Response to referee #1

Major comments

1.1 Give some overall view and more physical interpretation

The paper directly dives into complicated statistical diagnostics. But before this, I think some overview would be useful. For example, a few **basic figures showing the maps** of simulated δ 18Oprecip anomalies for a few key periods would be useful before showing the regional averages. When describing the results of the statistical methods, it would be useful to better guide the reader in the **physical interpretation of the figures**: what does it mean when values are more negative, positive, larger, smaller... (more details in minor comments.) At the end of each sub-section, a few sentences would be useful to **summarize the results** in terms of physical understanding of the processes driving the isotopic variability. A statistical analysis is not enough to identify causality and thus isotopic "drivers", so the discussion should rely more on the huge body of literature devoted to the interpretation of water isotopic records in monsoon regions.

The focus of this paper is not to use isotope-enable models directly to explain observed changes in the speleothem records but rather to use statistical approaches to explore patterns in the observations and model outputs, on the assumption that consistency between the two reveals physically plausible explanations of regional speleothem changes. We will modify the introduction to make the logic of using statistical analyses combining observations and model outputs clearer (please see specific modifications below). Since we are using previously published model results, it does not seem necessary to include a separate section describing these results. However, we will include anomaly maps of simulated $\delta^{18}O_{precip}$ and speleothem $\delta^{18}O$ data from SISAL v2 in the supplementary material and refer to these maps in the main text.

Although the statistical approaches we are using (PCoA, multiple regression, z-scores) are not commonly applied to speleothem records, they are standard techniques for analysing other kinds of environmental data. However, to guide readers through these statistical analyses, we will revise the text in the methods and results sections to clearly describe what each analysis or figure shows, as follows:

Section 2.3 (line 167): "PCoA results were displayed as a biplot, where sites ordinated close to one another (i.e., with similar PCoA scores) show similar trends and sites ordinated far apart have dissimilar trends."

Section 3.1 (line 268): "PCoA shows the (dis)similarity of Holocene $\delta^{18}O_{spel}$ evolution across individual records, and thus allows an objective regionalisation of these records."

Section 3.2 (line 280): "To investigate the causes of glacial-interglacial shifts in $\delta^{18}O$, we compare simulated and observed regional $\delta^{18}O$ signals during the LIG, LGM and MH with shifts in climate variables (precipitation and temperature)."

Section 3.4 (line 319): "The MLR analyses of simulated $\delta^{18}O_{precip}$ trends identify the impact of an individual climate variable on $\delta^{18}O_{precip}$ in the absence of changes in other variables."

We agree with the reviewer that statistical relationships do not necessarily indicate causal relationships. Generally, explanations of the causes of observed $\delta^{18}O_{spel}$ variability either rely on modern $\delta^{18}O$-climate observations and assume that these are constant through time (e.g. Sinha et

al., 2015), or interpret changes by comparison with other palaeoclimate reconstructions (e.g. Maher, 2008; Ward et al., 2019). However, δ^{18}O-climate relationships may not have remained constant through the past and cross-comparison between palaeoclimate reconstructions is complicated by the fact that different archives record climate in different and non-linear ways. We therefore tackle this problem with a data-model approach that has two main advantages:

- *By using a large number of coexistent speleothem records to identify the large-scale coherent trends, we reduce the impact of non-climatic factors (i.e. soil and karst processes) on δ^{18}O$_{spel}$. This approach focuses on the trends consistent across records that are inherited from δ^{18}O$_{precip}$.*
- *We use model simulations that explicitly include water isotope physics to reproduce large-scale orbital trends in δ^{18}O. These therefore provide a physically plausible explanation of δ^{18}O trends under past climate conditions. Congruence between the observed and simulated trends suggests that the drivers of regional changes in the model world are plausible explanations of these changes in the observations. Multiple regression analysis is a convenient way of exploring the various drivers of δ^{18}O.*

Our approach to explain past isotope changes in terms of specific climate drivers is robust as it takes into account large-scale trends in δ^{18}O$_{spel}$ using a known understanding of isotope physics.

To clarify how our approach investigates the underlying mechanisms of δ^{18}O$_{spel}$ trends, we will amend the introduction (from line 79):

"These interpretations generally rely on modern δ^{18}O$_{precip}$-climate observations, which may not have remained constant through time. The sources of δ^{18}O variability can also be explored using isotope-enabled climate models (e.g. Hu et al., 2019), which incorporate known isotope effects and therefore provide plausible explanations for δ^{18}O$_{spel}$ trends."

And from line 91:

"In this study, we combine speleothem δ^{18}O records from version 2 of the Speleothem Isotopes Synthesis and Analysis (SISAL) database with isotope-enabled palaeoclimate simulations from two climate models to investigate the plausible mechanisms driving changes in δ^{18}O in monsoon regions through the Holocene (last 11,700 years) and between interglacial (mid-Holocene and Last Interglacial) and glacial (Last Glacial Maximum) states."

Given the inherent limitations, discussed above, in interpreting the speleothem records based on modern relationships and/or comparison with other reconstructions, our statistical approach offers new insights into the interpretation of regional changes. Nevertheless, we have included a discussion (from line 380 to 396) of how our results fit with existing literature.

1.2 Evaluate and discuss the model realism and robustness

The models are used in the regression analysis but what is the realism of the simulations? To what extent can they be trusted? Some comparison between SISAL and the models are shown in the figures, but the variables and diagnostics are different, so it's hard to compare (more details in minor comments). The observations and simulations should be compared in a more rigorous way. Also, figure 1 could be redone with the models, as an additional check of the realism of the simulations. An entire sub-section should be devoted to model evaluation. Every time it is possible, both models should be used for the same diagnostics to assess to robustness. It's a great opportunity to have two

models, and it should be used more systematically. After the evaluation section, the reader should have a clear opinion on what feature in the simulations can or cannot be trusted. Then when the regression analysis is performed, there should be some discussion on what specific results can be trusted or not.

Climate models produce internally physically consistent changes in the simulated variables and our goal in this paper is not to evaluate the model simulations as such. This has been done to a greater or lesser extent in previous publications (e.g. LeGrande and Schmidt, 2009; Wackerbarth et al., 2012, Gierz et al., 2017; Werner et al., 2018; Comas-Bru et al., 2019). Here we assume that the broadscale trends shown in these simulations are robust and that they can be used to diagnose what factors might contribute to observed changes in speleothem δ^{18}O between glacial and interglacial states and through the Holocene. We will make this clearer by rewriting the sentence in line 101 to explain this logic as follows:

"We exploit the fact that models produce internally physically consistent changes to explore potential and plausible causes of the trends observed in speleothem records across specific monsoon regions, using multiple regression analysis."

Since we are using previously published simulations, our description of the models focuses on the model set-up boundary conditions. However, we agree that it would be worthwhile to expand these descriptions in order to comment on their reliability on the basis of previously published analyses and will amend the model description text. We will also include figures showing relevant model outputs in Supplementary.

We agree that directly comparing multiple models would be a good way to test the robustness of our findings, but this is currently not possible. There are only a few isotope-enabled palaeoclimate simulations and they often use different modelling protocols. Here, for example, the only time period which was run by both models was the mid-Holocene (6 ka) and the experimental protocols by each modelling group were different. This makes it difficult to isolate the reasons behind any differences between the two simulations (~0.5‰, line 425). This is why we decided to focus on comparing glacial-interglacial trends using the ECHAM simulations, and the trends through the Holocene using the GISS simulations. We will re-order and rewrite the model description section (section 2.2; from line 121) to make this logic clearer as follows:

"There are relatively few paleoclimate simulations made with models that incorporate oxygen isotope tracers, and the available simulations do not necessarily focus on the same periods or use the same modelling protocols. Here we use simulations of opportunity from two isotope-enabled climate models: ECHAM5 (version 5 of the European Centre for medium range weather forecasting model in HAMburg) and GISS E-R (Goddard Institute for Space Studies Model version E-R). The ECHAM5 simulations provide an opportunity to examine large-scale changes between glacial and interglacial states, using simulations of the MH, LGM and LIG. The GISS Model E-R Ocean-Atmosphere Coupled General Circulation Model was used to investigate the evolution of δ^{18}O evolution during the Holocene, using eight time slice (9 ka, 6 ka, 5 ka, 4 ka, 3 ka, 2 ka, 1 ka and 0 ka) experiments. Although simulations of the MH 6ka time slice are available with both models, there are differences in the protocol used for the two experiments which preclude direct comparison of the simulations.

The ECHAM5-wiso MH experiment (Wackerbarth et al., 2012; Werner, 2019) was forced by orbital parameters (based on Berger and Loutre, 1991) and greenhouse gas (GHG) concentrations (CO_2 = 280 ppm, CH4 = 650 ppb, N2O = 270 ppb) appropriate to 6 ka. Changes in sea-surface temperature (SST) and sea-ice were derived from a transient Holocene simulation (Varma et al., 2012). The

control simulation for the MH experiment was an ECHAM-wiso simulation of the period 1956-1999 (Langebroek et al., 2011), using observed SSTs and sea-ice cover. This control experiment was forced by SSTs and sea-ice only, with atmospheric circulation free to evolve. The ECHAM5-wiso LGM experiment (Werner, 2019; Werner et al., 2018) was forced by orbital parameters (Berger and Loutre, 1991), GHG concentrations (CO_2 = 185 ppm, CH_4 = 350 ppb, N_2O = 200 ppb), land-sea distribution and ice sheet height and extent appropriate to 21 ka; SST and sea-ice cover were prescribed from the GLAMAP dataset (Schäfer-Neth and Paul, 2003). Sea surface water and sea-ice $\delta^{18}O$ were uniformly enriched by 1 ‰ at the start of the experiment. The control simulation for the LGM experiment used present-day conditions, including orbital parameters and GHG concentrations set to modern values, and SSTs and sea-ice cover from the last 20 years (1979-1999). Both the MH and LGM simulations were run at T106 horizontal grid resolution, approximately 1.1°by 1.1°. Comparison of the MH and LGM simulations with speleothem data globally (Comas-Bru et al., 2019; Fig. S1 and Fig. S2) show that the ECHAM model reproduces the broadscale spatial gradients and the sign of isotopic changes at the majority of cave sites (MH: 72%; LGM: 76%). However, the changes compared to present are generally more muted in the simulations than shown by the speleothem records.

The LIG experiment (Gierz et al., 2017a, 2017b) was run using the ECHAM5/MPI-OM Earth System Model, with stable water isotope diagnostics included in the ECHAM5 atmosphere model (Werner et al., 2011), the dynamic vegetation model JSBACH (Haese et al., 2012) and the MPI-OM ocean/sea-ice module (Xu et al., 2012). This simulation was run at T31L19 horizontal grid resolution, approximately 3.75° by 3.75°. The LIG simulation was forced by orbital parameters derived from Berger and Loutre (1991) and GHG concentrations (CO_2 = 276 ppm, CH_4 = 640 ppb, N_2O = 263 ppb) appropriate to 125 ka, but it was assumed that ice sheet configuration and land-sea geography is unchanged from modern and therefore no change was made to the isotopic composition of sea water. The LIG simulation is compared to a pre-industrial (PI) control with appropriate insolation, GHG and ice sheet forcing for 1850 CE. The sign of simulated isotopic changes in the LIG is in good agreement with ice core records from Antarctica and Greenland and speleothem records from Europe, the Middle East and China (Gierz et al., 2017b) although, as with the MH and LGM, the observed changes tend to be larger than the simulated changes (Fig. S3).

There are GISS ModelE-R (LeGrande and Schmidt, 2009) simulations for eight time slices during the Holocene (9 ka, 6 ka, 5 ka, 4 ka, 3 ka, 2 ka, 1 ka and 0 ka). The 0 ka experiment is considered as the pre-industrial control (ca 1880 CE). Orbital parameters were based on Berger and Loutre (1991) and GHG concentrations were adjusted based on ice core reconstructions (Brook et al., 2000; Indermühle et al., 1999; Sowers, 2003) for each time slice. A remnant Laurentide ice sheet was included in the 9 ka simulation, following Licciardi et al. (1998), and the corresponding adjustment was made to mean ocean salinity and ocean water $\delta^{18}O$ to account for this (Carlson et al., 2008). The ice sheet in all the other experiments was specified to be the same as modern, and therefore no adjustment was necessary. The simulations were run using the M20 version of GISS ModelE-R, which has a horizontal resolution of 4° by 5°. Each experiment was run for 500 years and we use the last 100 simulated years for the analyses. Comparison of the simulated trends in $\delta^{18}O$ show good agreement with Greenland ice core records, marine records from the tropical Pacific and Chinese speleothem records (LeGrande and Schmidt, 2009). However, as is the case with the ECHAM simulations, the model tends to produce changes less extreme than shown by the observations (Fig. S4, S5 and S6)."

Minor comments

- l 48: "The temperature effects stem from the temperature dependance of oxygen isotope fractionation during condensation and ..." -> "The temperature effects stem from the oxygen isotope fractionation during condensation and ...". The contribution of the temperature dependance of the fractionation coefficient in the temperature effect is small (e.g. realistic results can be obtained even with constant isotopic fractionation: Galewsky and Hurley (2010)).

We will reword this sentence as follows: "The temperature effect stems from the cooling required for progressive rainout during Rayleigh distillation (Dansgaard, 1964; Rozanski et al., 1993)."

- l 59: "depleted" -> "enriched"? Actually, it depends depleted or enriched compared to what, but the specificity of evapo-transpiration is to be enriched relatively to the overlying water vapor, and thus to have an enriching effect of the water vapor (Gat and Matsui (1991)).

We will reword this sentence as follows: "The isotopic composition of atmospheric water vapour may also be modified by precipitation recycling over land, since evapotranspiration returns moisture from precipitation back to the atmosphere thereby reducing the $\delta^{18}O_{precip}$/distance gradient along an advection path that occurs with Rayleigh distillation (Gat, 1996; Salati et al., 1979)."

- l 64: you can also add Caley et al. (2014) in the citations.

This section of the introduction summarises the various ways speleothem $\delta^{18}O$ records were interpreted in the original publications. Caley et al. (2014) does not publish or interpret a new speleothem d18O record but is a model-based analysis of factors driving changes in the Asian monsoon $\delta^{18}O_{spel}$ using an isotope-enabled model. We do not think it is relevant to cite it here.

- l 189: define "OIPC": is it the dataset described above?

Yes, this is the data set described in line 185-186. We apologise for not naming it there and will amend the text to do so, as follows: "... using as reference the Online Isotopes in Precipitation Calculator (OIPC: Bowen, 2018; Bowen and Revenaugh, 2003), a global gridded dataset of interpolated mean annual precipitation-weighted $\delta^{18}O_{precip}$ data."

- Figure 3: I have trouble reading this figure. For δ18O, is it possible to have the same y-scale for Δδ 18Oprecip and Δδ 18Ospel? This would allow a direct visual comparison of these 2 quantities. I also have trouble seeing whether anomalies are negative or positive: could you draw an horizontal line to indicate the 0? The 0 line could be shared for all potted variables. In addition, why do you compare observed Δδ 18Ospel to simulated Δδ 18Oprecip? Why not converting simulated Δδ 18Oprecip into δ 18Ocalcite for a more rigorous comparison?

In figure 3, each variable has different units and axes have been adjusted so that glacial-interglacial patterns are aligned for easier reading of trends, rather than comparison of quantitative values. Adding zero lines for each variable would make the figure more difficult to read. However, we will modify it so that boxes are grouped together by variable ($\delta^{18}O_{spel}$, $\delta^{18}O_{precip}$, precipitation, temperature) instead of by time period. We will order each group by time slice (MH, LGM, LIG). We think that this will make it easier for readers to see and interpret the trends. The $\delta^{18}O_{precip}$ has not been converted to its speleothem-equivalent (i.e., $\delta^{18}O_{calcite}$) as this requires knowing mean cave temperature which would have to be estimated by using model-simulated temperature, thereby adding more uncertainty to the data.

- l 270: "consistently low PCoA1 scores": what does it physically mean?

We will expand this text as follows: "The PCoA scores differentiate records geographically (Fig. 2a): southern hemisphere monsoon regions such as the southwestern South American Monsoon (SW-SAM) and South African Monsoon (SAfM) are characterised by low PCoA1 scores, whilst northern hemisphere monsoons such as the Indian Summer Monsoon (ISM) and the East Asian Monsoon (EAM), are characterised by higher PCoA1 scores. This indicates that regions can be differentiated based on their temporal evolution as captured by the first PCoA axis."

- l 300: "The regional composites are z-scores, i.e. anomalies with respect to the base period (3000-7000 yr BP)." Are these just anomalies or true z-scores? Please clarify how you calculate those z-scores and what they physically mean. And why using z-scores in the first place? Why not just simple anomalies?

Speleothem $\delta^{18}O$ values are converted to z-scores when constructing regional composites as this method standardises both the mean and the variance (unlike anomalies). We will emphasise this in the text adding the equation for the calculation of z-scores in the methods section to make it clearer, as follows (from line 219):

"The $\delta^{18}O$ data for individual speleothems were transformed to z-scores, so all records have a standardised mean and variance:

$$z\text{-}score_i = \left(\delta^{18}O_i - \overline{\delta^{18}O}_{(base\ period)}\right)/s\delta^{18}O_{(base\ period)}$$

Where $\overline{\delta^{18}O}$ is the mean and $s\delta^{18}O$ is the standard deviation of $\delta^{18}O$ for a common base period. A base period of 7,000 to 3,000 years BP was chosen to maximise the number of records included in each composite."

In the results section 3.3 (from line 300) when describing what the z-scores show, we will reword to more clearly state that z-scores show a standardised mean and variance with respect to the base period. We believe this will address the reviewer's concerns by allowing readers to interpret the regional speleothem composites of fig 4:

"The regional composites are expressed as z-scores, i.e. changes with respect to the mean and variance of $\delta^{18}O$ for the base period (3000-7000 yr BP)."

- Fig 4: what are the units of the plotted variables? Please add the units on the y-labels. I have trouble to compare the simulated and observed δ 18O: please use similar diagnostics and units for both. For example, convert precip δ 18O into calcite δ 18O for the model, and use simple δ 18O anomalies for the speleothem observations.

We will add units to the axis labels of fig 4 (W m^{-2} for insolation, ‰ for $\Delta\delta^{18}O_{precip}$, z-scores are unitless).

The goal of this figure is to compare the large-scale (regional) temporal trends in observed $\delta^{18}O_{spel}$ and simulated $\delta^{18}O_{precip}$, rather than to make a direct quantitative comparison. The z-scores used for the speleothem composite trends standardise the variance of the records and are unitless. Anomalies are used for simulated $\delta^{18}O_{precip}$ without a conversion to $\delta^{18}O_{calcite}$ as the latter would require

information on the cave temperature which could only be inferred using simulated air temperature, which in turn would add more uncertainty to the comparison.

- Fig 6: can you explain better how these diagrams should be interpreted? What do they physically mean?

We define partial residual plots in the methods section (line 263). However, we will modify the text (at line 324) to provide a physical interpretation of these plots:

"The global model for the Holocene (1 to 9ka) $\delta^{18}O_{precip}$ trends has a pseudo-R^2 of 0.80 and shows statistically significant relationships between the anomalies in $\delta^{18}O_{precip}$ and anomalies in regional precipitation, temperature and surface wind direction (Table 3). The partial residual plots (Fig. 6) show there is a strong negative relationship with regional precipitation (t value = -8.75) and a strong positive relationship (t value = 8.03) with surface wind direction over the moisture source region, an index of changes in either source area or moisture pathway. This indicates that increases in regional precipitation alone will lead to a decrease in $\delta^{18}O$ while changes in source area/moisture pathway, in the absence of changes in other variables, will lead to a significant change in $\delta^{18}O$. The relationship with temperature over the moisture source region is weaker, but positive (t value = 2.05), i.e. an increase in temperature over the moisture source region will lead to an increase in $\delta^{18}O$ if there are no changes in other climate variables. Precipitation recycling is not significant in this global analysis."

- Fig 6, section 3.4: on which model was this regression analysis done? GISS or ECHAM? More generally, why doing each diagnostic with only one model? Why not doing each diagnostic with each model (when the period of interest is available), to assess the robustness of the results?

As explained in our answer to major comment 1.2, there are only a few isotope-enabled palaeoclimate simulations, and they use different protocols even when they run simulations for the same time period. Thus, it is difficult to compare the simulations or assess their robustness because there are multiple possible causes for any differences between them. We use ECHAM for glacial-interglacial shifts and GISS for Holocene evolution. This has been clarified in our proposed amendment to the text. We will also amend the methods section describing our statistical analyses of simulated $\delta^{18}O$ and climate variables to clarify which models are being used. Amendments are for line 174:

"We examined glacial-interglacial shifts in $\delta^{18}O_{spel}$ observations and in annual precipitation-weighted mean $\delta^{18}O_{precip}$ from ECHAM-wiso in regions influenced by the monsoon. We focus on regional differences between MH, LGM and LIG with respect to the present-day for speleothems or the control simulation experiment for model outputs."

And line 238:

"We investigate the drivers of regional $\delta^{18}O_{precip}$, and by extension $\delta^{18}O_{spel}$, through the Holocene using multiple linear regression (MLR) of annual precipitation-weighted mean $\delta^{18}O_{precip}$ anomalies and climate variables from GISS modelE-R. Climate variables were chosen to represent the four potential large-scale drivers of regional changes in the speleothem $\delta^{18}O$ records."

- l 380: "drivers" -> "meteorological variables". This is just a statistical analysis, so no causality can be identified, so the meteorological variables cannot be assumed to be drivers.

Please see comment above related to this. We will change "drivers" to "climate variables"

- l 389: "changes in precipitation amount" -> "changes in local precipitation amount": changes in upstream precipitation amount has been shown to be very important in previous studies (e.g. Battisti et al. (2014)) but were not analyzed here.

Simulated precipitation changes in this study are regional averages over the monsoon regions, thus they are not equivalent to "local precipitation amount". To clarify this point, we will revise the occurrences of "precipitation amount" throughout to "regional precipitation", including at line 389:

"Changes in regional precipitation do not seem to explain the observed changes in $\delta^{18}O$spel in the EAM during the Holocene."

- Table 1: too many digits in the numbers.

We will reduce values in table 1 to two decimal places

Refs:

Comas-Bru, L., Harrison, S. P., Werner, M., Rehfeld, K., Scroxton, N. and Veiga-Pires, C.: Evaluating model outputs using integrated global speleothem records of climate change since the last glacial, Clim. Past., 15(4), 1557-1579, https://doi.org/10.5194/cp-15-1557-2019, 2019.

Gierz, P., Werner, M. and Lohmann, G.: Simulating climate and stable water isotopes during the Last Interglacial using a coupled climate-isotope model, J. Adv. Model. Earth Syst., 9(5), 2027–2045, https://doi.org/10.1002/2017MS001056, 2017.

Hu, J., Emile-Geay, J., Tabor, C., Nusbaumer, J. and Partin, J.: Deciphering oxygen isotope records from Chinese speleothems with an isotope-enabed climate model, Paleoceanogr. Paleoclimatology, 43(12), 2098-2112, https://doi.org/ 10.1029/2019PA003741, 2019.

LeGrande, A. N. and Schmidt, G. A.: Sources of Holocene variability of oxygen isotopes in paleoclimate archives, Clim. Past, 5(3), 441–455, https://doi.org/10.5194/cp-5-441-2009, 2009.

Maher, B. A.: Holocene variability of the East Asian summer monsoon from Chinese cave records: A re-assessment, Holocene, 18(6), 861–866, https://doi.org/10.1177/0959683608095569, 2008.

Sinha, A., Kathayat, G., Cheng, H., Breitenbach, S.F.M., Berkelhammer, M., Mudelsee, M., Biswas, J. and Edwards, R.L.: Trends and oscillations in the Indian summer monsoon rainfall over the last two millennia, Nat. Commun., 6(6309), https://doi.org/10.1038/ncomms7309, 2015.

Ward, B., Wong, C., Novello, V., McGee, D., Santos, R.V., Silva, L.C.R., Cruz, F.W., Wang, X., Edwards, R.L. and Cheng, H.: Reconstruction of Holocene coupling between the South American Monsoon System and local moisture variability from speleothem $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ records, Quat. Sci. Rev., 210, 51-63, http://doi.org/10.1016/j.quascirev.2019.02.019, 2019.

Wackerbarth, A., Langebroek, P. M., Werner, M., Lohmann, G., Riechelmann, S., Borsato, A. and Mangini, A.: Simulated oxygen isotopes in cave drip water and speleothem calcite in European caves, Clim. Past, 8(6), 1781–1799, https://doi.org/10.5194/cp-8-1781-2012, 2012.

Werner, M., Jouzel, J., Masson-Delmotte, V. and Lohmann, G.: Reconciling glacial Antarctic water stable isotopes with ice sheet topography and the isotopic paleothermometer, Nat. Commun., 9(1), 1–10, https://doi.org/10.1038/s41467-018-05430- y, 2018.

Response to referee #2

Major comments

1) The introduction does not reflect the current understanding of the speleothem δ18O, particularly in the East Asian monsoon domain. For example, it basically follows the previous misunderstanding(s) from modeling and other research communities, especially on orbital-scale, that the speleothem δ18O was interpreted as a rainfall amount proxy by the Chinese speleothem community over the past two decades. In fact, the mainstream idea from the speleothem community has never been the 'amount affect' (e.g., Cheng et al., 2019), and therefore, one of main scientific issues addressed here is groundless.

We understand from the reviewer's concerns that the introduction requires rewording to make clear how Chinese speleothem records have been interpreted as monsoon signals in the literature, i.e. as an upstream precipitation signal (Hu et al., 2008; Yuan et al., 2004) or a rainfall seasonality signal (Cheng et al., 2006, 2009; Wang et al., 2001).Other interpretations of Chinese monsoon $\delta^{18}O_{spel}$ have included rainfall source changes (Tan 2009, 2011, 2014) or local rainfall changes in specific areas (Cai et al., 2010; Tan et al., 2015). Changes to the text to address this are proposed under comment 3.

However, we would like to clarify that our focus here is not simply on the East Asian monsoon domain but rather to investigate all regional monsoons with sufficient speleothem data available in the SISALv2 database. Our discussion in the introduction was to highlight the fact that multiple mechanisms have been proposed in the existing literature to explain $\delta^{18}O_{spel}$ trends in monsoon regions, and we use the East Asian monsoon region as one example of this. We also provide examples from other regions, including the Indonesian-Australian monsoon (line 74 et seq.) and the South American monsoon (line 76 et seq.). To better emphasise this point, we will expand our discussion of other regions in the introduction (under comment 3), rather than mostly discussing the interpretation of East Asian speleothems. It is not uncommon in the literature to propose only one (dominant) mechanism to explain $\delta^{18}O_{spel}$ variability when in reality there could be several mechanisms acting in combination. Furthermore, the proposed mechanisms are often based on modern-day observed relationships, which may not have remained constant in the past. In this study, we utilise model simulations that incorporate known isotope effects/physics, under considerably different conditions to today and we use multiple regression analysis to account for multiple possible isotope drivers in combination. We will clarify this point when describing the aims of this paper in the introduction as follows (after line 91):

"In this study, we combine speleothem δ^{18}O records from version 2 of the Speleothem Isotopes Synthesis and Analysis (SISAL) database with isotope-enabled palaeoclimate simulations from two climate models to investigate the plausible mechanisms driving changes in δ^{18}O in monsoon regions through the Holocene (last 11,700 years) and between interglacial (mid-Holocene and Last Interglacial) and glacial (Last Glacial Maximum) states.

And after line 98: We use isotope-enabled model simulations to investigate the main drivers of $\delta^{18}O_{spel}$ variability in regions where the models reproduce the large-scale δ^{18}O changes shown by observations. We exploit the fact that models produce internally physically consistent changes to explore potential and plausible causes of the trends observed in speleothem records across specific monsoon regions, using multiple regression analysis."

2) The authors mentioned that "a composite record can minimize the influence of site specific karst and cave processes" (with real spatial variations?). However, the results and/or assumptions from the PCoA method are tentative, which lacks a underlying mechanism. The same monsoon system (e.g., the ISM and EAM boxes in the figure 1) could have different speleothem δ18O patterns on orbital-scale, as illustrated by a number of modeling results (e.g., Liu et al., 2014; Battisti et al., 2014).

The reviewer has misunderstood the purpose of the PCoA analysis. We use PCoA to investigate the (dis)similarity of Holocene $\delta^{18}O_{spel}$ trends in order to be able to determine whether there is any large-scale coherency between individual monsoon speleothem records and thus whether it is possible to group records based on the similarity of their Holocene trends in a quantitative and objective way. By showing that records show geographic coherency, we are able to construct regional composites which we subsequently use to study mechanisms through multiple regression.

We will clarify the purpose of the PCoA by amending the text from line 268:

"PCoA shows the (dis)similarity of Holocene $\delta^{18}O_{spel}$ evolution across individual records, and thus allows an objective regionalisation of these records."

3) Lines 66-68: This is really a misleading statement. I suggest that the authors should read the original papers they cited here more carefully (as well as Cheng et al., 2016, 2019; Zhang et al., 2018; Zhang et al., 2020) and quote the original statements in these papers if necessary. For example, Cheng et al. (2009) (cited in the sentence) clearly asserted: "Thus, neither the temperature- δ18O relationship, commonly used to interpret ice-core data, nor the interpretation based on the "amount effect" is justified".

Our purpose here, as explained above, was simply to demonstrate that there are alternative interpretations of the records from specific regions rather than to review the literature from any one region exhaustively. However, we will expand this text to reflect what these various papers meant when discussing summer precipitation changes (from line 66):

"In the East Asian monsoon, for example, speleothem δ^{18}O records have been interpreted as a summer monsoon signal, manifested as either a change in the amount of water vapour removed along the precipitation trajectory (Yuan et al., 2004), and/or as a change in the contribution of summer precipitation to annual totals (Cheng et al., 2006, 2009, 2016; Wang et al., 2001), based on the relationship between modern $\delta^{18}O_{precip}$ and climate. Other interpretations of Chinese monsoon

δ¹⁸O$_{spel}$ have included rainfall source changes (Tan 2009, 2011, 2014) or changes in monsoon precipitation amount (Cai et al., 2010; Tan et al., 2015). Maher (2008) interpreted δ¹⁸O$_{spel}$ as reflecting changes in moisture source area, based on differences between δ¹⁸O$_{spel}$ and loess/palaeosol records of rainfall and the strong correlation between East Asian and Indian monsoon speleothems. Maher and Thompson (2012) used a mass balance approach to show that the changes in precipitation (either local or upstream) or rainfall seasonality required to reproduce δ¹⁸O$_{spel}$ trends would be unreasonably large. They therefore argued that changes in moisture source were required to explain shifts in δ¹⁸O both on glacial/interglacial time scales and during interglacials. There are also multiple interpretations of the causes of δ¹⁸O$_{spel}$ variability in other monsoon regions. In the Indonesian-Australian monsoon region, for example, δ¹⁸O$_{spel}$ variability has been interpreted as a precipitation amount signal (Carolin et al., 2016; Krause et al., 2019) or a precipitation seasonality signal (Ayliffe et al., 2013; Griffiths et al., 2009), based on modern δ¹⁸O$_{precip}$ and climate observations (Cobb et al., 2007; Moerman et al., 2013), and/or as a moisture source/trajectory signal (Griffiths et al., 2009; Wurtzel et al., 2018). South American speleothem records have been interpreted as records of monsoon intensity, due to changes in the amount of precipitation over the region (Cruz et al., 2006; Wang et al., 2006; Cheng et al., 2013), changes in the degree of upstream precipitation and evapotranspiration (Cheng et al., 2013) or changes in the ratio of precipitation sourced from the low-level jet versus the Atlantic (Cruz et al., 2005; Wang et al., 2006). In the Indian monsoon region, speleothem δ¹⁸O records are interpreted primarily as an amount effect signal (Berkelhammer et al., 2010; Fleitmann et al., 2004), supported by δ¹⁸O$_{precip}$/climate observations (e.g. Battacharya et al., 2003). However, other studies have suggested that δ¹⁸O$_{precip}$ changes in this region are driven primarily by large-scale changes in monsoon circulation and hence, Indian monsoon δ¹⁸O$_{spel}$ should be interpreted as a moisture source/trajectory signal (Breitenbach et al., 2010; Sinha et al., 2015). "

4) Lines 229-236 and figure 4: What are the simulated precipitation δ18O values in the EAM, ISM, IAM, SW-SAM domains? Are they amount-weighted annual mean precipitation δ18O values, annual mean precipitation δ18O values or only summer (MJJAS) mean precipitation δ18O values? In addition, please give the boundary coordinate of these monsoon regions (the EAM, ISM, IAM, SW-SAM. . .) for the calculations. Give a detail explanation about the δ18O amplitude differences between observation and model results in the figure 4 if significant.

All simulated δ¹⁸O$_{precip}$ values are annual precipitation-weighted δ¹⁸O anomalies with respect to a control simulation. We will amend the text as follows:

Line 174: "We examined glacial-interglacial shifts in δ¹⁸O$_{spel}$ observations and in annual precipitation-weighted mean δ¹⁸O$_{precip}$ from ECHAM-wiso in regions influenced by the monsoon. We focus on regional differences between MH, LGM and LIG with respect to the present-day for speleothems or the control simulation experiment for model outputs."

Line 229: "We calculated Holocene regional composites from annual precipitation-weighted mean δ¹⁸O$_{precip}$ anomalies simulated by the GISS model."

Line 238: "We investigate the drivers of regional δ¹⁸O$_{precip}$, and by extension δ¹⁸O$_{spel}$, through the Holocene using multiple linear regression (MLR) of annual precipitation-weighted mean δ¹⁸O$_{precip}$ anomalies and climate variables from GISS modelE-R. Climate variables were chosen to represent the four potential large-scale drivers of regional changes in the speleothem δ¹⁸O records.

Figure 1: Spatial distribution of speleothem records used is this study. Colours indicate the sites used in Principal Coordinates Analysis and Redundancy Analysis (PCoA, RDA) to separate monsoon regions, and sites not used in PCoA and RDA but used in subsequent analyses. The individual regional monsoons are shown by boxes: CAM = Central American Monsoon (latitude: 10 to 33°; longitude: -115 to -58°) , SW-SAM = southwestern South American Monsoon (latitude: -10° to 0°; longitude: -80° to -64° and latitude: -30° to -10°; longitude -68° to -40°), NE-SAM = northeastern South American Monsoon (latitude: -10° to 0°; longitude: -60° to -30°), SAfM = southern African Monsoon (latitude: -30° to -17°; longitude: 10° to 40°), ISM = Indian Summer Monsoon (latitude: 11° to 32°; longitude: 50° to 95°), EAM = East Asian Monsoon (latitude: 20° to 39°; longitude: 100° to 125°), IAM = Indonesian-Australian Monsoon (latitude: -24° to 5°; longitude: 95° to 135°). Source region limits used in the multiple linear regression analysis are also shown. The background carbonate lithology is from the World Karst Aquifer Mapping (WOKAM) project (Goldschneider et al., 2020).

5) Lines 376-379: ". . .there is little different in the δ18O values between the MH and the LIG in the ISM and EAM regions. . .", "Given that the increase in summer insolation is much larger during the LIG than the MH, this finding is again consistent with the idea that other factors play a role in modulating the monsoon response to insolation forcing". What are the other factors and the processes? Moisture source and/or pathway? Or some kind of thresholds (e.g., Cheng et al., 2012; Cai et al., 2015)? In addition, the summer insolation is indeed higher during the LIG than during the MH, but the monsoon circulation or intensity is influenced by the temperature (thus pressure?) gradient between land and sea as well. What is the difference of the land-sea temperature (pressure) gradients for the MH and the LIG periods? Or monsoon circulation scales? A more comprehensive discussion of the issue with a help of climate models would be very welcome.

An in-depth discussion of the influences on the East Asian and Indian monsoons is beyond the scope of this paper, since it requires consideration of the monsoons as an integral part of the global atmospheric overturning circulation (see e.g. Schneider et al., 2014; Biasutti et al., 2018; Seth et al., 2019) and associated energy, angular momentum, and moisture budgets. Given that the monsoons cannot simply be considered as regional land-sea breeze circulations, analysis of the land-sea temperature/pressure gradients in the MH and LIG would be insufficient. Our point here was to support the idea, expressed in this paragraph, that there is no simple correspondence between insolation forcing and monsoon response. We have argued that land and ocean feedbacks might have played a role in modulating the response to insolation changes during the Holocene - and the pattern of change through the LIG would also support this. We will clarify our argument about the role of insolation on monsoon changes as follows (line 377):

"The evolution of regional monsoons during the LIG shows patterns similar to those observed during the Holocene, including the lagged response to insolation and the persistence of wet conditions after peak insolation. This is again consistent with the idea that internal feedbacks play a role in modulating the monsoon response to insolation forcing. We have also shown that there is little difference in the isotopic values between the MH and the LIG in the ISM and EAM regions, which is also observed in individual speleothem records (Kathayat et al., 2016; Wang et al., 2008). Given that the increase in summer insolation is much larger during the LIG than the MH, this finding indicates that other factors play a role in modulating the monsoon response to insolation forcing and may

reflect the importance of global constraints on the externally-forced expansion of the tropical circulation (Biasutti et al., 2018)."

6) The main conclusion is that "East Asian monsoon speleothem δ18O evolution through the Holocene relates to changes in atmospheric circulation (i.e. changes in moisture pathway and/or source). Changes in precipitation amount are the predominant driver of Holocene δ18Ospel evolution in the Indian, southwestern South American and Indonesian-Australian monsoons, although changes in atmospheric circulation also contribute in the Indian and Indonesian-Australian monsoon regions and changes in precipitation recycling in southwestern South America". This conclusion is not well supported and problematic as well. First, the 'amount effect' discussed here is not the same 'amount effect' as conventionally defined in the tropics (see Zhao et al., 2019 for instance). The authors implies that the local rainfall amount drive the orbital-scale variations in speleothem δ18O value. They really need to provide a mechanism/calculation for the Indian, southwestern South American and Indonesian-Australian monsoon systems to explain how the oxygen isotopic fractionation under different conditions of rainfall amounts at each cave site could result in the observed δ18Ospel changes on orbital-scale without significant monsoon circulation (including the moisture pathway and/or source) changes. On the other hand, the "East Asian monsoon speleothem δ18O evolution through the Holocene relates to changes in atmospheric circulation" is just a reinforcement of the previous view on the East Asian monsoon evolution inferred by speleothem δ18O records published in a large number of speleothem works over the past two decades. In short, it is the monsoon circulation that to first order drives the orbital δ18Ospel changes, not only for the East Asian monsoon, but also (most likely) for other monsoon systems.

The reviewer is correct that we are talking about regionally averaged precipitation changes rather than changes in what is conventionally understood as the precipitation amount effect and we will change this wording to "changes in regional precipitation". However, the reviewer has misunderstood the purpose of our analyses. We acknowledge that there may be concurrent changes in multiple factors. However, the aim of multivariate analysis is to separate out the various $\delta^{18}O_{precip}$-climate relationships and investigate which variables (it may be a combination of several) are most important in a given region. For example, if circulation and regional precipitation were changing in a way that both together can explain $\delta^{18}O_{precip}$ trends, the MLR model would show this. On the other hand, if there was a significant change in atmospheric circulation (and hence source area) without a corresponding change in regional precipitation, there would be conformity with the global circulation relationship, but not with precipitation, as is seen for the EAM. This would be possible if circulation changes drove a precipitation change outside of the region where the speleothem sites are located, as has been proposed by several papers focusing on the EAM (e.g. Hu et al., 2008; Liu et al., 2014). In all cases, these varying atmospheric and/or precipitation changes are underpinned by the physics incorporated in the climate model simulations. One point that the reviewer appears to have missed is that our aim is not to disprove or reinforce conclusions based on previous EAM speleothem studies. Rather we are using multivariate analysis to provide an alternative way of examining the potential causes of observed changes, independently from the assumption that underpins most interpretations in the literature that modern relationships provide a robust guide to what has happened in the past. By analysing the individual effects of different variables using isotope-enabled models that reproduce

the large-scale monsoon trends shown by the observations, we are able to determine what factors are important in a robust way.

We will make the following amendments to the text in order to make the purpose of our analyses clearer:

In the introduction, we will reword to make clearer the goal of this study (under comment 1).

In the results, we will more clearly state what the multiple linear regression shows (line 319):

"The MLR analyses of simulated $\delta^{18}O_{precip}$ trends identify the impact of an individual climate variable on $\delta^{18}O_{precip}$ in the absence of changes in other variables."

When discussing the multiple linear regression (line 389):

Changes in regional precipitation (where the cave sites are located) do not seem to explain the observed changes in $\delta^{18}O_{spel}$ in the EAM during the Holocene, where Holocene $\delta^{18}O_{precip}$ evolution is largely driven by changes in atmospheric circulation (indexed by changes in surface winds). This is consistent with existing studies that emphasise changes in moisture source and/or pathway rather than local precipitation changes (Maher, 2016; Maher and Thompson, 2012; Tan, 2014; Yang et al., 2014).

7) Please illustrate the x- and y-axes of the figure 2a in the section 3.1 or describe them in the section 2.3. In the section 3.1, the authors illustrated that Southern Hemisphere monsoon regions are characterized by low PCoA1 scores, while Northern Hemisphere monsoon regions are characterized by higher PCoA1 scores. Please explain these terms in the context of instrumental data or modern climatology, which may be more interesting for the paleoclimate community.

The aim of the PCoA is to investigate the (dis)similarity of Holocene $\delta^{18}O_{spel}$ trends amongst speleothem sites. We then use RDA to investigate whether the distribution of site (dis)similarity relates to geographic location (latitude and longitude). This allows us to investigate whether there is a regional and global-scale coherency to Holocene $\delta^{18}O_{spel}$ records, and thus to regionalise the records based on the observations themselves rather than any assumption of regional synchroneity in the speleothem records. We make no assumption that these trends are related to modern climatology, or that regions should be defined on the basis of their modern climatology. We have modified our description of the purpose of the PCoA analyses (in response to comments from reviewer 1) and this will hopefully make the purpose of these analyses clearer.

8) The authors used the anomaly for comparison from different model results. However, readers might also want to see a detailed comparison between model results, particularly between the model results from this study and those from previous studies.

There are relatively few isotope-enabled palaeoclimate simulations, and they are generally run under different protocols/boundary conditions, thus precluding a rigorous comparison between them since it is difficult to attribute differences to model structure or experimental protocol. Furthermore, an analysis of model-based results per se is not the goal of this paper. In response to comments by reviewer 1, we have included anomaly maps of simulated $\delta^{18}O_{precip}$ from the simulations we are using for our analyses in the supplementary material.

9) Lines 397-400 and the figure 3: "The LGM is characterised by a similar orbital configuration to today, however global ice volume was at a maximum and GHG concentrations were lower than present. The δ18Ospel anomalies are more positive during the LGM than the MH or LIG, suggesting drier conditions in the ISM, EAM and IAM, supported by simulated changes in δ18Ospel and precipitation (Fig. 3)." This sentence is again misleading. While the authors highlighted a similar orbital configuration between the LGM and today, they actually discussed the issue related to a comparison of the LGM with the MH or LIG, presumably implying that they have similar orbital configurations. The LGM (21±1ka) is near a Northern Hemisphere insolation minimum whereas the MH/LIG are near the insolation maxima. As such the related discussions should be rephrased, and so does the related conclusion, since the insolation difference should be taken into account together with GHG and the global ice volume, because one could also argue that the δ18Ospel just follows the insolation with effect to a lesser extent from GHG and the global ice volume.

We agree that it is not ideal to describe the LGM boundary conditions with respect to the modern day when the purpose of this paragraph is to contrast the LGM signals with those of the MH and LGM, so we will rephrase this to read (line 397):

The LGM is characterised by lower northern hemisphere summer insolation, globally cooler temperatures, expanded global ice volumes and lower GHG concentrations than either the MH or the LIG.

And rephrase the conclusion (line 405) as:

Enriched $\delta^{18}O_{precip}$ and $\delta^{18}O_{spel}$ values during the LGM must therefore be caused by a significant decrease in atmospheric moisture and precipitation that resulted from the cooler conditions.

Minor comments

Lines 97, 112 and 160, 'the Principal Coordinate Analysis (PCoA)', the abbreviation occurred three times, keep the first one.

We will amend the text so that PCoA is only defined at its first mention (at line 97).

Line 121, please give the full name of the climate models: ECHAM5 and GISS E-R

We will modify the text to define these acronyms as follows:

"Here we use simulations of opportunity from two isotope-enabled climate models: ECHAM5 (version 5 of the European Centre for medium range weather forecasting model in HAMburg) and GISS E-R (Goddard Institute for Space Studies Model version E-R)."

Line 163, '. . .missing data that . . .' , 'that' should be 'than'?

Yes, we will correct the text here.

Line 189, what is the 'OIPC'?

OIPC is the data set described in line 185-186. We apologise for not naming it there and will amend the text to do so, as follows: "... using as reference the Online Isotopes in Precipitation Calculator

(OIPC: Bowen, 2018; Bowen and Revenaugh, 2003), a global gridded dataset of interpolated mean annual precipitation-weighted $\delta^{18}O_{precip}$ data. "

Lines 268-277, the abbreviations (EAM, SW-SAM, SAfM, CAM, IAM) occurred too late in the section 3.1, it's better put them in the introduction.

The regional monsoons, and their abbreviations are not introduced until section 3.1, as the results from PCoA justify our grouping of the data in regional monsoons. We therefore introduce them here. However, abbreviations are also available in the caption of Figure 1, which is first cited in line 116.

Line 358 'southern China Sea' should be 'South China Sea'.

We will amend the text accordingly

Figure 5, the time series for Dongge Cave can be replaced by a high-resolution timeseries, please double check with the database.

We use speleothem records from the SISALv2 database because these have been standardised, quality-controlled, and the age models have been verified. The higher resolution records of the LIG from Dongge cave (Kelly et al., 2006) are not in the SISAL database.

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A data-model approach to interpreting speleothem oxygen isotope records from monsoon regions on orbital timescales

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Abstract. Reconstruction of past changes in monsoon climate from speleothem oxygen isotope ($\delta^{18}O$) records is complex because $\delta^{18}O$ signals can be influenced by multiple factors including changes in precipitation, precipitation recycling over land, temperature at the moisture source and changes in the moisture source region and transport pathway. Here, we analyse >150 speleothem records from version 2 of the Speleothem Isotopes Synthesis and Analysis (SISAL) database to produce composite regional trends in $\delta^{18}O$ in monsoon regions; compositing minimises the influence of site-specific karst and cave processes that can influence individual site records. We compare speleothem $\delta^{18}O$ observations with isotope-enabled climate model simulations to investigate the specific climatic factors causing these regional trends. We focus on differences in $\delta^{18}O$ signals between interglacial (mid-Holocene and Last Interglacial) and glacial (Last Glacial Maximum) states, and on $\delta^{18}O$ evolution through the Holocene. Differences in speleothem $\delta^{18}O$ between the mid-Holocene and Last Interglacial in the East Asian and Indian monsoons are small, despite the larger summer insolation values during the Last Interglacial. Last Glacial Maximum $\delta^{18}O$ values are significantly less negative than interglacial values. Comparison with simulated glacial-interglacial $\delta^{18}O$ shows that changes are principally driven by global shifts in temperature and regional precipitation. Holocene speleothem $\delta^{18}O$ records show distinct and coherent regional trends. Trends are similar to summer insolation in India, China and southwestern South America, but different in the Indonesian-Australian region. Redundancy analysis shows that 37% of Holocene variability can be accounted for by latitude and longitude, supporting the differentiation of records into individual monsoon regions. Regression analysis of simulated precipitation $\delta^{18}O$ and climate variables show that global Holocene monsoon $\delta^{18}O$ trends are driven by changes in precipitation, atmospheric circulation and (to a lesser extent) source area temperature, whilst precipitation recycling is non-significant. However, there are differences in regional scale mechanisms; there are clear relationships between changes in precipitation and in $\delta^{18}O$ for India, southwestern South America and the Indonesian-Australian regions, but not for the East Asian monsoon. Changes in atmospheric circulation contributes to $\delta^{18}O$ trends in the East Asian, Indian and Indonesian-Australian monsoons, and a weak source area temperature effect is observed over southern and central America and Asia. Precipitation recycling is influential in southwestern South America and southern

Africa. Overall, our analyses show that it is possible to differentiate the impacts of specific climatic mechanisms influencing precipitation $\delta^{18}O$ and use this analysis to interpret changes in speleothem $\delta^{18}O$.

1 Introduction

35   The oxygen isotopic ($\delta^{18}O$: $^{18}O/^{16}O$ ratio relative to a standard, in permil, ‰) composition of speleothems is widely used to infer past regional climates (Bar-Matthews et al., 1997; McDermott, 2004; Wang et al., 2008). Speleothem oxygen isotope ($\delta^{18}O_{spel}$) signals are inherited from $\delta^{18}O$ in precipitation ($\delta^{18}O_{precip}$) above the cave, which in turn is determined by the initial $\delta^{18}O$ of water vapour as it evaporates at the oceanic moisture source region, the degree of rainout and evaporation from source to cave site and air temperature changes encountered throughout the moisture transport pathway (Fairchild and Baker, 2012; Lachniet, 2009). Understanding the effects and contribution of each of these climate processes to $\delta^{18}O_{precip}$ (and therefore $\delta^{18}O_{spel}$) is essential to inferring palaeoclimate from speleothem $\delta^{18}O$ records.

Initial $\delta^{18}O$ is determined by oceanic $\delta^{18}O$ at the evaporative moisture source region (Craig and Gordon, 1965), which varies spatially (LeGrande and Schmidt, 2006) and through time (e.g. Waelbroeck et al., 2002). During evaporation from the moisture source, ^{16}O is preferentially incorporated into the vapour, whilst subsequent fractionation during atmospheric transport occurs by Rayleigh distillation; As air masses cool and moisture condenses, heavier ^{18}O is enriched in the liquid phase and removed by precipitation. With progressive rainout along a moisture pathway, precipitation becomes gradually more depleted (Dansgaard, 1964). Within this framework, $\delta^{18}O_{precip}$ is controlled by two variables: temperature and the amount of precipitation along a moisture pathway. The temperature effect stems from the ~~temperature dependence of oxygen isotope fractionation during condensation and the~~ cooling required for <u>progressive</u> rainout <u>during Rayleigh distillation</u> (Dansgaard, 1964; Rozanski et al., 1993). The temperature-$\delta^{18}O$ impact is dominant at mid to high latitudes<u>,</u> whilst ~~at tropical latitudes,~~ observations suggest that changes in upstream and local precipitation dominate changes in the $\delta^{18}O_{precip}$ signal~~.~~ <u>at tropical latitudes</u>. The negative relationship between local precipitation and $\delta^{18}O_{precip}$, often referred to as the "amount effect" (Bailey et al., 2018; Dansgaard, 1964), results from the re-evaporation and diffusive exchange between precipitation and water vapour during deep convective precipitation (Risi et al., 2008). However, Rayleigh distillation is complicated by changes in atmospheric circulation and moisture recycling. Changes in the area from which the moisture is sourced will modify $\delta^{18}O_{precip}$ because the initial $\delta^{18}O$ values differ between sources (Cole et al., 1999; Friedman et al., 2002), whilst changes in the moisture transport pathway and/or distance between source and cave site can result in differing degrees of fractionation associated with condensation and evaporation (Aggarwal et al., 2012; Bailey et al., 2018). The isotopic composition of atmospheric water vapour may also be modified by precipitation recycling over land, since evapotranspiration returns <u>moisture</u>~~depleted $\delta^{18}O$~~ from precipitation back to the atmosphere thereby reducing the $\delta^{18}O_{precip}$/distance gradient along an advection path that occurs with Rayleigh distillation (Gat, 1996; Salati et al., 1979).

Speleothem $\delta^{18}O$ records from monsoon regions show multi-millennial variability that has been interpreted as documenting the waxing and waning of the monsoons in response to changes in summer insolation, manifested as either a change in the absolute amount of precipitation (Cai et al., 2012; Cheng et al., 2006) or a change in the ratio of more negative $\delta^{18}O$ summer precipitation to less negative $\delta^{18}O$ winter precipitation (Cruz et al., 2005; Dong et al., 2010; Wang et al., 2001). However, the multiplicity of processes that influence $\delta^{18}O$ before incorporation in the speleothem make it difficult to attribute the climatic causes of changes in individual speleothem records unambiguously. In the East Asian monsoon, for example, speleothem $\delta^{18}O$ records have been interpreted as a summer monsoon signal, manifested either as a change in the amount of water vapour removed along the moisture trajectory (Yuan et al., 2004), and/or as a change in the contribution of summer precipitation to annual totals ~~regional rainfall signals~~ (Cheng et al., 2006, 2009, 2016; Wang et al., 2001~~; Yuan et al., 2004~~) based on the relationship between modern $\delta^{18}O_{precip}$ and climate. Other interpretations of Chinese monsoon $\delta^{18}O_{spel}$ have included rainfall source changes (Tan 2009, 2011, 2014) or local rainfall changes (Cai et al., 2010; Tan et al., 2015). ~~However,~~ Maher (2008) interpreted $\delta^{18}O_{spel}$ as reflecting changes in moisture source area, based on differences between $\delta^{18}O_{spel}$ and loess/palaeosol records of rainfall and the strong correlation between East Asian and Indian monsoon speleothems. Maher and Thompson (2012) used a mass balance approach to show that ~~the~~ the changes in precipitation (either local or upstream) or rainfall seasonality required to reproduce $\delta^{18}O_{spel}$ trends would be unreasonably large. ~~the observed changes in $\delta^{18}O_{spel}$ was unreasonably large even if the effects of seasonality and temperature were considered~~. They therefore argued that changes in moisture source were required to explain shifts in $\delta^{18}O$ both on glacial/interglacial time scales and during interglacials. There are also multiple~~conflicting~~ interpretations of the causes of $\delta^{18}O_{spel}$ variability in other monsoon regions. In the Indonesian-Australian monsoon region, for example, $\delta^{18}O_{spel}$ variability has been interpreted as ~~both~~ ~~either~~ a precipitation amount signal (Carolin et al., 2016; Krause et al., 2019) or a precipitation seasonality signal (Ayliffe et al., 2013; Griffiths et al., 2009), based on modern $\delta^{18}O_{precip}$ and climate observations (Cobb et al., 2007; Moerman et al., 2013), and/or as a moisture source/trajectory signal~~result of changing moisture source regions~~ (Griffiths et al., 2009; Wurtzel et al., 2018). South American speleothem records have been interpreted as records of monsoon intensity, ~~either~~ due to ~~a~~ changes in the amount of precipitation over the region (Cruz et al., 2006; Wang et al., 2006; Cheng et al., 2013), changes in the degree of upstream precipitation and evapotranspiration (Cheng et al., 2013)~~based on relationships derived from instrumental data (Hardy et al., 2003)~~ or ~~due to~~ changes in the ratio of precipitation sourced from the low-level jet versus the Atlantic (Cruz et al., 2005; Wang et al., 2006). In the Indian monsoon region, speleothem $\delta^{18}O$ records are interpreted primarily as an amount effect signal (Berkelhammer et al., 2010; Fleitmann et al., 2004), supported by $\delta^{18}O_{precip}$/climate observations (e.g. Battacharya et al., 2003). However, other studies have suggested that $\delta^{18}O_{precip}$ changes in this region are driven primarily by large-scale changes in monsoon circulation and hence, Indian monsoon $\delta^{18}O_{spel}$ should be interpreted as a moisture source/trajectory signal (Breitenbach et al., 2010; Sinha et al., 2015).

These interpretations generally rely on modern $\delta^{18}O_{precip}$-climate observations, which may not have remained constant through time~~through the past~~. The sources of $\delta^{18}O$ variability can also be explored using isotope-enabled climate models (e.g. Hu et

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95 al., 2019), which. ~~These models~~ incorporate known isotope effects and therefore provide plausible explanations for $\delta^{18}O_{spel}$ trends. ~~The sources of $\delta^{18}O_{spel}$ variability have been explored using isotope-enabled climate models.~~ Modelling studies ~~(e.g. LeGrande and Schmidt, 2009; Lewis et al., 2010; Pausata et al., 2011)~~ suggest that ~~past~~ changes in East Asian monsoon $\delta^{18}O_{precip}$ (during Heinrich events: Lewis et al., 2010; Pausata et al., 2010, and on orbital timescale: Battisti et al., 2014; LeGrande and Schmidt, 2009) do not reflect local rainfall variability ~~in the East Asian monsoon~~ but instead reflect changes in

100 $\delta^{18}O$ of vapour delivered to the~~a~~ region~~, consistent with the interpretation by Maher (2008)~~. Variability in the $\delta^{18}O$ of vapour delivered to East Asia on orbital timescales has been diagnosed as due ~~attributed~~ to changes in precipitation upstream of the region (Battisti et al., 2014), changes in moisture source location (Hu et al., 2019; Tabor et al., 2018) or changes in the strength of monsoon winds (LeGrande and Schmidt, 2009; Liu et al., 2014). $\delta^{18}O_{precip}$ variability in the East Asian monsoon during Heinrich events has also been attributed to non-local isotope fractionation (Lewis et al., 2010; Pausata et al., 2011). Modelling

105 results ~~generally support the qualitative interpretations of speleothem records from other regions,~~ suggesting that changes in precipitation amount are the predominant source of $\delta^{18}O$ variability in the Indian monsoon during the Holocene (LeGrande and Schmidt, 2009) and in the glacial (Lewis et al., 2010), and in the South American and Indonesian/Australian regions during Heinrich events (Lewis et al., 2010) and the Last Interglacial (Sjolte and Hoffman, 2014).

In this study, we combine speleothem $\delta^{18}O$ records from version 2 of the Speleothem Isotopes Synthesis and Analysis (SISAL)

110 database with isotope-enabled palaeoclimate simulations from two climate models to investigate the plausible mechanisms driving~~causes of~~ changes in $\delta^{18}O$ in monsoon regions through the Holocene (last 11,700 years) and between interglacial (mid-Holocene and Last Interglacial) and glacial (Last Glacial Maximum) states. We compare $\delta^{18}O_{spel}$ signals across geographically separated cave sites to extract a regional signal, thus minimising the influence of karst and in-cave processes, such as the mixing of groundwaters from different precipitation events or changes in cave ventilation, that can be important for the $\delta^{18}O_{spel}$

115 of individual records. We use Principal Coordinate Analysis (PCoA) to identify regions with geographically coherent $\delta^{18}O_{spel}$ records, and then examine how these regions behave on glacial-interglacial time scales and through the Holocene. We use isotope-enabled model simulations~~, where the processes that influence $\delta^{18}O_{precip}$ are explicitly simulated~~and large-scale $\delta^{18}O$ ~~observations are broadly reproduced~~, to investigate the main drivers of ~~regional~~ $\delta^{18}O_{spel}$ variability in regions where the models reproduce~~s~~ the large-scale $\delta^{18}O$ changes shown by observations. We exploit the fact that models produce internally physically

120 consistent changes to explore potential and plausible causes of the trends observed in speleothem records across specific monsoon regions, using multiple regression analysis. ~~We also use multiple regression analysis to identify the likely causes of the observed $\delta^{18}O_{precip}$ changes and trends in specific monsoon regions.~~

2. Methods

2.1 Speleothem oxygen isotope data

125 Speleothem $\delta^{18}O$ records were obtained from the SISAL (Speleothem Isotopes Synthesis and Analysis) database (Atsawawaranunt et al., 2018; Comas-Bru et al., 2020a, 2020b). Records were selected based on the following criteria:

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- They are located in monsoon regions, between 35°S and 40°N;
- The mineralogy is known ~~but~~ and does not vary (i.e. between calcite and aragonite) through time, because oxygen isotope fractionation during speleothem precipitation is different for calcite and aragonite;

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- For the interglacial-glacial analyses, the records contain samples in the periods used for the model simulations: mid-Holocene (MH, 6,000±500 years BP, 6ka), Last Glacial Maximum (LGM, 21,000±1,000 years BP, 21ka) or the Last Interglacial (LIG, 125,000±1,000 years BP, 125 ka), where BP (before present) is 1950 CE;
- For the ~~Principal Coordinate Analysis (~~PCoA~~)~~, the records have a temporal coverage of at least 4,000 years in the Holocene;

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- For Holocene trend analyses, speleothems have a record of the period from 7,000 to 3,000 years BP;
- They are the most recent update of the record from a site available in version 2 of the SISAL database.

This resulted in the selection of 125 records from 44 sites for the PCoA analysis, 64 records from 38 sites for the analysis of MH, LGM and LIG signals and 79 records from 40 sites for the Holocene trend analysis (Fig. 1). Although the SISALv2 database contains multiple age models for some sites, we use the published age models given by the original authors for all
140 records.

2.2 Climate model simulations

There are relatively few paleoclimate simulations made with models that incorporate oxygen isotope tracers, and the available simulations do not necessarily focus on the same periods or use the same modelling protocols. Here, ~~w~~We use simulations of opportunity from two isotope-enabled climate models: ECHAM5 (version 5 of the European Centre for medium range weather
145 forecasting model in HAM~~amb~~urg) and GISS E-R (Goddard Institute for Space Studies Model version E-R). The ECHAM5 simulations provide an opportunity to examine large-scale changes between glacial and interglacial states, using simulations of the MH, LGM and LIG. The GISS Model E-R Ocean-Atmosphere Coupled General Circulation Model was used to investigate the evolution of δ^{18}O evolution during the Holocene, using eight time slices (9 ka, 6 ka, 5 ka, 4 ka, 3 ka, 2 ka, 1 ka and 0 ka) experiments. Although simulations of the MH 6ka time slice are available with both models, there are differences in
150 the protocol used for the two experiments which preclude direct comparison of the ~~results~~simulations. ~~We used published simulations from the ECHAM5 climate model, run either in atmosphere-only (Werner et al., 2011) or in fully coupled mode (Gierz et al., 2017b; Werner et al., 2016) to investigate δ¹⁸O signals during the MH, LGM and LIG. The GISS Model E-R Ocean-Atmosphere Coupled General Circulation Model (LeGrande and Schmidt, 2009) was used to investigate Holocene δ¹⁸O evolution as multiple isotope-enabled Holocene time-slice experiments exist. There are relatively few paleoclimate simulations~~
155 ~~made with models that incorporate oxygen isotope tracers, and the available simulations do not necessarily use the same modelling protocols. However, two sets of isotope-enabled simulations used here generally conform with PMIP (Paleoclimate Modelling Intercomparison Project) protocols, using reasonable and recommended boundary conditions (trace gases, orbital configuration etc).~~

The ECHAM5-wiso MH experiment (Wackerbarth et al., 2012; Werner, 2019) was forced by orbital parameters (based on Berger and Loutre, 1991) and greenhouse gas (GHG) concentrations (CO_2 = 280 ppm, CH_4 = 650 ppb, N_2O = 270 ppb) appropriate to 6 ka. Changes in sea-surface temperature (SST) and sea-ice were derived from a transient Holocene simulation (Varma et al., 2012). The control simulation for the MH experiment was an ECHAM-wiso simulation of the period 1956-1999 (Langebroek et al., 2011), using observed SSTs and sea-ice cover. This control experiment was forced by SSTs and sea-ice only, with atmospheric circulation free to evolve. The ECHAM5-wiso LGM experiment (Werner, 2019; Werner et al., 2018) was forced by orbital parameters (Berger and Loutre, 1991), GHG concentrations (CO_2 = 185 ppm, CH_4 = 350 ppb, N_2O = 200 ppb), land-sea distribution and ice sheet height and extent appropriate to 21 ka; SST and sea-ice cover were prescribed from the GLAMAP dataset (Schäfer-Neth and Paul, 2003). Sea surface water and sea-ice $\delta^{18}O$ were uniformly enriched by 1 ‰ at the start of the experiment. The control simulation for the LGM experiment used present-day conditions, including orbital parameters and GHG concentrations set to modern values, and SSTs and sea-ice cover from the last 20 years (1979-1999). Both the MH and LGM simulations were run at T106 horizontal grid resolution, approximately 1.1° by 1.1°. Comparison of the MH and LGM simulations with speleothem data globally (Comas-Bru et al., 2019; SI Fig. S1 and, SI Fig. S2) show that the ECHAM model reproduces the broadscale spatial gradients and the sign of isotopic changes at the majority of cave sites (MH: 72%; LGM: 76%). However, the changes compared to present are generally more muted in the simulations than shown by the speleothem records.

The LIG experiment (Gierz et al., 2017b, 2017a) was run using the ECHAM5/MPI-OM Earth System Model, with stable water isotope diagnostics included in the ECHAM5 atmosphere model (Werner et al., 2011), the dynamic vegetation model JSBACH (Haese et al., 2012) and the MPI-OM ocean/sea-ice module (Xu et al., 2012). This simulation was run at a T31L19 horizontal grid resolution, approximately 3.75° by 3.75°. The LIG simulation was forced by orbital parameters derived from the Berger and Loutre (1991) solution and GHG concentrations (CO_2 = 276 ppm, CH_4 = 640 ppb, N_2O = 263 ppb) appropriate to 125 ka, but it was assumed that ice sheet configuration and land-sea geography is unchanged from modern and therefore no change was made to the isotopic composition of sea water. The LIG simulation is compared to a pre-industrial (PI) control with appropriate insolation, GHG and ice sheet forcing for 1850 CE. The sign of simulated isotopic changes in the LIG is in good agreement with ice core records from Antarctica and Greenland and speleothem records from Europe, the Middle East and China (Gierz et al., 2017b) although, as with the MH and LGM, the observed changes tend to be larger than the simulated changes (SI Fig. S3).

There are GISS ModelE-R (LeGrande and Schmidt, 2009) simulations for eight time slices during the Holocene (9 ka, 6 ka, 5 ka, 4 ka, 3 ka, 2 ka, 1 ka and 0 ka). The 0 ka experiment is considered as the pre-industrial control (ca 1880 CE). Orbital parameters were based on the Berger and Loutre (1991) solution and GHG concentrations were adjusted based on ice core reconstructions (Brook et al., 2000; Indermühle et al., 1999; Sowers, 2003) for each time slice. A remnant Laurentide ice sheet was included in the 9 ka simulation, following Licciardi et al. (1998), and the corresponding adjustment was made to mean ocean salinity and ocean water $\delta^{18}O$ to account for this (Carlson et al., 2008). The ice sheet in all the other experiments was

specified to be the same as modern, and therefore no adjustment was necessary. The simulations were run using the M20 version of GISS ModelE-R, which has a horizontal resolution of 4° by 5°. Each experiment was run for 500 years and we use the last 100 simulated years for the analyses. Comparison of the simulated trends in δ^{18}O show good agreement with Greenland ice core records, marine records from the tropical Pacific and Chinese speleothem records (LeGrande and Schmidt, 2009). However, as is the case with the ECHAM simulations, the model tends to produce changes less extreme ~~changes~~ than shown by the observations (~~SI~~ Fig~~ure~~ S4, S5 and S6).

2.3 Princip~~alle~~ Coordinate Analysis and Redundancy Analysis

We used ~~Principle Coordinate Analysis (~~PCoA~~)~~ to identify regionally coherent patterns in the speleothem δ^{18}O records for the Holocene. PCoA is a multivariate ordination technique that uses a distance/dissimilarity matrix to represent inter-object (dis)similarity in reduced space (Gower, 1966; Legendre and Legendre, 1998). Speleothem records from individual sites are often discontinuous; missing data is problematic for many ordination techniques. PCoA is more robust to missing data tha~~nt~~ other methods (Kärkkäinen and Saarela, 2015; Rohlf, 1972). We used a correlation matrix of speleothem records as the (dis)similarity measure. The temporal resolution of speleothem records was first standardised by calculating a running average mean with non-overlapping 500-year windows. This procedure produces a single composite record when there are several records for a given site. PCoA results were displayed as a biplot, where sites ordinated close to one another (i.e., with similar PCoA scores) show similar Holocene trends and sites ordinated far apart have dissimilar trends. We used the 'broken stick' model (Bennett, 1996) to identify which PCoA axes were significant. We used redundancy analysis (RDA: Legendre and Legendre, 1998; Rao, 1964) with latitude and longitude as predictor variables to identify if PCoA (dis)similarities were related to geographical location, and Principal Components Analysis (PCA) to identify the main patterns of variation. As these explanatory variables are not dimensionally homogeneous, they were centred on their means and standardised to allow direct comparison of the gradients. PCoA and RDA analyses were carried out using the 'vegan' package in R (Oksanen et al., 2019).

2.4 Glacial-interglacial changes in δ^{18}O

We examined glacial-interglacial shifts in $^{18}O_{spel}$ observations and in annual precipitation-weighted mean $\delta^{18}O_{precip}$ from ECHAM-wiso in regions influenced by the monsoon~~.~~ We focus~~ing~~ on regional differences between MH, LGM and LIG with respect to the present-day for speleothems or the control simulation experiment for model outputs. Comas-Bru et al. (2019) have shown that differences in speleothem δ^{18}O data between the 20th century and the pre-industrial period (i.e. 1850±15 CE) are within the temporal and measurement uncertainties of the data, and thus the use of different reference periods (i.e. PI for the ECHAM LIG experiment, 20th century for ECHAM MH, LGM experiments) should have little effect on our analyses. We used mean site $\delta^{18}O_{spel}$ values for each period for the regions identified in the PCoA analysis. Where there are multiple speleothem δ^{18}O records for a site in a time period, they were averaged to calculate mean $\delta^{18}O_{spel}$. Three sites above 3500m were excluded from the calculation of the means because high elevation sites have more negative δ^{18}O values than their low-elevation counterparts and their inclusion would distort the regional estimates.

There are relatively few speleothems covering both the present-day and the period of interest (i.e., MH, LGM or LIG), precluding the calculation of $\delta^{18}O_{spel}$ anomalies from the speleothem data. We therefore calculated anomalies with respect to modern (1960-2017 CE) using as reference the Online Isotopes in Precipitation Calculator (OIPC: Bowen, 2018; Bowen and Revenaugh, 2003), a global gridded dataset of interpolated mean annual precipitation-weighted $\delta^{18}O_{precip}$ data (Bowen, 2018; Bowen and Revenaugh, 2003). This dataset combines data from 348 stations from the Global Network of Isotopes in Precipitation (IAEA/WMO, 2018), covering part or all of the period 1960-2014, and other records available at the Water Isotopes Database (Waterisotopes Database, 2017).

OIPC $\delta^{18}O_{precip}$ was converted to its speleothem equivalent assuming that: (i) precipitation-weighted mean annual $\delta^{18}O_{precip}$ is equivalent to mean annual drip-water $\delta^{18}O$ (Yonge et al., 1985) and (ii) precipitation of calcite is consistent with the empirical speleothem-based kinetic fractionation factor of Tremaine et al. (2011) and precipitation of aragonite follows the fractionation factor from Grossman and Ku (1986), as formulated by Lachniet (2015):

$$\delta^{18}O_{calcite\_SMOW} = w\delta^{18}O_{precip\_SMOW} + \left(\left(\frac{16.1 \cdot 1000}{T}\right) - 24.6\right) \qquad \text{(T in K)} \qquad (1)$$

$$\delta^{18}O_{aragonite\_SMOW} = w\delta^{18}O_{precip\_SMOW} + \left(\left(\frac{18.34 \cdot 1000}{T}\right) - 31.954\right) \qquad \text{(T in K)} \qquad (2)$$

where $\delta^{18}O_{calcite\_SMOW}$ and $\delta^{18}O_{aragonite\_SMOW}$ are the speleothem isotopic composition for calcite and aragonite speleothems with reference to the V-SMOW standard (in permil); $w\delta^{18}O_{precip}$ is the OIPC precipitation-weighted annual mean isotopic composition of precipitation with respect to the V-SMOW standard and T is the mean annual cave temperature (in degrees Kelvin). We used the long-term (1960-2016) mean annual surface air temperature from the CRU-TS4.01 database (Harris et al., 2014) at each site as a surrogate for mean annual cave air temperature. The resolution of the gridded data means that $w\delta^{18}O_{precip\_SMOW}$ and T may be the same for nearby sites.

We use the V-SMOW to V-PDB conversion from Coplen et al. (1983), which is independent of speleothem mineralogy:

$$\delta^{18}O_{PDB} = 0.97001 \cdot \delta^{18}O_{SMOW} - 29.29 \qquad (3)$$

where $\delta^{18}O_{PDB}$ is relative to the V-PDB standard and $\delta^{18}O_{SMOW}$ is relative to V-SMOW standard.

Average uncertainties in the speleothem age-depth models are ~50 years during the Holocene. This interval is smaller than the time windows used in this analysis, and age uncertainty is therefore expected to have a negligible impact on the results. We investigated the influence of age uncertainties on the LGM and LIG $\delta^{18}O_{spel}$ anomalies by examining the impact of using different window widths ($\pm 500, \pm 700, \pm 1000, \pm 2000$ years) on the regional mean $\delta^{18}O_{spel}$ anomalies.

We used anomalies of $w\delta^{18}O_{precip}$, mean annual surface air temperature (MAT) and mean annual precipitation (MAP) from the ECHAM5-wiso simulations to investigate the drivers of glacial-interglacial $\delta^{18}O_{spel}$ variability. Values were calculated from land grid cells (>50% land) $\pm 3°$ around each speleothem site. This distance was chosen with reference to the coarsest resolution simulation (LIG, ca. 3.75 x 3.75°). Gridded values of MAT and MAP were weighted by the proportion of each grid cell that lies within $\pm 3°$ of the site and linear distance-weighted means were calculated for each site and time slice. We only considered

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regions with at least one speleothem record for each of the three time periods, although these were not required to be the same sites, and where the observed shifts in $\delta^{18}O_{spel}$ were in the same direction and of a similar magnitude to the simulated $w\delta^{18}O_{precip}$.

2.5 Holocene and Last Interglacial regional trends

Regional s~~S~~peleothem $\delta^{18}O$ changes through the Holocene were examined by creating composite time-series for each region identified in the PCoA analysis with at least four Holocene records (> 5000 years long). Regional composites were constructed using a 4-step procedure, modified from Marlon et al. (2008): (i) the $\delta^{18}O$ data for individual speleothems were transformed ~~standardised~~ to z-scores, so all records have a standardised mean and variance:

$$z\text{-}score_i = \left(\delta^{18}O_i - \overline{\delta^{18}O}_{\,(base\ period)} \right) / s\delta^{18}O_{\,(base\ period)} \quad\quad\quad\quad (4)$$

Where $\overline{\delta^{18}O}$ is the mean and $s\delta^{18}O$ is the standard deviation of $\delta^{18}O$ for a common base period. A base period of 7,000 to 3,000 years BP was chosen to maximise the number of records included in each composite.~~, by subtracting the mean $\delta^{18}O_{spel}$ for a defined base period (7,000 to 3,000 years BP) then dividing these anomalies by the base period standard deviation. The base period was chosen to maximise the number of records included in each composite;~~ (ii) the standardised data for a site were re-sampled by applying a 100-year non-overlapping running mean with the first bin centred at 50 years BP, in order to create a single site time series while ensuring that highly resolved records do not dominate the regional composite; (iii) each regional composite was constructed using locally weighted regression (Cleveland and Devlin, 1988) with a window width of 3,000 years and fixed target points in time; and (iv) confidence intervals (5th and 95th percentiles) for each composite were generated by bootstrap resampling by site over 1,000 iterations. There are too few sites to construct regional composites for the LIG and thus the trends in $\delta^{18}O_{spel}$ were examined using records from individual sites covering the period 130-116 ka BP. We calculated Holocene regional composites from annual precipitation-weighted mean~~simulated~~ $\delta^{18}O_{precip}$ anomalies simulated by~~from~~ the GISS model. Simulated $\delta^{18}O_{precip}$ trends were calculated using linear distance-weighted mean $\delta^{18}O_{precip}$ values from land grid cells (>50% land) within ±4° around each site. This distance was determined by the grid resolution of the model. Regional composites were then produced using bootstrap resampling in the same way as for the speleothem data. The simulated anomalies are relative to the control run rather than the specified base period used for the speleothem-based composites, so absolute values of simulated and observed Holocene trends are expected to differ. Preliminary analyses showed that neither the mean values nor trends in $\delta^{18}O_{precip}$ were substantially different if the sampled area was reduced to match the sampling used for the ECHAM-based box plot analysis, or was increased to encompass the larger regions shown in Fig. 1 and used in the multiple regression analysis.

2.6 Multiple regression analysis

We investigate the drivers of regional $\delta^{18}O_{precip}$, and by extension $\delta^{18}O_{spel}$, through the Holocene using multiple linear regression (MLR) of annual precipitation-weighted mean~~simulated~~ $\delta^{18}O_{precip}$ anomalies and climate variables from GISS model E-R. Climate variables were chosen to represent the four potential large-scale drivers of regional changes in the speleothem

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δ^{18}O records. Specifically, we use changes in mean precipitation and precipitation recycling over the monsoon regions, and changes in mean surface air temperature and surface wind direction over the moisture source regions. Whereas the influence of changes in precipitation, recycling and temperature are relatively direct measures, the change in surface wind direction over the moisture source region is used as an index of potential changes in the moisture source region and transport pathway. The boundaries of each monsoon region (Fig. 1) were defined to include all the speleothem sites used to construct the Holocene δ^{18}O$_{spel}$ composites. Moisture source area limits (Fig. 1) were defined based on moisture tracking studies (Bin et al., 2013; Breitenbach et al., 2010; D'Abreton and Tyson, 1996; Drumond et al., 2008, 2010; Durán-Quesada et al., 2010; Kennett et al., 2012; Nivet et al., 2018; Wurtzel et al., 2018) and GISS simulated summer surface winds. All climate variables were extracted for the summer months, defined as May to September (MJJAS) for northern hemisphere regions and November to March (NDJFM) for southern hemisphere regions (Wang and Ding, 2008). Only grid cells with >50% land were used to extract variables over monsoon regions and only grid cells with <50% land were used to extract variables over moisture source regions. The inputs to the MLR for each time interval were calculated as anomalies from the control run.

Precipitation recycling was calculated as the ratio of locally sourced precipitation versus total precipitation. Although the GISS E-R mid-Holocene experiment explicitly estimates recycling using vapour source distribution tracers (Lewis et al., 2014), this was not done for all the Holocene time slice simulations. Therefore, we calculate a precipitation recycling index (RI), following Brubaker et al. (1993):

$$RI = \frac{P_R}{P} = \frac{E}{2Q_H + E} \qquad (53)$$

Where locally sourced (recycled) precipitation (P$_R$) is estimated using total evaporation over a region (E) and total precipitation (P) is estimated as the sum of total evaporation and net incoming moisture flux integrated across the boundaries of the region (Q$_H$). RI therefore expresses the change in the contribution of local, recycled precipitation independently of any overall change in precipitation amount.

We use pseudo-R^2 to determine the goodness-of-fit for the overall MLR model, and t values (the regression coefficient divided by its standard error) to determine the strength of each relationship. Partial residual plots were used to show the relationship between each predictor variable and δ^{18}O$_{precip}$ when the effects of the other variables are removedheld constant. All statistical analyses were performed in R (R Core Team, 2019) and plots were generated using ggplot (Wickham, 2016).

3 Results

3.1 Principalle Coordinate Analysis and Redundancy Analysis

PCoA shows the (dis)similarity of Holocene δ^{18}O$_{spel}$ evolution across individual records, and thus allows an objective regionalisation of these records. The first two PCoA axes are significant, according to the broken stick test, and account for 65% and 20% of δ^{18}O$_{spel}$ variability respectively (Table 1). The PCoA scores differentiate records geographically (Fig. 2a): southern hemisphere monsoon regions such as the southwestern South American Monsoon (SW-SAM) and South African

Monsoon (SAfM) are characterised by consistently low PCoA1 scores over time, whilst northern hemisphere monsoons such as the Indian Summer Monsoon (ISM) and the East Asian Monsoon (EAM), are characterised by higher PCoA1 scores. This indicates that regions can be it is possible to differentiated the regions based on their temporal evolution as captured by the first PCoA axis. Most southern hemisphere regions also have lower PCoA2 scores although this is not consistent over time. Speleothem records from Central America (CAM) and Indonesian-Australian monsoon (IAM) have PCoA scores intermediate between the northern and southern hemisphere regions. PCoA clearly separates the South American records into a northeastern region (NE-SAM) with scores similar to other northern hemisphere monsoon regions and a southwestern region (SW-SAM), with scores similar to other southern hemisphere regions. The RDA supports a geographical control on the (dis)similarity of speleothem $\delta^{18}O$ records over the Holocene (Fig. 2b). RDA1 explains 37% of the variability and is significantly correlated with both latitude and longitude (Table 2).

3.2 Regional interglacial-glacial differences

To investigate the causes of glacial-interglacial shifts in $\delta^{18}O$, we cCompareison of simulated and observed regional $\delta^{18}O$ signals during the LIG, LGM and MH with shifts in climate variables (precipitation and temperature). allows the causes of glacial-interglacial shifts to be investigated. Regional scale signals of $\delta^{18}O_{spel}$ and simulated $\delta^{18}O_{precip}$ anomalies shown in Fig. 3 are consistent with large-scale glacial-interglacial shifts in monsoon climate (precipitation and temperature). Only the ISM, EAM and IAM regions have sufficient speleothem data (i.e. at least one record from every time period) to allow comparisons across the MH, LGM and LIG (Fig. 3) and have similar shifts in observed $\delta^{18}O_{spel}$ and simulated $\delta^{18}O_{precip}$. The regional mean $\delta^{18}O_{spel}$ anomalies calculated for different time windows ($\pm 500, \pm 700, \pm 1000, \pm 2000$ years) vary by less than 0.35‰ for the LGM (ISM: <0.16‰, EAM: <0.35‰, IAM: <0.22‰) and 0.48 ‰ for the LIG (ISM: <0.16‰, EAM: <0.48‰, IAM: <0.11‰), indicating that age uncertainties have a minimal impact on these mean values. The most positive $\delta^{18}O_{spel}$ anomalies in all three regions occur at the LGM, with more negative anomalies for the MH and LIG. The simulated $\delta^{18}O_{precip}$ anomalies show a similar pattern, more positive anomalies during the LGM than during the MH or the LIG. Although the mean values of the simulated $\delta^{18}O_{precip}$ anomalies differ from the observed $\delta^{18}O_{spel}$ anomalies, they are consistent with the observations within the range of their uncertainties. The differences in regional $\delta^{18}O_{spel}$ anomalies between MH and LIG differ across the three regions. In both the ISM and the EAM, differences in $\delta^{18}O_{spel}$ values between the MH and LIG are small (Fig. 3a, 3b), although ISM LIG $\delta^{18}O_{spel}$ values are slightly more negative than MH values. In the IAM, MH values are less negative than the LIG (Fig. 3c). However, there are only a limited number of speleothem records from the ISM and IAM during the LIG, so the apparent differences between the two intervals in these regions may not be meaningful. The glacial-interglacial changes in $\delta^{18}O_{precip}$ are consistent with the simulated temperature and precipitation changes, with warmer and wetter conditions during interglacials and cooler and drier conditions during the LGM in all three regions. Differences in simulated precipitation between the MH and the LIG could help explain the differences between $\delta^{18}O_{spel}$ in the ISM and IAM, since the LIG is wetter than the MH in the ISM and drier than the MH in the IAM. However, the LIG is also drier than the MH in the EAM, a feature that appears inconsistent with the lack of differentiation between the $\delta^{18}O$ signals in this region.

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3.3 Regional-scale interglacial δ¹⁸O evolution

There are four underlined monsoon regions with sufficient data to examine regional Holocene $\delta^{18}O$ trends: EAM, ISM, SW-SAM and IAM (Fig. 4). The IAM region has the fewest records (n=7) whilst the EAM has the largest number (n=14). The regional composites are expressed as z-scores, i.e. changes~~anomalies~~ with respect to the mean and variance of $\delta^{18}O$ for the base period (3000-7000 yr BP). The confidence intervals on the regional composites are small for all regions, except SW-SAM in the early Holocene. The EAM and ISM regions (Fig. 4 ~~left~~a-e) show the most positive $\delta^{18}O_{spel}$ z-scores around 12 ka followed by a rapid decrease towards their most negative values at ~9.5 ka and ~9 ka, respectively. The $\delta^{18}O_{spel}$ z-scores in the EAM are relatively constant from 9.5 to ~7 ka, whereas this plateau is present but less marked in the ISM. There is a gradual trend towards more positive $\delta^{18}O_{spel}$ z-scores towards the present in both regions thereafter. The SW-SAM records (Fig. 4i) have their most positive $\delta^{18}O_{spel}$ z-scores in the early Holocene with a gradual trend to more negative scores towards the present. By contrast, the IAM z-scores (Fig. 4g) are most positive at 12ka, gradually decreasing until ca 5 ka and are relatively flat thereafter.

There are insufficient data to create composite curves for the LIG, but individual records from the four regions (Fig. 5) show similar features to the Holocene trends. Records from the ISM and EAM (Fig. 5 left), for example, are characterised by an initial sharp decrease in $\delta^{18}O_{spel}$ values of about 4 ‰ between 130-129 ka and then most of the records (Dykoski et al., 2005; Kathayat et al., 2016; Wang et al., 2008) show little variability for several thousand years. Despite the fact that the Tianmen record (Cai et al., 2010, 2012) shows considerable variability between 123-127 ka, there is nevertheless a similar plateau in the average observed value before the rapid change to less negative values after 127 ka. Similar to the Holocene, the SW-SAM record (Cheng et al., 2013) shows increasingly negative $\delta^{18}O_{spel}$ values through the LIG. The trend shown for Whiterock cave (Carolin et al., 2016) also shows similar features to the IAM Holocene composite, with a gradual trend towards more negative values initially and a relatively complacent curve towards the end of the interglacial (Fig. 5 right).

3.4 Multiple regression analysis of Holocene δ¹⁸Oₚᵣₑᵢₚ

The MLR analyses of simulated $\delta^{18}O_{precip}$ trends identify the impact of an individual climate variable on $\delta^{18}O_{precip}$ in the absence of changes in other variables. ~~Multiple regression analysis allows simulated $\delta^{18}O_{precip}$ trends to be investigated as a function of multiple climate variables. This method identifies and constrains relationships between a predictor variable (e.g., precipitation, temperature or atmospheric circulation) and $\delta^{18}O_{precip}$ whilst taking into account the effects of the other variables.~~ The global ~~multiple linear regression~~ model ~~that includes all of~~for the Holocene (1 to 9ka) $\delta^{18}O_{precip}$ ~~regional~~ trends has a pseudo-R^2 of 0.80 and shows statistically significant relationships between the anomalies in $\delta^{18}O_{precip}$ and anomalies in regional precipitation ~~amount~~, temperature and surface wind direction (Table 3). The partial residual plots (Fig. 6) show ~~s the nature of the relationship with each variable when the other variables are held constant. T~~there is a strong negative relationship with regional precipitation ~~amount~~ (t value = -8.75) and a strong positive relationship (t value = 8.03) with surface wind direction over the moisture source region, an index of changes in either source area or moisture pathway. This indicates that increases in regional

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precipitation alone will lead to a decrease in δ^{18}O while changes in source area/moisture pathway, in the absence of changes in other variables, will lead to significant change in δ^{18}O. The relationship with temperature over the moisture source region is weaker, but positive (t value = 2.05), i.e. an increase in temperature over the moisture source region will lead to an increase in δ^{18}O if there are no changes into other climate variables. Precipitation recycling is not significant in this global analysis.

There are too few data points to make regressions for individual monsoon regions, but the distribution of data points for each region in the partial residual plots (Fig. 6) is indicative of the degree of conformity to the global model. Data points from the ISM, SW-SAM, IAM and SAfM are well aligned with the global relationship with regional precipitation amount (Fig. 6a), indicating that precipitation amount is an important control on changes in δ^{18}O$_{precip}$ in these regions. The NE-SAM, EAM and CAM values deviate somewhat from the global relationship and, although there are relatively few points, this suggests that changes in precipitation are a less important influence on δ^{18}O$_{precip}$ changes in these regions. The impact of temperature changes (Fig. 6b) in the ISM, EAM and SW-SAM is broadly consistent with the global model. The slope of the relationship with temperature is negative for the IAM and NE-SAM, and since this is physically implausible it suggests that some factor not currently included in the MLR is influencing these records. However, the inconsistencies between the regional signals helps to explain why the global relationship between anomalies in temperature and δ^{18}O$_{precip}$ is weak (Fig. 6b) and probably reflects the fact that tropical temperature changes during the Holocene are small. Data points from the EAM, ISM and IAM are well aligned with the global relationship between changes in δ^{18}O$_{precip}$ and changes in wind direction (Fig. 6c), indicating that changes in source area or moisture pathway are an important control on changes in δ^{18}O$_{precip}$ in these regions. However, values for CAM, SW-SAM, NE-SAM and SAfM deviate strongly from the global relationship. Recycling does not appear to be an important contributor to changes in δ^{18}O$_{precip}$ except in SW-SAM and SAfM (Fig. 6d).

4 Discussion

We have shown that it is possible to derive an objective regionalisation of speleothem records based on PCoA of the oxygen-isotope trends through the Holocene (Fig. 2). This approach separates out regions with a distinctive northern hemisphere signal (e.g. ISM, EAM, NE-SAM) from regions with a distinctive southern hemisphere signal (e.g. SW-SAM, SAfM), reflecting the fact that the evolution of regional monsoons in each hemisphere follows, to some extent, insolation forcing. It also identifies regions that have an intermediate pattern (e.g. IAM). The robustness of the regionalisation is borne out by the fact that Holocene composite trends from each region have tight confidence intervals (Fig. 4), showing that the signals of individual records across a region show broad similarities. The monsoon regions identified by PCoA are consistent with previous studies (Wang et al., 2014). The tracking of northern hemisphere insolation is a recognised feature of monsoon systems in India and China (see reviews by Kaushal et al., 2018; Zhang et al., 2019). The separation of speleothem records from NE-SAM from those in SW-SAM is consistent with the precipitation dipole that exists between northeastern Brazil (Nordeste) and the continental interior (Berbery and Barros, 2002; Boers et al., 2014). The anti-phasing of speleothem records from the two regions during the Holocene has been recognised in previous studies (Cruz et al., 2009; Deininger et al., 2019). The intermediate nature of

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the records from the maritime continent is consistent with the fact that the Indonesian-Australian (IAM) summer monsoon is influenced by cross-equatorial air flow and hence can be influenced by northern hemisphere conditions (Trenberth et al., 2000). Palaeoenvironmental records from this region show mixed signals for the Holocene: some have been interpreted as showing enhanced (Beaufort et al., 2010; Mohtadi et al., 2011; Quigley et al., 2010; Wyrwoll and Miller, 2001) and others reduced precipitation (Kuhnt et al., 2015; Steinke et al., 2014) during the early and mid-Holocene. Modelling studies have shown that this region is highly sensitive to SST changes in the Indian Ocean and Ssouthern China Sea, which in turn reflect changes in the northern hemisphere winter monsoons. Although most climate models produce a reduction in precipitation across the IAM during the mid-Holocene in response to orbital forcing, this is less than might be expected in the absence of ocean feedbacks associated with changes in the Indian Ocean (Zhao and Harrison, 2012).

The separation of northern and southern monsoon regions is consistent with the idea that changes in monsoon rainfall are primarily driven by changes in insolation (Ding and Chan, 2005; Kutzbach et al., 2008). Indeed, regional $\delta^{18}O_{spel}$ composites from the EAM, ISM and SW-SAM show a clear relationship with the long-term trends in local summer insolation (Fig. 4). Similar patterns are seen in individual speleothem records from each region confirming that the composite trends are representative. However, the composite trends are not an exact mirror of the insolation signal over the Holocene. For example, the ISM and EAM composites show a more rapid rise during the early Holocene than implied by the insolation forcing. The maximum wet phase in these two regions lasts for ca 3,000 years, again contrasting with the gradual decline in insolation forcing after its peak at ca 11 ka. Both the rapid increase and the persistence of wet conditions for several thousand years is also observed in other palaeohydrological records across southern and central China, including pollen (Zhao et al., 2009; Li et al., 2018) and peat records (Hong et al., 2003; Zhou et al., 2004). These features are also characteristic of lake records from India (Misra et al., 2019). The lagged response to increasing insolation is thought to be due to the presence of northern hemisphere ice sheets in the early Holocene (Zhang et al., 2018). The persistence of wetter conditions through the early and mid-Holocene is thought to reflect the importance of land-surface and ocean feedbacks in sustaining regional monsoons (Dallmeyer et al., 2010; Kutzbach et al., 1996; Marzin and Braconnot, 2009; Rachmayani et al., 2015; Zhao and Harrison, 2012). The evolution of regional monsoons during the LIG shows patterns similar to those observed during the Holocene, including the lagged response to insolation and the persistence of wet conditions after peak insolation. This is again consistent with the idea that internal feedbacks play a role in modulating the monsoon response to insolation forcing. We have also shown that there is little difference in the isotopic values between the MH and the LIG in the ISM and EAM regions, which is also observed in individual speleothem records (Kathayat et al., 2016; Wang et al., 2008). Given that the increase in summer insolation is much larger during the LIG than the MH, this finding indicates is again consistent with the idea that other factors play a role in modulating the monsoon response to insolation forcing and may illustratereflect the importance of global constraints on the externally-forced expansion of the tropical circulation (Biasutti et al., 2018).

Global relationships between $\delta^{18}O_{precip}$ and climate variablesits drivers (precipitation amount, temperature and surface wind direction; Fig. 6) are consistent with existing studies: a strong relationship with precipitation and a weaker temperature effect

has been widely observed at tropical and sub-tropical latitudes in modern observations (Dansgaard, 1964; Rozanski et al., 1993). The significant global relationship between $\delta^{18}O_{precip}$ and surface winds supports the idea that changes in moisture source and pathway are also important for explaining $\delta^{18}O$ variability over the Holocene. The multiple regression analysis also provides insights into the relative importance of different influences at a regional scale. In the ISM, the results support existing speleothem studies that suggest changes in precipitation amount (Cai et al., 2015; Fleitmann et al., 2004) and to a lesser extent moisture pathway (Breitenbach et al., 2010) drive $\delta^{18}O_{spel}$ variability. The $\delta^{18}O$ variability in the IAM region through the Holocene also appears to be strongly driven by changes in precipitation and moisture pathway, consistent with the interpretation of Wurtzel et al. (2018). Changes in regional precipitation (where the cave sites are located) ~~amount~~ do not seem to explain the observed changes in $\delta^{18}O_{spel}$ in the EAM during the Holocene, where Holocene $\delta^{18}O_{precip}$ evolution is largely driven by changes in atmospheric circulation (indexed by changes in surface winds). This is consistent with existing studies that emphasise changes in moisture source and/or pathway rather than local precipitation changes~~a precipitation amount~~ (Maher, 2016; Maher and Thompson, 2012; Tan, 2014; Yang et al., 2014). Speleothem $\delta^{18}O$ records in the SW-SAM clearly reflect regional-scale changes in precipitation, consistent with interpretations of individual records (Cruz et al., 2009; Kanner et al., 2013). However, this is a region where changes in precipitation recycling also appears to be important, perhaps unsurprisingly given that recycling presently contributes ca 25-35% of the precipitation over the Amazon (Brubaker et al., 1993; Eltahir and Bras, 1994).

The LGM is characterised by lower northern hemisphere summer insolation, globally cooler temperatures~~a similar orbital configuration to today~~, ~~however~~ expanded global ice volume~~s was at a maximum~~ and lower GHG concentrations ~~were lower~~ than either the MH or the LIG~~present~~. The $\delta^{18}O_{spel}$ anomalies are more positive during the LGM than the MH or LIG, suggesting drier conditions in the ISM, EAM and IAM, supported by simulated changes in $\delta^{18}O_{precip}$ and precipitation (Fig. 3). Cooler SSTs of approximately 2°C (relative to the MH and LIG) in the ISM and EAM and of approximately 3°C in IAM source areas, together with a ca 5% decrease in relative humidity (Yue et al., 2011) would result in a water vapour $\delta^{18}O$ signal at the source ca 1 ‰ more depleted than seawater. This depletion results from the temperature dependence of equilibrium fractionation during evaporation and kinetic isotope effects related to humidity (Clark and Fritz, 1997). This fractionation counteracts any impact from enriched seawater $\delta^{18}O$ values during the LGM (ca. +1 ‰ relative to the MH or LIG; Waelbroeck et al., 2002). Enriched $\delta^{18}O_{precip}$ and $\delta^{18}O_{spel}$ values during the LGM must therefore be caused by a significant decrease in atmospheric moisture and precipitation that resulted from the cooler conditions.

We have used version 2 of the SISAL database (Atsawawaranunt et al., 2018; Comas-Bru et al., 2020a) in our analyses. Despite the fact that SISALv2 includes more than 70% of known speleothem isotope records, there are still too few records from some regions (e.g. Africa, the Caribbean) to make meaningful analyses. The records for older time periods are also sparse. There are only 14 records from monsoon regions covering the LIG in SISALv2, for example. Nevertheless, our analyses show that there are robust and explicable patterns for most monsoon regions during the Holocene and sufficient records to make meaningful analyses of the LGM and LIG. Whilst there is a need for the generation of new speleothem records from key regions such as

northern Africa, further expansion of the SISAL database will certainly provide additional opportunities to analyse the evolution of the monsoons through time.

The impact of age uncertainties, included in SISALv2, are not taken into account in our analyses. Age uncertainties during the Holocene are smaller than the interval used for binning records and the width of the time windows used, and thus should not
485 have a significant effect on our conclusions. The mean age uncertainty at the LGM and LIG is ca 430 and 1140 years, respectively. However, varying the window length for the selection of LGM and LIG samples from ±500 to ±2000 years, thereby encompassing this uncertainty, has a negligible effect (<0.5 ‰) on the average $\delta^{18}O$ values. Thus, the interglacial-glacial contrast in regional $\delta^{18}O_{spel}$ is also robust to age uncertainties.

Isotope-enabled climate models are used in this study to explain observed regional-scale trends in $\delta^{18}O_{spel}$. There is a limited
490 number of isotope-enabled models, and there are no simulations of the same time period using the same experimental protocol. Although there are simulations of the MH from both ECHAM5-wiso and GISS, for example, these models have different grid resolutions and used different boundary conditions. This could help to explain why the two models yield different estimates of the change in regional $\delta^{18}O_{precip}$ (of 0.5 ‰) at the MH. However, both models show trends in $\delta^{18}O_{precip}$ that reproduce the observed changes in regional $\delta^{18}O_{spel}$ (Figs 3 and 4), and this provides a basis for using these models to explore the causes of
495 these trends on different timescales. The failure to reproduce the LGM $\delta^{18}O_{spel}$ signal in SW-SAM in the ECHAM5-wiso model, which precluded a consideration of interglacial-glacial shifts in this region, is a common feature of other isotope-enabled simulations (Caley et al., 2014; Risi et al., 2010).

This study illustrates a novel data-model approach to investigate the drivers of $\delta^{18}O_{spel}$ under past conditions, by comparing composite regional records and then using multiple linear regression of isotope-enabled palaeoclimate simulations to determine
500 the role of individual climate drivers of these trends. This obviates the need to use modern $\delta^{18}O_{precip}$-climate relationships to explain changes under conditions considerably different from today or to rely on coherency between different palaeohydrological archives which may respond to different climate variables. This model interrogation approach could be employed to address questions about the regional drivers of speleothem records outside the monsoon regions.

5 Conclusions

505 Geographically distributed speleothem $\delta^{18}O$ records and isotope-enabled climate models can be used together to understand the drivers of $\delta^{18}O_{spel}$ on orbital timescales. Speleothem records, objectively grouped into monsoon regions by record correlation and multivariate ordination techniques, show regional trends that are consistent with changes in summer insolation but modulated by land-surface and ocean feedbacks Glacial $\delta^{18}O_{spel}$ signals are best explained by a large decrease in precipitation, as a consequence of lower atmospheric moisture content driven by global cooling. The evolution of $\delta^{18}O_{spel}$
510 through the Holocene across the global monsoon domain is driven by changes in precipitation, atmospheric circulation and temperature. At the regional scale, our analyses support the increasing number of studies suggesting that East Asian monsoon

speleothem $\delta^{18}O$ evolution through the Holocene relates to changes in atmospheric circulation (i.e. changes in moisture pathway and/or source). Changes in regional precipitation ~~amount~~ are the predominant driver of Holocene $\delta^{18}O_{spel}$ evolution in the Indian, southwestern South American and Indonesian-Australian monsoons, although changes in atmospheric circulation also contribute in the Indian and Indonesian-Australian monsoon regions and changes in precipitation recycling in southwestern South America.

Code and data availability

The SISAL (Speleothem Isotopes Synthesis and AnaLysis) database version 2 is available through the University of Reading Research Archive at http://dx.doi.org/10.17864/1947.256~~http://dx.doi.org/10.17864/1947.242~~. The ECHAM5-wiso MH and LGM simulations are available at https://doi.org/10.1594/PANGAEA.902347. The ECHAM LIG simulation is available at https://doi.pangaea.de/10.1594/PANGAEA.879229. The OIPC mean annual $\delta^{18}O_{precip}$ data are available at http://wateriso.utah.edu/waterisotopes/pages/data_access/ArcGrids.html. CRU-TS4.01 mean annual temperature data are available at http://doi.org/10/gcmcz3. The GISS simulations and code used to generate the figures in this paper are~~is~~ available at https://doi.org/10.5281/zenodo.3875496.

Author contributions

The study was designed by SP, SPH, LCB and NK. MW and ALG provided climate model outputs. SP ran the analyses. The first draft of the manuscript was written by SP, SPH and LCB and all authors contributed to the final draft.

Competing interests

The authors declare no competing interests.

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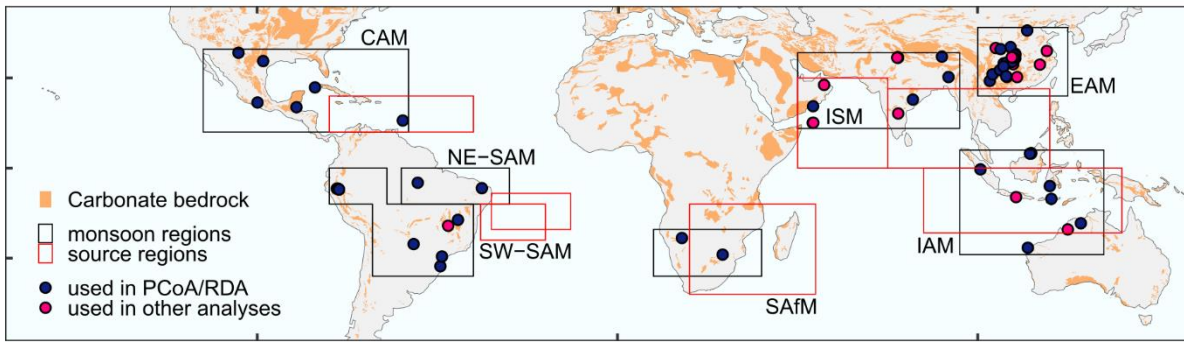
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Figure 1: Spatial distribution of speleothem records used is this study. Colours indicate the sites used in Principalle Coordinates Analysis and Redundancy Analysis (PCoA, RDA) to separate monsoon regions, and sites not used in PCoA and RDA but used in subsequent analyses. The individual regional monsoons are shown by boxes: Boxes used in the allocation of sites into individual regional monsoons are shown: CAM = Central American Monsoon latitude: 10 to 33°; longitude: -115 to -58°), SW-SAM = southwestern South American Monsoon latitude: -10° to 0°; longitude: -80° to -64° and latitude: -30° to -10°; longitude -68° to -40°), NE-SAM = northeastern South American Monsoon (latitude: -10° to 0°; longitude: -60° to -30°), SAfM = southern African Monsoon (latitude: -30° to -17°; longitude: 10° to 40°), ISM = Indian Summer Monsoon (latitude: 11° to 32°; longitude: 50° to 95°), EAM = East Asian Monsoon (latitude: 20° to 39°; longitude: 100° to 125°), IAM = Indonesian-Australian Monsoon (latitude: -24° to 5°; longitude: 95° to 135°). Source region limits used in the multiple linear regression analysis are also shown. The background carbonate lithology is from the World Karst Aquifer Mapping (WOKAM) project (Goldschneider et al., 2020)(Chen et al., 2017).
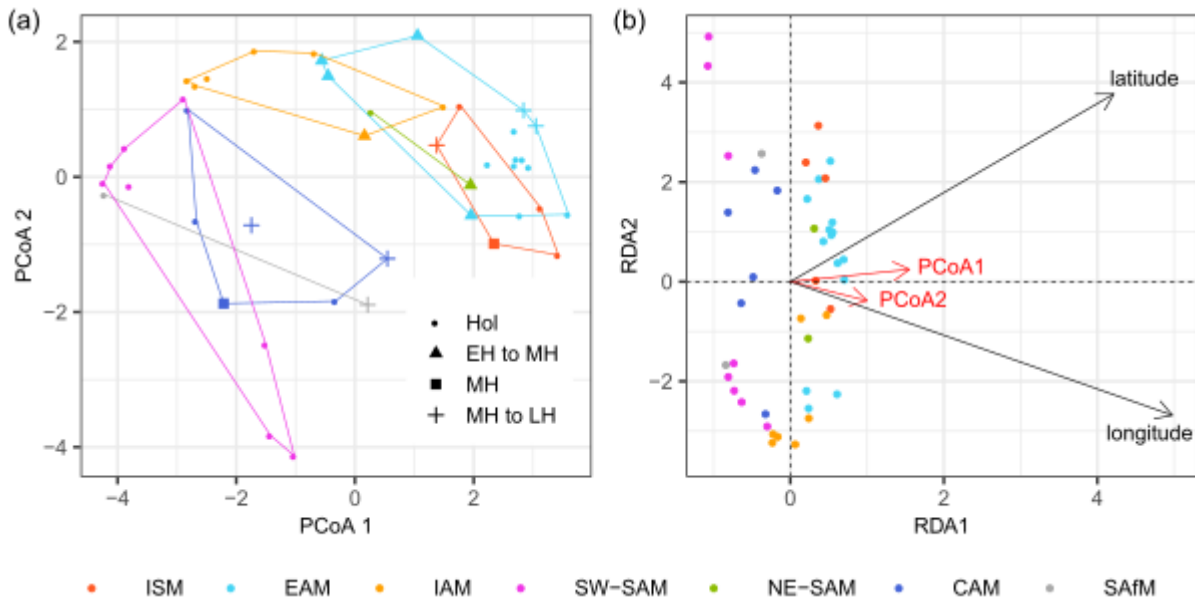

Figure 2: Results of Principalle Coordinate Analysis (PCoA) and Redundancy analysis (RDA). (a) PCoA biplot showing the loadings of each site on the first 2 axes, which represent 85% of the total variance. Shapes indicate the Holocene coverage of each site, where sites with a coverage ≥ 8000 years represent most or all of the Holocene (Hol). Sites with a temporal coverage of < 8000 years are coded to show whether they represent the early to mid-Holocene (EH to MH, record midpoint > 8,000 years BP), the mid Holocene (MH, record midpoint between 8,000 and 5,000 years BP), or the mid to late-Holocene (LH to MH, midpoint <5,000 years BP). (b)

RDA triplot, where the response variables are the PCoa1 and PCoA2 axes explained by latitude and longitude. The direction of the PCoA axes have been fixed so that they align with the explanatory variables.


(a) EAM

(b) ISM

(c) IAM

MH   LGM   LIG
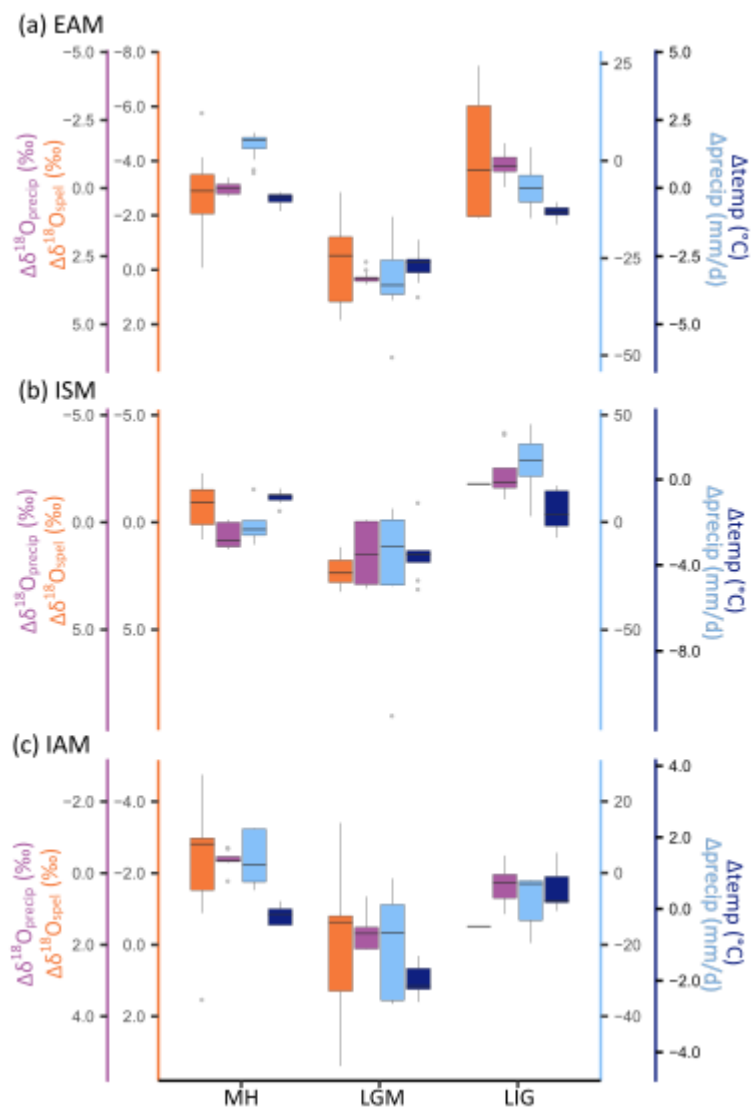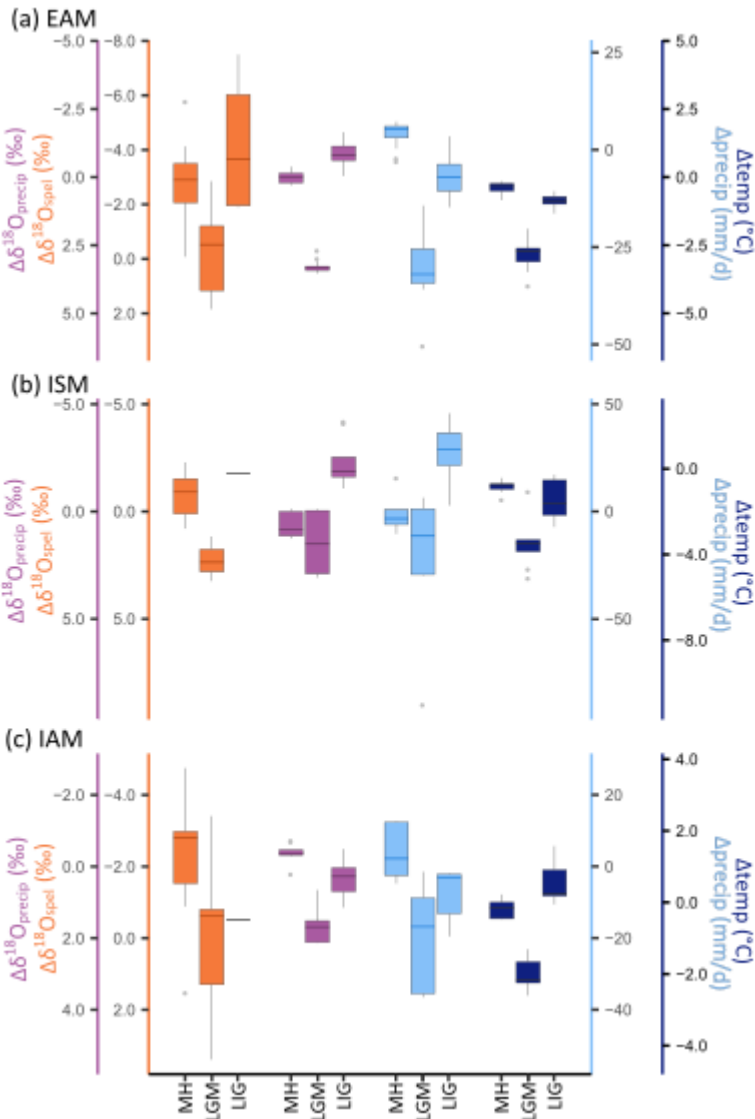
Figure 3: Speleothem δ¹⁸O anomalies compared to anomalies of δ¹⁸O$_{precip}$, precipitation and temperature from the ECHAM simulations for the (a) East Asian (EAM), (b) Indian (ISM) and (c) Indonesian-Australian (IAM) monsoons. The boxes show the median value (line) and the interquartile range, and the whiskers shown the minimum and maximum values, with outliers represented by grey dots. Note that the isotope axes are reversed, so that the most negative anomalies are at the top of the plot, to be consistent with the assumed relationship with the direction of change in precipitation and temperature.

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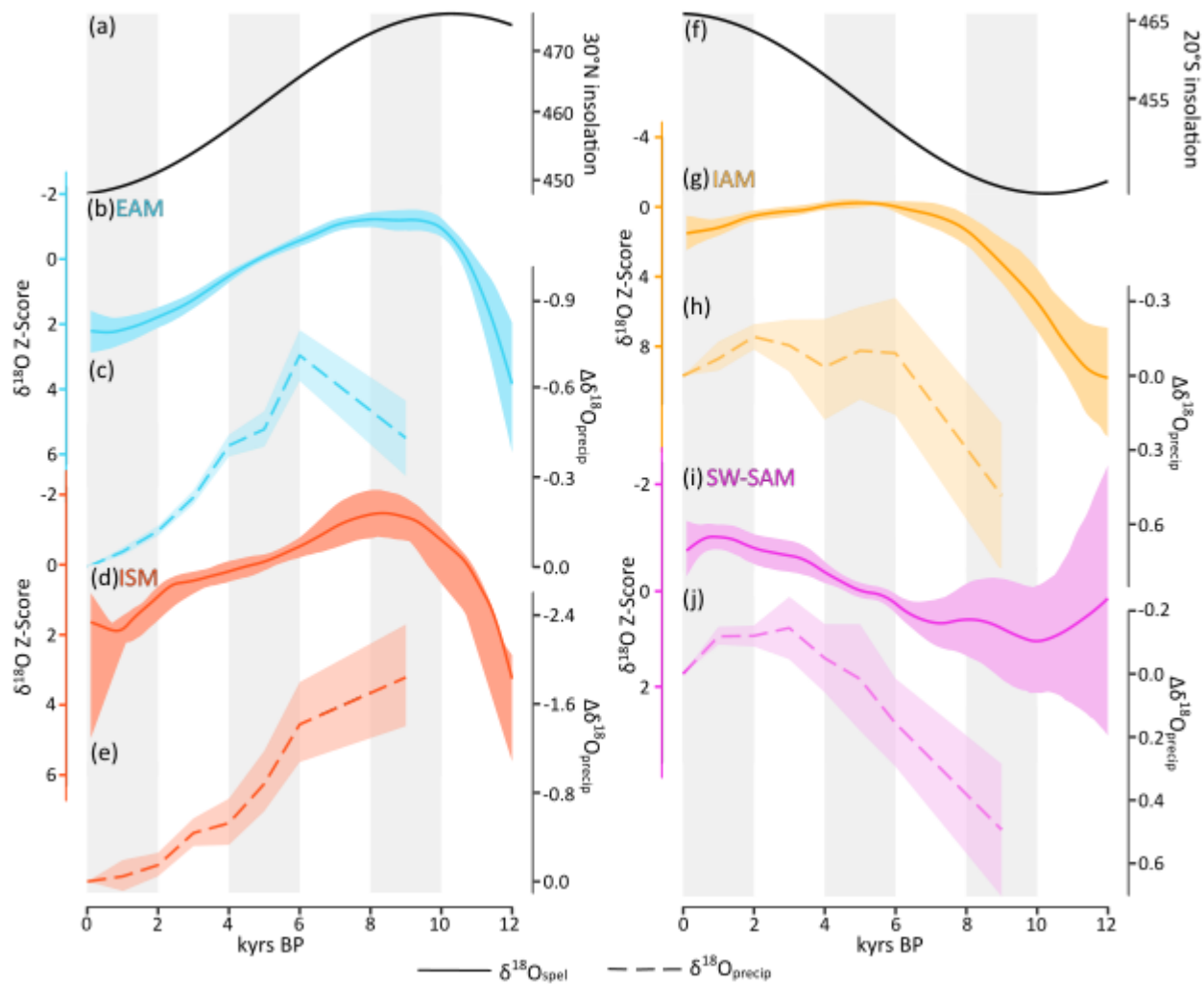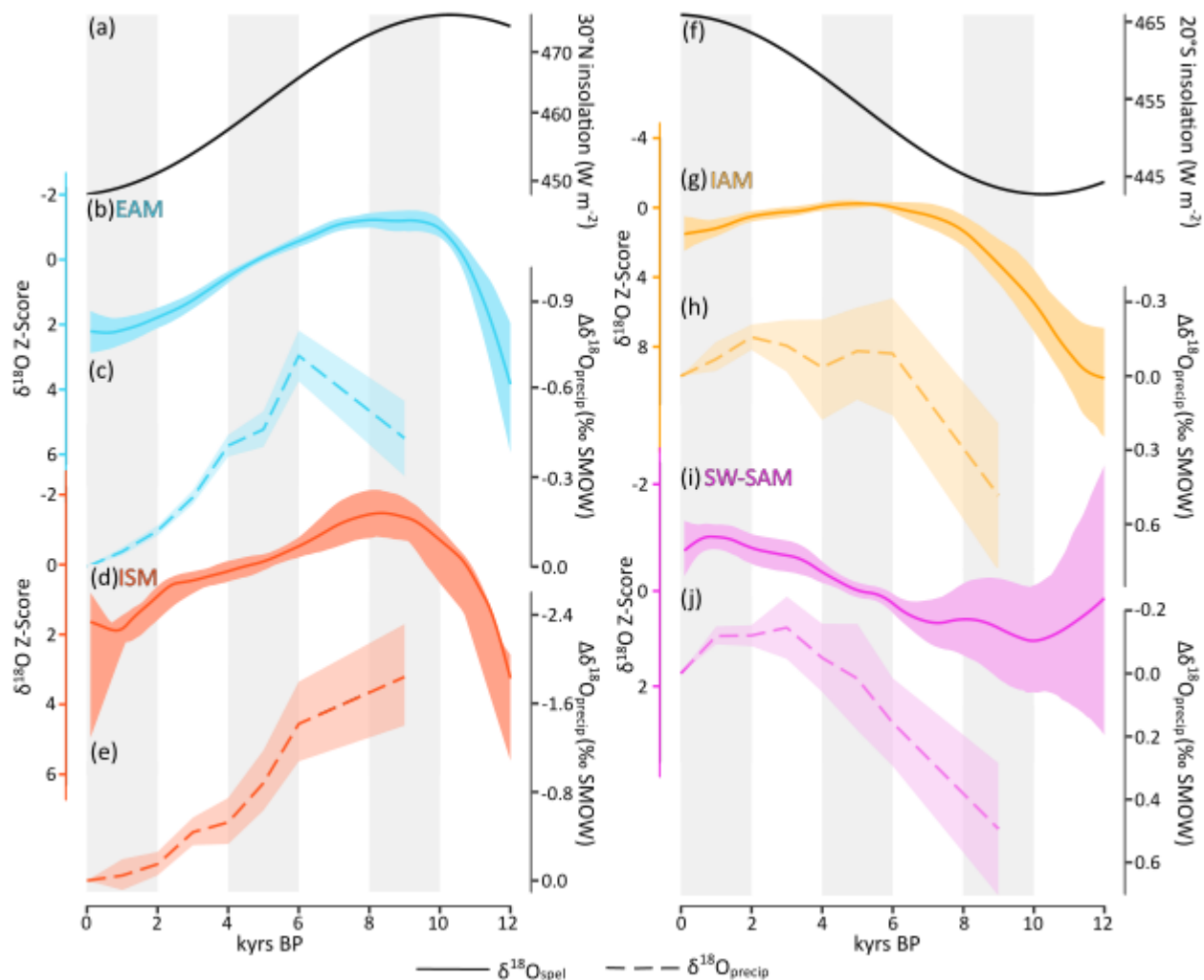
Figure 4: Evolution of regional speleothem δ18O signals through the Holocene compared to δ18Oprecip simulated by the GISS model. The left panel shows northern hemisphere monsoons (EAM = East Asian Monsoon; ISM = Indian Summer Monsoon) and summer (May through September) insolation at 30° N (Berger, 1978). The right panel shows southern hemisphere monsoons (SW-SAM = southwest South American Monsoon; IAM = Indonesian-Australian Monsoon) and summer (November through March) insolation for 20° S (Berger, 1978). The speleothem δ18O changes are expressed as z-scores, with a smoothed loess fit (3,000 year window), and confidence intervals obtained by bootstrapping by site. δ18Oprecip values are expressed as anomalies from the pre-industrial control simulation. Note that the isotope axes are reversed, so that the most negative anomalies are at the top of the plot, to be consistent with the assumed relationship with the changes in insolation.
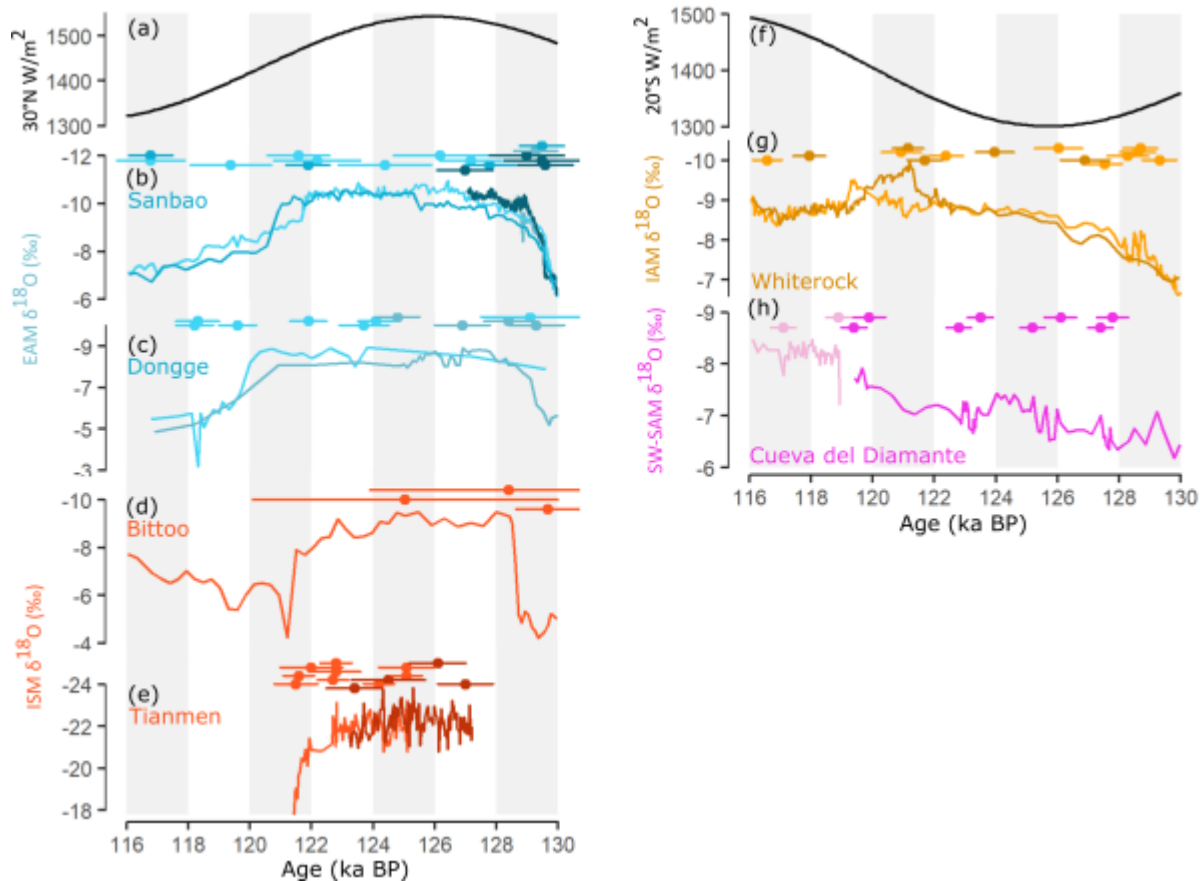
Figure 5: Comparison of changes in summer insolation and Last Interglacial δ¹⁸O_{spel} records from the (b,c) East Asian Monsoon (EAM), (d,e) Indian Summer Monsoon (ISM), (g) southwest South American Monsoon (SW-SAM) and (h) Indonesian-Australian Monsoon (IAM) regions. The U/Th dates and uncertainties are shown for each record. The summer insolation curves (Berger, 1978) are for May through September at 30° N in the northern hemisphere (a) and for November through March for 20° S in the southern hemisphere (f). Note that the isotope axes are reversed, so that the most negative anomalies are at the top of the plot, to be consistent with the assumed relationship with the changes in insolation.
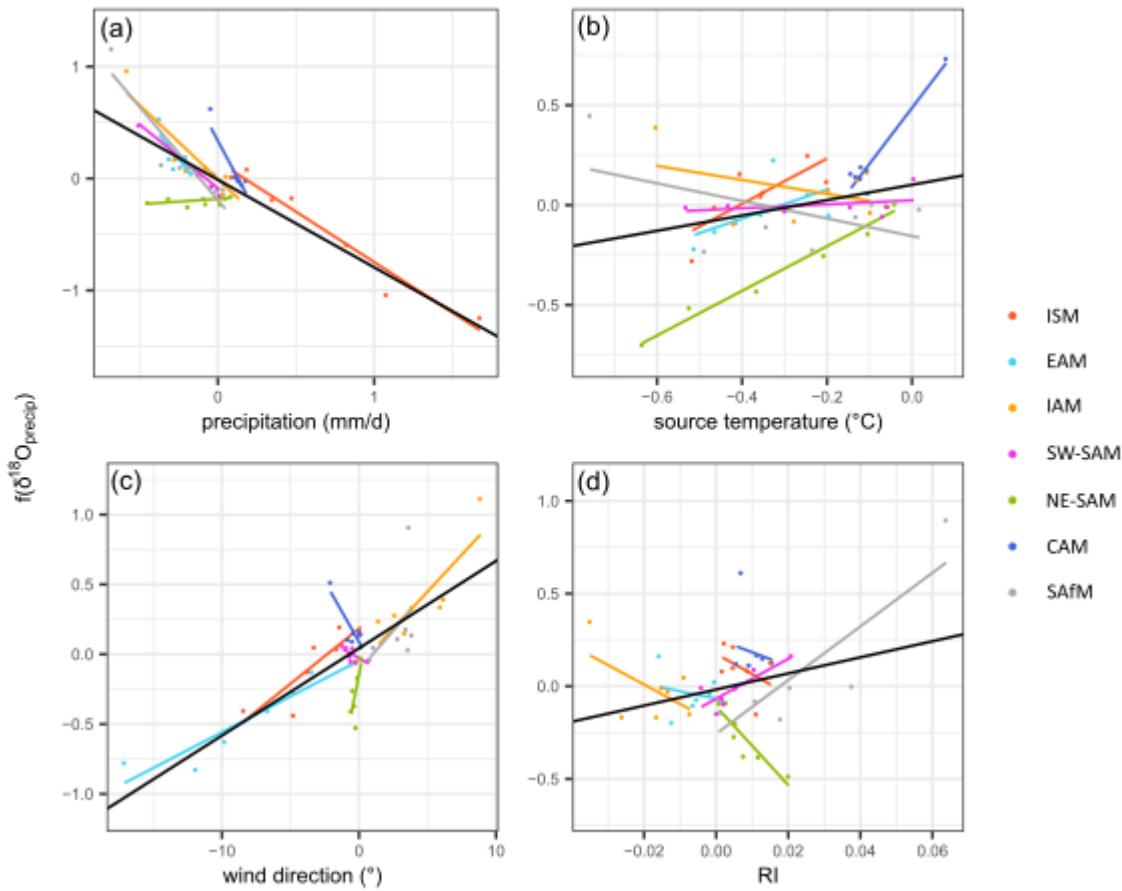
Figure 6: Partial residual plots from the multiple linear regression analysis, showing the relationship between anomalies in simulated $\delta^{18}O_{precip}$ and the four predictor variables, after taking account of the fitted partial effects of all the other predictors. The simulated $\delta^{18}O_{precip}$ are anomalies relative to the pre-industrial control simulation, and are annual values weighted by precipitation amount. The predictor variables are: precipitation in the delineated monsoon region (mm/d), temperature in the source region (°C), surface wind direction over the source region (°) as an index of potential changes in source region and the ratio of precipitation recycling to total precipitation over the monsoon region (RI, unitless). The predictor variables are summer mean values, where summer is defined as May to September for northern hemisphere monsoons and November to March for southern hemisphere monsoons.

	PCoA1	PCoA2	PCoA3	PCoA4	PCoA5
Eigenvalue	**269.06586**	**85.22482**	16.81258	10.25072	5.552469
Explained (%)	**64.87459**	**20.554915**	4.0543797	2.471623	1.3438794
Cumulative (%)	**64.87459**	**85.42374**	89.487753	91.954916	93.278795

Table 1: Results of the Principalle Coordinates Analysis (PCoA). Significant axes, as determined by the broken stick method (Bennett, 1996), are shown in bold.

	RDA1	RDA2
Latitude	**0.88**	-0.47
Longitude	**0.75**	0.67
Eigenvalue	0.73	0.04
Explained (%)	36.7	2.2

Table 2: Results of the redundancy analysis (RDA). Variables that are significantly correlated (P <0.01) with the RDA axes are shown in bold.

	Regression coefficient	T value
Regional precipitation	**-0.16**	-8.75
Source area temperature	**0.39**	2.05
Wind direction	**0.06**	8.03
Precipitation recycling	4.34	1.92

Table 3: Results of the multiple linear regression analysis. Significant relationships (P > 0.01) are shown in bold.