi et al., 2012a]) or from a "observation simulator" ([Field et al., 2012]).

For paleo-climate applications, forward-proxy models of isotopic archives can be plugged to GCMs or ICMs (e.g. [Caley and Roche, 2013] for marine carbonates).

To summarize, **many observational and modeling tools are now available to study water isotopes.** I think that **the limiting factor now is how to make use of these tools to improve our understanding** of isotopic controls. In the next section we argue that we need to develop new interpretative frameworks.

1.3.3 Interpretative frameworks

Isotopic fractionation

Isotopic fractionation occurs due to the difference in molecule mass (equilibrium fractionation) and diffusivity (kinetic fractionation). Equilibrium fractionation coefficients are well know ([Merlivat and Nief, 1967, Majoube, 1971a, Majoube, 1971b]), except for ice-vapor fractionation at very low temperature ([Ellehoej, 2011]). Kinetic fractionation, which occurs during evaporation from liquid water in unsaturated conditions ([Merlivat and Jouzel, 1979]) or during condensation of ice crystal in supersaturated conditions ([Jouzel and Merlivat, 1984]) is more uncertain. Key uncertainties are the estimation of molecular diffusivities ([Merlivat and Jouzel, 1979, Cappa et al., 2003]), of the supersaturation in cold conditions ([Ciais and Jouzel, 1994, Casado et al., 2016]), of the effective relative humidity around re-evaporating rain drops ([Stewart, 1975, Bony et al., 2008]) or of a coefficient involved in the evaporation of bare soil that depends on the flow regime (turbulent, laminar and stagnant) ([Mathieu and Bariac, 1996, Braud et al., 2009b, Braud et al., 2009a]).

These uncertainties affect mainly second-order parameters such as d-excess, but to a lesser extent δD or $\delta^{18}O$. **Outside very low temperatures (polar regions, near the tropopause) and areas of strong rain re-evaporation in dry environments, the effect of individual phase changes on the water vapor δD are relatively well known.**

q-δD diagrams

The effect of individual phase changes on the water vapor δD can be visualized in q-δD diagrams popularized by [Worden et al., 2007] (figure 1.5), where q stands for specific humidity.

In these diagrams, **Rayleigh distillation** represents the dehydration of an air parcel through condensation. In a case where the condensate precipitates immediately, the remaining water vapor becomes more and more depleted following a logarithmic curve (blue on figure 1.5). In a case where the precipitation re-evaporates partially, the remaining water vapor becomes more and more depleted following a line that lie below Rayleigh (purple curves on figure 1.5). As an air parcel is dehydrated by mixing with a very dry (distilled) air parcel, the remaining water vapor becomes more and more depleted following a hyperbolic curve (orange on figure 1.5) that lies above Rayleigh.

Therefore, the advantage of these diagrams is to visualize the added value of δD compared to q only: different dehydrating processes may have the same effect on q but different effects on δD. Similarly, different moistening processes may have the same effect on q but different effects on δD. The joint q-δD distribution allows us to **discriminate between different moistening and dehydrating processes** ([Noone, 2009, Galewsky and Hurley, 2010, Noone, 2012, Galewsky and Samuels-Crow, 2014]). For example, these diagrams have been used to try and quantify the contribution of rain re-evaporation to tropical humidity ([Worden et al., 2007]), to try and understand moistening processes during the Madden-Julian Oscillation ([Berkelhammer et al., 2012, Tuinenburg et al., 2015], section 2.1.7) or to discriminate between different hypothesis explaining the dryness of the air during a stratospheric intrusion event ([Galewsky and Samuels-Crow, 2014]).

Beyond q-δD diagrams?

We know how each individual phase change and mixing event impacts water vapor δD. The difficulty is how to combine a large number of individual events to reproduce the complex reality. How much of the each air mass condenses, is mixed, is re-moistened with surface evaporation, rain re-evaporation, condensate detrainment? In addition, mixing usually involves more than two air parcels. In fact, air parcels involved are many, since air parcels at each level undergo mixing with air parcels at levels below and/or above and/or to the South, North, West and East, where the condensation, mixing and re-moistening history is different. How could we account for this complexity? q-δD diagrams cannot increase indefinitely in complexity. A better understanding requires to bridge the gap between this simple framework and more complex numerical modeling such as GCM modeling. I think that our progress is currently limited by the lack of an intermediate framework or model. This document will not provide any answer but I plan to work on it in the future (proposal A.4).

1.4 Document outline

In chapter 2, I will review processes controlling the water vapor isotopic composition, with a focus on those for which I have contributed to the understanding. In chapter 3, I try to apply this understanding to better evaluate the representation of convective and cloud processes in GCMs. In chapter 4, I try to apply this understanding to better evaluate the simulated climate response to past climate variations.

Chapter 2

Understanding controls on the water vapor isotopic composition

This chapter is devoted to the understanding of processes controlling the water vapor isotopic composition in the tropical troposphere, both in nature and in models. This is not a comprehensive review: rather, I focus on processes for which I have contributed to the understanding, i.e. convective and cloud processes (section 2.1) and continental recycling (section 2.2).

Process understanding was gained using the **simple** $q - \delta D$ **framework** (section 1.3.3), a single-column model ([Bony et al., 2008, Risi et al., 2008a]), the **LMDZ** GCM ([Hourdin et al., 2006]) in which I have implemented water isotopes ([Risi et al., 2010b]) or the LMDZ GCM coupled to the ORCHIDEE land surface model ([Krinner et al., 2005]) in which I have also implemented water isotopes ([Risi et al., 2016]).

2.1 Impact of convective and cloud processes

2.1.1 Distillation in convective ascents

The impact of convective and cloud processes on water vapor δD can be understood using the $q - \delta D$ diagram (figure 2.1). Boundary-layer water vapor (point 1 on figure 2.1) lifted by convective updrafts up to the upper troposphere (point 5) undergoes Rayleigh distillation (blue). To first order, vertical profiles of water vapor δD in the tropical environment are set by convective detrainment. Since water vapor δD decreases as air rises in convective updrafts, δD decreases with altitude in the environment.

2.1.2 In the upper troposphere: impact of convective condensate detrainment

In the upper troposphere, convective detrainment is a mixture of the distilled water vapor with condensate. Condensate is enriched: in the upper troposphere, nearly all the initial water vapor has been converted into condensate, so mass conservation implies that averaged over the vertical, condensate has a δD similar to that in the boundary layer vapor. Therefore, when distilled water vapor in the upper troposphere (e.g. point 3) mixes with condensate, δD increases following a mixing line connected with point 1 (cyan curve on figure 2.1). Consequently, even for a small moistening effect, condensate detrainment has a strongly enriched signature.

The strong enriched signature of convective condensate detrainment had been demonstrated by many studies based on satellite observations ([Moyer et al., 1996, Kuang et al., 2003, Nassar et al., 2007]), in-situ observations ([Webster and Heymsfield, 2003, Hanisco et al., 2007, Sayres et al., 2010]) and modeling ([Smith et al., 2006, Bony et al., 2008, Blossey et al., 2010]). The role of the shape of the mixing curve was pointed out by [Dessler and Sherwood, 2003] to explain the strong impact of of detrainment on δD even for a very small moistening (cyan on figure 2.1).

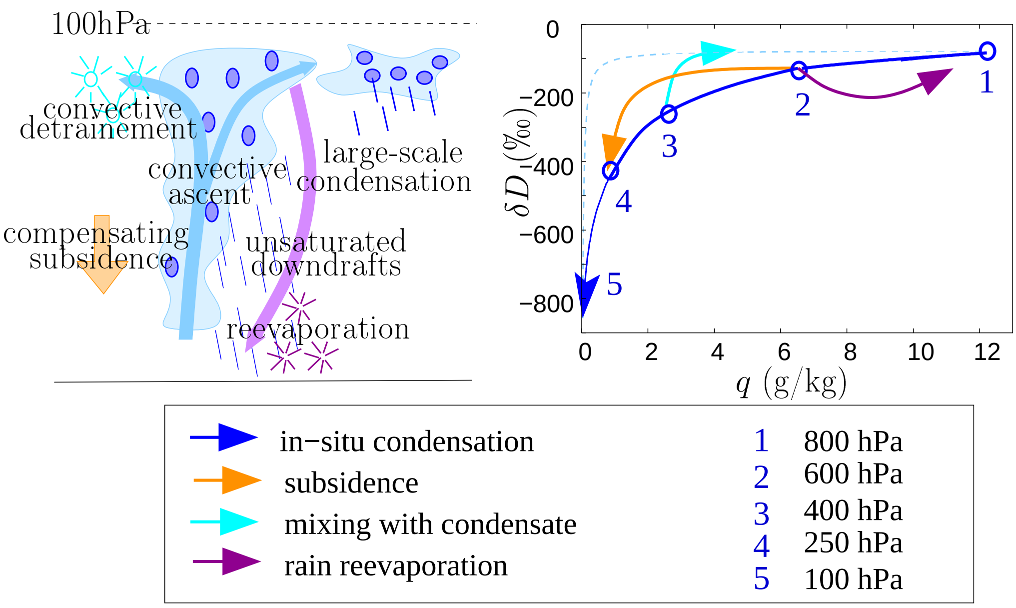

Figure 2.1: Effect of moistening and dehydrating processes associated with tropical deep convection, as illustrated in a q-δD diagram. Point 1 represents the boundary layer vapor. Points 2, 3, 4 and 5 represent distilled boundary layer vapor at different tropospheric levels, following a Rayleigh line.

In the upper troposphere, some in-situ condensation may also occur in anvils or in cirrus clouds. In this case, δD decreases following the same Rayleigh line (blue). All in all, the impact of detrainment on δD dominates over that of in-situ condensation since TES and MIPAS observations suggest that convective regions are associated with more enriched upper-tropospheric δD values (figure 2.2a).

Factors controlling the transport of water vapor from the upper troposphere to the lower stratosphere have been debated. In particular, the low humidity of stratospheric water vapor may be explained either by the low temperature in convective overshoots or by in-situ condensation during slow ascent through the tropical tropopause layer (TTL, ≃14-19 km) [Sherwood and Dessler, 2000, Holton and Gettelman, 2001]). The relative contributions of these two processes is difficult to quantify. The large-scale contribution of the direct injection of ice crystals into the TTL by convective detrainment is unknown ([Folkins et al., 1999, Gettelman et al., 2002, Khaykin et al., 2009]). Water isotope measurement are consistent with some convective injection of ice water through the tropopause layer. Could we use water isotopes to quantify the contribution of direct convective ice detrainment to the moistening of the tropical tropopause layer and lower stratosphere? So far, quantitative estimations have remained a challenge (section 1.2.3).

2.1.3 In the lower troposphere, impact of downdrafts and rain re-evaporation

In the middle or lower troposphere, air can be dehydrated by subsidence in meso-scale descents, convective downdrafts or, in models, by the compensating subsidence. Subsidence at a given level (e.g. point 2) acts as mixing with air at a higher level (e.g. point 4). Therefore, δD decreases following a mixing line connected with point 4.

Rain re-evaporation has a complex effect on water vapor. When rain re-evaporation is partial, light isotopes evaporate first so δD decreases in the water vapor. However, if rain re-evaporation is total, it plays a role similar to mixing with condensate, so δD increases in the water vapor. Therefore, **rain re-evaporation can have either a depleting or an enriching effect**, depending on the fraction of rain drops that re-evaporate (purple curve on figure2.1). At the scale of the tropics, [Worden et al., 2007, Field et al., 2010], using observations and model simulations respectively, argue for a depleting effect. In the Sahel, an arid region, [Risi et al., 2010a] argues for an enriching effect.

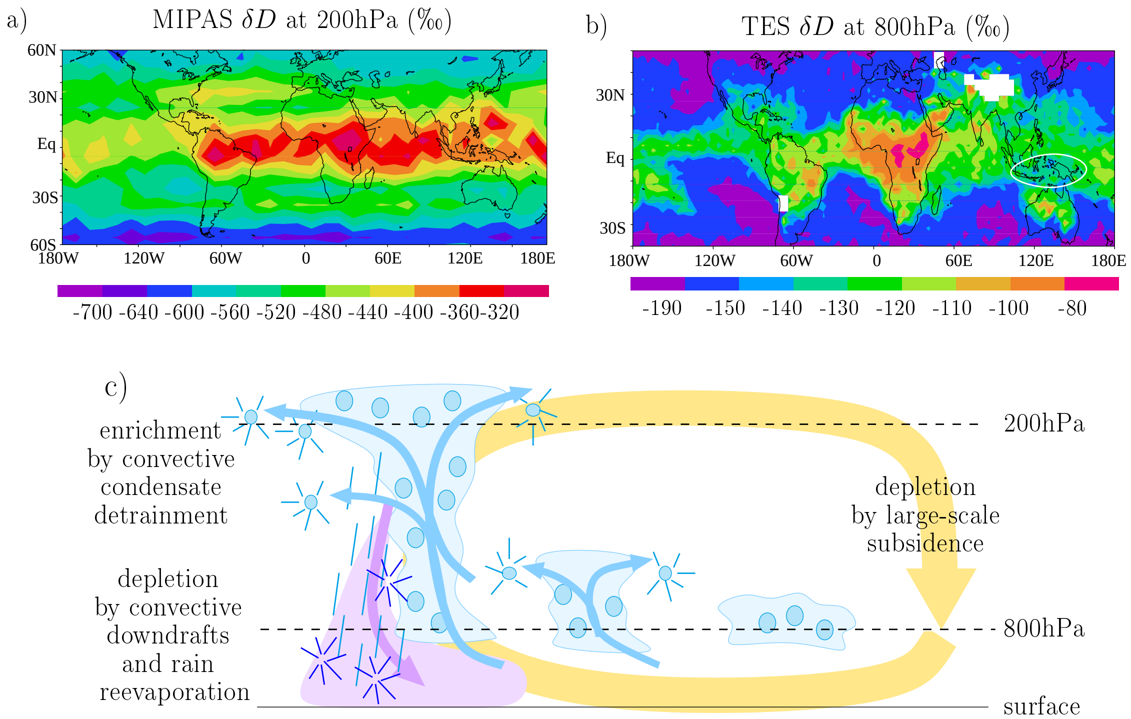

Figure 2.2: Annual mean maps of water vapor δD measured at 200hPa by MIPAS (a) and at 800hPa by TES (b). The white circle shows the Maritime Continent where the amount effect (section 2.1.3) can be observed the most clearly. c) Schematics explaining to first order the distribution of δD at these two levels, with deep convection enriching the water vapor by convective detrainment in the upper troposphere and depleting it by unsaturated downdrafts and rain re-evaporation in the lower troposphere.

In the case where the air is saturated, rain cannot re-evaporate but diffusive exchanges between the rain and the ambient vapor may occur. Since the rain falls from higher altitudes, it is more depleted than if it was in isotopic equilibrium with the ambient vapor. Therefore, diffusive exchanges act to deplete the vapor ([Lawrence et al., 2004, Lee and Fung, 2008]).

In the lower troposphere, deep convective regions are associated with more depleted δD values (figure 2.2b, mainly over the Maritime Continent). It has long been observed that regions and seasons with greater precipitation amount have more depleted precipitation ([Dansgaard, 1964]): this is called the "**amount effect**". The lower-tropospheric water vapor behaves in a similar way. The depletion is explained by a combination of subsidence in unsaturated downdrafts, meso-scale descents and compensating subsidence, of rain re-evaporation and of diffusive exchanges ([Risi et al., 2008a, Field et al., 2010, Kurita, 2013]), in proportions that remain to be quantified.

Note that some studies have explained the amount effect by the vertically-integrated water vapor budget ([Lee et al., 2007, Moore et al., 2014]). Deep convection is associated with convergence of air masses. As convergence increases, more of the vapor originates from the surrounding boundary layer vapor and less originates from fresh surface evaporation. The former being more depleted, as convergence increases, the column-integrated water vapor becomes more depleted. This explanation does not contradict the effect of subsidence, rain re-evaporation and of diffusive exchanges, which are necessary to explain why the surrounding boundary layer vapor is more depleted than fresh surface evaporation. It is just an another angle of view of the same processes.

2.1.4 In the middle troposphere

In the mid-troposphere, the impact of convection is probably a mixture of the effects occurring in the lower and in the upper troposphere, in proportions that depend on the depth of the convective mixing and on the height of convective detrainment ([Lacour, 2015], section 2.1.5).

2.1.5 Impact of convective depth

While precipitation amount is high in many areas of the inter-tropical convergence zone (ITCZ), the expected decrease in lower-tropospheric water vapor δD associated with the increase in precipitation amount (i.e. amount effect) can be seen very clearly only over the Maritime Continent (figure 2.2b,). Does the amount effect depend on the character of deep convection? Several studies suggest that the contributions of deep and shallow convection to precipitation could play a role ([Lee et al., 2015, Sutanto et al., 2015]). Here we test the hypothesis that the amount effect depends on the depth of deep convection.

As a proxy for precipitation amount, or convective intensity, we use the monthly large-scale vertical velocity (ω) at 500 hPa (ω_{500}) ([Bony et al., 2004]). We refer to the amount effect as "strong" if δD steeply decreases with increased ascent (i.e. negative ω_{500}). Conversely, we refer to the amount effect as "weak" if δD weakly decreases with increased ascent, or even increases. As a proxy for the convective depth, we use the pressure level at which the ω profile exhibits the maximum ascent. For example, ω peaking in the upper troposphere is consistent with deep convection, whereas ω peaking in the lower troposphere is consistent with shallow convection.

In LMDZ, locations and months where ω peaks in the upper troposphere, the amount effect is the strongest (figure 2.1.5a, blue). In contrast, in locations and months where ω peaks in the lower troposphere, the amount effect is the weakest (figure 2.1.5a, green). We can explain this behavior as follows (figure 2.1.5b): as deep convection increases, condensation, downdrafts and rain re-evaporation deplete the lower mid-troposphere efficiently. In contrast, as shallow convection increases, more HDO-enriched condensate is detrained into the lower mid-troposphere.

This property simulated by LMDZ is consistent with TES (figure 2.1.5c) and with IASI observations ([Lacour et al., 2016]). This explains why in the lower troposphere, strong convection is associated with depleted water vapor over the Maritime Continent, where convection is the deepest, whereas it is associated with enriched water vapor over equatorial Africa, where shallow convection is very active (figure 2.2b).

This work hs not been published yet.

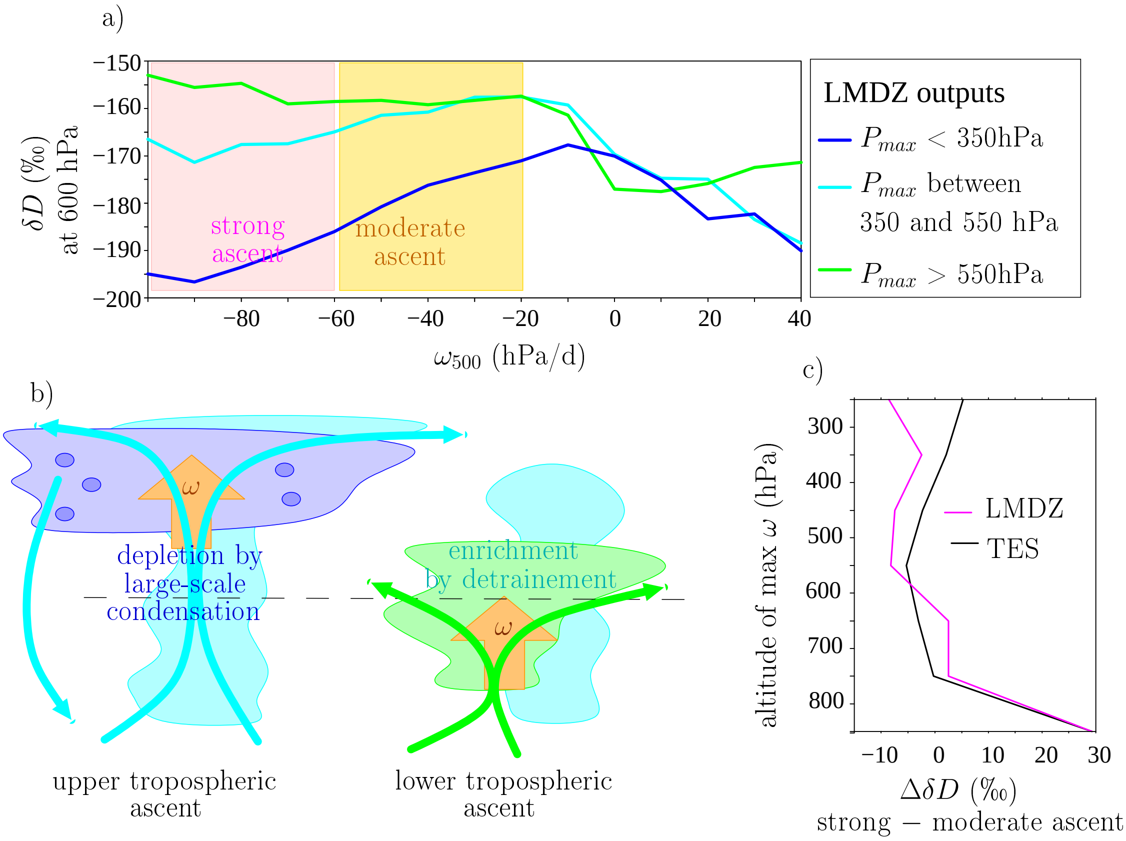

Figure 2.3: a) Monthly water vapor δD simulated by LMDZ at 600hPa as a function of the monthly large-scale vertical velocity at 500 hPa (ω), over tropical oceans. Results are classified into 3 categories depending on the shape of the vertical profile of ω: maximum ascent above 350 hPa (blue), between 350 and 550 hPa (cyan) or below 550 hPa (green). We see that δD decreases more strongly with increased ascent when the ascent peaks in the upper troposphere. b) Processes explaining this effect. c) Difference of δD at 600 hPa between the "strong" and "moderate" ascent regimes (as defined on panel a), for LMDZ and for the TES data.

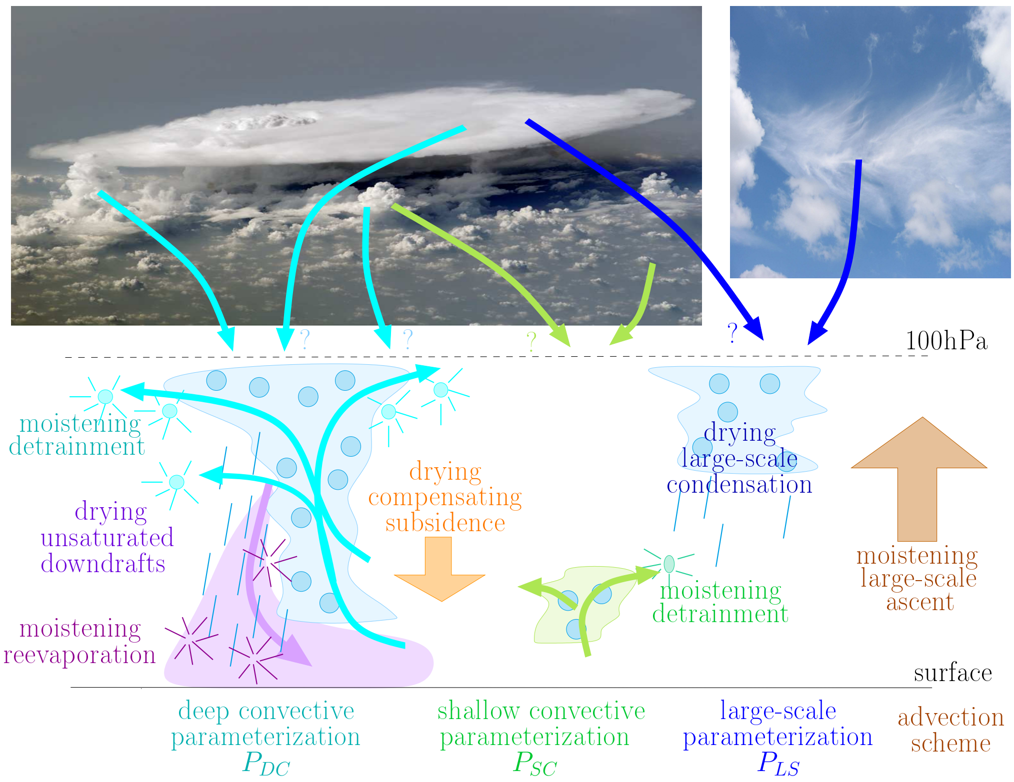

Figure 2.4: Partitioning of precipitation between deep convection (P_{DC}), shallow convection (P_{SC}) and large-scale precipitation (P_{LS}).

2.1.6 In models, impact of convective vs large-scale precipitation partitioning

In GCMs, precipitation is produced mainly by two kinds of parameterizations (Figure 2.4). First, **deep convection** represents the precipitation occurring in convective updrafts at the sub-grid scale. Second, the **large-scale condensation** scheme represents the precipitation produced at the grid scale, when the relative humidity at a given level exceeds a threshold. Partitioning between these two schemes is arbitrary ([Lawrence and Salzmann, 2008]). Its is model and resolution-specific. For example, some convective updrafts could be resolved at the grid-scale and therefore be dealt by the large-scale condensation parameterization. In contrast, some anvils could be represented by convective mixtures in the convection scheme (e.g. with the Emanuel scheme, [Emanuel, 1991]). Nevertheless, the partitioning between deep convection and large-scale condensation has an important impact on the simulated transport of water vapor and chemical species ([Rasch et al., 1997]), on the cloud vertical distribution and on the intra-seasonal variability of precipitation ([Kim et al., 2012]). Therefore, for a given model in a given configuration, this partitioning must be evaluated.

The isotopic composition of water vapor in the middle to upper troposphere is particularly sensitive to this partitioning (figure 2.5). We compare four versions of LMDZ (described in [Risi et al., 2012b]) in which parameters involved in the convective parameterization, in the large-scale condensation scheme or in the large-scale advection scheme were perturbed (figure 2.5). The proportion of the precipitation occurring as deep convection varies from 18% in the cyan version to 85% in the blue version. To investigate the strength of the "amount effect" in the upper troposphere, we plot water vapor δD at 400 hPa as a function of monthly large-scale vertical velocity at 500 hPa (ω_{500}) over all tropical oceans (figure 2.5a). Some LMDZ versions have a weak amount effect, i.e. δD decreases weakly from moderate to strong ascent regimes (red and blue versions). These versions

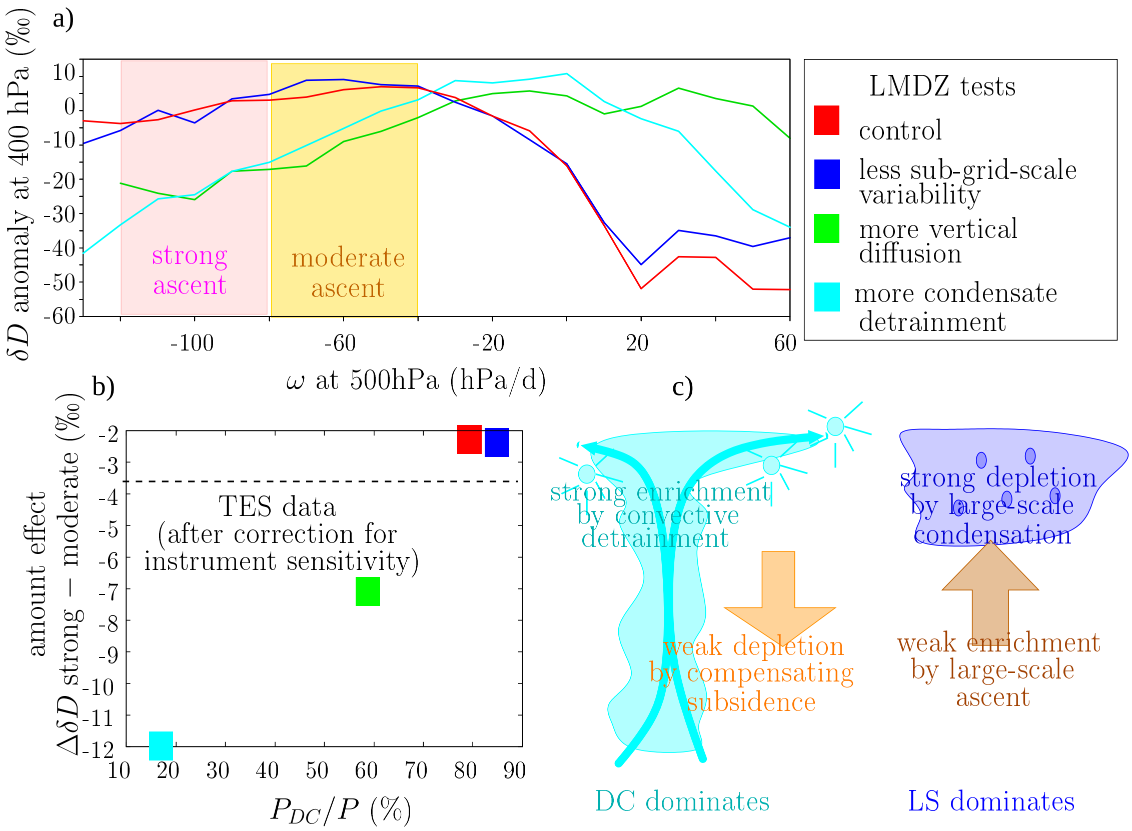

Figure 2.5: a) Monthly water vapor δD at 400 hPa (anomaly with respect to its tropical average) as a function of monthly vertical velocity of large-scale 500 hPa (ω_{500}) over tropical oceans, for four LMDZ versions. Whatever the version of the model, the water vapor is more depleted in subsidence regimes and in regimes of strong ascent characterized by intense convection. The depletion observed when the convection is more intense is called the "amount effect". The magnitude of the amount effect depends on the model version. b) The magnitude of the amount effect (quantified by the δD difference between regimes of strong and moderate ascent) as a function of the proportion of the precipitation (P) produced by deep convection (P_{DC}) to the expense of large-scale condensation. The amount effect is stronger as more precipitation is produced by large scale condensation. c) Schematic to explain this effect.

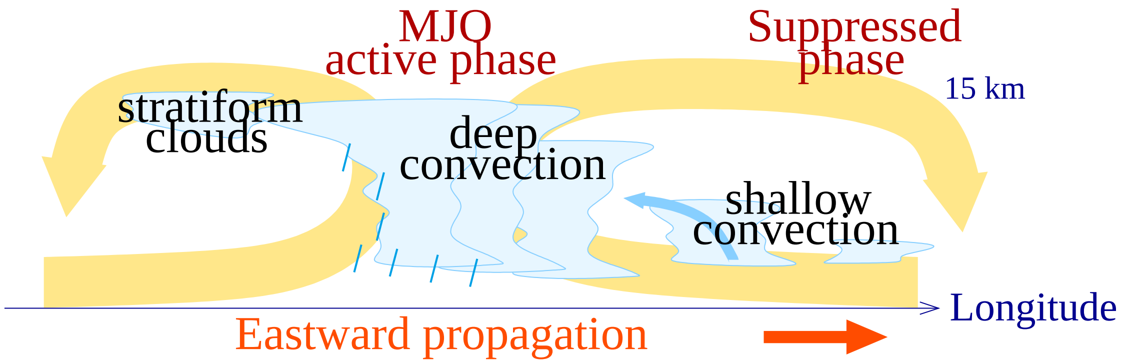

Figure 2.6: Sequence of cloud types associated with the propagation of an MJO event.

happen to be those with the strongest proportion of precipitation occurring as deep convection (figure 2.5b). In contrast, the other versions have a strong amount effect, i.e. δD decreases steeply from moderate to strong ascent regimes (green and cyan versions). These versions happen to be those with a lower proportion of precipitation occurring as deep convection. In fact, there is a very good relationship between the strength of the amount effect and the proportion of precipitation occurring as deep convection (figure 2.5b): **precipitation depletes the upper-tropospheric water vapor all the more as it occurs as large-scale precipitation at the expense of deep convective precipitation**.

When deep convection dominates, the largest terms in the water budget are the moistening by deep convective detrainment compensated by the dehydration by the compensation subsidence (figure 2.5c). In contrast, when large-scale precipitation dominates, the largest terms in the water budget are the moistening by large-scale vertical advection of water vapor compensated by the dehydration by large-scale condensation. According to the $q-\delta D$ framework (section 1.3.3), condensate detrainment has a strong enriching effect. For a given moistening, convective detrainment enriches the water vapor more than vertical advection. Therefore, when the detrainment-subsidence balance dominates over the advection-condensation balance, the source of water vapor is much more enriched. This contributes to the more enriched δD when deep convection dominates. In addition, according to the $q - \delta D$ framework, for a given dehydrating effect, subsidence depletes the water vapor less than condensation. This also contributes to the more enriched δD when deep convection dominates.

This property could open the door to the use of isotopic observations to constrain the partitioning between deep convection and large-scale condensation parameterizations. This work has not been published yet.

2.1.7 Convective processes during the Madden-Julian Oscillation

What is the MJO? Why is it so difficult to simulate by GCMs?

The Madden-Julian Oscillation (MJO, [Madden and Julian, 1971, Madden and Julian, 1972]) is the largest mode of intra-seasonal variability in the tropical atmosphere ([Zhang, 2005]). With a periodicity of 30-90 days, it exhibits a large area of strong convection that initiates over the Western Indian Ocean and propagate westwards until it fades away in the central Pacific (figure 2.6). It involves a tight coupling between large-scale circulation, convective, cloud and radiative processes, and air-sea interactions. For example, the large-scale dynamical response of the atmosphere to convective heating leads to large-scale convergences that feeds back positively onto convection ([Ghil and Mo, 1991]). The greenhouse effect of anvil clouds associated with convection can also contribute to the heating ([Kim et al., 2015]). Shallow and mid-level convective moistening play a key role to pre-condition the atmosphere to the active phase of the MJO ([Inness et al., 2001, Benedict and Randall, 2007, Cai et al., 2013]). Moisture-convection feedbacks may contribute to slowing down the MJO propagation ([Bony and Emanuel, 2005]). A recharge-discharge mechanism, in which shallow convection progressively moistens the free troposphere until a threshold is reached

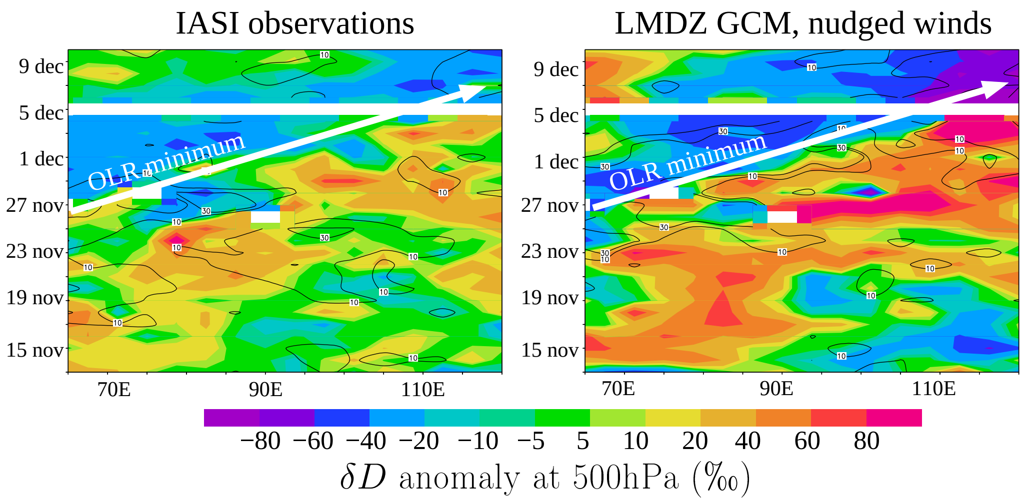

Figure 2.7: Propagation of the November 2011 MJO event as observed by the IASI satellite instrument (a) and as simulated by the LMDZ model when nudged by reanalysis winds (b). Water vapor δD and q at 500 hPa are plotted as colors and contours respectively, as a function of longitude and time. The white arrow illustrate the propagation of the minimum in OLR (Out-going Long-wave Radiation) which tracks the convective maximum.

that allows for the maintenance of deep convection, may explain the slow periodicity ([Maloney, 2009]). Air-sea coupling may also play a role ([Waliser et al., 1999, DeMott et al., 2014]).

To reproduce the MJO, a GCM must therefore have a set of physical parameterizations that can simulate all these feedbacks. It is therefore not surprising that for decades, GCMs have had difficulties to simulate the MJO ([Slingo et al., 1996, Lin and Coauthors, 2006, Kim et al., 2009, Hung et al., 2013]).

Joint $q - \delta D$ evolution during observed MJO events

Since these feedbacks at play during the MJO involve convective and cloud processes that are known to impact water stable isotopes, could isotopic measurement help to understand the reason for the model difficulties? This was the motivation of my CONVISO project (2013-2017, funded by the French National Research Agency), in collaboration with my post-doc Obbe Tuinenburg. As a first step, the impact of convective and cloud processes on water vapor δD during MJO events needs to be understood, both in nature and in models.

Using TES observations, q and δD during composites of MJO events can be analyzed ([Berkelhammer et al., 2012]). With the unprecedented spatio-temporal coverage offered by IASI, q and δD during individual MJO events can be also documented. For example, figure 2.7a shows q and δD observed at 500 hPa during the November 2011 MJO event that took place during the CINDY-DYNAMO campaign ([Yoneyama et al., 2013]). Observed q is maximum 0 to 1 day before the convective maximum, whereas observed δD is minimum 3 days after after the convective maximum.

For more representative statistics, we plot the joint $q - \delta D$ evolution for 7 MJO events in the central Indian Ocean, including the November 2011 event (figure 2.8). In observations, on average, the $q - \delta D$ evolution follows a circular clockwise shape: a moistening before the active phase, a depletion during the active phase, then a drying and finally a re-enrichment (figure 2.8a).

Joint $q - \delta D$ evolution during MJO events simulated by LMDZ

LMDZ captures the observed $q - \delta D$ cycle for the CINDY-DYNAMO case, notably the δD minimum lagging the convective maximum by several days (figure 2.7b). LMDZ captures the circular clockwise shape to some extent for about one third of MJO events (figure 2.8b). But for many other events, the shape is simply a "return trip" following a diagonal: simultaneous moistening and depletion up to the convective maximum, followed by a drying and re-enrichment after the

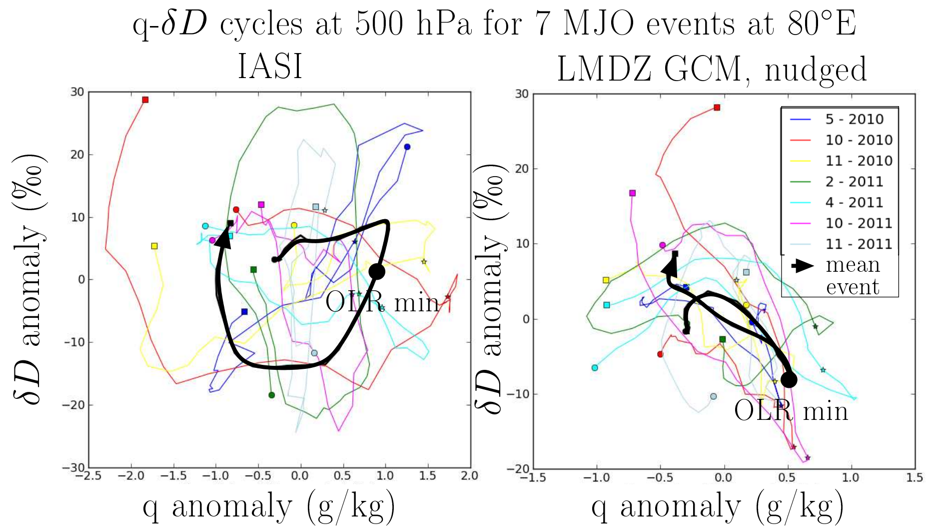

Figure 2.8: δD anomaly as a function of q anomaly at 500 hPa as observed by IASI (a) and simulated by LMDZ (b) for 7 MJO events at 80°E (in the central Indian Ocean) during the period 2010-2011. The black line shows the average over the different events. Stars represent values on the day of maximum precipitation, circles represent values 17 days before this maximum, and squares represent values 17 days after this maximum. The date (month and year) of the start of each event is listed in the legend.

convective maximum (figure 2.8b). This suggests that LMDZ does not reproduce the sequence of moistening and drying processes properly during these MJO events.

What determines the $q - \delta D$ shape in LMDZ? To address this question, dq/dt and $d\delta D/dt$ tendencies from the different parameterizations were analyzed in detail ([Tuinenburg et al., 2015]).

What determines the q-δD shape in LMDZ for well-simulated events and in nature?

For events showing the circular clockwise shape, there is first a moistening by large-scale convergence that occurrs without affecting δD. Then, during the few days surrounding the convective maximum, both convection and large-scale condensation deplete the water vapor (consistent with the amount effect), while strong large-scale ascent supplies enough water vapor for q to remain constant. Then, the large-scale ascent weakens and becomes to weak to compensate for the drying by convection and large-scale condensation, so q decreases. Finally, large-scale convergence recycles the air with fresh and enriched vapor, so δD increases back to its initial value.

Since LMDZ captures the observed $q - \delta D$ evolution well for these events, we propose that a similar sequence of processes occurs in nature. The $q - \delta D$ evolution suggests that before the active phase, **humidity builds up mainly due to large-scale convergence**. If the humidity build-up was due to shallow convection ([Benedict and Randall, 2007, Maloney, 2009]), we would expect a simultaneous increase in δD, which is not the case here. More analysis is needed to check that the observed $q - \delta D$ evolution reflects a dominant role of large-scale convergence over shallow convection in the moisture build-up.

What determines the q-δD shape in LMDZ for poorly-simulated events? Implications to understand model difficulties in simulating the MJO

For events showing the diagonal return-trip shape, the processes at play are the same but the time sequence is different. In particular, convection triggers sooner, as soon as q increases, leading the δD to decrease in concert with the q increase. Then, convection and large-scale condensation weaken sooner while the large-scale convergence recycles the air sooner, leading the δD to increase in concert with the q decrease.

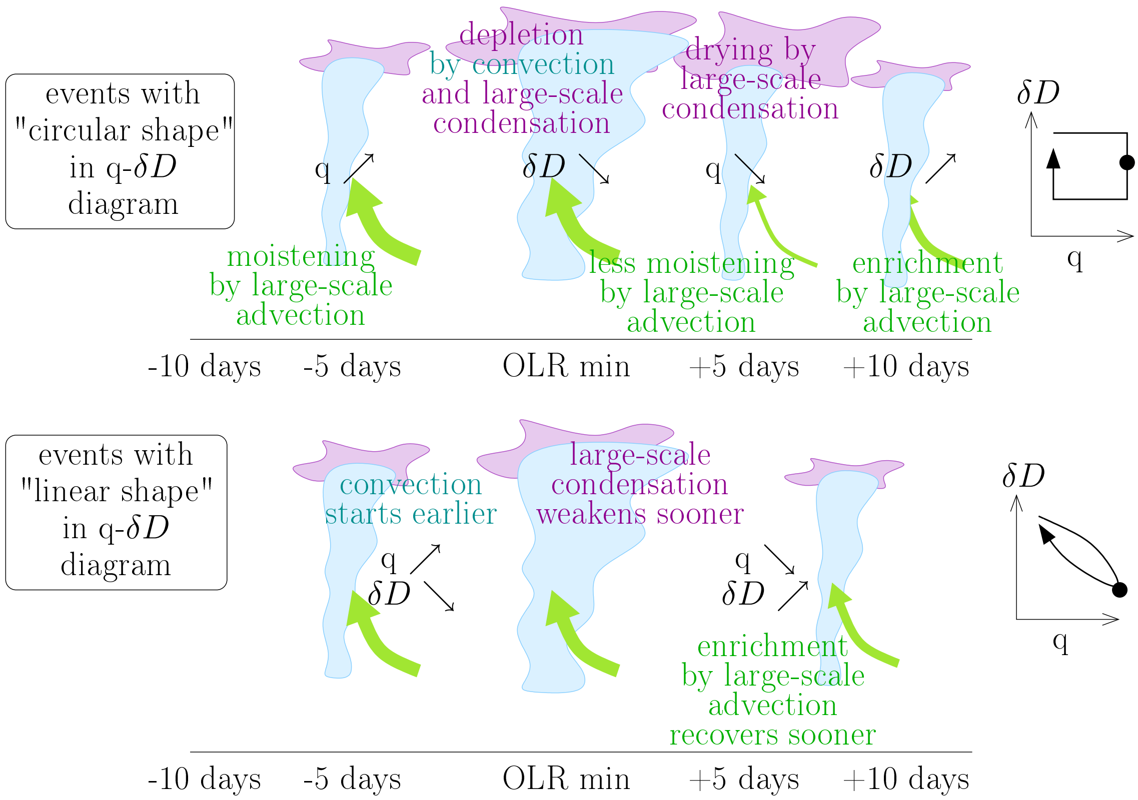

Figure 2.9: Schematic showing the sequence of moistening and dehydrating process simulated by LMDZ during two kinds of MJO events: events showing the circular, clockwise shape (a) and those showing the diagonal return-trip shape (b) in the $q-\delta D$ diagram. Blue and purple clouds represent deep convective clouds and large-scale condensation clouds respectively. Green arrows represent large-scale advection.

Since LMDZ does not capture the observed $q-\delta D$ evolution for these events, we hypothesize that the simulated sequence of events reflects LMDZ faults. These faults might be responsible for the poor MJO simulation when LMDZ is run in a free mode. In particular, the fact that **convection triggers too soon**, before the moisture has the time to build up, is a possible cause for the poor MJO simulation by models ([Kim et al., 2012, Kim et al., 2014]). This could be due to the underestimated sensitivity of convection to tropospheric humidity of most convective parameterizations ([Derbyshire et al., 2004]).

These results suggest that the joint $q-\delta D$ evolution during MJO events is useful for model evaluation. Work is in progress to test this hypothesis by comparing sensitivity tests with LMDZ (proposal A.7).

2.2 Impact of continental recycling

2.2.1 Impact of continental recycling on water vapor δD and d-excess

Water vapor δD is all the more enriched as recycling of precipitation through evapo-transpiration is high. This property may contribute to the maximum enrichment that is observed by satellites over tropical land masses ([Worden et al., 2007, Brown et al., 2008, Frankenberg et al., 2009]). This property is also the basis for studies trying to quantify continental recycling ([Salati et al., 1979, Risi et al., 2013]). Water vapor from vegetation transpiration is much more enriched than oceanic evaporation. This is because lighter isotopes evaporate more easily from free liquid surfaces. In contrast, transpiration is not associated with fractionation relatively to soil water, because there is no fractionation during root extraction ([Washburn and Smith, 1934, Barnes and Allison, 1988, Flanagan and Ehleringer, 1991]) and all water extracted by the root needs to be transpired shortly after. Soil water originates from precipitation, which is to first order at equilibrium with the ambient vapor ([Field et al., 2010]), and thus more enriched than the ambient vapor (figure 2.10).

Evaporation from bare soil is characterized by a strong signature in the second-order parameter d-excess ($d=\delta D-8\cdot\delta^{18}O$, section 1.2.1) (figure 2.10). This is because kinetic fractionation during the evaporation of soil water is very strong ([Mathieu and Bariac, 1996, Braud et al., 2009b, Braud et al., 2009a]). As kinetic fractionation increases, the diffusivity coefficients become important and the evaporation of HDO is favored by its high diffusivity. This property is the basis of studies trying to partition continental recycling into its transpiration and evaporation components ([Gat and Matsui, 1991]).

2.2.2 Implications to quantify continental recycling

These two properties could in theory be exploited to quantify both the continental recycling (with $\delta^{18}O$ or δD) and its components (with d-excess). In practice, it is difficult. First, d-excess cannot presently be measured by satellite, and in-situ measurements with sufficient precision remain scarce ([Noone et al., 2012, Tremoy et al., 2012]). Satellites can measure δD from space with reasonable precision and spatio-temporal coverage, but lack absolute calibration. Placing satellite measurements on figure 2.10 in an attempt to quantify continental recycling is thus not applicable.

In [Risi et al., 2013], I rather tried to focus on temporal variations of water vapor δD. I explored the potential of satellite isotope measurements to quantify the contribution of continental recycling to the intra-seasonal variability in water vapor in different regions. I compared estimates of this contribution by directly tracking water in LMDZ-ORCHIDEE or by isotopic methods. I showed that unfortunately, it is difficult to get quantiative estimates of the recycling contribution from isotopic methods, for two reasons. First, uncertainties associated with isotopic satellite observations lead to contradictory results between the TES and GOSAT datasets. Second, and more fundamentally, the isotopic composition of water vapor is sensitive to other processes than continental recycling, such as the intensity and depth of the turbulent and convective mixing (section 2.1). It is **very difficult to disentangle the different processes and extract the continental recycling component only**.

With the growing deployment of in-situ instruments that measure surface water vapor d-excess (section 1.3.1), will we be able to use this parameter? Even if the case, the same limitation

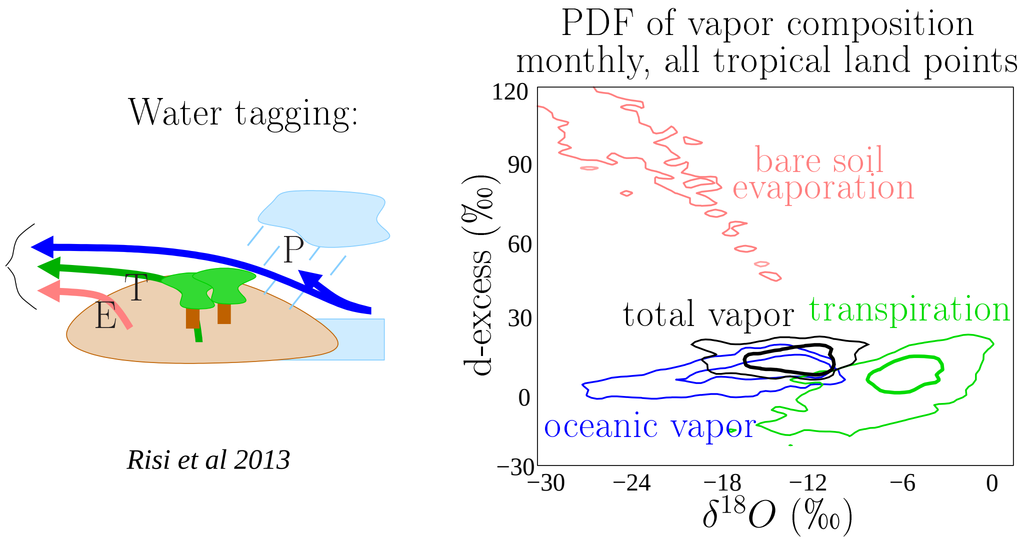

Figure 2.10: Probability density function of near-surface water vapor δD and d-excess (black) simulated by LMDZ for all tropical land locations. Using water tagging, water vapor was decomposed into three components depending on their evaporative origins: vapor originating from ocean evaporation (blue), vapor originating from land surface transpiration (green) and vapor originating from bare soil evaporation (pink).

associated with our incomplete understanding will remain. In the case of d-excess, uncertainties reside in atmospheric processes but also in land surface hydrological processes. In particular, a multi-layer soil model is necessary to simulate the d-excess of the evapo-transpiration flux ([Risi et al., 2016]).

Chapter 3

Using water stable isotope measurements to evaluate and improve physical parameterizations

In this chapter, I try to use the understanding reviewed in chapter 2 to better evaluate and improve the representation of convective and cloud processes in GCMs, trying to go beyond the usual suggestion that water isotope measurement *have the potential* to help evaluate and improve models.

3.1 Overview

3.1.1 A necessary condition: the representation of isotopic processes is less uncertain than that of physical processes

Using isotopes for improving the representation of physical processes by models is possible only if the representation of isotopic processes is less uncertain than the representation of physical processes. In section 1.3.3, we argue that isotopic fractionation during individual phase changes is well known, at least for δD in the tropical troposphere. The major uncertainty in the simulation of δD is thus how to combine all phase changes of the water cycle and all advection and mixing events between air parcels to reproduce the complex reality.

3.1.2 Water isotopes to understand the cause of model biases

When there is a well-known model bias in the simulation of a meteorological variable M, water isotope measurements can help discriminate between different hypotheses to explain this bias. To do so, we can plot the bias for an isotopic variable I as a function of the bias for M, for different sensitivity tests or for different models (figure 3.1, squares). Using some theory, we can predict how the bias in I should relate to the bias in M depending on the reason for the bias M (figure 3.1, arrows). If the different model simulations plot along one of these theoretical lines, this suggests the most likely reason for the bias in M. I will illustrate this method by my study on the middle-upper tropospheric subtropical bias in relative humidity ([Risi et al., 2012b], section 3.2).

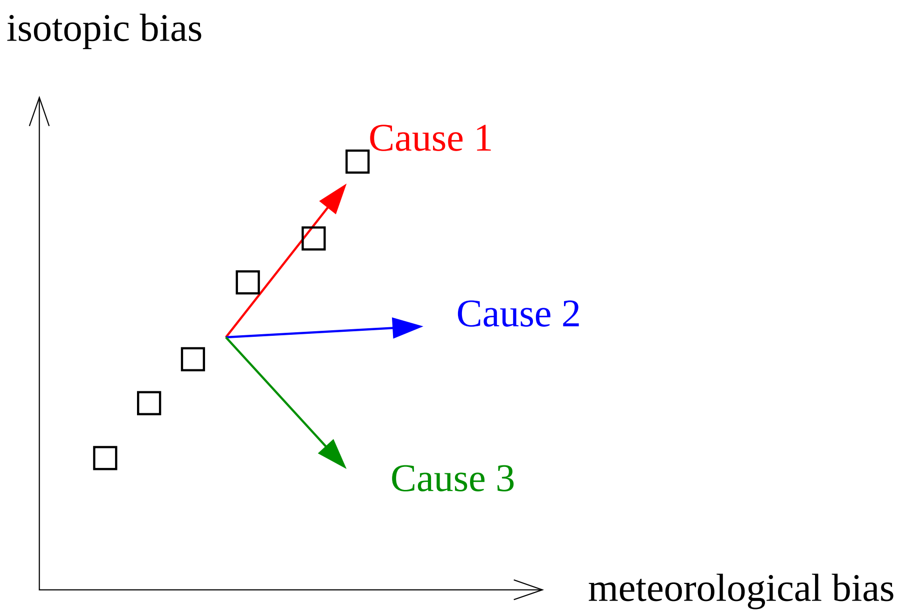

Figure 3.1: Method to discriminate between different hypotheses for processes causing a meteorological bias in a GCM. Arrows represent the expected relationship between the meteorological bias and an isotopic bias depending on the reason for these biases. Squares represent different models or sensitivity tests.

3.1.3 "Isotopic mysteries": water isotopes to detect shortcomings in the representation of physical processes

GCMs exhibit some common model biases for water isotopes. For example, all isotopic GCMs underestimate the latitudinal gradient in upper tropospheric water vapor δD ([Risi et al., 2012a]). They all underestimate the variability in water vapor and precipitation d-excess at the daily scale in polar regions ([Steen-Larsen et al., 2013]). They all underestimate the isotopic depletion recorded at LGM in tropical archives ([Jouzel et al., 2000, Werner et al., 2001, Risi et al., 2010b, Jasechko et al., 2015]). Do these biases reveal a common problem of GCMs to represent some physical processes? Does it mean some key processes are missing?

If a climate simulation seems correct when compared to classical climate variables, it could be at the price of compensating errors. Such errors may compoensate for q and other hydrological variables, but not always for water isotopes. Therefore, water isotopes are helpful to detect shortcomings in the representation of physical processes. This opportunity has remained largely unexplored so far. I present a possible explanation for underestimated latitudinal gradient in upper tropospheric δD (section 3.3).

3.1.4 Water isotopes to tune physical parameters

GCMs rely on physical parameterizations, in which empirical model parameters are tuned to optimize the global simulation ([Mauritsen et al., 2012, Hourdin et al., 2012]). Several studies have shown the sensitivity of simulated water isotopic composition of vapor or precipitation to model parameters in cloud or convective parameterizations ([Schmidt et al., 2005], [Bony et al., 2008], [Lee et al., 2009], [Field et al., 2014]). This suggests that water isotopic measurements could help constrain some tunable model parameters.

However, processes controlling water isotopes are complex. Several model parameters can impact the water isotopic composition in the same way, so the model tuning is an under-constrained problem. Ideally, it is recommended to combine several variables, and not only isotopic variables, in the tuning procedure. It is also important to understand exactly to which physical process water isotopes are sensitive, beyond individual model parameters.

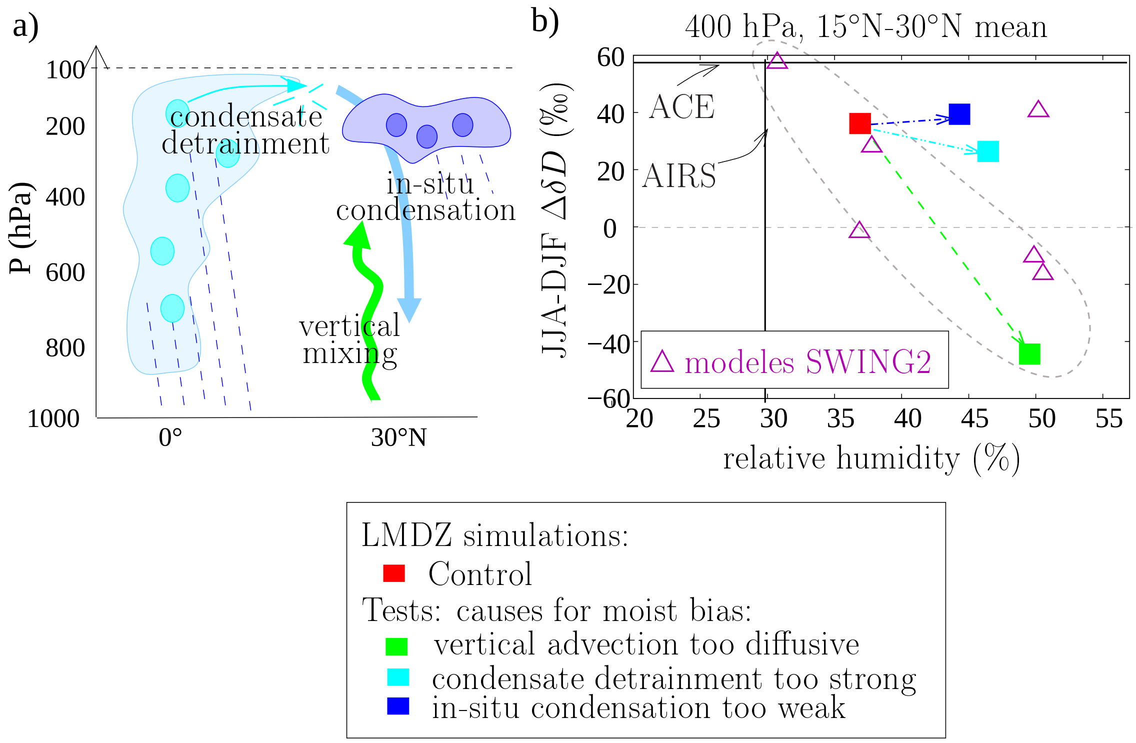

Figure 3.2: a) Hypotheses for processes that could explain the moist bias in the subtropical middle and upper troposphere shared by all GCMs: insufficient condensation (blue), excessive convective detrainement (cyan), or excessive vertical diffusion (green). b) Seasonality ofδD as a function of relative humidity in the subtropical mid-troposphere, for observations (black), sensitivity tests with the LMDZ GCM (full colored squares) and for SWING2 models (purple empty squares) ([Risi et al., 2012b]). Relative humidity and isotopic observations from AIRS and ACE respectively are also shown.

In section 3.4, I present a preliminary study using water vapor δD measurements to evaluate the convective depth (section 3.4).

3.2 Understanding the bias in middle and upper-tropospheric subtropical humidity

This is an example of how water isotope measurements could be used to understand a well-known meteorological bias, using the methodology explained in section 3.1.2.

For decades, climate models have suffered from a wet bias in the middle and upper tropical and subtropical troposphere ([Soden and Bretherton, 1994, Salathe and D, 1995, Chen et al., 1996, Roca et al., 1997, Chen et al., 1998, Allan et al., 2003, Brogniez et al., 2005, Pierce et al., 2006, John and Soden, 2007, Sherwood et al., 2010, Chung et al., 2011]). The bias can be explained by many different hypotheses, such as insufficient condensation, excessive convective detrainement, or excessive vertical diffusion ([Risi et al., 2012b], figure 3.2a). To test whether these different hypotheses are associated with specific isotopic signatures, 3 sensitivity tests with LMDZ were performed:

1. one with a modified parameter in the large-scale condensation scheme leading to reduced large-scale condensation (blue),
2. another with reduced precipitation efficiency in the convective scheme leading to enhanced convective detrainment (cyan),
3. another with a modified advection scheme leading to enhanced vertical diffusion (green).

In each of these tests, the moist bias is enhanced in the same way, so humidity measurements alone cannot discriminate what is the cause of the moist bias. In contrast, the effect on δD depends on the test, especially the effect on the seasonality of subtropical mid-tropospheric δD (figure 3.2b). The δD seasonality is not affected much when large-scale condensation is reduced or when detrainment is enhanced, but it is strongly reduced, or even reversed, when the vertical diffusion is increased (figure 3.2b, green).

Relative humidity observations by AIRS show that the control LMDZ simulation (red) is already too moist (figure 3.2b), consistent with the common moist bias shared by all GCMs. In addition, ACE observations show that the δD seasonality simulated by the control version of LMDZ is lower than observed. More precisely, the control LMDZ simulation plots along the line linking the observations and the excessive diffusion simulation. This suggests that in the control LMDZ simulation, the cause of the moist bias is a vertical advection that is already too diffusive.

To identify the reason of the moist bias in GCMs, the available SWING2 models were plotted on figure 3.2b (purple triangles). Most models plot along a line linking the control and the excessive diffusion simulation. Models exhibiting the largest moist bias have the lowest (or most reversed) seasonality. This suggests that **the most frequent reason for the moist bias in most of these models is excessive vertical diffusion**. For example, this could be due to insufficient vertical resolution.

3.3 Identifying a missing process controlling upper-tropospheric convective detrainment

All GCMs from the SWING2 archive systematically underestimate the zonal gradient in upper tropospheric water vapor δD ([Risi et al., 2012a]). Does this indicate that a process in the upper troposphere is missing in all those GCMs? This study has not been published yet.

A systematic bias in the simulated upper tropospheric δD

In MIPAS observations, annual-mean upper-tropospheric water vapor δD at 200 hPa is maximum in deep convective regions, i.e. in the inter-tropical convergence zone (ITCZ), especially above equatorial land masses (equatorial South America and Africa) (figure 3.3a). This probably reflect the enriching effect of condensate detrainment by deep convection (section 2.1.2). Subtropical regions are comparatively more depleted, especially the regions undergoing strong subsidence at the East of oceanic basins (off the coast of California, Peru and Angola). We can quantify this pattern by calculating the difference $\Delta\delta D$ between annual zonal-mean δD in the equatorial region (15°S-15°N) and annual zonal-mean δD in the tropical region as a whole (30°S-30°N). In MIPAS observations, δD is 50‰ more enriched in the equatorial region than in the tropical region as a whole ($\Delta\delta D = 50$‰). ACE observations are qualitatively consistent ($\Delta\delta D$=30‰.

LMDZ cannot capture this pattern (figure 3.3b). It does capture a local maximum of upper-tropospheric water vapor δD above equatorial land masses, but it is much weaker. Over ocean, LMDZ does not capture the observed enrichment above the inter-tropical convergence zone. In LMDZ, $\Delta\delta D$ is only 13‰.

All other GCMs from the SWING2 database share the same difficulties as LMDZ (figure 3.3c), with $\Delta\delta D$ ranging from -14‰ to 18‰.

Upper tropospheric δD reflects the moistening impact of convective detrainment

To try and understand the reason for the disagreement, we compared different sensitivity tests (the same as in section 3.2, [Risi et al., 2012b]). We notice that simulated $\Delta\delta D$ is very sensitive to the contribution of upper tropospheric convective detrainment to the water budget (figure 3.3c). In sensitivity tests where convective detrainment moistens the upper troposphere the most, $\Delta\delta D$ is the largest. This is consistent with the hypothesis that the equatorial maximum in δD is associated with a maximum in convective detrainement. The absence of any relationship between Δq (defined

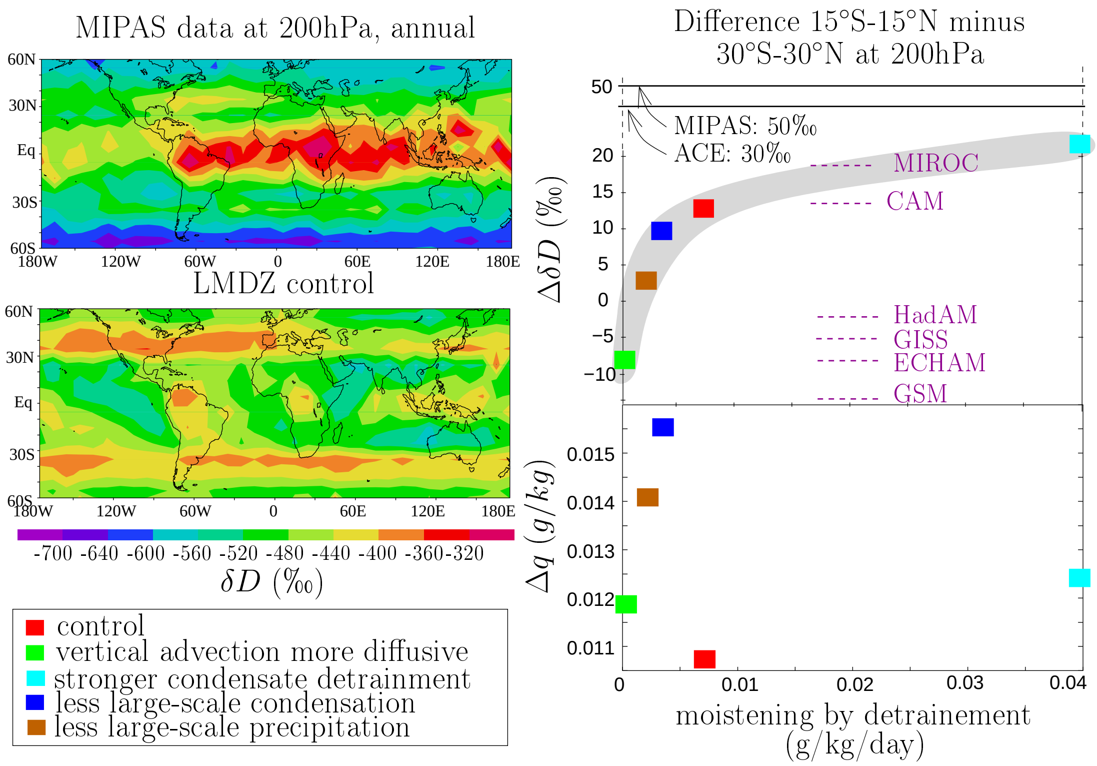

Figure 3.3: a) Annual-mean distribution of water vapor δD observed at 200 hPa by MIPAS. b) Same as (a) but simulated by LMDZ. The model outputs were co-located and averaging kernels were applied. c) Difference of δD ($\Delta\delta D$) between the tropical average (30°S-30°N) and the equatorial average (15°S-15°N) as a function of moistening by convective detrainment at 200 hPa simulated by LMDZ, for 5 sensitivity tests. $\Delta\delta D$ is a proxy for the contrast between the equatorial, deep convective region, where observed δD is enriched, and the subtropical regions, where observed δD is depleted. $\Delta\delta D$ observed by MIPAS and ACE and simulated by other GCMs from the SWING2 database are also shown. The moistening by detrainment is not available as a diagnostic in the SWING2 archive, so only the $\Delta\delta D$ values are shown. d) Same as (c) but for q instead of δD.

similarly as $\Delta\delta D$) and moistening by detrainement supports the advantage of δD to quantify the effect of convective detrainment (figure 3.3d).

In spite of the link between δD and the moistening by detrainement, none of the sensitivity tests is able to simulate a $\Delta\delta D$ as strong as observed. The relationship with moistening by detrainment seems to saturate. In one of the tests (cyan on figure 3.3c), the convective precipitation efficiency (i.e. the proportion of the condensate that falls as precipitation, ϵ_p, prescribed as horizontally constant in LMDZ) was divided by 2, leading the moistening by detrainment being multiplied by a factor of 5 compared to the control version. Even in this extreme test, $\Delta\delta D$ remains limited to 20‰ only. This suggests that simply tuning parameters is not sufficient to capture the observed δD distribution.

We note that the range of $\Delta\delta D$ values for the 6 other GCMs coincides with the range of $\Delta\delta D$ values simulated by the different model versions of LMDZ (figure 3.3c), suggesting that our sensitivity tests allow us to sample the diversity of behaviors in GCMs. A common process may be missing in all models.

The missing effect of updraft velocity on precipitation efficiency

To increase $\Delta\delta D$, we would need to decrease ϵ_p in the most convective regions only. To do so, the convective precipitation efficiency ϵ_p, prescribed as horizontally constant in GCMs, should vary with convective intensity. We expect that when updrafts are more vigorous, the fall speed of ice

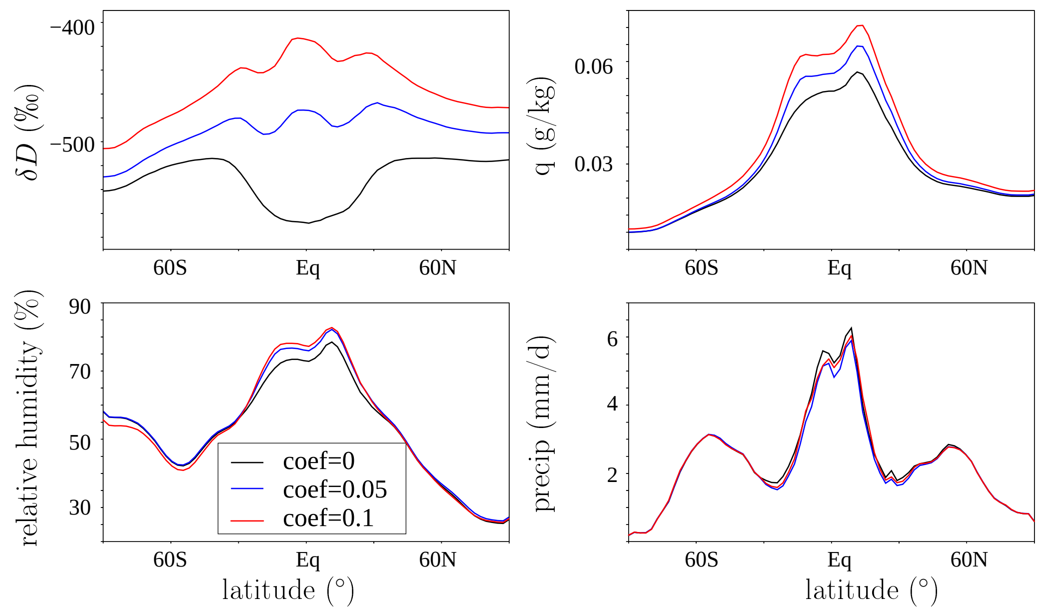

Figure 3.4: a) Annual, zonal mean δD at 200 hPa simulated by LMDZ for 3 values of $coef$. b) Same for q.

crystals is smaller, so a larger proportion of condensate should detrain rather than precipitate. So ϵ_p should be smaller. We hypothesize that neglecting this effect in GCMs is responsible for the difficulties to simulate upper-tropospheric δD.

To test this hypothesis, we implemented this effect in LMDZ. In LMDZ ϵ_p varies linearly with height from 0 at cloud base to ϵ_p^{max} at cloud top, where ϵ_p^{max} is a global constant. To simulate the effect of updraft velocity on ϵ_p, we replace ϵ_p^{max} by $\epsilon_p^{max,eff}$ calculated as:

$$\epsilon_p^{max,eff} = \epsilon_p^{max} - coef \cdot \sqrt{CAPE}$$

where $CAPE$ is the convective available potential energy. $\sqrt{CAPE}$ is an approximation for updraft velocity ([Williams and Renno, 1993]). $coef$ is a tunable parameter: when set to 0, $\epsilon_p^{max,eff} = \epsilon_p^{max}$ as in the control version. As $coef$ increases, the impact of updraft velocity on ϵ_p increases.

Although the impact of this modification on simulated upper-tropospheric q is small, the impact on upper-tropospheric δD is very large (figure 3.4). With $coef$ set to 0.1, $\Delta\delta D$ reaches 30‰, consistent with observations.

Therefore, upper tropospheric water vapor δD measurements suggest that in all models, the sensitivity of precipitation efficiency to updraft velocity is missing. Adding this process is the only way to capture the observed δD distribution. The global impact of this modification remains to be studied.

3.4 Evaluating the convective mixing depth

[Sherwood et al., 2014] has shown that the relative proportion of shallow to deep convective mixing in the tropical atmosphere simulated by climate models was a key factor to determine the cloud response to sea surface temperature changes, and thus the climate sensitivity. Models with the strongest proportion of shallow convective mixing are the most sensitive. They split this proportion into a sub-grid scale component, directly associated with convective parameterizations, and a large-scale component, representing the depth of the large-scale tropical overturning circulation. They estimate the large-scale component using vertical profiles of large-scale vertical velocity, $\omega(z)$, from reanalyses. Although reanalyses are constrained by a variety of observations, $\omega(z)$ remains a model product, since it has to balance the latent heating generated by shallow and deep convective parameterizations.

Here we attempt to provide an independent, observation-based constrain for the depth of the large-scale component of tropical mixing using water vapor isotope measurements. Indeed, both observations and model results show that the water vapor δD in the lower middle troposphere is sensitive to the depth of deep convection (section 2.1.5, [Lacour et al., 2016]). Since to first order in cloudy regions latent heat release by condensation balances the adiabatic cooling by large-scale ascent ([Emanuel et al., 1994]), the convective depth is tied to the $\omega(z)$ profile. The amount effect, defined here as the decrease of lower-tropospheric water vapor δD when the regime is more ascending, is weak when $\omega(z)$ peaks in the lower troposphere, whereas it is strong when $\omega(z)$ peaks in the upper troposphere.

Does this property linking the shape of $\omega(z)$ to the magnitude of the amount effect also holds when inter-comparing SWING2 models? SWING2 models show a wide spread in their simulation of the amount effect (figure 3.5a). In some models, such as LMDZ-free, the amount effect is weak, i.e. simulated δD at 600 hPa decreases weakly when the large-scale velocity is more ascending (figure 3.5a, black). In average over the regimes of strong ascent, these models happen to simulate $\omega(z)$ profiles peaking in the lower troposphere (figure 3.5b). In other models, such as GSM, the amount effect is strong (figure 3.5a, purple). These models happen to simulate $\omega(z)$ profiles peaking in the upper troposphere. There is a significant relationship between the magnitude of the amount effect and bottom-heaviness the of $\omega(z)$ profiles (figure 3.5c), consistent with the expected effect of convective depth on the magnitude of the amount effect. Therefore, we propose that the magnitude of the amount effect, observed by IASI and TES, provides an independent observational constrain for the depth of the tropical mixing.

This study has not been published yet.

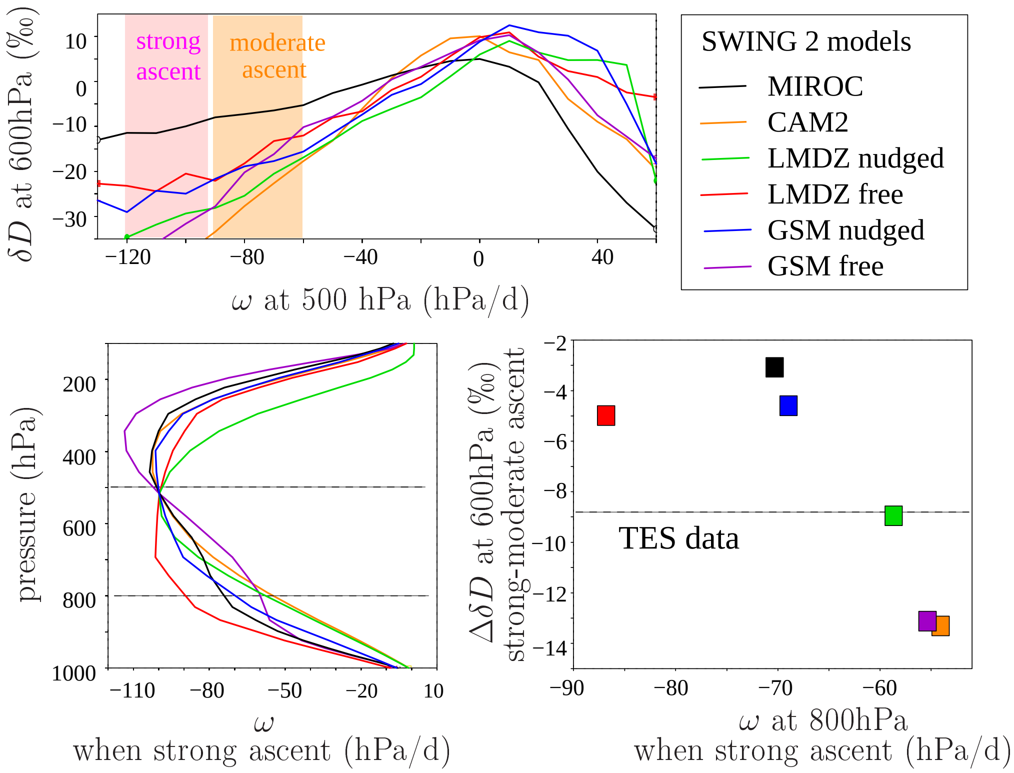

Figure 3.5: a) Monthly water vapor δD at 600 hPa as a function of monthly large-scale vertical velocity at 500 hPa, over the tropical oceans, for 6 SWING2 simulations. b) Shape of $\omega(z)$ profiles in average over the regimes of strong ascent, for the same SWING2 models. c) Proxy for the magnitude of the amount effect (difference of water vapor δD at 600 hPa between regimes of strong and moderate ascent, as defined on panel a) and a proxy for the bottom-heaviness of $\omega(z)$ profiles (ω at 800 hPa in average over regimes of strong ascent) for the same SWING2 models. The magnitude of the amount effect observed by TES is also shown.

Chapter 4

Using water isotope measurements to evaluate the capacity of climate models to simulate climate variations

To assess our credibility in climate change projections, observational tests of the models are necessary ([Hall and Qu, 2006, Boé et al., 2009, Brient and Bony,]). Testing the capacity of climate models to represent climate change in response to an external forcing in the context of past climates is a possible solution ([Braconnot et al., 2007, Schmidt, 2010, Schmittner et al., 2011, Braconnot et al., 2012, Hargreaves et al., 2012, Schmidt et al., 2014]). If a model is better for the past, is it more credible for the future? To address this question, we investigate links between the model behavior for past climates and for global warming projections. For example, links between tropical-mean temperature change during the last glacial maximum (LGM) and the future climate sensitivity ([Hargreaves et al., 2012]) have been evidenced. In section 4.1, I report my contribution to this topic by showing a link between precipitation changes in the future and during the Mid-Holocene.

Even if the similarity between physical processes involved in past and future climate changes remain to be verified ([Yoshimori et al., 2009, Yoshimori et al., 2011]), the effective use of paleo-climates to constrain future climate changes relies on the existence of precise and well dated paleo-climatic archives ([Hargreaves et al., 2013]). Traditional paleo-climatic information, such as pollen data, is sparse over most tropical continental regions ([Farrera et al., 1999, MARGO project members, 2008, Bartlein et al., 2010]). Could we use paleo archives of precipitation isotopic composition for this purpose? In section 4.2, I discuss the paleo-climatic interpretation of paleo-climatic isotopic archives in tropical regions.

4.1 Using paleo-climate archives to evaluate the credibility of future projections

4.1.1 Necessary conditions

To use paleo-climate archives to evaluate the credibility of future projections, 3 conditions are necessary:

1. there should be an inter-model statistical link between the behavior for past climates and for future projections,
2. this statistical link should be explained by robust physical processes,
3. paleo-climate data should be available to constrain the model behavior for past climates.

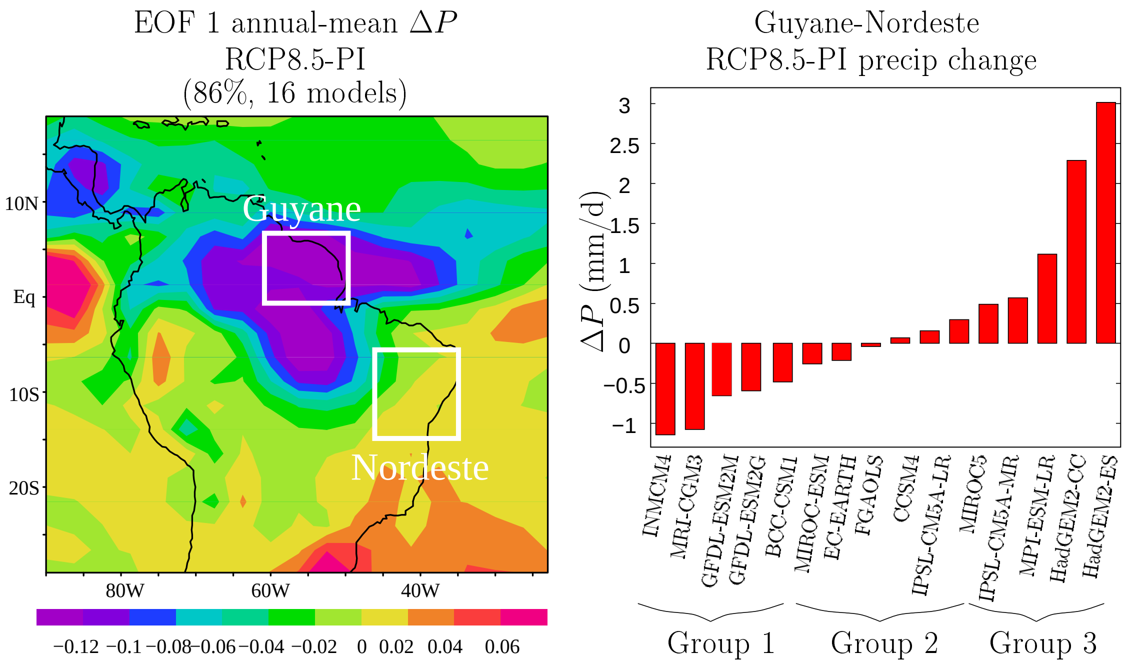

Figure 4.1: a) EOF analysis of the multi-annual-mean precipitation change ΔP between the RCP8.5 scenario and pre-industrial simulations, for 16 CMIP5 models. The number of models is used instead of the traditional time dimension of the EOF. The map of the 1st EOF shows a dipole pattern between the Northern and the Eastern parts of South America. We define the "Guyana" region and the "Nordeste" regions as highlighted with white boxes. b) RCP8.5 minus pre-industrial ΔP difference between the Guyana and Nordeste region, for the difference models. Positive values indicate that in RCP8.5, precipitation increases in Guyana and decrease in Nordeste, reflecting a Northward shift in the ITCZ. Conversely, negative values indicate that in RCP8.5, precipitation increases in Nordeste or decreases in Guyana, reflecting a Southward shift in the ITCZ. Models are classified into 4 groups based on these values.

The CMIP5 (Coupled Model Intercomparison Project version 5, [Taylor et al., 2012]) offers a unique opportunity to check the first two conditions: for the first time, the same models are used for present, past and future simulations.

4.1.2 Pilot study on precipitation changes in tropical South America

I present here my contribution to checking the first condition, focusing on precipitation changes in tropical South America. Results were reported in [Schmidt et al., 2014].

We use CMIP5 simulations for pre-industrial (PI), the most pessimistic climate change scenario (RCP8.5) and Mid-Holocene (MH) simulations.

To extract the most common patterns of future precipitation change, we performed an EOF (empirical orthogonal function) analysis over the different models for multi-annual mean of precipitation change (ΔP) from pre-industrial to RCP8.5. The first EOF shows a dipole pattern between Northern and Eastern South America (figure 4.2a), indicating that some models project an increase in precipitation in the Northern part and a decrease in precipitation in the Eastern part whereas other models simulate the opposite. Hereafter we refer to the Northern and Eastern part as "Guyana" and "Nordeste".

The precipitation change difference between "Guyana" and "Nordeste", ΔP_{G-N}, exhibits a very large spread across models, from -1 mm/d to +3 mm/d (figure 4.2b). We classify models into 3 groups based on ΔP_{G-N}. Group 1 models project a precipitation decrease in Guyana and a precipitation increase in Nordeste (figure 4.2a), consistent with a Northward shift of the ITCZ (Inter-tropical Convergence Zone). In contrast, group 3 models project a precipitation increase in

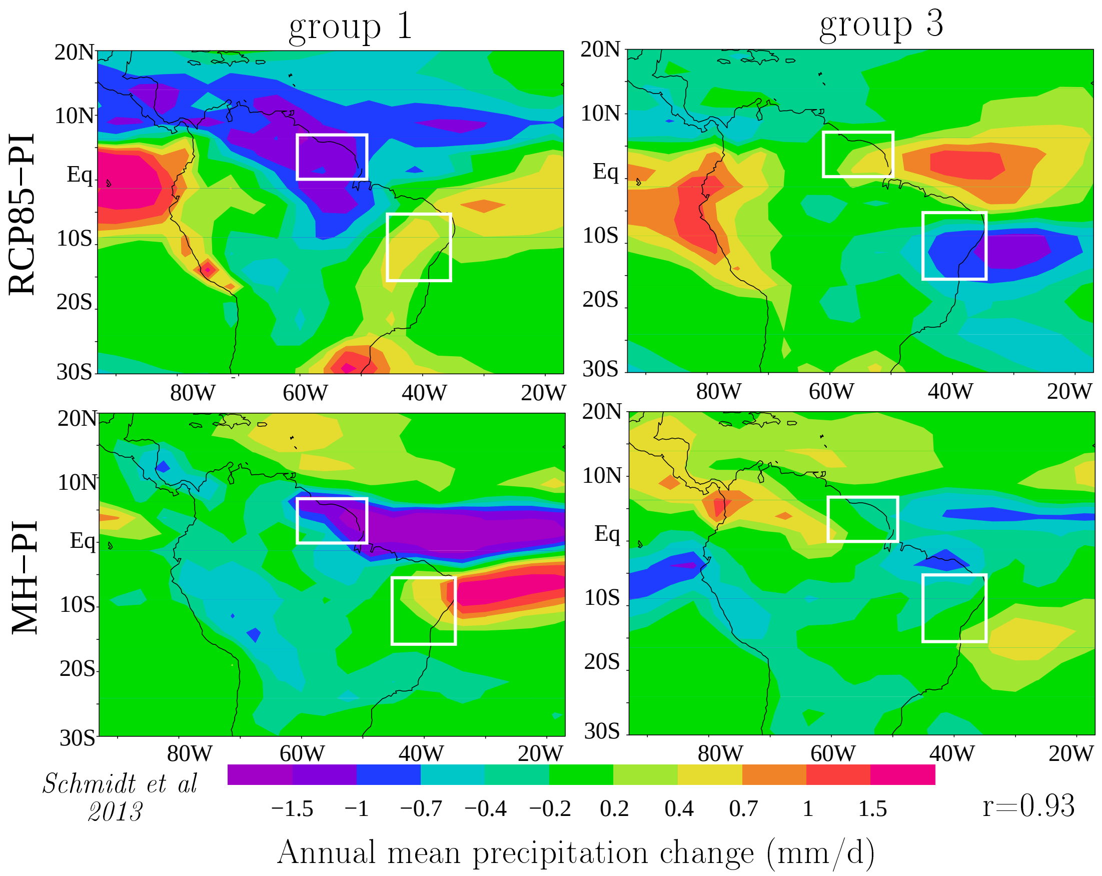

Figure 4.2: (a-b) Multi-annual-mean precipitation change ΔP between the RCP8.5 scenario and pre-industrial simulations, for models classified as group 1 (a) and group 3 (b). The classification is explained in figure 4.1. (c-d): as a-b but for ΔP between the Mid-Holocene and pre-industrial.

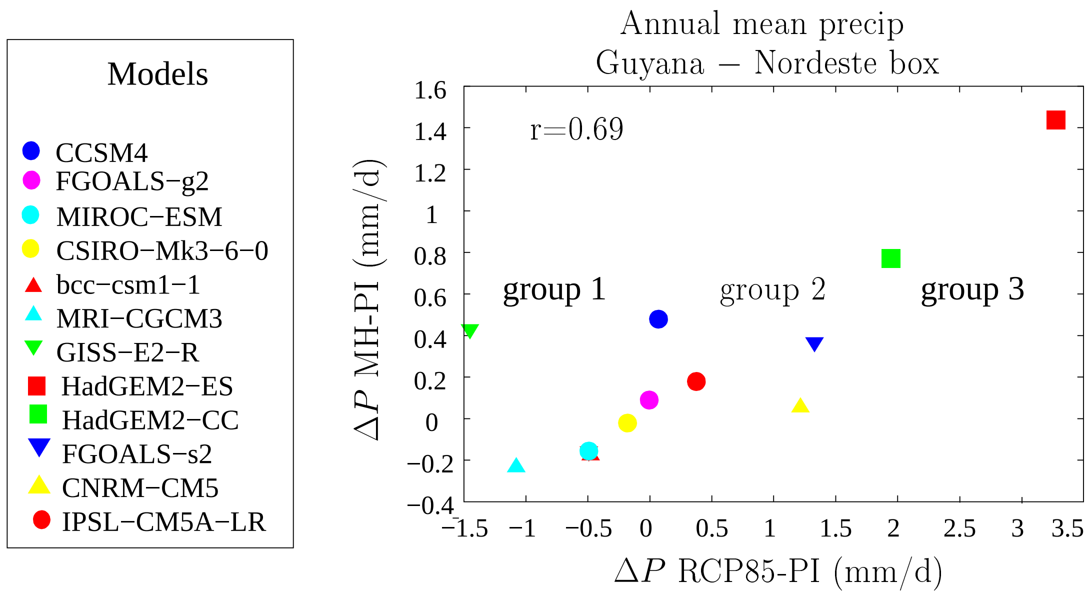

Figure 4.3: e) "Guyana" −"Nordeste" difference of ΔP from pre-industrial to Mid-Holocene as a function of the "Guyana" −"Nordeste" difference of ΔP from pre-industrial to RCP8.5.

Guyana and a precipitation decrease in Nordeste (figure 4.2b), consistent with a Southward shift of the ITCZ.

For the MH, group 1 models simulate a precipitation increase in Nordeste and a precipitation decrease in Guyana (figure 4.2c), similar to the precipitation change projected for the future climate. In contrast, group 3 models simulate a tripole pattern that reflects a widening of the ITCZ (figure 4.2d). Therefore, the behavior of a model for the MH determines its behavior for the future climate. Tob be more quantitative, we plot ΔP_{G-N} for the MH as a function of ΔP_{G-N} for the future climate (figure 4.2). We find a significant correlation (r=0.69), suggesting that paleo archives of ΔP_{G-N} for the MH could constrain ΔP_{G-N} for the future climate.

Physical processes explaining this common behavior between MH and future climate still need to be understood. To this aim, the decomposition methodology into dynamical and physical components ([Bony and Bellon, 2013]) and a moist static energy budget ([Oueslati et al., 2016]) could be applied (proposal A.9).

Are group 1 or group 3 models more realistic? Group 3 models more frequently show the double ITCZ problem for present-day simulations. However, the mechanisms linking the double-ITCZ problem for present-day and the behavior for past future changes still need to be understood.

This "pilot study" is encouraging about the potential of paleo precipitation archives to provide constrains on future precipitation changes. In the future, we plan to extend this to all tropical regions and to deepen our understanding of physical processes (proposal A.9).

4.2 Interpretation of paleo-climatic isotopic archives in the tropics

4.2.1 Introduction

The isotopic composition of precipitation can be archived in many archives such as ice cores ([Thompson et al., 2000]) in speleothems ([McDermott, 2004]), shelly fossils ([Eagle et al., 2013]), cellulose ([McCarroll and Loader, 2004]) or ground waters ([Gasse, 2000]), leaf waxes ([Tierney et al., 2008, Collins et al., 2013]). In the tropics however, the interpretation of water isotopic records has been debated. They were first interpreted in terms of **temperature** ([Thompson et al., 1989, Thompson et al., 1998]), as was the case for polar ice cores ([Jouzel, 2003]). This interpretation is supported by the fact that all tropical ice core records show a common depletion at LGM ([Thompson et al., 2000], figure 4.4a), suggesting that a global-scale feature is recorded and supporting this interpretation. In addition, the amplification of temperature changes with altitude may explain the strongest signals in high altitude records ([Poulsen and Jeffery, 2011]).

In contrast, most recent studies have interpreted tropical isotopic records as **precipitation** proxies ([Vuille et al., 2003]), consistent with the effect of convection on water vapor and precipitation ("amount effect", section 2.1). This effect has been observed in present-day precipitation at the spatial ([Dansgaard, 1964, Rozanski et al., 1993]), seasonal ([Dansgaard, 1964, Rozanski et al., 1993, Vimeux et al., 2005]), intra-seasonal ([Vimeux et al., 2011, Risi et al., 2008b, Gao et al., 2011]) and inter-annual time ([Hoffmann, 2003]) scales. Convection may therefore also play a role at the paleo-climatic scale. Finally, some authors point out the importance of the large-scale circulation and origin of air masses ([LeGrande and Schmidt, 2009]).

The goal of this study is to try and understand what controls isotopic variations recorded in tropical archives: temperature, precipitation, a bit of both, or none of them?

4.2.2 LMDZ paleo simulations

To this purpose, a series of LMDZ simulations were performed for 11 different "climates". Each climate is characterized by sea surface temperatures (SSTs), sea ice cover, ice sheets, orbital parameters and various greenhouse gas concentrations. The 11 climates include Last Glacial Maximum (LGM), Mid-Holocene (MH), CO_2 doubling with SSTs projected by different CMIP5 models, last inter-glacial... In addition, some of these climates were simulated with 4 different model physics (the same sensitivity tests as described in section 3.2), and some climates were also simulated with the zoom functionality to refine the resolution over South America or over the Tibetan Plateau.

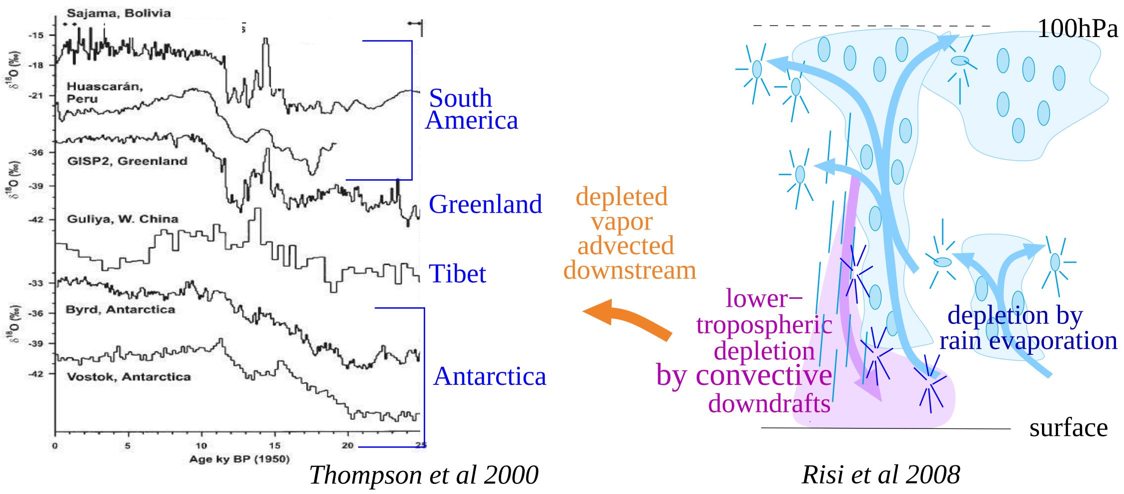

Figure 4.4: a) Ice core isotopic records on different continents, from 25 000 years to today. All records show the LGM depletion. b) Processes by which convection impacts the isotopic composition recorded in tropical archives. Convective downdrafts and rain re-evaporation deplete the lower-tropospheric water vapor. This depleted water vapor is advected downstream, imprinting a depleted signature into the subsequent precipitation.

These two regions concentrate most of tropical ice core records and the mountainous conditions require high resolution ([Yao et al., 2013, Eagle et al., 2013]).

As an example, we evaluate the LGM simulations using available paleo-climatic archives of precipitation $\delta^{18}O$ ($\delta^{18}O_p$). LMDZ simulates more depleted $\delta^{18}O_p$ during the LGM, and the depletion is all the largest as the elevation is high, consistent with the altitude amplification effect described by [Poulsen and Jeffery, 2011]. However, the depletion observed in Andean ice cores is strongly underestimated, even in the high-resolution simulation. The difficulty to simulate the LGM depletion as large as observed in the tropics is common to all GCMs ([Jouzel et al., 2000, Werner et al., 2001, Jasechko et al., 2015]) and the cause remains a mystery.

4.2.3 Interpreting $\delta^{18}O_p$ changes using a simple Rayleigh distillation model

To interpret $\delta^{18}O_p$ changes observed in paleo-climatic archives, the simulated $\delta^{18}O_p$ changes at some sites is decomposed into different components using a simple Rayleigh distillation model. This method is described in [Botsyun et al., 2016, Gao et al., 2016]:

$$R_p = (R_p - R_v) + R_v$$

and

$$R_v = R_{v0} \cdot \left(\frac{h \cdot q_s(T)}{h_0 \cdot q_s(T_0)} \right)^{\alpha - 1} + r$$

where R_p and R_v are the water isotopic ratio in precipitation and in water vapor; h_0, T_0 and R_0 are arbitrary initial conditions (whose choice has little impact on the final results) for relative humidity, temperature and water vapor isotopic ratio; h and T are relative humidity and temperature at the site of interest, q_s is the saturation specific humidity as a function of temperature, r is a residual term.

Variations in R_p are decomposed into 4 terms:

1. a term representing variations in the precipitation-vapor difference, $R_p - R_v$, which reflects condensation and post-condensation processes,
2. a term representing variations in local temperature T,

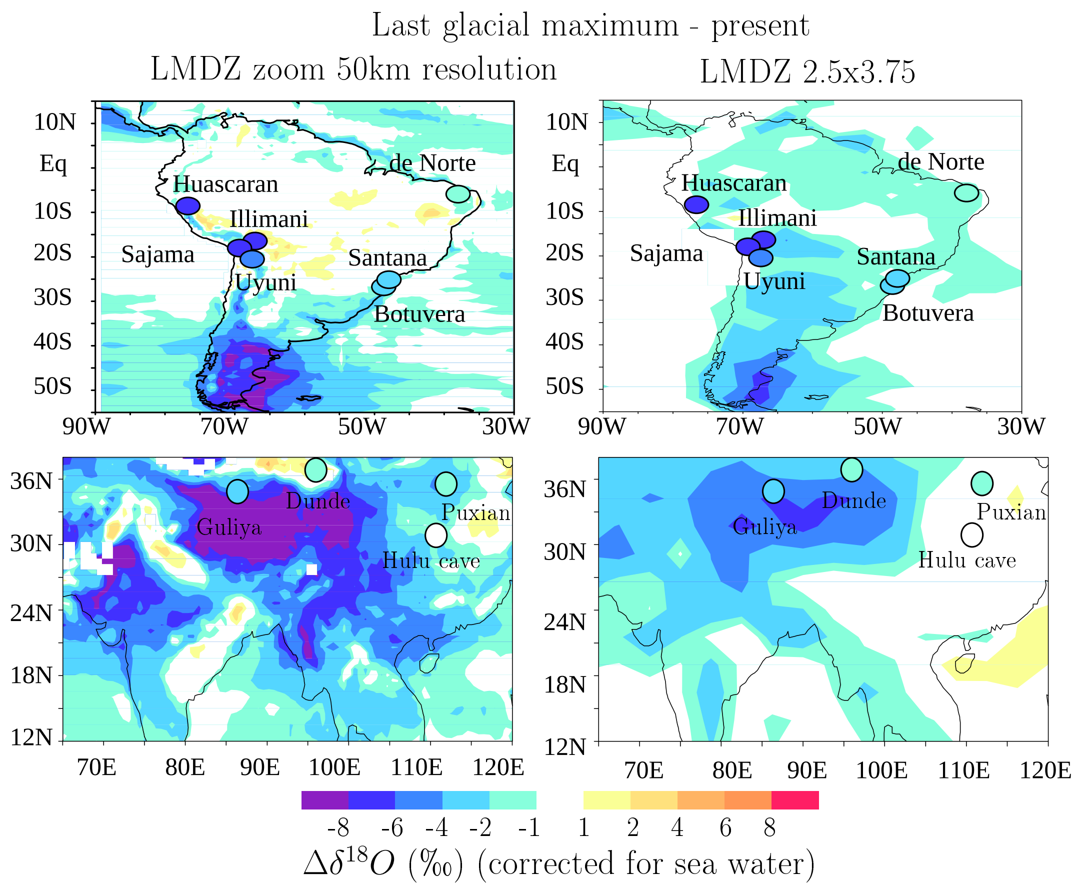

Figure 4.5: Evaluation of the simulated precipitation $\delta^{18}O$ changes from present-day to LGM in South America (a-b) and over the Tibetan Plateau (c-d) for the standard resolution simulation (a,c) and for the high-resolution simulations (b-d). Some available paleo-climatic archives are indicated for comparison.

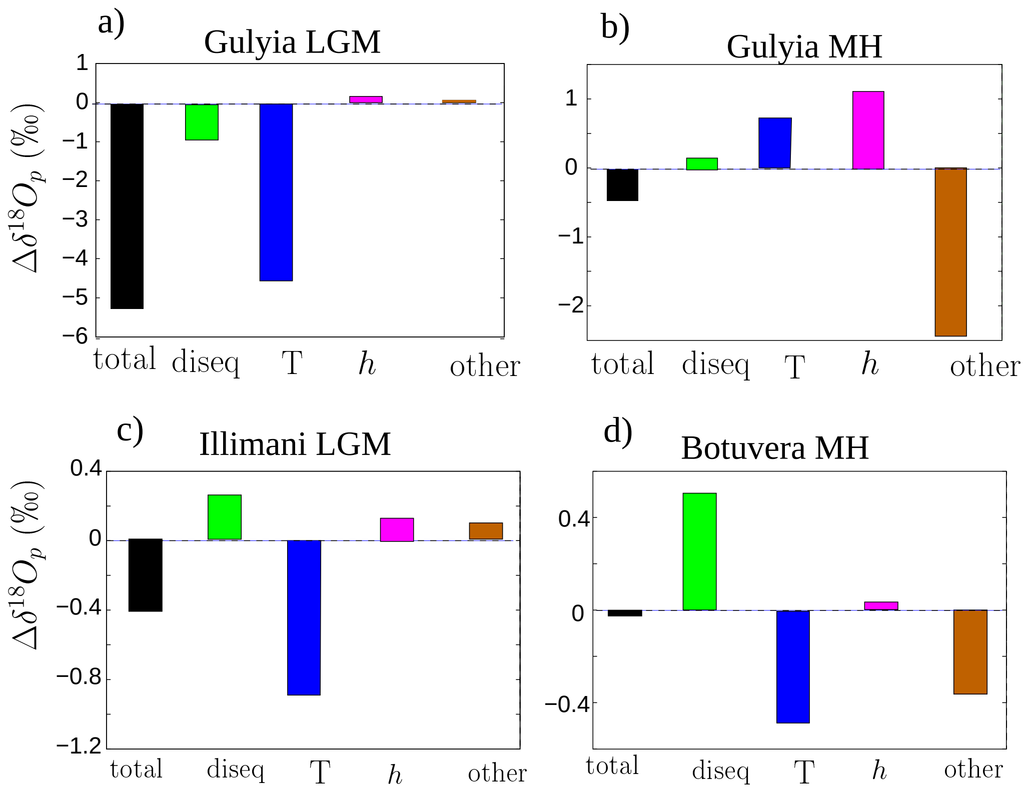

Figure 4.6: Decomposition of the total simulated $\delta^{18}O_p$ change (black) into 4 contributions: condensation and post-condensation effects (green), local temperature (blue), local relative humidity (pink) and a residual that includes the effect of deep convection along air mass trajectories (brown). The decomposition is shown for the Guliya ice core during the LGM (a) and the MH (b), for Illimani for the LGM (c) or for the Botivera cave for the MH (d).

3. a term representing variations in local relative humidity h,

4. a term representing variations in the residual term, r , encompassing changes in moisture source properties, deep convection and continental recycling along back-trajectories.

During the LGM, at most sites, the simulated $\delta^{18}O_p$ changes are dominated by the effect of local temperature changes (figure 4.6a,c). This supports the interpretation of $\delta^{18}O_p$ in terms of temperature ([Thompson et al., 2000]). In contrast, during the MH, the simulated $\delta^{18}O_p$ changes are dominated by local temperature, condensation and post-condensation processes, or by the residual term, depending on the sites (figure 4.6b,d). For example, for the Guliya ice core, the dominance of the residual term, which probably represents the impact of deep convective activity along monsoon flow trajectories, is consistent with several studies arguing for interpreting Asia ice cores in terms of past monsoon intensity ([Pausata et al., 2011]).

To conclude, what controls $\delta^{18}O_p$ changes may depend on the past climatic epoch of interest. **For epochs with large temperature changes, the temperature effect may dominate, but for epochs with smaller temperature changes, the convective effect may dominate**.

4.2.4 Inter-climate correlations with sea surface temperature and deep convection

Correlation between $\delta^{18}O_p$ in tropical ice cores and sea surface temperature

To check whether large-scale temperature changes impact the $\delta^{18}O_p$ in tropical ice cores at the paleo-climatic time scales, we calculated correlations between tropical-mean sea surface temperature (SST) changes and $\delta^{18}O_p$ changes, across the 11 climates. For the standard version of LMDZ, the correlation is significant in most regions of South America and Asia, even at low elevations (figure 4.7a,c). This supports the interpretation of tropical records in terms of temperature.

How robust is it with respect to model physics?

To check whether this results is robust with respect to the model physics, we compared different perturbed-physics versions of LMDZ. At the Guliya ice core for example, all perturbed-physics versions of LMDZ show a significant correlation with SST (figure 4.7b). The dominance of the temperature control is thus robust to the model physics in this region. In contrast, the Illimani ice core is in a "pocket" where the standard version of LMDZ simulates no correlation with SST. At Illimani, only one of the LMDZ versions simulates a correlation with temperature (figure 4.7d). This suggests that at some sites, **the effect of temperature is sensitive to the model physics**. For example, a possible impact of convective entrainment on simulated isotopic lapse rates has been eveidenced ([Tripati et al., 2014]).

Correlation between $\delta^{18}O_p$ in tropical ice cores and upstream precipitation

To check whether deep convection impacts the $\delta^{18}O_p$ in tropical ice cores at the paleo-climatic time scales, we calculated correlations between regional precipitation changes and $\delta^{18}O_p$ changes at a site of interest across the 11 climates. For the standard version of LMDZ, at Guliya ice core, there is a strong anti-correlation with precipitation in North Western India, up to -0.9 (figure 4.8a). This region is on the way of monsoon flow trajectories. The importance of deep convection upstream air mass trajectories is consistent with paleo-climatic interpretation of $\delta^{18}O_p$ in terms of precipitation and with the isotopic controls observed at the daily time scale at Lhasa ([Gao et al., 2013]). For the Illimani ice core, there is a very strong anti-correlation with precipitation in Eastern Brazil (figure 4.8c), below -0.9 in some places. This region is also on the way of monsoon flow trajectories. The importance of deep convection upstream air mass trajectories is also consistent with the isotopic controls observed at the daily time scale in Bolivia ([Vimeux et al., 2005]).

How robust is it with respect to model physics?

To check whether this results is robust with respect to the model physics, we plotted $\delta^{18}O_p$ changes as a function of precipitation averaged over the upstream region for different perturbed-physics versions of LMDZ (figure 4.8b,c). For both Guliya and Illimani, the correlation is very good for the standard version of LMDZ, but breaks down for other versions. Therefore, **the importance of upstream convection in controlling $\delta^{18}O_p$ changes is sensitive to the model physics**.

Summary

To summarize, depending on model versions, LMDZ can simulate either a dominance of temperature or of upstream convection in controlling $\delta^{18}O_p$ changes in paleo-climatic archives. What model version is the most credible? Could daily water vapor observations help us to discriminate between different model versions? This is what we propose to do in the future (proposal A.6).

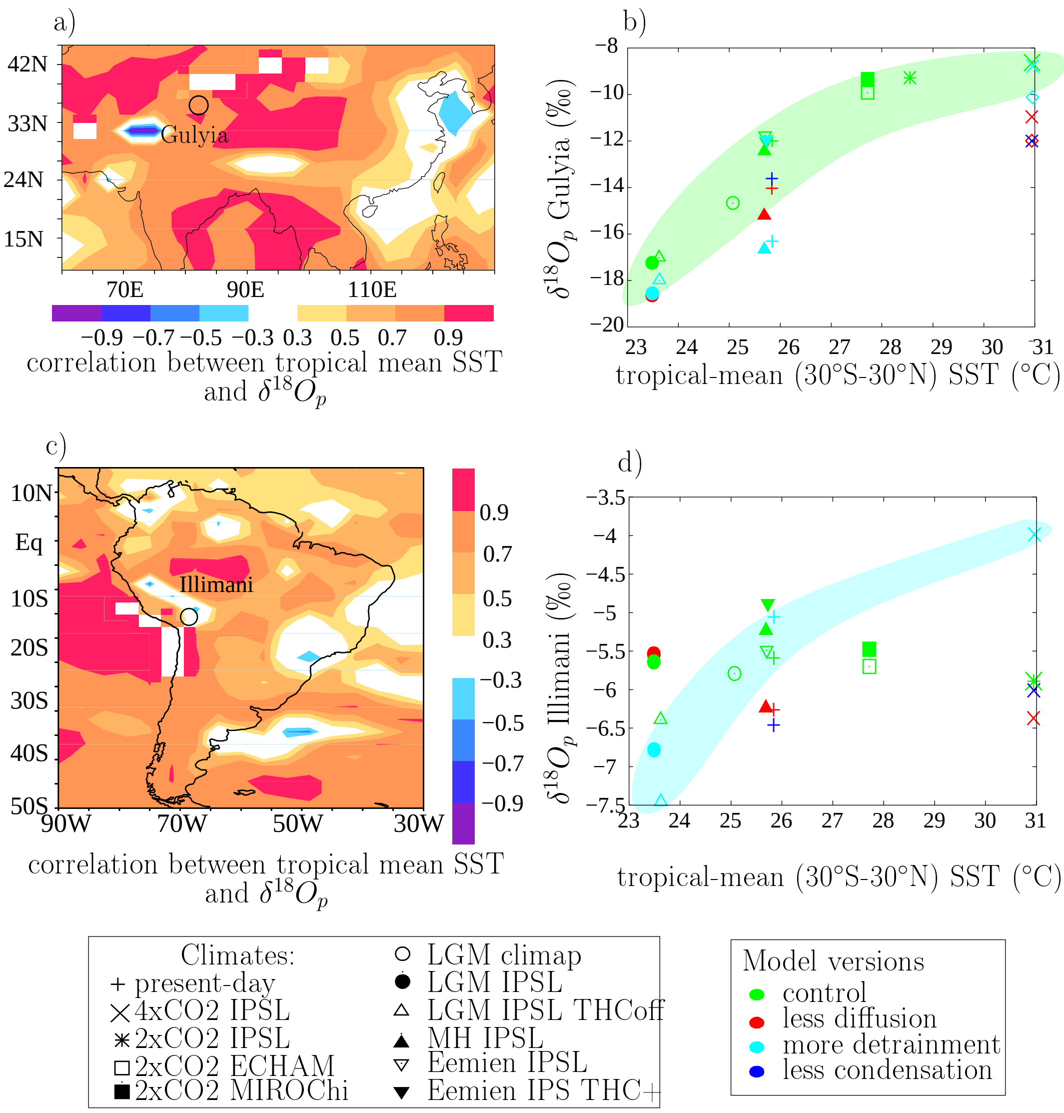

Figure 4.7: a) Correlation between tropical-mean sea surface temperature (SST) changes and simulated $\delta^{18}O_p$ in each grid box across the 11 climates for the standard version of LMDZ, over Asia. b) Simulated $\delta^{18}O_p$ change at Guliya ice core as a function of tropical-mean SST, for the 11 different climates (markers) and the 4 LMDZ versions (colors). c) Same as a but over South America. d) Same as b but at the Illimani ice core.

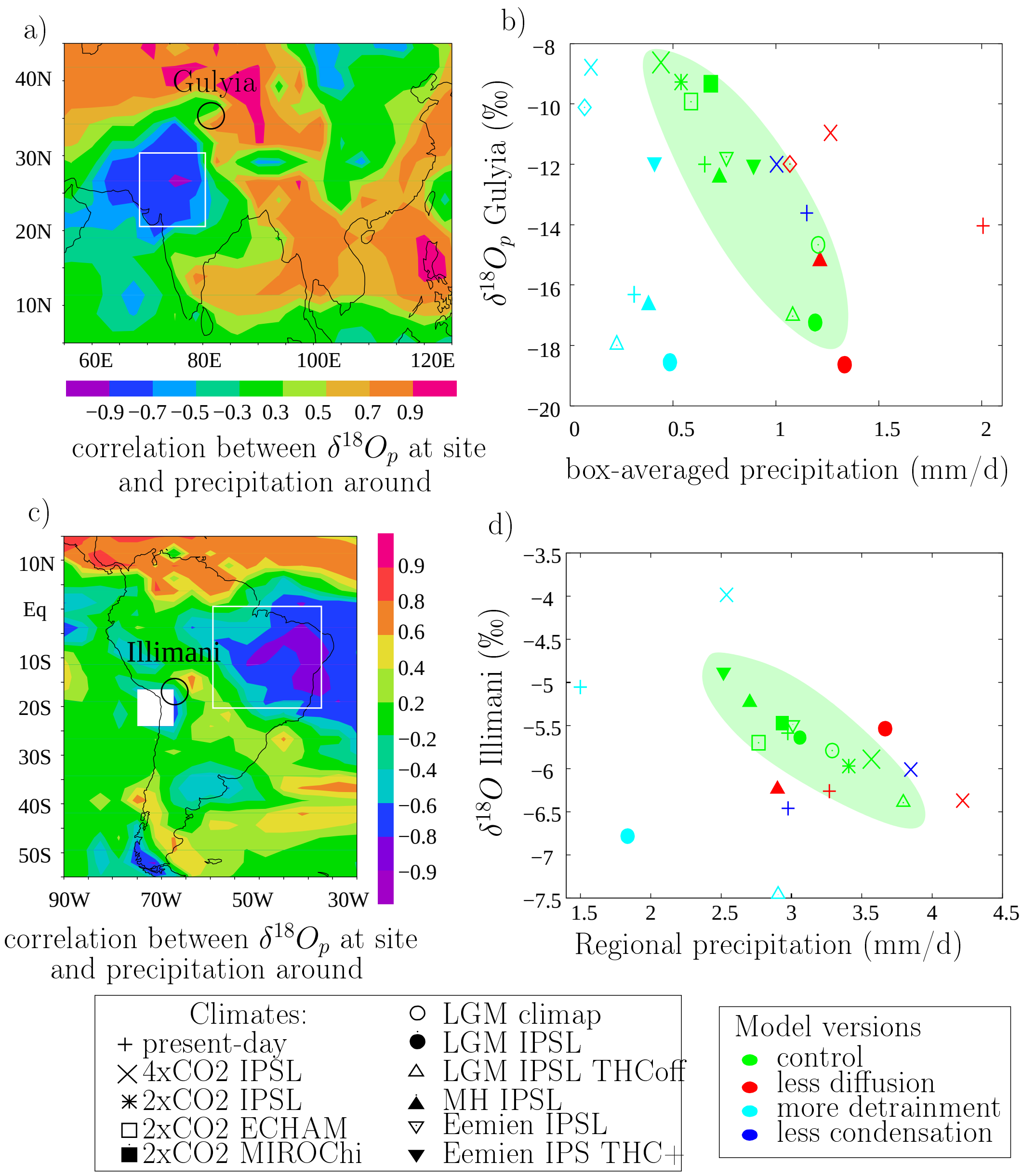

Figure 4.8: a) Correlation between the simulated $\delta^{18}O_p$ at the Guliya ice core site and the precipitation rate in each grid box across the 11 climates for the standard version of LMDZ. b) Simulated $\delta^{18}O_p$ change at Guliya ice core as a function of precipitation change averaged over the white box, for the 11 different climates (markers) and the 4 LMDZ versions (colors). c) Same as a but for the Illimani ice core site. d) Same as b but at the Illimani ice core.

Chapter 5

Conclusion and perspectives

5.1 Conclusion

Water isotope measurements are potentially useful to quantify the relative importance of moistening and dehydrating processes (e.g. mixing, convective processes, continental recycling), to detect and understand biases in models, and to evaluate and improve models.

The main difficulty is to go beyond the demonstration that water isotopes are "potentially useful", and actually use water isotope measurements quantitatively. Studies actually proposing model improvement based on water isotopes (chapter 3) are few, and even among those, none have lead to effective changes in climate models. We still have a long way to go before water isotopes become part of the model development process; I think that this is because we, as isotopists, still have a lot of work to do to demonstrate the real added value of water isotopes.

A major limiting factor in doing so is that **what controls the water isotopic composition is not fully understood yet**. I think that more efforts should be put into understanding what controls water vapor isotopes, before we pretend to use them.

When trying to gain better understanding of isotopic controls, another limiting factor is the big **gap between complex numerical modeling and the simple $q-\delta D$ theoretical framework** (section 1.3.3). Intermediate models would be necessary.

A great amount of measurements and modeling tools are now available, which gives hope that there will be a lot of progress in isotopic research in the years to come. Unfortunately, many of these measurements and modeling tools have been under-exploited so far. A limiting factor is that we are missing human work force to analyze all the already-available measurements and simulations.

Finally, consolidating the link between isotopists and model developers is crucial. It ensures that isotopic research addresses questions that are relevant to model developers and that isotopic constraints translate into model parameterization improvement. My position at LMD, near the LMDZ development team, is ideal for this purpose, although I would like to make a better job in the future collaborating with the development team.

5.2 Perspectives

Here are some suggestions to try to overcome the limiting factors described above, together with my plans to contribute to this effort. I give a series of 9 proposals for master's students, PhD students, post-docs or myself, to make my plans more concrete.

5.2.1 Better understand the isotopic response to convective and cloud processes

First, I plan to consolidate the understanding of the isotopic response to convective and cloud processes (section 2.1). To understand how these processes impact isotopic spatio-temporal variations with time and space (section 2.1), I plan to document isotopic variations associated with different phases of the convective life and with different degrees of convective aggregation, by co-locating various satellite datasets (proposals A.1 and A.2). One possible application is to identify what are the moistening and drying processes responsible for the observed effect of convective aggregation on tropospheric humidity ([Tobin et al., 2012]).

The impact of rain re-evaporation on water vapor isotopic composition remains uncertain (section 2.1.3). I would like to combine the surface meteorological data, satellite observations and continuous isotopic measurements available at Niamey in the Sahel ([Tremoy et al., 2012]) to quantify the effect of rain re-evaporation on both q, water vapor δD and d-excess in cold pools of squall lines (proposal A.3). One possible application is to quantify the contribution of rain re-evaporation to intra-seasonal variations in Sahelian humidity ([Poan et al., 2013]).

To evaluate the isotopic response to convective and cloud processes simulated by GCMs, I'd like to use high-resolution (CRM or LES) simulations, where convection is explicitly represented (proposal A.4). Such simulations are ideal for a detailed process study, and could be compared with GCM or SCM simulations using the conditional sampling methodology ([Couvreux et al., 2010]). Inter-comparing SCM from different GCMs, by including isotopes into some GEWEX case studies or by extending the SWING initiative to SCM simulations, would be very interesting.

5.2.2 Better understand what controls the isotopic composition at the large-scale

From a more fundamental point of view, to better understand what controls the isotopic composition at the large-scale, bridging the gap between complex numerical modeling by GCMs and the simple $q - \delta D$ framework would be useful. In an attempt to do so, I'd like to investigate processes along air mass trajectories in the LMDZ GCM. The Lagrangian framework would allow for a easier comparison with the theoretical lines in the $q - \delta D$ framework. Based on this study, a new set of lines, representing better the diversity and complexity of atmospheric processes, will be develop (proposal A.5).

With paleo-climatic applications in mind, to check to what extent our understanding of isotopic controls at the daily scale helps us understanding the isotopic controls at the paleo time scale, processes at play at the two time scales will be compared in LMDZ simulations (proposal A.6). Inter-comparing different GCMs, by including isotopes into CMIP6 or by extending the SWING initiative to paleo simulations, would be very useful for this purpose.

5.2.3 Work more closely with model evaluation and improvement issues

The goal of the CONVISO project was to help evaluate and improve the simulation of the MJO by GCMs. Most of the time of this project has been used so far in understanding processes controlling the isotopic composition (section 2.1.7), which is a necessary first step. Now, to what extent could water isotope measurements provide process-oriented diagnostics that can predict the realism of the MJO simulation? I'd like to address this issue by comparing different sensitivity tests with LMDZ (proposal A.7), and, ideally, comparing different isotopic GCMs as well. Extending the SWING initiative to provide daily outputs would be useful for this purpose.

When trying to apply water isotope measurements to evaluate models, we need to keep in mind that for many processes, water isotopes may not be the best tools. Combining different isotopic and chemical tracers, such as CO, O_3, ^{10}Be (proposal A.8), would be maximize the probability to develop useful constraints on parameterizations ([Folkins et al., 2006, Liu et al., 2009]).

Finally, I think comparing simulated and observed paleo-climatic variations is a very powerful test the model capacity to respond to climate forcing, in a way that integrates the effect of all parameterizations. I'd like to continue to investigate the link between past and future precipitation variations and develop pertinent diagnostics, possibly based on water isotopes (proposal A.9).

Appendix A

Some proposals for future students, post-docs or myself

A.1 Moistening and dehydrating processes as a function of convective aggregation

Context and science questions

A significant part of the spread in future precipitation projections can be attributed to the representation by climate models of atmospheric convection. It is thus important to understand the role of convection on its environment. This role may depend on the degree of aggregation. Convective systems can be either aggregated as large meso-scale systems (figure A.1a) or isolated (figure A.1b). [Tobin et al., 2012] developed a diagnostic to quantify the degree of aggregation of convection, and showed that more aggregated systems are associated with lower tropospheric relative humidity. Why? Do convective processes depend on the degree of aggregation of convection? Does convection moistens its environment more when it is less aggregated, and if so, by what mechanism? By stronger convective detrainment? Or more efficient re-evaporation of the raindrops?

What impact does this effect has on the climate and its variability? The degree of aggregation of convection is not considered in current climate models ([Tobin et al., 2013]). To what extent does this lead models to ignore some important moisture-convection feedbacks?

The isotopic composition of water vapor, observed by satellite, can give some information about moistening and dehydrating process. For example, convective detrainment enriches vapor water while moderate re-evaporation of raindrops depletes it. The goal of this study is to document the impact of convective aggregation on the tropospheric specific humidity (q) and water vapor δD in the environment and to link it with processes by which aggregation impacts tropospheric humidity.

Where we stand

Tropospheric profiles of q and water vapor δD observed by TES were co-located with the degree of aggregation at the $10° \times 10°$ scale diagnosed by [Tobin et al., 2012] and with precipitation observed by TRMM. For a given precipitation rate, more aggregated convection is associated with drier free troposphere (consistent with previous studies, [Tobin et al., 2012]), with more enriched lower troposphere (suggesting less rain evaporation) and with more depleted middle and upper troposphere (suggesting less condensate detrainment or more subsidence) (internship by Natacha Legrix).

Although LMDZ does not explicitly represents convective aggregation, when nudged to reanalyses winds, it captures most of these features to some extent. This suggests that at least some of the observed q and δD signal is not a direct consequence of convective aggregation. Convective aggregation may be associated with some meteorological situations that impact q and δD.

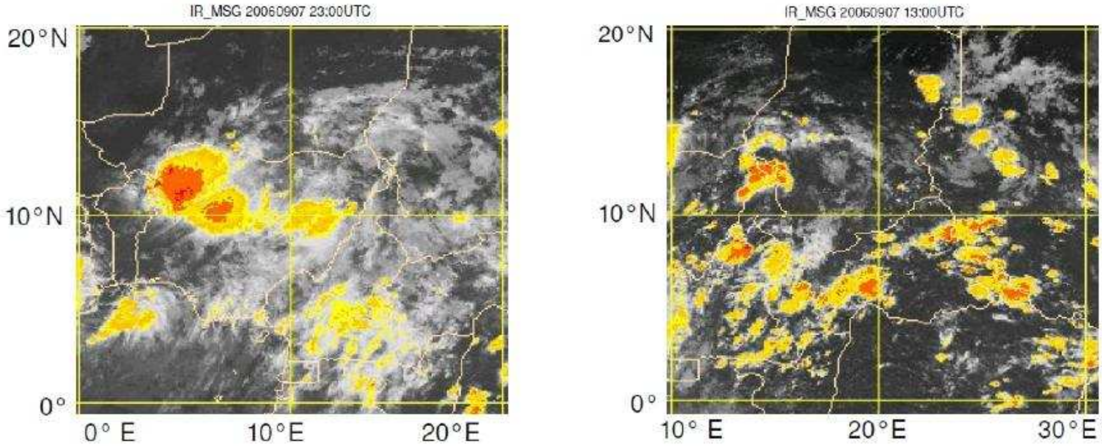

Figure A.1: Satellite images illustrating different states of convective aggregation: aggregated (a) and scattered (b).

Road map

- To isolate the direct impact of convective aggregation on observed q and δD, we will investigate the influence of other factors, such as large-scale vertical velocity profiles, sea surface temperature, proximity with land or moisture origin.
- To assess whether convective aggregation impacts its environment or whether meteorological situation favors some states of convective aggregation, we will look at temporal evolution of convective aggregation, q and δD.
- To check the robustness of q and δD signals and to document full tropospheric profiles, we will extend the analysis to the IASI, GOSAT and MIPAS datasets.
- Once the real impact of convective aggregation on tropospheric q and δD profiles is documented, we will investigate the convective processes at play and conclude on the mechanisms linking convective aggregation to tropospheric humidity.
- Finally, we will discuss the consequence for climate models to neglect the effect of convective aggregation.

A.2 Moistening and dehydrating processes as a function of the convective life cycle

Context and science questions

A significant part of the spread in future precipitation projections can be attributed to the representation by climate models of atmospheric convection. It is thus important to understand the role of convection on its environment. Meso-scale convective systems (MCS) often follow a typical life cycle, in which they initiate as small isolated cumulonimbus, organize into larger convective clusters that may propagate (figure A.1b), and then dissipate. Convective and cloud processes depend on the phase of the life cycle ([Sherwood and Wahrlich, 1999, Bouniol et al., 2016]). How do moistening and dehydrating processes vary with the phase of the life cycle? Can climate model capture these different processes?

Since the isotopic composition of water vapor can give some information about moistening and dehydrating process, the goal of this study is to document the tropospheric specific humidity (q) and water vapor δD along the convective life cycle and to link it with convective processes at play in the different phases.

Where we stand

In collaboration with R. Roca, we used MCS properties (duration, track, size and brightness temperature along their track) derived from geostationary satellites and coputed using a tracking algorithm ([Fiolleau and Roca, 2013a]) to build composite life cycles ([Fiolleau and Roca, 2013b]). TES, GOSAT and IASI q and δD observations in the vicinity of MCSs were selected and composited for different phases of the life cycle and for different precipitation rates as measured by TRMM. A preliminary analysis with a few months of data over Western Africa suggests that results are noisy, not robust across datasets and inconsistent with expectations, maybe due to insufficient data sampling (internship by Florentin Breton).

Road map

- To increase the data sampling, the analysis should be extended to several years and over a larger spatial domain. Since q and δD observations are clear-sky only, the co-location algorithm to select observations in the vicinity of MCSs may have to be refined.
- To isolate the direct impact of local convective processes, we must quantify and subtract the effect of the fact that some phases are preferentially sampled in some regions, in some seasons or during some hours of the day. We must also remove the effect of instrument sensitivity, by comparing LMDZ results with or without account for this effect.
- For a given precipitation rate, how do q and δD tropospheric profiles vary along the convective life cycle? What can we deduce about the convective processes at play?
- Can LMDZ, when nudged to reanalyses winds, capture the observed q and δD evolution? If not, what processes are missing, and what is the impact on the model capacity to simulate some meteorological conditions?

A.3 Role of rain re-evaporation: case study in the Sahel

Context and science questions

Rain re-evaporation is among the micro-physical processes whose parameterization rely on several empirical and unknown parameters. How much of the rain drops re-evaporate? How does this depend on the type (e.g. size, propagation, organization, life cycle) of convective systems and on environmental conditions (e.g. relative humidity profile)? Do climate models represent this process well? What is the role of rain re-evaporation on tropospheric humidity? Studies have suggested that rain re-evaporation could contribute to intra-seasonal variations of tropospheric humidity in the Sahel ([Poan et al., 2013]): could we quantify this contribution?

Tropospheric water vapor δD observed by TES has been used to attempt to quantify the proportion of water vapor that originates from rain re-evaporation ([Worden et al., 2007]). However, this study relied on the approximation that the fraction of the rain that re-evaporates is very small, which is often not valid. In the general case, the δD of the vapor evaporating from rain drops may have either an enriching or a depleting effect ([Risi et al., 2010a]) depending on the fraction of the drop that re-evaporates. Tropospheric water vapor δD observations are thus more difficult to interpret.

The goal of this study is to develop a model for predicting tropospheric water vapor δD, or d-excess, as a function of rain re-evaporation. Then, could this model be used to address the above-mentioned science questions?

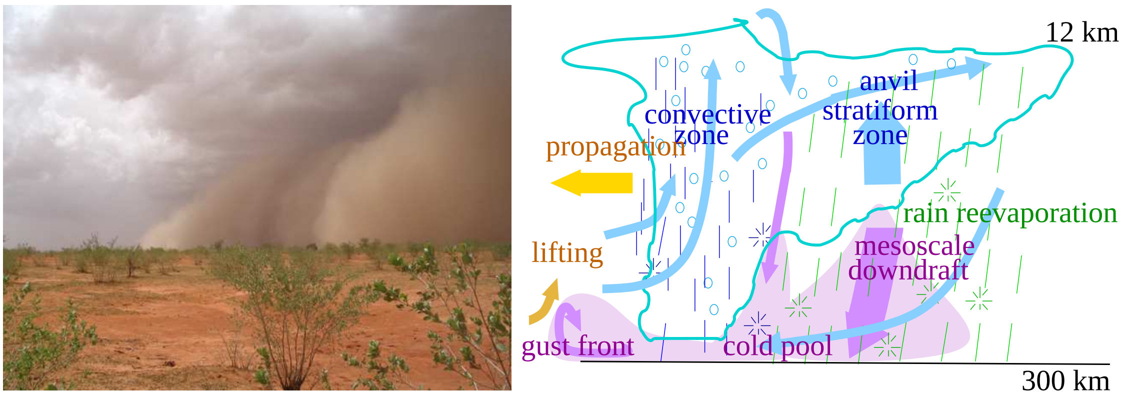

Figure A.2: a) Picture of the gust front of a squall line in Mali. The gust front materializes the border of the cold pool. b) Processes leading to the cold pool in a typical squall line.

Where we stand

As a case study, we investigated cold pools of convective systems in the Sahel (internship of Yannick Lamarre in collaboration with F. Vimeux). These cold pools are driven by rain re-evaporation that cools the air (figure A.2). We used 3 years of continuous in-situ measurements of surface water vapor δD and d-excess in Niamey (Niger), in the Sahel ([Tremoy et al., 2012, Tremoy et al., 2014]), combined with meteorological data. All convective systems during these 3 years were identified based on field notes. Convective system properties, including size, propagation, organization and life cycle, were examined based on Meteosat images every 15 minutes. Cold pools were identified based on observed temperature drops characteristic of cold pools. Temperature, humidity (q), δD and d-excess variations associated with each cold pool were estimated. Based on a moist static energy budget and on twice-daily radio-soundings, the fraction of the water vapor originating from rain re-evaporation, q_{ev}, was calculated.

Road map

- What controls q_{ev}? Among possible factors, cloud system properties based on Meteosat images and tropospheric relative humidity profiles based on radio-soundings will be investigated.

- How do surface water vapor δD and d-excess vary with q_{ev}? What are the other factors controlling δD and d-excess? A model for predicting tropospheric water vapor δD and d-excess as a function of rain re-evaporation will be developped.
- Using this model, we will revisit the interpretation of the TES data in terms of rain re-evaporation. We will also use the IASI data. What is the contribution of rain re-evaporation to tropospheric moisture? How does it depend on convective and environmental properties?
- Using this model, we will also try and quantify the contribution of rain re-evaporation to intra-seasonal variations of tropospheric humidity in the Sahel.
- Finally, we will use δD measurements to evaluate the representation of rain re-evaporation in the LMDZ model.

A.4 Understanding the isotopic signature of convective processes using Cloud Resolving Model simulations

Context and science questions

Our understanding of how convection impact the isotopic composition of water vapor is still incomplete. Relying on convective parameterizations may contribute to the gap between understanding based on GCMs and reality. The goal of this study is to use Cloud Resolving Models (CRMs) or Large-Eddy Simulations (LES) (figure A.3) to understand the isotopic response to convection and evaluate its simulation in GCMs.

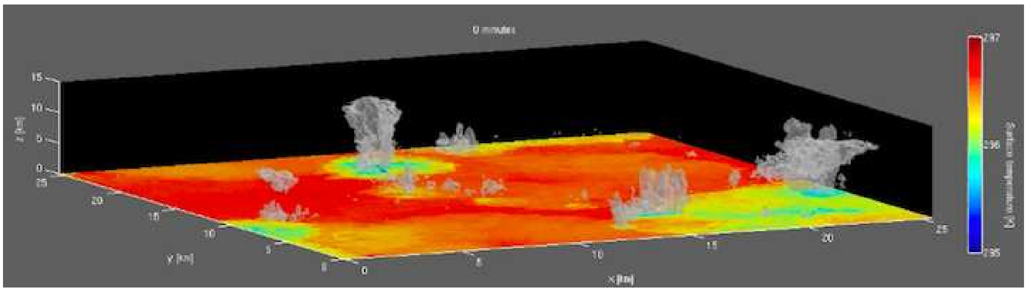

Figure A.3: Illustration of convective clouds simulated by a Large-Eddy Simulation (LES) ([Muller and Held, 2012]).

Where we stand

Using the SAM CRM, [Moore et al., 2014] performed a series of CRM simulations in Radiative-Convective Equilibrium (RCE) spanning different precipitation rates. Although they studied the amount effect, processes were not examined in detail and the specific advantages of a CRM simulation were not exploited. I would like to analyze these simulations in more details, in collaboration with Peter Blossey.

Road map

- Single Column Model (SCM) simulations equivalent to the CRM simulations will be performed with LMDZ. How do the simulated δD profiles compare with those from the CRM simulations?

- Criteria for identifying unsaturated downdrafts, saturated downdrafts and updrafts in CRM simulations will be developed. The isotopic composition of the water vapor and condensate in each of these drafts will be diagnosed, following the conditional sampling methodology ([Couvreux et al., 2010]).

- How does the isotopic composition in the different drafts in CRM simulations compare with the isotopic composition computed as internal variables by the different components of the convective parameterization in the SCM?

- To understand what controls the isotopic composition of tropospheric water vapor, water and isotopic budget will be performed at different levels, in both the CRM and SCM. Where CRM and SCM show different results, are physical parameterizations or isotopic parameterizations responsible for the difference?

A.5 Bridging the gap between theoretical Lagrangian models and reality

Context and science questions

Much of the understanding on what controls the isotopic composition of precipitation has been gained thank to the use of theoretical curves in the $q-\delta D$ framework ([Worden et al., 2007, Noone, 2012]). However, the simplicity of this framework makes it difficult to connect with GCM simulations and with reality, where processes are much more complex, diverse and inter-mingled (figure A.4). The goal of this study is to try to bridge the gap between the $q-\delta D$ framework and GCM simulations.

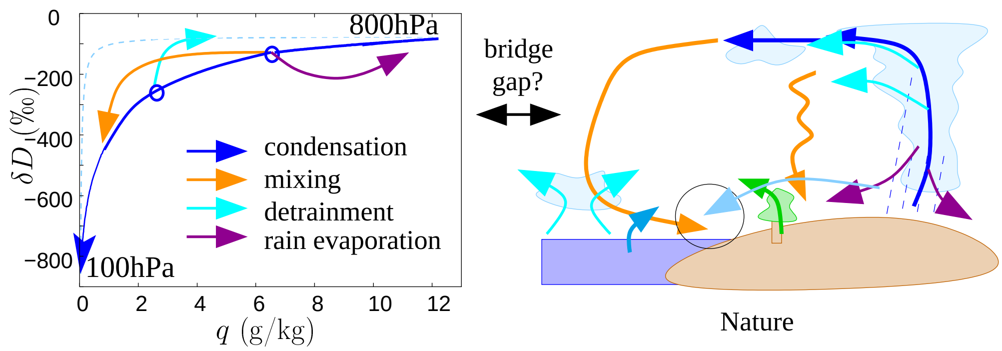

Figure A.4: a) Idealized $q-\delta D$ framework. b) Diversity and complexity of processes in a GCM and in reality.

Where we stand

During his post-doc, Obbe Tuinenburg connected the $q-\delta D$ evolution simulated by LMDZ during MJO events to the simulated tendencies (i.e. the effect of each physical parameterization on q and δD) ([Tuinenburg et al., 2015]). During his PhD thesis, You He looked at tendencies simulated by LMDZ along back-trajectories to interpret the simulated isotopic variability observed at Lhasa (Tibet) ([He et al., 2015]). These studies showed that looking at LMDZ tendencies is useful to interpret $q-\delta D$ variations at the daily scale.

Road map

- Identify a few sites of interest and calculate back-trajectories
- Look at physical tendencies simulated by LMDZ along back-trajectories and plot them in the $q-\delta D$ diagram. How does this relate to theoretical curves? Should we create new ones?
- Can we develop a new framework that relates the $q-\delta D$ at the site of interest with the sum of all tendencies that affected q and δD along the trajectories?

A.6 Temperature or precipitation: how should we interpret paleo-climate archives of precipitation isotopes?

Context and science questions

In polar regions, paleo-climate archive of precipitation isotopic composition ($\delta^{18}O$) have long been interpreted in terms of changes in temperature. In the tropics, the interpretation of such archives (Andean glaciers and Tibetan, speleothems...) is more controversial. Inspired by studies in polar regions, the tropical $\delta^{18}O$ signal was first interpreted in terms of the temperature change ([Thompson et al., 1989, Thompson et al., 1998, Thompson et al., 2000]). This interpretation is supported by the fact that all tropical ice cores exhibit similar $\delta^{18}O$ variations over last 25 000 years, with depleted values during the last glacial maximum as in polar cores (figure 4.4a). However, at inter-annual to daily scales in present-day observations, precipitation $\delta^{18}O$ anti-correlates with local ([Dansgaard, 1964]) or upstream ([Vimeux et al., 2005, Gao et al., 2013]) precipitation amount. This is because atmospheric convection depletes the lower-tropospheric water vapor (figure 4.4b). Thus many authors now interpret the isotopic variations in tropical paleo-climatic records in terms of regional precipitation variations ([Vuille and Werner, 2005, Pausata et al., 2011]).

The aim of this study is to understand what controls the isotopic variations in tropical rainfall at the paleo-climatic scale (thousands of years): temperature, precipitation, both?

Where we stand

A series of simulations were performed with LMDZ for different climates and for different versions of the model physical package (section 4.2). In some model versions, the temperature is the dominant control, while in other versions, the regional precipitation is the dominant control. Which version is more credible?

During his PhD thesis, You He used daily in-situ and satellite data to investigate how convective processes impact the isotopic composition of water vapor along back-trajectories and controls the isotopic signal recorded in the precipitation at Lhasa ([He et al., 2015]). How can this improved understanding at the daily scale be used for paleo-climatic interpretations?

Road map

- It is likely that results will depend on the region of interest. If the case, the same road map will be repeated for each region.
- Two model versions will be identified: one that produces a dominant temperature control, and another that produces a dominant precipitation controls, at the paleo-climatic time scale.
- How do these model versions simulate the control of precipitation $\delta^{18}O$ at the daily time scale? Could daily observations be used to discriminate which one is the most realistic?
- If precipitation is more depleted for a given paleo-climatic epoch, is it because it is systematically more depleted, or because there is a higher frequency of depleted events, or because depleted events are more intense? The daily probability distribution of precipitation $\delta^{18}O$ will be compared for different epochs. Based on the results, can we interpret some paleo-climatic variations in the light of our increased knowledge of daily processes at present-day?
- To better understand processes controlling the precipitation $\delta^{18}O$, processes along back trajectories will be analyzed and compared between different model versions and model epochs. Are there processes that are common to all time scales?

A.7 Does the simulation of moistening and dehydrating processes associated with convection determines the quality of MJO simulation?

Context and science questions

The Madden-Julian Oscillation (MJO) is the main mode of intra-seasonal variability in the tropical troposphere. GCMs have persistent difficulties to simulate this mode of variability ([Slingo et al., 1996, Lin and Coauthors, 2006, Kim et al., 2009, Hung et al., 2013]). What determines the capacity of GCMs to simulate well or not the MJO?

The MJO results from feedbacks between convective processes and the large-scale circulation (figure A.5). In models, convective processes respond to the large-scale dynamical forcing. In turn, convective processes affect temperature and humidity profiles, which can feedback on convective processes or on the large-scale circulation. Therefore, the capacity of a model to simulate the MJO may depend on several factors:

1. the response of convective and cloud parameterizations to the dynamical forcing in terms of moistening and dehydrating processes (green on figure A.5),

2. the response of convective and cloud parameterizations to the dynamical forcing in terms of heating and cooling profiles (orange),

3. the dynamical response to heating and cooling profiles (purple).

In the case of hypothesis (1), water vapor isotope measurements could help evaluate the moistening and dehydrating processes at play. This hypothesis would be consistent with the suggested importance of shallow convection to moisten the free troposphere and pre-condition the atmosphere to deep convection ([Maloney, 2009]).

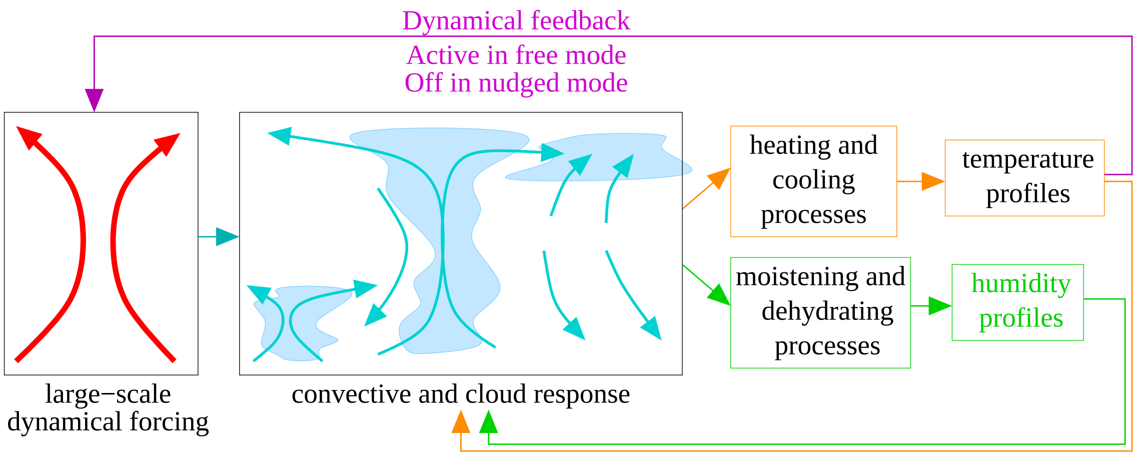

Figure A.5: Feedback between the large-scale dynamics and convective and cloud processes: the large-scale dynamics acts as a forcing to convective and cloud processes simulated by physical parameterizations. These parameterizations represent heating and cooling processes that impact large-scale temperature profiles, and moistening and dehydrating processes that impact large-scale q profiles. These profiles can feedback onto convective and cloud processes. In free-running simulations, temperature profiles also feed-back onto the large-scale dynamics. In nudged simulations, the large-scale dynamics is imposed, cutting the feedback loop.

Where we stand

To test hypothesis (1) (i.e. that a good simulation of moistening and dehydrating processes is crucial for a good simulation of the MJO), we compared different versions of LMDZ with perturbed physical parameters. The simulation of moistening and dehydrating processes as a response to the dynamical forcing is diagnosed using $q - \delta D$ cycles simulated when winds are nudged towards reanalyses ([Tuinenburg et al., 2015]). The simulation of the MJO is diagnosed in free-running simulations using published diagnostics ([Kim et al., 2009, Waliser et al., 2008]).

During her internship, Ella Tribes compared two versions of LMDZ: the standard version with a poor MJO simulation, and a version with modified entrainment in which the MJO is improved. This improvement is associated with a moister middle troposphere, which is due to a smaller contribution of the deep convective scheme to precipitation.

Road map

- After identifying other perturbed-physics versions of LMDZ in which the MJO is improved, Ella Tribes's analysis will be redone: are improvements in the MJO simulation determined by the convective response to dynamical forcing in terms of moistening and dehydrating processes?
- To better understand how precipitation anomalies are amplified more or less strongly by convective-dynamical feedbacks depending on simulations, **hindcasts** of different MJO events and with different model versiosn of LMDZ will be performed
- To extend the perturbed-physics experiments accounting for model architecture, several isotopic GCMs will be compared in nudged and free modes. An **isotopic GCM intercomparison study** has started involving the GISS, GSM and ECHAM models, in collaboration with R. Field (GISS), K. Yoshimura (U. Tokyo), and M. Werner (AWI Germany) respectively. Ella Tribes's analysis will be redone.
- If in some models or model versions MJO simulation improvements result from improvements in the convective response to dynamical forcing in terms of moistening and dehydrating processes, then we will try to develop process-oriented diagnostics for the MJO simulation based on water isotope measurements. In contrast, if MJO simulation improvements always result from improvements in the convective response to dynamical forcing in terms of heating and cooling processes, then we will communicate on this result and my research on isotopes during the MJO will stop there.

A.8 Multi-tracer signature of convection

Context and science questions

Water vapor isotope measurements are potentially useful to evaluate the representation of some convective or cloud processes in atmospheric models, but they cannot solve everything. The problem is often under-constrained. For example, δD in the upper troposphere increases when the detrainment of condensate increases (section 2.1.2). But δD cannot make the difference between an increase in condensate detrainment associated with an increase in the condensate load of the detraining air, or an increase in the detrainment flux. A low-tropospheric air tracer, such as carbon monoxide (CO), would be useful to estimate the detrainment flux.

How could we combine different isotopic and chemical tracers (CO, ozone O_3, Beryllium Be...) to optimally constrain the different components of convective schemes?

Where we stand

To my knowledge, the only study combining water isotope measurements with other chemical measurements is [Liu et al., 2009], who investigate ozone variability and use water isotopes are used to infer the origin of air masses.

I find the study by [Folkins et al., 2006] very inspiring: he simulates simultaneously H_2O , CO, O_3 and nitric acid (HNO_3) profiles in single-column and 3D versions of GCMs. He investigates the sensitivity of these profiles to convective parameterizations and discusses how these combined chemical tracers could provide constrains on convective parameterizations.

Road map

- To simulate both humidity, water isotopes and chemical tracer profiles, a single-column model simulation with LMDZ equipped with both isotopes and the **chemical module INCA** (Interaction with Chemistry and Aerosols , [Hauglustaine et al., 2004]) will be set up. Simulations will be evaluated to first order using averaged climatologies.

- A simple, theoretical framework will be developed to interpret simultaneously the humidity, water isotopes and chemical tracer profiles. Ideally, for n tracers, we would end up with a model with n parameters quantifying key processes in convective and cloud parameterizations (e.g. precipitation efficiency, rain reevaporation fraction...). The advantages of combining different tracers will be discussed.

- Sensitivity tests to model parameters will help develop and validate the theoretical framework.

- For more realistic simulation and rigorous model-data comparisons, 3D LMDZ simulations equipped with both isotopes and the chemical module INCA will be performed. Outputs will be compared with TES or IASI data, which simultaneously retrieve water vapor, water isotopes and a variety of chemical tracers.

A.9 Link between past and future behavior of climate models for tropical precipitation changes

Context and science questions

Climate models show persistent spread in their simulation of future tropical precipitation changes. Could we use paleo-climatic archives of precipitation changes to assess which model simulates the most credible future precipitation changes?

The goal of this study is to establish links between the behavior of climate model for past climate changes and that for future climate changes, and to use these links to develop paleo-climatic constrains for future precipitation changes.

Where we stand

As a pilot study, CMIP5 simulations for pre-industrial, Mid-Holocene and RCP8.5 scenario were compared over tropical South America. We showed that models that shift the ITCZ to the North in Mid-Holocene also do so in future climate projections (section 4.1, [Schmidt et al., 2014]).

During his internship, Benedikt Rakotonirina-Hess tried to extend this study to Last Glacial Maximum (LGM) simulations and to the entire globe. The main conclusion was that tropical precipitation changes are usually not zonally consistent, so such studies should remain at the regional scale. He also examined the physical and dynamical components of precipitation changes, following the decomposition by [Bony and Bellon, 2013]. He found that when there is a link between past and future changes, this link is explained mainly by the dynamical component of precipitation changes.

Road map

- CMIP5 simulations for pre-industrial, Mid-Holocene and $4 \times CO_2$ scenario will be intercompared over different tropical land regions. To search for statistical links between past and future behavior, EOF analysis will be used to identify the main patterns of precipitation changes depending on models.

- When statistical links are identified, we will try to explain it in terms of physical processes, using the decomposition into dynamical and physical components ([Bony and Bellon, 2013]). The dynamical component will be further decomposed based on the moist static energy budget ([Oueslati et al., 2016]).

- To check our interpretation of physical processes, we may run sensitivity tests with LMDZ simulations. For example, if the impact of cloud radiative effects on ITCZ shifts ([Hwang and Frierson, 2013]) is confirmed, simulations with and without cloud radiative effects can be performed.

- When a statistical link is identified for a given region, a given precipitation pattern and a given paleo-climatic epoch, and if the underlying physical processes are understood, we will check whether paleo-climatic archives for past precipitation changes (isotopic or non-isotopic) are available. If the case, we will propose a paleo-climatic constrain on that specific regional past precipitation change pattern and discuss its implications for future climate change.

Appendix B

Curriculum Vitae

Last update: October 20th, 2016

B.1 Contact information

Adress : LMD
case postale 99,
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75252 Paris Cedex 05
Phone: 01-44-27-52-62
Fax: 01-44-27-62-72
Email : Camille.Risi@lmd.jussieu.fr
Web : http://www.lmd.jussieu.fr/˜crlmd

B.2 Education

2005-2006: Master “Ocea , Atmosphere, Climate and Remote Sensing” at the Université Pierre and Marie Curie (UPMC) (mention ■Très Bien■).

2004-2005: Preparation of the “Agrégation” (national competitive exam for posts in the teaching staff of french high schools and Universities) in Biology-Geology; accepted 4 th

2002-2004: “Licence” and ■Maîtrise” of Earth Sciences at the Ecole Normale Supérieure (ENS) of Paris.

2000-200: “Classes préparatoires” of Biology, Geology, Physics and Chemistry in Paris: preparation of national competitive exams to enter the ENS and engineering schools; accepted at the ENS of Paris.

2000: “Baccalauréat” in Sciences, Mathematics speciality (mention ■Très Bien■).

B.3 Reaseach

B.3.1 Research experience

2011-: Research Scientist at the Laboratoire de Météorologie Dynamique (LMD) , Paris

2010-2011: post-doc at the University of Colorado at Boulder (USA) at the Cooperative Institute for Research in Environmental Science, advised by David Noone

2006-2009: PhD at LMD, Paris, co-advised by Sandrine Bony and Jean Jouzel, untitled “Water stable isotopes: applications to study the water cycle and climate variations”

2006: 5-month Master internship at LMD advised by Sandrine Bony, on the relationship between atmospheric convection and the isotopic composition of tropical water.

2004: 6-month “Maîtrise” internship at the Massachusetts Institute of Technology, advised by Kerry Emanuel, on the simulation of tropical cyclones tracks using a statistical model

	Articles in international journals (accepted or published)	including article written as a first author	Articles in national journals (accepted or published)
total	69	10	5
2016	10	0	1
2015	13	0	1
2014	9	0	2
2013	13	2	0
2012	8	2	0
2011	7	0	1
2010	5	4	0
2009	0	0	0
2008	3	2	0
2007	0	0	0
2006	1	0	0

Table B.1: Summary table of publications for different years.

2003: 5-week “Licence” internship at LMD, advised by Frédéric Hourdin, on boundary layer parametrization.

B.3.2 Publications

- 69 publications in international journals with review comitees (tables B.1 and B.2, list in appendix C)
- 5 publications in national journals with review comitees (list in appendix D)

B.3.3 Funded projects

- **2013-2018**: Principal Investigator of the “ANR jeune chercheur-jeune chercheuse” project CONVISO

B.4 Advised students and post-docs

So far, I have advised 1 post-doc, co-advised 3 other post-docs, co-advised a PhD student and 1 Masters’ student, advised 12 bachelor students and advised one “mission doctorale”.

1. Florentin Breton, **2016**, “M1” internship, on the isotopic signature of the convective life cycle
2. Natacha Legrix, **2016**, “L3” internship, on the isotopic signature of the convective aggregation
3. Ella Tribes, **2016**, “M1” internship, on the impact of the representation of convective processes on the simulation of the MJO
4. Rongrong Zhang, **2016**, “M1” internship, on what controls the isotopic variations simulated during the last glacial maximum
5. Benedikt Rakotonirina-Hess, **2014**, “M1” internship, on past and future tropical precipitation changes in CMIP5 models
6. Yannick Lamarre, **2014**, “M1” internship, on cold pools in Niamey
7. Anis Bouchelit, **2014**, “M1” internship, on diabatic heating profiles simulated by LMDZ-1D
8. Obbe Tuinenburg, **2013-2015**: post-doc on water isotopes during MJO events

Journal	Number of articles
J. Geophys. Res.	20
Clim. Past	13
Atmos. Chem. Phys.	7
Earth Planet. Sci. Lett.	5
Atmos. Meas. Tech.	4
The Cryosphere	4
Clim. Dyn.	3
P. Natl. Acad. Sci. USA	2
Geophys. Res. Lett.	2
Rev. Geophysics	2
Water Ressources Res.	2
Nature Comm.	1
Quat. J. R. Meteorol. Soc.	1
Quat. Sci. Rev.	1
Tellus	1
Hydrology: Current Research	1
La Météorologie	5

Table B.2: Summary table of publications in different journals.

9. Alexandre Cauquoin, **2013-2015**: post-doc co-advised with Amaelle Landais on tritium modeling in LMDZ

10. Marion Saint-Lu, **2012-2013**: "Mission Doctorale" (PhD student) on the educational software SimClimat

11. Francesca Guglielmo, **2012-2013**: post-doc co-advised with Catherine Ottlé on the modeling of soil water isotopes in Siberia using the ORCHIDEE land surface model

12. Victor Gryazin, **2012-2013**: post-doc co-advised with Jean Jouzel on model-data comparison in Siberia

13. You He, **2012-2014**: PhD student co-advised with Valérie Masson-Delmotte and Gao Jing, on the interpretation of daily isotopic variability observed at Lhasa.

14. Tiana Jacquemart and Julie Meyer, **2012**, tandem "L3" internship, on daily isotopic variability observed at Darwin during the TWPice campaign and simulated by LMDZ in a single-column

15. Gaelle Benoit, **2012**, "L3" internship, on isotopic modeling using single-column LMDZ radiative-convective equilibrium

16. Lorraine Desbordes, **2012**, "M1" internship, on the influence of irrigation on precipitation distribution

17. Magali Hug, **2012**, "M1" internship, on water vapor isotopic variations observed by TES during El-Nino events

18. Guillaume Tremoy, **2009**, "M2" internship co-advised with Françoise Vimeux, on the interpretation of Andean ice core records

19. Marie Vicompte, **2007**, "L3" internship co-advised with Sandrine Bony, on Sahelian squall lines

B.5 Awards

- **2010**: "Prix André Prud'homme" of the "Société Météorologique de France" for my PhD thesis
- **2010**: "Prix Le Monde de la recherche universitaire" for my PhD thesis
- **2010**: "Prix de la meilleure thèse" of the EADS fundation for my PhD thesis.
- **2016**: "Médaille de bronze" of CNRS.

B.6 Teaching

- **2015**, **2016**: Course on climate modeling for Masters' students in "Sciences et Génie de l'Environnement" of Paris 7.
- **2014**: Advised students in the "Science diffusion" section of UPMC.
- **2013**: Course on Climate at Ecole Polytechnique Féminine.
- **2011-2012**: Training students who prepare the competitive exam to become Biology-Geology teachers at UPMC.
- July **2012**: Teaching a workshop on Climate during the Summer school of "KIC-Climat" in Polytechnique.
- **2006-2009**: Teaching at Université Pierre et Marie Curie (UPMC) in Physics and Geology.

B.7 Outreach activities

B.7.1 Educational software on climate: SimClimat

http://www.lmd.jussieu.fr/˜crlmd/simclimat

This educational software simulates climate given user-chosen parameters. Through a friendly interface, the user chooses the length of the simulation (from 100 years to a few billions years) and the initial conditions, and tests the influence of various parameters involved in climate: astronomic forcing, atmospheric composition, carbone cycle, climatic feedbacks (ice albedo, vegetation, ocean, water vapor). Simulation results (such as temperature, sea level and ice cover), calculated by a physical climate model, appear on the interface, through curves and images (figure).

Work on this software included:

- 2006-2008: software development
- tests in high schools in the Paris region
- advised "mission doctorale" of Marion Saint-Lu to improve the software
- contribution to a "hackaton" as part of the "train du climat" (october 2015).

B.7.2 Contribution to the "La Météorologie" outreach journal

La Météorologie is a journal on meteorology and climate targeting students, teachers and enthusiasts.

- **2012-**: member of the editorial comitee
- wrote 5 articles (list in chapter D)
- wrote 3 short articles and 1 editorial

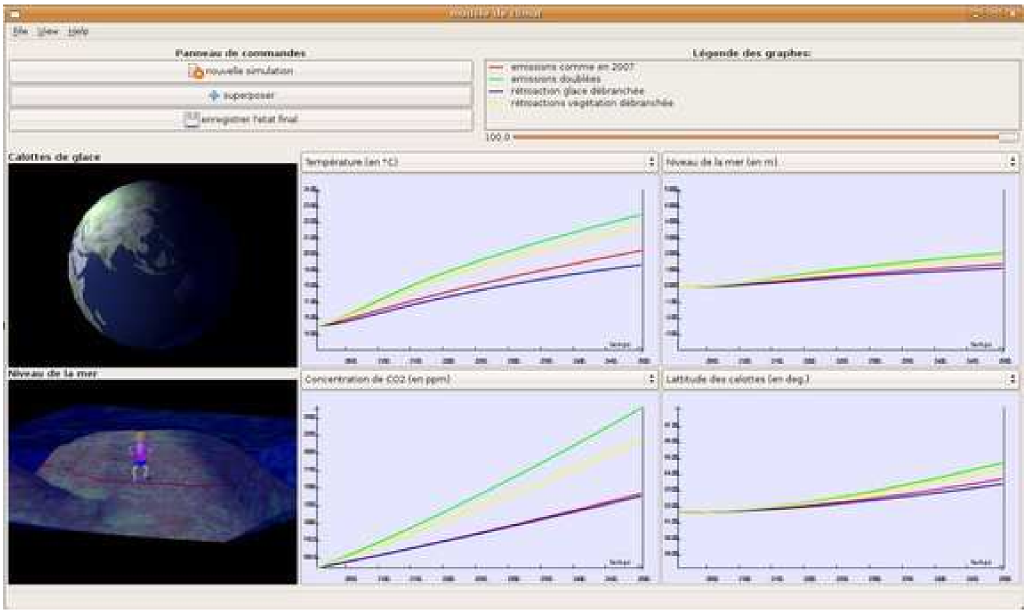

Figure B.1: Snapshot of the SimClimat graphical interface.

B.7.3 Round tables and outreach workshops

- April **2015**: Invited to a round table on multi-disciplinarity in climate sciences at the ”disciplines sans frontières” workshop in Paris
- March **2015**: Invited to a round table at the workshop ”Educate and train on climate change” on ”How to conduct a debate in an educational context?” at the “Conseil Economique, Social et Environnemental de Paris”.
- January **2015**: Participation in the workstop ”Climatologie, météorologie et enseignement(s)” at the “Forum des ressources pour l’éducation au développement durable”, Amiens

B.7.4 Presenting experiments in science festivals

During science festivals, I usually present educational experiments on clouds and atmospheric physics, play with the SimClimat software or discuss about climate.

- October **2016**: Participation in the IPSL tent during the the Fête de la Science, UPMC.
- March **2015**: Participation in the IPSL tent at the Forum de la Météorologie, Paris
- October **2014**: Participation in the INSU tent at the “cité de sciences de la Vilette” during the Fête de la science
- October **2014**: Participation in the Fête de la Science at LMD
- April **2014**: Participation in the IPSL tent at the Forum de la Météorologie, Paris
- October **2013**: Participation in the Fête de la Science at LMD, UPMC
- May-June **2013**: Participation in the CNRS and UPMC tents at “salon des Jeux et de la culture mathématiques” at the “cité des sciences de la Villette”.
- March **2013**: Participation in the “Bar des sciences” ”Météo, mode d’emploi” at the Chamarande castle.
- November **2013**: Participation in the “forum Climat du Grand Troyes” in Troyes

- October **2013**: Participation in the INSU tent during the “Forum de la Météorologie”, Paris.
- October **2012**: Participation in the Fête de la Science at LMD
- December **2012**: organisation of the visit of a secondary school class at LMD
- Novembre **2012**: Participation in the “forum Climat du Grand Troyes” in Troyes
- October **2011**: Participation in the Fête de la Science at LMD
- October **2011**: Participation in the INSU tent during the “Forum de la Météorologie”, Paris.
- October 2**009**: Participation in the Fête de la Science at LMD
- October **2009**: Participation in the INSU tent during the “Forum de la Météorologie”, Paris.
- October **2008**: Participation in the Fête de la Science at LMD
- October **2008**: Participation in the INSU tent during the “Forum de la Météorologie”, Brussels.
- October **2008**: Participation in a tent on climate change at the Fête de l’Essone.
- March **2008**: Presentation of educational experiments at the “commission Education Formation” of the “Conseil Supérieur de la Météorologie”, Ecole Polytechnique.
- October **2007**: Participation in the Fête de la Science at LMD
- October **2007**: Participation in the INSU tent during the “Forum de la Météorologie”, Paris.
- Summer **2007**: Participation in a tent on volcanoes at the Paris-Montagne festival
- October **2006**: Participation in the Fête de la Science at LMD
- Summer **2006**: Participation in a tent on climate change at the Paris-Montagne festival

B.7.5 Conferences in Universities

- September **2016**: Conference on climate change during the inauguration of the new campus of Jussieu

B.7.6 Conferences in primary and secondary schools

- December **2014**: Organizing the visit of LMD by a secondary school class of Lycée Voltaire.
- December **2014**: Conference in a secondary school, Confolens.
- November **2014**: Conference at lycée Voltaire in Paris
- September **2014**: Debate with Roland Schlisch at lycée Louis Le Grand following the projection of the moovie ”Entérrés volontaires”
- April **2014**: Climate activities for secondary school pupils as part of the “Sciences Ouvertes” association.
- March **2014**: Conference in a secondary school, La Ferté-sous-Jouarre
- March **2014**: Conference in a primary school, Nogent-sur-Seine, as part of the “festival des Sciences et Techniques de l’Aube” and the “Terre Avenir” association.
- April **2012**: Conference to a network of primary schools, Nogent-sur Oise.
- February **2012**: Conference in a secondary school, Argenteuil.
- March **2009**: Conference in a secondary school as part of “1000 ambassadrices pour les Sciences à Paris”
- Décembre **2007**: Conference in a secondary school, lycée Soeur Rosalie-Louise de Marillace, Paris

B.7.7 Conference in jails

- February **2016**: Conference on climate change at the "Maison d'Arrêt des Hauts de Seine", Nanterre.

B.7.8 Radio and TV

- May **2016**: "E=M6" show on M6 channel on super-cell thunderstorms
- December **2015**: Discussion on climate change on "Radio Grenouille" as part of the COP21.
- April **2015**: Interview for an audio podcast on a blog for scientific information "indesciences"
- March **2015**: Video on "climate modeling and pedagogy" as a resource for the education for sustainable development of the "Education Nationale".
- October **2014**: Show "3 minutes for the planet" on "Radio Classique".
- October **2014**: Debate on the science diffusion on the "Aligre FM" radio

B.7.9 Teacher training

- February **2016**: Training animators for the association "Pionniers de France" to take part at the "Planète mômes" festival
- October **2015**: Course on climate modeling as part of the training "Formaterre" for secondary school teachers.
- May **2015**: course on climate as part of a training for secondary school teachers.
- January **2015**: Round table on "Climatology, meteorology and teaching(s)" at the "forum sur l'éducation au développement durable" for secondary school teachers.
- August **2012**: workshop on climate change in the Summer school of mathematics (for teachers) at Sourdun, for secondary school teachers

B.7.10 Miscellaneous

- Hosted 5 secondary school pupils at LMD
- Helped many secondary school pupils and students for their science projects ("TPEs" and "TIPEs")

B.8 Administrative and miscelaneuous activities

B.8.1 Member of PhD committees

1. Maximilien Bolot, Approche théorique de la distribution des isotopologues stables de l'eau dans l'atmosphère tropicale, de l'échelle convective aux grandes échelles. ENS Paris, **2013**
2. J-L Lacour, Estimations du profil du rapport isotopique δD de la vapeur d'eau dans la troposphère à partir des spectres mesurés par le sondeur satellitaire IASI dans l'infrarouge thermique : Méthodologie d'inversion et analyses des premières distributions spatiales. ULB (Belgium), **2014**

B.8.2 Expertise of research articles and proposals

- expertise of research articles: > 40 articles including Nature, JGR, GRL, Clim Dyn...
- expertise of proposals: 5: ECOS-Sud, NSF, ANR, LEFE (x2).

B.8.3 Miscellaneous

- 2011-: Responsible for GENCI project 0292 for HPC resources: managing accounts, writing proposals for computational time and reports on computational activities
- 2013-: Member of the “MJO Task Force”
- 2015: Member of the working group on Water vapor for a “Phase-0” study at CNES.

Appendix C

List of international publications

Last update: October 20th, 2016

1. Galewsky, J, Steen-Larsen, H C, Field, R, Worden, J, Risi, C. Stable isotopes in atmospheric water vapor and applications to the hydrologic cycle. **Accepted** by Reviews of Geophysics.

2. Cauquoin, A., Jean-Baptiste, P., Risi, C., Fourré, E., Stenni, B., and Landais, A. Modeling the global bomb-tritium transient signal with an Atmospheric General Circulation Model: a promising method to evaluate the dynamics of the hydrological cycle in the models and its link with stratospheric air intrusions. **Accepted** by J.Geophys. Res.,

3. C. Risi S. Bony, J. Ogée, T. Bariac, Naama Raz-Yaseef and Lisa Wingate, Welker, J, Knohl, A, Kurz-Besson, C, Leclerc, M, Zhang, G, Buchmann, N, Santrucek, J, Hronkova, M, David, T, Peylin, P, Guglielmo, F. The water isotopic version of the land-surface model ORCHIDEE: implementation, evaluation, sensitivity to hydrological parameters. **Accepted** by Hydrology: Current Research.

4. Stenni, B, Scarchilli, C, Masson-Delmotte, V, Schlosser, E, Ciardini, V, Dreossi, G, Grigioni, P, Bonazza, M, Cagnati, A, Karlicek, D, Risi, C, Udisti, R, Valt, M. Three-year monitoring of stable isotopes of precipitation at Concordia Station, East Antarctica, **accepted**, The Cryosphere.

5. Bolliet, T., Brockmann, P., Masson-Delmotte, V., Bassinot, F., Daux, V., Genty, D., ... and Risi, C. (**2016**). Water and carbon stable isotope records from natural archives: a new database and interactive online platform for data browsing, visualizing and downloading. Climate of the Past, 12(8), 1693-1719., doi:10.5194/cp-12-1693-2016

6. Scheepmaker, R., aan de Brugh, J., Hu, H., Borsdorff, T., Frankenberg, C., Risi, C., Hasekamp, O., Aben, O., Landgraf, J. (**2016**). HDO and H2O total column retrievals from TROPOMI shortwave infrared measurements, Atmos. Meas. Tech., 9, 3921-3937, 2016, doi:10.5194/amt-9-3921-2016

7. Ritter, F, Steen-Larsen, H C, Werner, M, Masson-Delmotte, V, Orsi, Anais, Behrens, M, Birnbaum, G, Freitag, J, Risi, C, and Kipfstuhl, S (**2016**). Isotopic exchange on the diurnal scale between near-surface snow and lower atmospheric water vapor at Kohnen station, East Antarctica, The Cryosphere, 10, 1647-1663, 2016 doi:10.5194/tc-10-1647-2016

8. Touzeau, A, A. Landais, B. Stenni, R. Uemura, K. Fukui, S. Fujita, S. Guilbaud, A. Ekaykin, M. Casado, E. Barkan, B. Luz, O. Magand, G. Teste, E. Le Meur, M. Baroni, J. Savarino, I. Bourgeois, and C. Risi (**2016**). Acquisition of isotopic composition for surface snow in East Antarctica and the links to climatic parameters. The Cryosphere, 10, 837-852, 2016, doi:10.5194/tc-10-837-2016

9. Oueslati, B, Bony, S, Risi, C, Dufresne, J-L (**2016**). Interpreting the inter-model spread in regional precipitation projections in the tropics: Role of surface evaporation and cloud radiative effects. *Clim. Dyn.* 47: 2801-2815. doi:10.1007/s00382-016-2998-6

10. Botsyun, S, Sepulchre, P, C. Risi, and Y. Donnadieu (**2016**). Impacts of Tibetan Plateau uplift on atmospheric dynamics and associated precipitation d18O . Clim. Past, 12, 1401-1420, 2016. doi:10.5194/cp-12-1401-2016

11. Retanauer C, A. Landais, T. Blunier, C. Bréant, M. Kageyama, M.-N. Woillez, C. Risi, V. Mariotti, and P. Braconnot (**2015**), Quantifying molecular oxygen isotope variations during a Heinrich Stadial. *Clim. Past* 11, 1527– 1551

12. Jasechko, S, Lechler, A, Pausata, F S R, Fawcett, P J, Gleeson, T, Cendron, D I, Galewsky, J, LeGrande, A N, Risi, C, Sharp, Z D, Welker, J M, Werner, M, Yoshimura, K (**2015**). Late-glacial to late-Holocene shifts in global precipitation delta18O. *Clim. Past* , 11, 1375-1393

13. Cauquoin, A., Jean-Baptiste, P., Risi, C., Fourré, E., Stenni, B., and Landais, A. (**2015**). The global distribution of natural tritium in precipitation simulated with an Atmospheric General Circulation Model and comparison with observations. *Earth Planet. Sci. Lett.*, 427, 160-170. doi:10.1016/j.epsl. 2015.06.043

14. Tuinenburg, O, C. Risi, J.-L. Lacour, M. Schneider, A. Wiegele, J. Worden, N. Kurita, J.P. Duvel, N. Deutscher, S. Bony, P.F. Coheur, C. Clerbaux (**2015**). Moist processes during MJO events as diagnosed from water isotopic measurements from the IASI satellite. *J. Geophys. Res.* 120(20):10,619.10,636.

15. V. Masson-Delmotte, H.C. Steen-Larsen, P. Ortega, D. Swingedouw, T. Popp, B.M. Vinther, H. Oerter, A.E. Sveinbjornsdottir, J.E. Box, S. Falourd, X. Fettweis, H. Gallée, E. Garnier, J. Jouzel, A. Landais, B. Minster, N. Paradis, A. Orsi, C. Risi, M. Werner, H. Gudlaugsdottir, and J.W.C. White. Recent changes in North West Greenland climate documented by NEEM shallow ice core data and simulations, and implications for past temperature reconstructions. by *The Cryosphere*, 9, 1481-1504

16. S S.P. Shen, T. Yao, C. Risi, N. Tafolla, Y He. (**2015**) Reconstruction of precipitation delta18O over the Tibetan Plateau since 1910. *J. Geophys. Res., 120, 10, 4878-4888*

17. He, Y, C. Risi, J. Gao, V. Masson-Delmotte, T. Yao, J. Worden, C. Frankenberg, and G. Cesana (**2015**). Impact of atmospheric convection on the isotopic composition of Tibetan precipitation using a combination of satellite data and atmospheric general circulation modeling. *J. Geophys. Res.* 120(9):3852–3871

18. Gao, J, et al. (**2015**) Southern Tibetan Plateau ice core $\delta^{18}O$ reflects abrupt shifts in atmospheric circulation in the late 1970s. *Clim Dyn.* 46(1):291-302

19. Benetti, M, Aloisi, G, Reverdin, G and Risi, C., Sèze, G. (**2015**), Importance of boundary layer mixing for the isotopic composition of surface vapor over the subtropical North Atlantic Ocean. *J. Geophys. Res.* 120(6):2190-2209, doi:10.1002/2014JD021947

20. Bonne, J-L, Steen-Larsen, H C, Risi, C, Werner, M, Sodemann, H, Lacour, J-L, Fettweis, X, Cesana, G, Delmotte, M, Cattani, O, Vallelonga, P, Kjaer, H A, Clerbaux, C, Sveinbjörnsdottir, A E and Masson-Delmotte, V. (**2015**). The summer 2012 Greenland heat wave: in-situ a,d remote sensing observations of water vapor isotopic composition during an atmospheric river event. *J. Geophys. Res.*, 120(7): 2970-2989

21. J.-L. Lacour, L. Clarisse, J. Worden, M. Schneider, S. Barthlott, C. Risi, C. Clerbaux, D. Hurtmans, and P.-F. Coheur (**2015**). Cross-validation of IASI/MetOp derived tropospheric deltaD with TES and ground-based FTIR observations. *Atmos. Meas. Tech.*, 8, 1447-1466, doi:10.5194/amt-8-1447-2015

22. Day, J, Fung, I and Risi, C. (**2015**). Weak Coupling of Juillet-Aout Precipitation between India and East Asia. *J. Climate*, 28, 4330-4356doi: http://dx.doi.org/10.1175/JCLI-D-14-00393.1

23. Pang, H, Hou, S, Landais, A et al. (**2015**). Spatial distribution of 17O-excess in surface snow along a traverse from Zhongshan station to Dome A, East Antarctica. *Earth Planet. Sci. Lett.*, 414:126-133 doi:10.1016/j.epsl.2015.01.014

24. Gryazin V, Risi C, Jouzel J, Kurita N, Worden J, Frankenberg C, Bastrikov V, Gribanov K, Stukova O (**2014**) . The added value of water isotopic measurements for understanding model biases in simulating the water cycle over Western Siberia. *Atmos. Chem. Phys.*, 2014. doi:10.5194/acp-14-9807-2014

25. J-L Bonne, V Masson-Delmotte, O Cattani, M Delmotte, C Risi, H Sodemann, HC Steen-Larsen (**201**4). The isotopic composition of water vapour and precipitation in Ivittuut, southern Greenland. *Atmos. Chem. Phys.*, 14, 4419-4439, 2014. doi:10.5194/acp-14-4419-2014

26. Pommier, M., Lacour, J. L., Risi, C., Bréon, F. M., Clerbaux, C., Coheur, P.-F., Gribanov, K., Hurtmans, D., Jouzel, J., and Zakharov, V. (**2014**). Observation of tropospheric deltaD by IASI over the Western Siberia: Comparison with a GCM. Accepted in *Atmospheric Measurement Techniques*. DOI: 10.5194/amtd-6-11055-2013

27. Liu, Z., Yoshimura, K., Bowen, G. J., Buenning, N. H., Risi, C., Welker, J. M., and Yuan, F. (**2014**). Paired oxygen isotope records reveal modern North American atmospheric dynamics during the Holocene. *Nature communications*, 5. doi:10.1038/ncomms4701

28. P. Ortega, D. Swingedouw, V. Masson-Delmotte, C. Risi, B. Vinther, P. Yiou, R. Vautard, K. Yoshimura (**2014**). Characterizing atmospheric circulation signals in Greenland ice cores: insights from a weather regime approach. *Clim. Dyn.,* 1-21. doi:10.1007/s00382-014-2074-z

29. M. Benetti, G. Reverdin, C. Pierre, L. Merlivat, C. Risi, F. Vimeux (**2014**). Deuterium excess in marine water vapor: Dependency on relative humidity and surface wind speed during evaporation. *J. Geophys. Res*, 119, 584-593, DOI: 10.1002/2013JD020535

30. M. Butzin, M. Werner, C. Frankenberg, J. Jouzel, V. Masson-Delmotte, C. Risi, and V. I. Zakharov (**2014**). Variations of oxygen-18 in West Siberian precipitation during the last 50 years. *Atmos. Chem. Phys.* 14, 5853-5869, doi:10.5194/acp-14-5853-2014

31. G Tremoy, F Vimeux, S Soumana, I Souley, C Risi, O Cattani, G Favreau, M Oi (**2014**). Clustering mesoscale convective systems with laser-based water vapor delta18O monitoring in Niamey (Niger). *J. Geophys. Res.*, DOI: 10.1002/2013JD020968

32. Steen-Larsen, H. C., Masson-Delmotte, V., Hirabayashi, M., Winkler, R., Satow, K., Prié, F., Bayou, N., Brun, E., Cuffey, K. M., Dahl-Jensen, D., Dumont, M., Guillevic, M., Kipfstuhl, S., Landais, A., Popp, T., Risi, C., Steffen, K., Stenni, B., and Sveinbjörnsdottír, A. E. (**2014**) What controls the isotopic composition of Greenland surface snow? *Clim. Past,* 10, 377-392, 2014, doi:10.5194/cp-10-377-2014

33. Jouzel, J, Delaygue, G, Landais, A, Masson-Delmotte, V, Risi, C, Vimeux, F (**2013**). Water isotopes as tools to document oceanic sources of continental precipitation. *Water Ressources Res.*, 49, 7469-7486, DOI: 10.1002/2013WR013508

34. G. A. Schmidt, J. D. Annan, P. J. Bartlein, B. I. Cook, E. Guilyardi, J. C. Hargreaves, S. P. Harrison, M. Kageyama, A. N. LeGrande, B. Konecky, S. Lovejoy, M. E. Mann, V. Masson-Delmotte, C. Risi, D. Thompson, A. Timmerman, L.-B. Tremblay, P. Yiou (**2013**). Using paleo-climate comparisons to constrain future projections in CMIP5. *Clim. Past* doi:10.5194/cp-10-221-2014

35. Gao, J., V. Masson-Delmotte, C. Risi, Y. He, T. Yao (**2013**). What controls southern Tibetan Plateau precipitation deltaO18 at seasonal and intra-seasonal scales? A case study at Lhasa and Nyalam. *Tellus*, http://dx.doi.org/10.3402/tellusb.v65i0.21043

36. T. Yao, V. Masson-Delmotte, J. Gao, L. Tian, W. Yu, X. Yang, C. Risi, C. Sturm, N. Kurita, Y. He, C Shi and S. Hou (**2013**). A Review of Stable Isotope Studies in precipitation from observation stations and Ice Cores in Tibetan Plateau and surrounding regions. *Rev. Geophysics* DOI: 10.1002/rog.20023

37. R. Winkler, A Landais, C Risi, M Baroni, A Ekaykin, J Jouzel, J-R Petit, F Prie, B Minster, S Falourd. (**2013**). Interannual variation of water isotopologues at Vostok indicates a contribution from stratospheric water vapor, *P. Natl. Acad. Sci. USA.*. doi: 10.1073/pnas.1215209110

38. Risi, C, Noone, D, Frankenberg, C, Worden, J. (**2013**). The role of continental recycling in intra-seasonal variations of continental moisture as deduced from model simulations and water vapor isotopic measurements. *Water Ressources Res.* DOI:10.1002/wrcr.20312

39. R. Eagle, C. Risi, J. Mitchell, J. Eiler, U. Seibt, D. Neelin, G. Li, A. Tripati (**2013**). High regional climate sensitivity over continental China inferred from glacial-recent changes in temperature and the hydrologic cycle, *P. Natl. Acad. Sci. USA.* doi: 10.1073/pnas.1213366110

40. L. C. Sime, C. Risi, J. C. Tindall, J. Sjolte, E. W. Wolff, V. Masson-Delmotte, E. Capron (**2013**). Warm climate isotopic simulations: What do we learn about interglacial signals in Greenland ice cores?. *Quaternary Science Reviews* doi:10.1016/j.quascirev.2013.01.009

41. H. C. Steen-Larsen, et al. (**2013**) Continuous monitoring of summer surface water vapour isotopic composition above the Greenland Ice Sheet. *Atmos. Chem. Phys.*, 13, 4815-4828, doi:10.5194/acp-13-4815-2013

42. M. Casado, P. Ortega, V. Masson-Delmotte, C. Risi, D. Swingedouw, V. Daux, D. Genty, F. Maignan, N. Viovy, O. Solomina, B. Vinther, P. Yiou (**2013**). Impact of precipitation intermittency on NAO-temperature signal in proxy records. *Clim. Past,* 9, 871-886, 2013 doi:10.5194/cp-9-871-2013

43. Frankenberg, C, Wunch, D, Toon, G, Risi, C, Scheepmaker, R, Lee, J-E, and Worden, J. (**2013**) Water vapor isotopologues retrievals from high resolution GOSAT short-wave infrared spectra, *Atmos. Meas. Tech.* 6, 263-274, 2013. doi:10.5194/amt-6-263-2013

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Appendix D

List of national publications

Last update: October 20th, 2016

1. C. Risi, V Journé, J-L Dufresne, J-Y Grandpeix, A Spiga, **2016**. Mise en évidence de la chaleur latente liée à l'évaporation et à la condensation de l'eau: Applications au fonctionnement des orages. *La Météorologie*, n°94
2. C. Risi, **2015**. SimClimat, un logiciel pédagogique de simulation du climat. *La Météorologie*, n°88
3. C. Risi, J-L Dufresne, J-Y Grandpeix, S Labetoulle, G Sèze, A Spiga, **2014**, Comment se forment les nuages? L'expérience du nuage dans une bouteille. *La Météorologie*, n°86
4. C. Risi and J-P Duvel, **2014**. L'oscillation de Madden Julian. *La Météorologie*. n°86
5. C. Risi and S. Bony, **2011**. Évaluer les modèles de climat et leurs projections: la valeur ajoutée de la composition isotopique de l'eau. *La Météorologie*, n°74

Appendix E

Articles attached at the end of this document

1. C. Risi, S. Bony and F. Vimeux, (**2008**), Influence of convective processes on the isotopic composition (d18O and dD) of precipitation and water vapor in the tropics: 2. Physical interpretation of the amount effect, *J. Geophys. Res.*

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