Point by point reply to the comments from Anonymous Referee #3

R3: *"This manuscript uses an isotope-enabled GCM to understand how the many factors that influence rainfall isotopes shape the* $\delta^{18}O$ *of rainfall in the Himalaya-Tibet Plateau (HTP) region as topography changes. This is the first manuscript I have seen examining this topic in the HTP region although other papers have used a similar approach in S. America and elsewhere. This is an important topic to understand as it underpins the field of isotope paleoaltimetry and inferences of the topographic history of the HTP region. The paper is well written and informative, especially the author's decomposition of the isotope changes into different processes."*

A: We thank Anonymous Referee #3 for this appreciation of our work.

R3: "*The major comments I have relate to the mismatch between the modern and model* $\delta^{18}O$ *gradients on Tibet and the details of the decomposition method. I recommend acceptance after major revisions. I think the fundamental study and analysis approach is sound. However, additional explanation of the methodology and exploration of the limitations of the model are needed to make the paper useful to readers. I am also not sure their analysis will end with the same conclusions if different methods for decomposing the isotope signal are used (precip weighting climate values, non- linear adiabats that change with temperature etc.)."*

A: We thank the Anonymous Referee #3 for this very constructive review. We will provide all necessary correction in the corrected manuscript version. Meanwhile, for the online discussion, we provide a detailed point-by-point response and a reworked methods section in the end of this document.

Response to Major comments:

R3: *"1. The methods for decomposing the rainfall isotope change into different components was difficult to follow. Additional details would greatly help the reader understand this process. I have a few specific notes about this below but encourage a more general re-thinking of this explanation to make it explicit how each of the terms are derived (particularly the partial differentials, reference values Rvo, To etc.)."*

A: Thank you, we agree that the theoretical framework for the precipitation decomposition method has to be better explained. In the revised version of the manuscript we rewrite the method part with a purpose to make it clearer for readers. Full rewriting of section 2.3 is underway to make clear how each term is calculated. You will find a reworked version of sections “1. Theoretical framework for the precipitation composition” and “2. Decomposition of precipitation composition differences” below point-by-point response. We will remove the partial differentials, because it is misleading. We don't calculate partial differentials, we calculate total differences. How each term is calculated as a difference will be explicated.

R3: *“1a. It is not directly clear how the different processes that contribute to rainfall isotope change are actually calculated. For example, how is dRvi (Eq 3) actually calculated? What region is used to define Rvi and Rvo so as to calculate dRvi? Likewise, where is q0 determined (eq 4), T0 (eq. 6) etc. A much more detailed explanation is needed I think. How is the change in elevation contribution to rainfall isotopes calculated (it depends upon dz times lapse rate but also depends upon the Rayleigh distillation, saturation relationship to T). What is the actual analytical term for the partial differential in Eg. 8 ?”*

A: We add more detailed explanations and a specific equation for each analytical term from the Eq. 8 (see the end of this document). We assume that the δ*Rvi* term is a residual part of the vapor isotopic difference that accounts for processes of deep convection and air mass mixing. Equations from section 2 (see the decomposition method below point-by-point response) show that we do not need to know δ*Rvi* to estimate its contribution to the total isotopic change between two cases. For parameters z0, q0 and T0 we took values over New Delhi region to be consistent with previous isotopic studies over the region, but the sensitivity to these arbitrary choices will be tested. Even if initial conditions for the Rayleigh distillation vary depending on the atmospheric circulation, on deep convective processes and on the site of interest, we keep the same reference values, and all variations in initial conditions are accommodated by δ*Rvi*.

R3: *“1b. Furthermore, it is not abundantly clear how the partial differentials in eq. 8 were evaluated. For example, dRp/dRvi (1st partial differential in Eq. 8) depends upon f, etc. (e.g. change in rainfall Rp depends upon initial vapor composition plus modification by isotope fractionation, the magnitude in per-mil also depends upon the initial vapor composition). All of these partial differentials need to be more explicitly evaluated analytically in the paper and the dependence on the sensitivity of the partial differentials to the state at which they are calculated has to be demonstrated as not a factor (as is assumed on p7 L5).”*

A: To make the decomposition processes clearer we address the reviewer to the reworked methodological part (see below point-by-point response), in which we added supplemental information and equations describing the evaluation of the five partial differentials.

We carried sensitivity tests for partial differentials: results are presented in two additional tables for the southern and northern regions (Tabl. 1 and Tabl. 2). The sensitivity to Rv0 will remain as a multiplying factor. In addition, in the new version we will provide a test of the sensitivity of decomposition terms to the state at which they are calculated.

	Northern Region			South region		
	T0	Rh0	Rv0	T0	Rh0	Rv0
$\Delta R_{p,\Delta z}$	0,15	0,33	0,667	0,12	0,25	0,51
$\Delta R_{p,\Delta\delta T_s}$	0,09	0,02	0,04	0.12	0.06	0,13
$\Delta R_{p,\Delta h}$	0	0,351	0,66	0	0,19	0,83
$\Delta R_{p,\Delta\delta R_{vi}}$	0	0	0.05	0	0	0,52
$\Delta R_{p,\Delta\varepsilon}$	0	0	0	0	0	0

Tabl. 1. INT-LOW Sensitivity of the decomposition terms (in ‰) to the change of 1°C of T0 and 10 % of Rh0 and 1 ‰ of Rv0.

	Northern Region			South region		
	T0	Rh0	Rv0	T0	Rh0	Rv0
$\Delta R_{p,\Delta z}$	0,36	0,6	1,4	0,3	0,59	1,2
$\Delta R_{p,\Delta\delta T_s}$	0,34	0,09	0,18	0,31	0,02	0,05
$\Delta R_{p,\Delta h}$	0	0,78	0,9	0	0,57	0,47
$\Delta R_{p,\Delta\delta R_{vi}}$	0	0	0,85	0	0	0,67
$\Delta R_{p,\Delta\varepsilon}$	0	0	0	0	0	0

Tabl. 2. MOD-INT Sensitivity of the decomposition terms (in ‰) to the change of 1° of T0, 0,1 of Rh0, 1 ‰ of Rv0

R3: *"1c. Decomposition of the absolute humidity term into temperature and relative humidity means that attribution of $\delta^{18}O$ changes to Ts or rh changes depends upon how well the model really captures those two values."*

A: We fully agree. The question of model validation has been also raised by the Anonymous reviewer #2. In the current manuscript version we compare MOD run outputs with rainfall data from the Climate Research Unit (CRU) [*New et al.*, 2002]. Corresponding figure is in the supplementary (Fig. S1). Also following reviewers recommendations we have added a comparison of humidity transport between LMDZ-iso MOD simulation outputs and ERA-40 re-analysis data [*Uppala et al., 2005*]. Our MOD simulation is preindustrial, consequently a comparison with modern data is expected to provide differences driven by the pre-industrial boundary conditions. Still comparing LMDZ-iso outputs with mean annual temperatures from CRU dataset [*New et al.*, 2002] (Fig. N1) and relative humidity from NCEP-DOE Reanalysis. Fig. N2 shows that LMDZ-iso model captures these variables reasonably well.

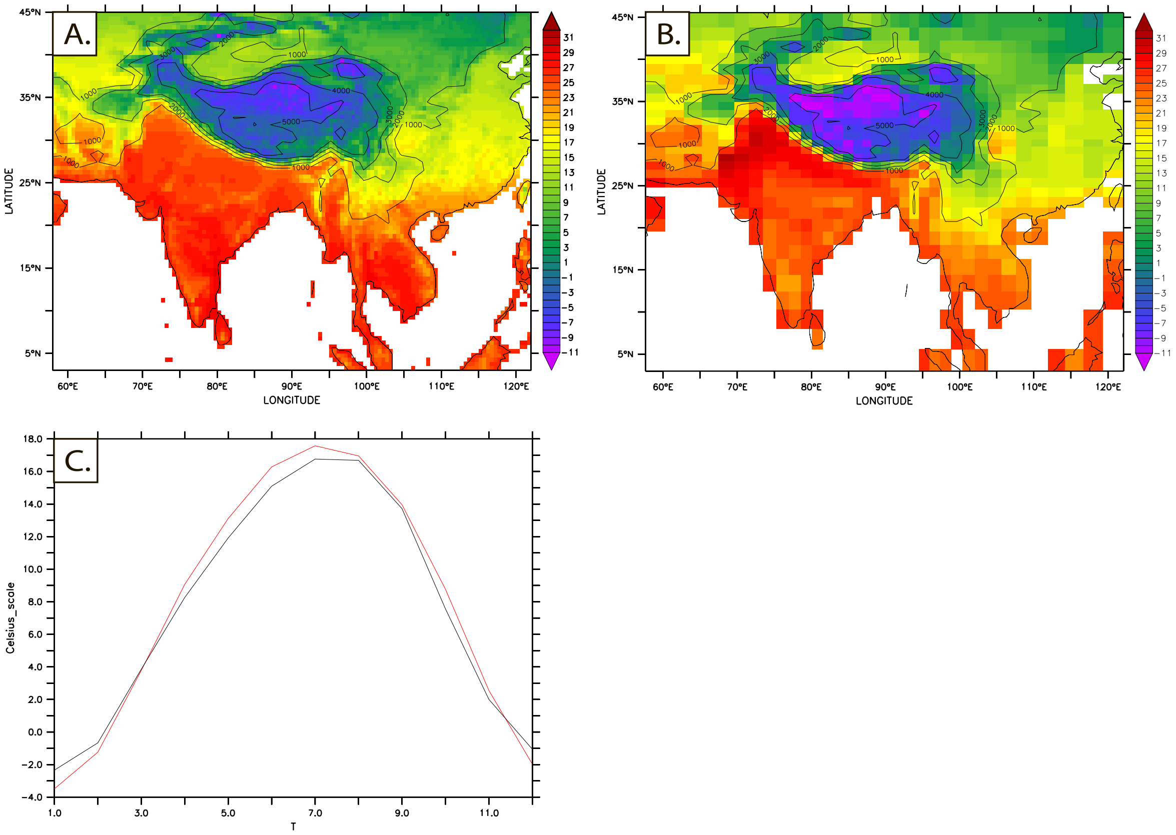

Fig. N1. Mean annual temperature from A) the Climate Research Unit (CRU) [*New et al.*, 2002] dataset and B) LMDZ-iso simulated for the MOD experiment. Figure (C) represent the seasonal cycles of temperature spatially averaged from 25°N to 40°N and from 70°E to 110°E for the MOD experiment (black) and for the CRU dataset (red).

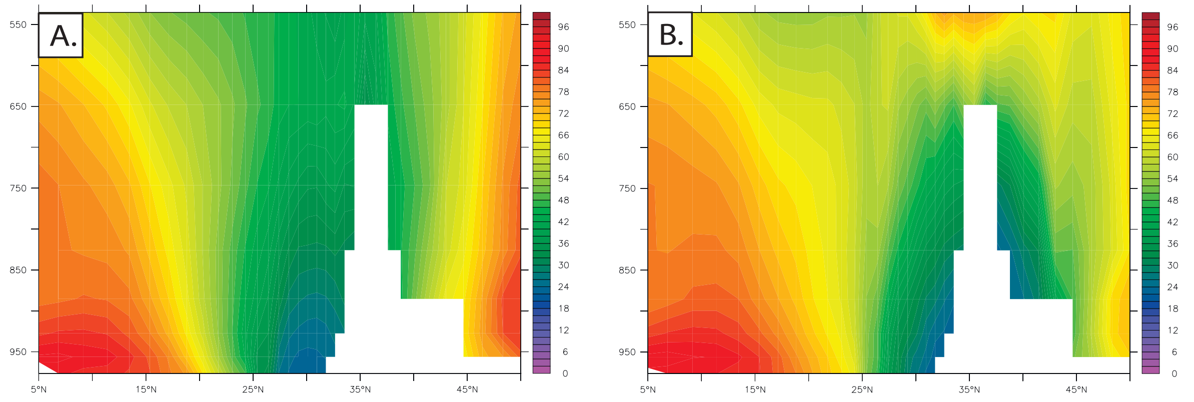

Fig. N2. Mean annual relative humidity profiles for A) NCEP-DOE Reanalysis and B) LMDZ-iso simulated for the MOD experiment.

R3: *"Another, and more fundamental way to partition the rainfall $\delta^{18}O$ changes would be to determine how much the specific humidity changes (or T*) contribute to the isotope changes. This gets at the transport and rainout process more directly. The local T and rh values are not really the source of the isotope changes, T* or q/qo (q[T*]/qo) are the actual cause of the changes (e.g. eq. 4). I would much rather the authors use T* or q/qo changes in their decomposition of the isotope changes (or at least do this in parallel to the T, rh decomposition as they are equivalent). In equation 8 this would combine the Dh and DdTs terms into a single q or T* term. Intuitively this makes more sense: Rp is a function of Rvi, epsilon, and q."*

A: One of the main purposes of the paper is to estimate the value of the $\frac{\partial R_p}{\partial z} \cdot \Delta z$ term. Altitude acts through temperature, this is why we chose to extract the temperature signal. We suggest to add an addition panel to the Fig. 8 and Fig. 9 that shows the part of the isotopic signal associated with the part of the change of the specific humidity with the uplift that is not associated with elevation (here Fig. N3). Nevertheless, we insist to keep all terms in the equation 8 as discussed below.

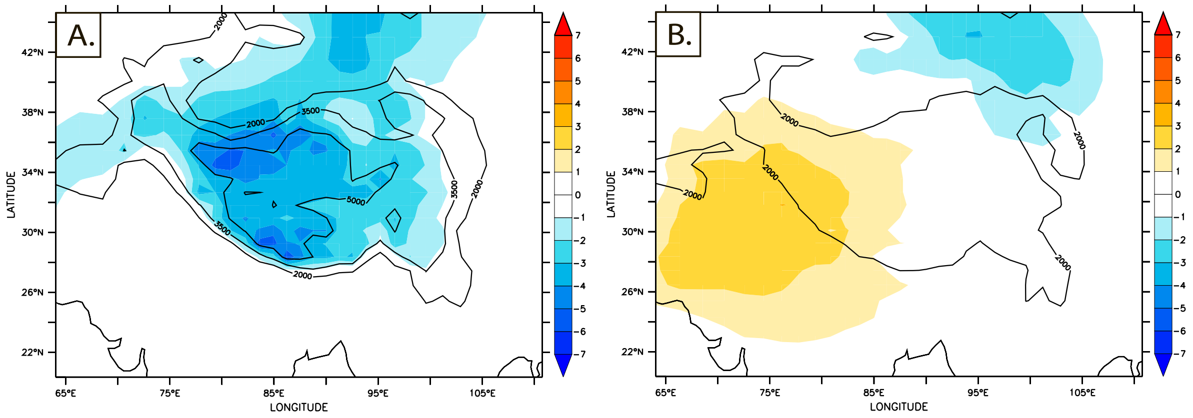

Fig. N3. The effect of specific humidity change

R3: *"1d. The dz term in Eq. 8 is really a temperature change (or T* change as is assumed in a Rowley type model - actually a Pierrehumbert-type model: Pierrehumbert, 1999 Huascaran d18O as an indicator of tropical climate during the Last Glacial Maximum). So, the q term could be decomposed into an elevation term and a non- elevation term. This would still keep the decomposition focused on the fundamental q/qo term rather than Ts and rh."*

A: Our model is equivalent to that of Rowley et al (2001) for $\delta R_{vi} = 0$ (i.e. neglecting the effects of mixing and deep convection on the initial water vapor), $\varepsilon = (a - 1)*R_v$ (i.e.

neglecting post-condensational effects), and h=1 (i.e. assuming the site of interest is inside the precipitating cloud). We agree that the most important point of the paper is the division of the total isotopic signal into an "elevation" and "non-elevation" term. However, we think that the relative humidity component is important from a physical point of view because it reflects the large-scale circulation: how high did the last saturation occur? Is the regime under large-scale ascent or descent? Actually we are showing an elevation term on Fig. 8 B and Fig. 9 B according to the uplift stage and non-elevation terms of Fig. 10 A and Fig. 11 A.

R3: *"1e. Are the values used to decompose the rainfall isotope changes precipitation weighted? That is, are q, T etc. weighted by the precipitation amount in a particular location? One of the reasons that there is such a strong relationship between isotopes and elevation is that the relationship is set only when it rains. This is a very small subset of all atmospheric conditions and thus much of the variability really doesn't matter. It only matters if it is actually raining and the isotope signal is being "sampled." In my opinion, rainfall weighted climate variables are essential for properly decomposing the isotope signal."*

A: In our calculations only δ^{18}O values are weighted by the precipitation amount, but the climatic variables are not weighted. We have now recalculated all contributions using precipitation-weighted variables. In doing so, we used monthly outputs, so that the effects of seasonality are taken into account by the precipitation weighting. However, we do not have the daily outputs. So the effects of precipitation intermittency at the daily time scale won't be taken into account. We will acknowledge this limitation in the revised manuscript. Also, we checked that the decomposition terms calculated for the summer period with a large number of days when it rains is not essentially different from those calculated using mean annual values (fig. N4)

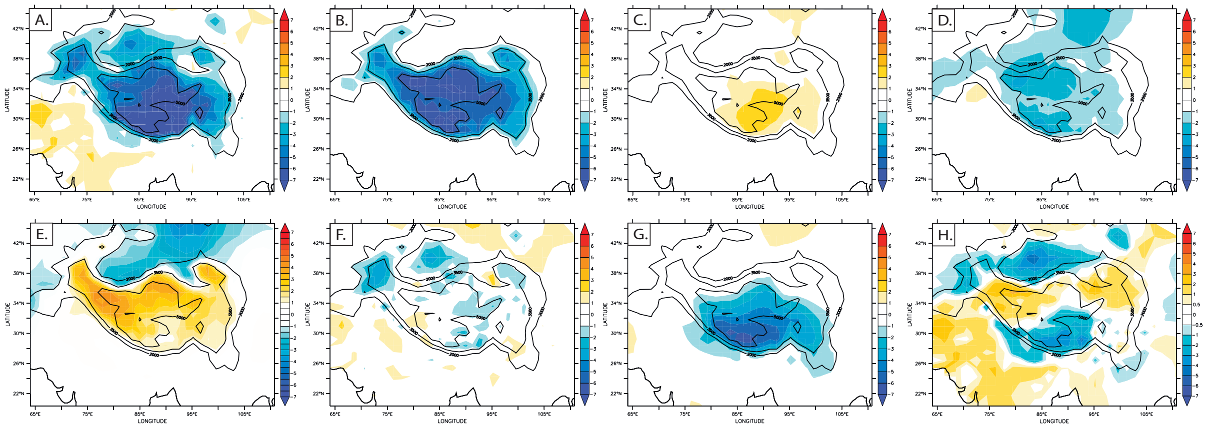

Fig N4. MJJAS total isotopic difference between MOD and INT experiments (ΔR_p) and spatial isotopic variations related to: (B) direct effect of topography changes, (C) effect of lapse rate change, associated with non-adiabatic effects, (D) effect of specific humidity, (E) effect of local relative humidity change, (F) effect of changes in post-condensational processes, (G) all other effect, (H) difference in $\delta^{18}O_p$ between MOD and INT experiments that is not related to direct effect of topography changes

R3: *"2. I think there are major and important differences in the modeled and observed rainfall isotopes (e.g. Fig. 7). This is particularly true for the northern plateau where isotope values in rainfall become so positive that they are nearly identical to low elevation rainfall south of the Himalaya. I disagree with the author's statement that these highly enriched isotope values are from surface processes (p9 l16-20). Bershaw et al, 2012 were discussing the Pamir region to the west and even in their data there is not evidence of non-equilibrium fractionation as would be expected from kinetic effects. Overall there is not a strong d- excess signal on the plateau as would be expected for evaporative processes thus the data indicate a robust positive isotope signal in rainfall values."*

A: We agree with the reviewer that over the northern part of the Plateau there are some model-data discrepancies that could not be explained by the surface processes. On the contrary, we would like to pay attention on the very good model-data fit of isotopic data over the northern-east slope of the TP (Bershaw et al., 2011). Over the northern margins of the TP, modelled δ^{18}O in precipitation is more negative than observations show. This model-data discrepancies may result from 3 types of uncertainties: 1) linked with the model resolution. Despite quite a high resolution that we are able to obtain with a zoomed grid, the relief could be not represented well at some parts of the TP, and 2) overestimation by the model of the westerlies flux (see the comparison with the ERA moisture transport) that probably lead to underestimation of δ^{18}O over the northern part of the TP. Our statement about the contribution of the surface processes to more positive values over the central part *(p9 l16-20)* of the TP (data of Quade) is consistent with the Quade explanation of the increased role of the continental recycling northward from the Himalayas crest.

R3: "*This feature, its interpretation, and whether and why it persists in the past are perhaps some of the major questions in Tibetan paleoaltimetry. If it is a persistent feature then ancient isotope values from central and northern Tibet that look like today's values may have come from modern-like elevations. If this is not a persistent feature (e.g. an arm of the Tethys north of Tibet would provide local moisture) then ancient isotope values that are the same as today may actually mean the site was at a low elevation.*"

A: We agree. In this paper we provide only sensitivity experiments with reduced topography. The influence of realistic paleogeography (eg. with the Tethys Sea, altered paleogeography) on the isotopic composition of precipitation is a topic of our further studies. However, the reviewer's point is very important and we discuss the possible sources of uncertainties while comparison with deep paleo data and further studies directions on p18 l7-10.

R3: "*My recommendations for this issue are twofold. First, I would like to see two scatterplots in addition to the heat map. The first would be the observed vs. modeled $\delta^{18}O$ rainfall from Fig. 7A along with RMSE estimates. This plot would show how well the model really captures isotope values regardless of location. The second would have latitude as an x-axis and actual observations of oxygen isotope values as the y axis along with values from the model as a continuous line. Values could be from a swath beginning at the south and extending north along the central axis of the plateau or projected in from the plateau. This plot would show how well the model gets the overall isotope gradient even if it doesn't get the absolute values correctly.*

Second would be a thorough discussion of what this mismatch means for interpretation of the model experiments. If the model is missing or underrepresents some moisture source or process that is important today on the northern plateau then what does this mean for the conclusions from the model experiments?"

A: Thank you for this recommendation. On the Fig. N5 observed vs. modeled δ^{18}O rainfall scatter plot is presented with a linear regression. Modeled vs observed data show quite a good correlation with a Person coefficient of 0,8646. Fig N6 shows a map of modeled δ^{18}O for the MOD experiment overploted by observed data values and a south-north transection (averaged between 70 and 100° E) of modelled values (black line) and projected in observed values of δ^{18}O. The general south-north isotopic gradient is simulated perfectly well by the model.

After the Himalayan crest δ^{18}Op values become more positive that is consistent with a South-North trend observed by Quade et al. [*Quade et al.*, 2007; *Bershaw et al.*, 2012b].

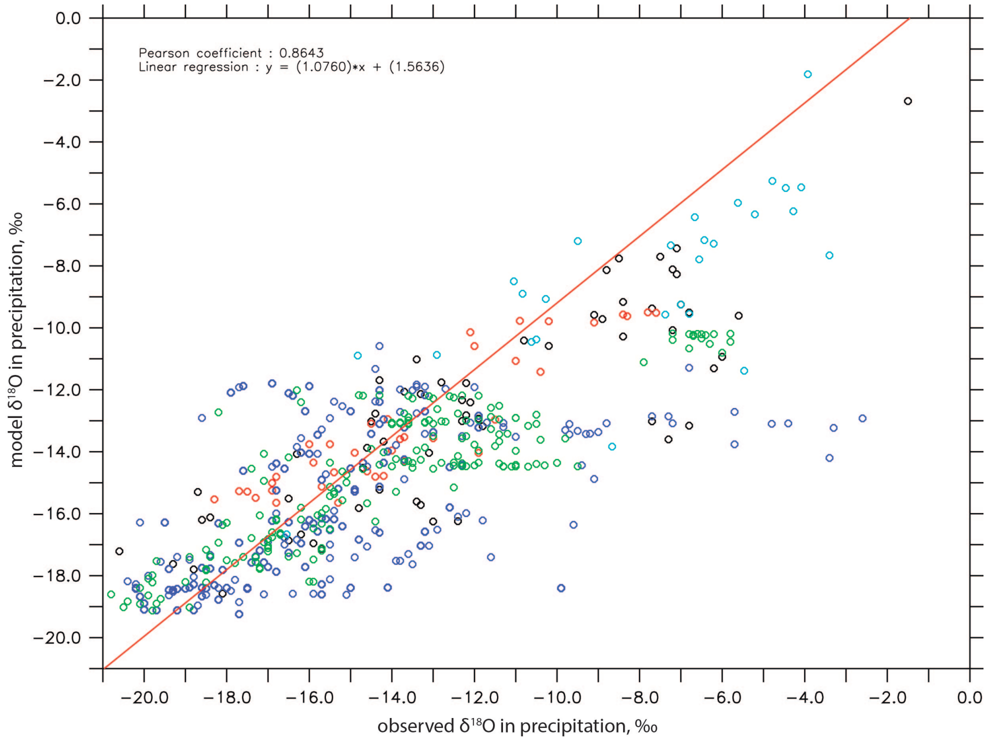

Fig. N5. Model vs observed δ^{18}O in precipitation. The colour of circles corresponds to the data set: red – Bershaw et al, 2012, blue – Quade et al, 2011, green – Hren et al, 2009, black – Caves et al, 2015, light blue show mean annual data from GNIP stations. Red line shows a linear regression.

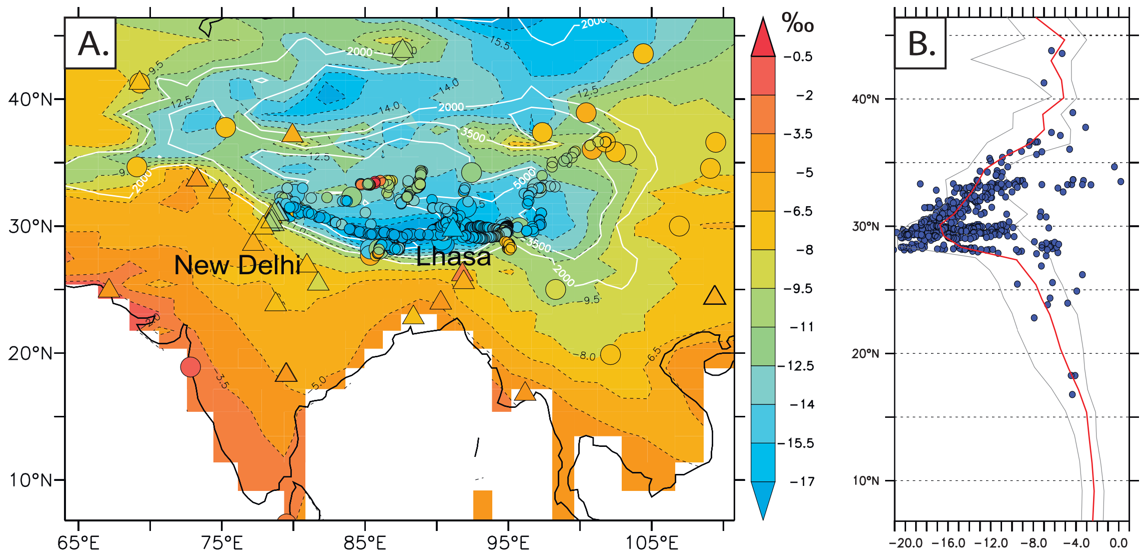

Fig. N6. A) Annual mean δ^{18}O in precipitation simulated by LMDZ-iso for the MOD case and B) S-N profiles of model simulated δ^{18}O in precipitation for the MOD case. Points correspond to present-day δ^{18}Op from published data (Bershaw et al, 2012, Quade et al, 2011, Hren et al, 2009, Caves et al, 2015), and mean annual data from GNIP stations). Solid black line shows model δ^{18}O values averaged between 70º E and 100º E. Grey lines show minimum and maximum values for the selected range of longitudes.

R3: *"3. Cite the original sources for precipitation isotope values on the plateau. These authors should get credit for the major amount of work it takes to generate this type of data and all the credit shouldn't go to the Caves 2015 compilation."*

A: Thank you for this remark. We have added the references to the original papers.

R3: *"One of the major conclusions of the paper is that there is a non-linear effect of elevation changes on isotope values. One expects a non-linear relationship between rainfall isotope values and elevation simply because of the non-linearity of (i) saturated adiabats, (ii) the saturation vapor pressure curve with temperature and (iii) the Rayleigh distillation process itself. Thus the null hypothesis is that isotope changes with elevation from low to intermediate elevations would be less than isotope changes from intermediate to high elevations. Whether the changes are greater than can be explained by the null hypothesis needs to be demonstrated. But, the qualitative observation itself is actually expected from theory. An additional plot would drive this home. What does a Rowley (Pierrehumbert) type model predict for isotope change with elevation and where do these GCM models plot? I would focus this plot on the Himalayan mountain region as this is where the simple model is most applicable. This would be an incredibly useful plot for folks that want to take lessons away from this paper. How similar/different are the results in this paper from a simple model that has been extensively used to reconstruct elevation?"*

A: We agree with the reviewer that our conclusion about non-linearity of the magnitude of the isotopic changes between the initial and the terminal stage of the uplift is a logical consequence of non-linearity of saturated adiabats, the saturation vapour pressure curve with temperature and the Rayleigh distillation process. However we find useful to stress this interesting characteristic for the geological community since such a conclusion has never been published before. The estimation whether the changes are greater than can be explained by the null hypothesis is out of scope of the paper and may be a subject of another study. Here we suggest an additional figure (Fig. N7) showing $\Delta(\delta^{18}O)$ vs elevation for MOD and INT simulations and Rowley type model (Rowley et al., 2001).

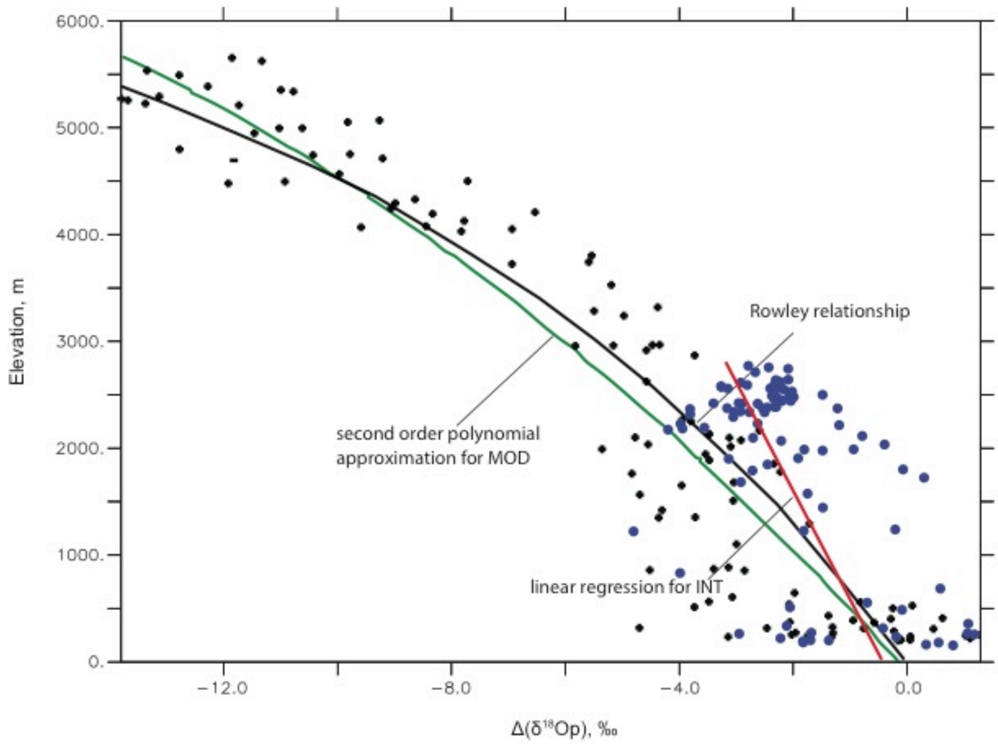

Fig. N7. Δ(δ^{18}O) vs elevation for MOD (black points) and INT (blue points) and isotopic gradients for the southern region (between 25°N and 30°N). Black line shows relationship from the empirical model (Rowley, 2001; Rowley and Garzione, 2007). Green line shows second order polynomial approximation of simulated MOD δ^{18}O values. Red line shows a linear regression for the INT δ^{18}O values.

R3 : *" There is a bit of a cottage industry in the isotope-enabled GCM field looking at how isotope paleoaltimetry does/doesn't work in different orogenic systems. I would encourage the authors not to fall into the trap of saying "its complicated and you need to take additional factors into account." (This is essentially what is said at the end of the abstract.) Rather, make the information accessible and useful to the readers. Be specific and helpful in the abstract and throughout so that it is directly clear to the readers what specific factors are actually important and how they should be accounted for when reconstructing paleoaltimetry."*

A: We totally agree. Text will be corrected accordingly. We also provide estimates of contribution of different decomposition terms to the total isotopic signal for locations where previous paleoelevation studies have been done in the Table. 3.

Response to Specific Comments:

R3: *"P2 L16-18 – Not all of these references are carbonates or oxygen isotopes"*

A: Thank you, these references has been replaced by: (Currie et al., 2005; DeCelles et al., 2007; Garzione et al., 2000; Rowley and Currie, 2006; Saylor et al., 2009; Xu et al., 2013; Li et al., 2015)

R3: "*P3 L4 – most studies do take into account changing seawater $\delta^{18}O$ either implicitly through normalization to a low elevation rainfall site or explicitly through correction using various estimates.*"
A: Our purpose here was to show how climate changes have been thought to change the δ^{18}O record. We don't discuss here where these corrections used to be applied to paleoelevation reconstructions. We suggest to modify this sentence in this way: "Moreover, it has been suggested that climate-driven changes in surface ocean δ^{18}O through the Cenozoic can also influence recorded values of precipitation δ^{18}O over the continent and an appropriate corrections has been applied in most modern studies."

R3: *"P3 L21 – Seems that studies of Ramstein and Fluteau should be mentioned here."*
A: Thank you, we have added this reference.

R3: *"P7 L5 – Is the assumption that sensitivity of the partial derivatives to state is not important ok? It seems this would be fairly easy to test by some simple calculations and then it could be definitively stated."*
A: We will address this point in the revised version of the manuscript.

R3: *"P11 L1 – Are the Ddts values precipitation weighted? In general are the climatic variables used in the decomposition precipitation weighted? They should be."*
A: Now we weight them by precipitation.

R3: *"P11 L7-10 – How much of the temperature changes are due to comparison to a constant adiabat for all experiments? The adiabats that matter are only ones when moisture is being transported on to the plateau (non-linear with elevation) and the slope should change with T and qo. Thus, inference of non-adiabatic temperature changes could simply reflect the way this is calculated and not actual changes."*
A: We thank reviewer for making a point on this. First of all we need to note that using of one mean lapse rate (in this paper equal to 5° km^{-1} based on the measurements of modern observed mean temperature lapse rate on the southern slope of the central Himalayas) is clearly idealized. In our calculations we neglect the non-linearity of lapse rate. We agree that in addition to the effect of climate changes, the lapse rate is also affected by the humidity and

temperature of the rising parcels. We will acknowledge this limitation in the revised version.

R3: *"P12 L28-30 – One expects a non-linear relationship between rainfall isotope values and elevation simply because of the non-linearity of (i) saturated adiabats, (ii) the saturation vapor pressure curve with temperature and (iii) the Rayleigh distillation process itself. Thus the null hypothesis is that isotope changes with elevation from low to intermediate elevations would be less than isotope changes from intermediate to high elevations. Whether the changes are greater than can be explained by the null hypothesis needs to be demonstrated. But, the qualitative observation itself is actually expected from theory."*

A: We agree. But interestingly, up to our knowledge, this null hypothesis that isotope changes with elevation from low to intermediate elevations would be less than isotope changes from intermediate to high elevations have never been discussed before in the paleoaltimetry literature. We agree with the reviewer that this conclusion is a logic consequence of non-linearity of saturated adiabats, the saturation vapour pressure curve with temperature and the Rayleigh distillation process. However we find useful to stress this interesting characteristic for the geological community. The estimation whether the changes are greater than can be explained by the null hypothesis is out of scope of the paper and may be a subject of another study.

R3: *"P14 L1 – Effects from post-condensation re-evaporation. This should have a distinct d-excess signal that should be evident in the model values. Examination of the d-excess signal spatially could directly answer this question."*

A: Unfortunately we didn't include hydrogen isotopes in the simulations presented in this paper. Zoomed experiments including the isotopes are very expensive in terms of computation time (about 700 days of single CPU core time per experiment with only oxygen isotopes) and the calculation time increases linearly with every one additional isotope. However, with the numerical simulation we have an access to $\delta^{18}O$ in both precipitation and vapour that gives a possibility to estimate the magnitude of post-condensation processes without appealing to the d-excess.

R3: *"P15 L1 – How do these results compare with those of Boos and Kuang (2010)?"*

A: Although our purpose totally differs from B&K2010, our results in terms of monsoon dynamics seem very consistent. The no-elevation run from B&K2010 depicts a weaker monsoon and lower rainfall over Asia (Fig3b, Fig4A in their study).

R3: *“P17 L15-16 – “Paleoelevation studies indicate the Himalayas attained their current elevation by the late Miocene.” This is not correct. Rowley and Currie (2006) and subsequent authors indicate earlier timing for modern elevations (middle Eocene or earlier).”*

A: Thank you for this correction. In fact, we show that for the southern part of the TP and Himalayas, paleoelevations based on stable oxygen isotopes measurement could be overestimated. We will modify our text accordingly.

1 Theoretical framework for the precipitation composition

Our goal is to understand why the precipitation composition varies depending on climate or geography, and to quantify the direct contribution of topography changes. To do so, we develop a theoretical expression for the precipitation composition.

To first order, the composition of the precipitation R_p follows that of the vapor R_v. Deviations from the vapor composition, $\epsilon = R_p - R_v$, are associated with local condensational or post-condensational process.

$$R_p = R_v + \epsilon \tag{1}$$

In an idealized framework of an isolated air parcel transported from an initial site at low altitude to the site of interest (figure **??**), the vapor composition can be predicted by Rayleigh distillation:

$$R_v = R_{vi} \cdot f^{\alpha-1} \tag{2}$$

where R_{vi} is the initial composition of the vapor at the initial site and f is the residual fraction of the vapor at the site of interest relatively to the initial site. We take the initial site as characterized by a temperature and humidity T_0 and q_0. Under these conditions, we note R_{v0} the theoretical isotopic composition that it would have if all the vapor originated from local evaporation over quiescent oceanic conditions. Depending on the atmospheric circulation, on deep convective and mixing processes and on the site of interest, the initial site may be characterized by a different isotopic composition:

$$R_{vi} = R_{v0} + \delta R_{vi} \tag{3}$$

The residual fraction f depends on the specific humidity q at the site of interest:

$$f = \frac{q}{q_0} \tag{4}$$

The air is not always saturated near the surface:

$$q = h \cdot q_s(T_s) \tag{5}$$

where h, T_s and are the relative humidity and air temperature near the surface of the site of interest. The air can be under-saturated because it can be considered as air that has been transported adiabatically from the area of minimum condensation temperature, T^* (Sherwood 1996, Galeswky et al 2005, Galewsky and Hurley 2010): $q = q_s\left(T^*\right)$.

The surface temperature can be predicted to first order by the adiabatic lapse rate, Γ, and is modulated by a non-adiabatic component, δT_s that represents processes such as large-scale circulation or radiation:

$$T_s = T_0 + \Gamma \cdot (z - z_0) + \delta T_s \tag{6}$$

where z and z_0 are the altitudes at the site of interest and at the initial site.

If we combine equations 1 to 6, we get that R_v is a function of δR_{vi}, ϵ, h, δT_s and z:

$$R_p = R_p\left(\epsilon, \delta R_{vi}, h, \delta T_s, z\right) \tag{7}$$

Parameters z_0, q_0, T_0 are reference values that are common to all sites of interest, all climates and geographies. Even if initial conditions for the Rayleigh distillation vary depending on the atmospheric circulation, on deep convective processes and on the site of interest, we keep the same reference values and all variations in initial conditions are accomodated by δR_{vi}.

This model is equivalent to that of Rowley et al for $\delta R_{vi} = 0$ (i.e. . neglecting the effects of mixing and deep convection on the initial water vapor), $\epsilon = (\alpha - 1) \cdot R_v$ (i.e. neglecting post-condensational effects), and $h = 1$ (i.e. assuming the site of interest is inside the precipitating cloud).

2 Decomposition of precipitation composition differences

Our goal is to understand why R_p varies from one climatic state to another. Let's refer to these climatic states using subscript 1 and 2 and to their difference using the Δ notation. We decompose $\Delta R_p = R_{p2} - R_{p1}$ into contributions from $\Delta\epsilon$, $\Delta\delta R_{vi}$, Δh, $\Delta\delta T_s$ and Δz:

$$\Delta R_p = \Delta R_{p,\Delta\epsilon} + \Delta R_{p,\Delta\delta R_{vi}} + \Delta R_{p,\Delta h} + \Delta R_{p,\Delta\delta T_s} + \Delta R_{p,\Delta z} \tag{8}$$

where $\Delta R_{p,\Delta\epsilon}$, $\Delta R_{p,\Delta\delta R_{vi}} \Delta R_{p,\Delta h}$, $\Delta R_{p,\Delta\delta T_s}$ and $\Delta R_{p,\Delta z}$ are respectively the contributions of $\Delta\epsilon$, $\Delta\delta R_{vi}$, Δh, $\Delta\delta T_s$ and Δz to ΔR_p. These terms are diagnosed from LMDZ outputs.

Although it may look complicated and fastidious, we present below how the different terms are calculated step-by-step. The physical meaning of the different terms and the use of LMDZ outputs is more apparent this way. However, the decomposition terms can also be written in a shorter and more formal way as detailed in table 8.

First, based on equation 1, the contribution of $\Delta\epsilon$ is estimated as:

$$\Delta R_{p,\Delta\epsilon} = R_{p2} - R_{v2} - (R_{p1} - R_{v1})$$

where R_{p1}, R_{p2}, R_{v1} and R_{v2} are the isotopic ratios in precipitation and in water vapor simulated by LMDZ for climatic states 1 and 2.

After isolating the effect of condensational and post-condensational processes, only the contribution of the water vapor isotopic ratio change, ΔR_v, remain. We then isolate the effect of Rayleigh processes. Based on equations 2, if $\delta R_{vi} = 0$, i.e. if the initial conditions were T_0, q_0 and R_{v0}, then the water vapor isotopic ratio change would be:

$$\Delta R_{p,q} = R_{v0} \cdot \left((q_2/q_0)^{\alpha_2 - 1} - (q_1/q_0)^{\alpha_1 - 1} \right)$$

where R_{v1} , R_{v2}, q_1, q_2 are all diagnosed from LMDZ simulations and α_1 and α_2 are calculated as functions of T_{s2} and T_{s1}, also diagnosed from LMDZ simulations. We estimate as a residual the contribution from all processes affecting the initial conditions of the Rayleigh distillation or disturbing the evolution of the water vapor isotopic ratio predicted by Rayleigh:

$$\Delta R_{p,\Delta\delta R_{vi}} = R_{v2} - R_{v1} - \Delta R_{p,q}$$

This way, we do not need to know δR_{vi} to calculate $\Delta R_{p,\Delta\delta R_{vi}}$. We do not need to know neither where the moisture comes from nor how convection and mixing modifies it along trajectories.

Then, based on equation 5, specific humidity q is decomposed into h and T_s terms. If h was identical for all climatic states, e.g. $h = h_1$, then the water vapor isotopic ratio change would be:

$$\Delta R_{p,T_s} = R_{v0} \cdot \left(\left(h_1 \cdot q_s\left(T_{s2}\right)/q_0\right)^{\alpha_2 - 1} - \left(h_1 \cdot q_s\left(T_{s1}\right)/q_0\right)^{\alpha_1 - 1} \right)$$

where T_{s1} and T_{s2} are diagnosed from LMDZ simulations. We calculate the contribution of h changes as:

$$\Delta R_{p,h} = \Delta R_{p,q} - \Delta R_{p,T_s}$$

Then, based on equation 6, we further decompose T_s into z and δT_s terms. If $\delta T_s = 0$, then the water vapor isotopic ratio change would be:

$$\Delta R_{p,z} = R_{v0} \cdot \left(\left(q_s\left(T_0 + \Gamma \cdot (z_2 - z_0)\right)/q_0\right)^{\alpha_2 - 1} - \left(q_s\left(T_0 + \Gamma \cdot (z_1 - z_0)\right)/q_0\right)^{\alpha_1 - 1} \right)$$

where z_1 and z_2 are the altitudes used to force LMDZ.

We finally calculate the contribution of δT_s changes as:

$$\Delta R_{p,\delta T_s} = \Delta R_{p,T_s} - \Delta R_{p,z}$$

Term written with differential format	Estimate of these terms	Physical meaning
ΔR_p	$R_p\left(\epsilon_2, \delta R_{vi2}, h_2, \delta T_{s2}, z_2\right) - R_p\left(\epsilon_1, \delta R_{vi1}, h_1, \delta T_{s1}, z_1\right)$	Total isotopic difference between state 1 and state 2
$\Delta R_{p,\Delta z}$	$R_p\left(0, 0, 1, 0, z_2\right) - R_p\left(0, 0, 1, 0, z_1\right)$	Direct effect of topography change
$\Delta R_{p,\Delta\delta T_s}$	$R_p\left(0, 0, 1, \delta T_{s2}, z_2\right) - R_p\left(0, 0, 1, \delta T_{s1}, z_1\right)$ $- \left(R_p\left(0, 0, 1, 0, z_2\right) - R_p\left(0, 0, 1, 0, z_1\right)\right)$	Effect of lapse rate change, associated with non-adiabatic effects, possibly due to changes in surface energy budget or in large-scale atmospheric stratification
$\Delta R_{p,\Delta h}$	$R_p\left(0, 0, h_2, \delta T_{s2}, z_2\right) - R_p\left(0, 0, h_1, \delta T_{s1}, z_1\right)$ $- \left(R_p\left(0, 0, h_1, \delta T_{s2}, z_2\right) - R_p\left(0, 0, h_1, \delta T_{s1}, z_1\right)\right)$	Effect of local relative humidity change, possibly due to large-scale circulation changes
$\Delta R_{p,\Delta\delta R_{vi}}$	$R_p\left(0, \delta R_{vi2}, h_2, \delta T_{s2}, z_2\right) - R_p\left(0, \delta R_{vi1}, h_1, \delta T_{s1}, z_1\right)$ $- \left(R_p\left(0, 0, h_2, \delta T_{s2}, z_2\right) - R_p\left(0, 0, h_1, \delta T_{s1}, z_1\right)\right)$	All other effects, including effects of deep convection, mixing, water vapor origin, continental recycling on the initial water vapor
$\Delta R_{p,\Delta\epsilon}$	$R_p\left(\epsilon_2, \delta R_{vi2}, h_2, \delta T_{s2}, z_2\right) - R_p\left(\epsilon_1, \delta R_{vi1}, h_1, \delta T_{s1}, z_1\right)$ $- \left(R_p\left(0, \delta R_{vi2}, h_2, \delta T_{s2}, z_2\right) - R_p\left(0, \delta R_{vi1}, h_2, \delta T_{s1}, z_1\right)\right)$	Effect of changes in condensational and post-condensational effects, possibly due to changes in rain reevaporation processes

Table 1: Table detailing how the different terms of the decomposition for ΔR_p, as written in equation 8, are estimated.