Anonymous referee #1

The paper has been improved by the authors and is now clearer and more accessible. The additional figures in the supplementary material are helpful. However, my major comment remains. I think that the wording could be modified to clarify the limitations of this study.

Major comment:

Correlations do not always indicate causality, even within the world of a GCM. Reading the response to reviewers, I understand that the authors want to document what climate variables are associated with changes in speleothem δ18O. Ideally, they would have used observations only, but they argue why they prefer to use a GCM: I think this an important argument that should be added in the introduction of the paper. I agree that a GCM provides a physically consistent framework, where all climate variables are available for analysis. However, the world of the GCM is extremely complex, almost as complex as reality. Analyzing GCM outputs to identify drivers of δ18O variations is thus extremely complex. This is why different authors in previous studies have developed decomposition methods to quantify the relative effects of different processes (e.g. [Botsyun et al., 2016, Tabor et al., 2018]). In absence of such decomposition methods, the drivers cannot be quantified. At best, you can look at how δ18O variations correlate with climate variables. This is what you do. This identifies concomitant changes, but not drivers. Some concomitant variations may be fortuitous, mediated by other variables, or may contribute to a small fraction of δ18O variations. I think this should be clarified in the paper. The wording "causes", "drivers", "explanations" should be avoided, for example:

- l 91: "provide plausible explanations for" -> "provide the changes in climate variables associated with"
- l 110: "main drivers of" -> "changes in climate variables associated with"
- l 112: "potential and plausible causes of" -> "trends in climate variables associated with"
- same l 232, 262, 467, 469, 475

We agree with the referee that concomitant changes in $\delta^{18}O$ and meteorological variables does not necessarily mean that one drives the other. However, it is worth noting that the relationships identified in this paper between $\delta^{18}O$ and climate variables are consistent with our theoretical understanding of oxygen isotope systematics in the hydrological cycle and are consistent with other papers (as discussed from l417 to l433). Therefore, some of these terms: "plausible explanations", "potential and plausible causes" are suitable here. We will, however, reword several sentences that the referee suggests:

From l109:

"We use isotope-enabled model simulations, to investigate the potential causes of $\delta^{18}O_{spel}$ variability in regions where the models reproduce the large-scale $\delta^{18}O$ changes shown by observations."

From l232:

"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 changes in $\delta^{18}O_{spel}$ between the MH, LGM and LIG, and their association with changes in climate."

From l262:

“We investigate the underlying relationships between regional $\delta^{18}O_{precip}$ (and by extension $\delta^{18}O_{spel}$) and monsoon climate through the Holocene using multiple linear regression (MLR)”.

From l467:

“This study illustrates a novel data-model approach to investigate the relationship between $\delta^{18}O_{spel}$ and monsoon climate under past conditions: We compare composite regional records and then use multiple linear regression of isotope-enabled palaeoclimate simulations to determine the change in individual climate variables associated with these trends.”

From l474:

“Geographically distributed speleothem $\delta^{18}O$ records and isotope-enabled climate models can be used together to understand the underlying relationships between $\delta^{18}O_{spel}$ and monsoon climate in the past and therefore elucidate possible drivers of $\delta^{18}O$ variability.”

In the discussion and conclusion, the main limitations of the approach should be recalled: (1) limitations associated with the GCM-observation mismatches, emphasizing the need for a thorough evaluation of the GCM simulations; and (2) limitations related to the correlation analysis that does not allow to identify drivers: a decomposition method would be necessary.

We will add this point when discussing the limitations in this study.

From l458:

“Isotope-enabled climate models are used in this study to explore observed regional-scale trends in δ18Ospel. There is a limited 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 δ18Oprecip (of 0.5 ‰) at the MH. However, both models show trends in δ18Oprecip that reproduce the observed changes in regional δ18Ospel (Figs 3 and 4), and this provides a basis for using these models to explore the potential causes of these trends on different timescales. The failure to reproduce the LGM δ18Ospel 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). Identifying the underlying relationships between $\delta^{18}O_{precip}$ and monsoon climate variables using multiple linear regression allows us to identify plausible mechanistic controls on $\delta^{18}O$ variability in the monsoon regions. Correlations between $\delta^{18}O$ and specific climate variables do not explicitly indicate causality. However, the relationships identified in the MLR model are consistent with the theoretical understanding of oxygen isotope systematics, and the findings of this paper are consistent with existing studies, suggesting that these relationships provide a plausible explanation for observed changes.”

Minor comments

• l 59: “reducing” -> “weakening” (because it’s probably negative)

We will amend the text as follows: “thereby minimising the $\delta^{18}O_{precip}$/distance gradient along an advection path…”

• l 345-347: “warmer and wetter”: this would have opposite effects on δ 18O. So what is δ 18O consistent with? Same for “cooler and drier”

In this analysis, periods of more negative $\delta^{18}O$ values are warmer and wetter, whilst the LGM has less negative $\delta^{18}O$ values and cooler and drier conditions. The observation made in this section is that shifts are seen in all three variables. In the discussion (l434-l443), we elaborate that overall cooler conditions during the LGM result in a significant decrease in atmospheric moisture and therefore precipitation. Our use of the word “consistent” is unsuitable here, as we do not elaborate until later in the paper. Therefore, we will amend the text (at l320-l322) as follows:

“Glacial-interglacial shifts are also seen in precipitation and temperature, with warmer and wetter conditions during interglacials and cooler and drier conditions during the LGM in all three regions.”

• l 462: this number is a local recycling ratio. For δ 18O, what matters is the total fraction of the precipitation that comes from continental recycling on any land grid box, and the number can be larger than 50% ([Yoshimura et al., 2004, Risi et al., 2013]). For the effect of continental recycling on paleo isotopic records, you may cite [Pierrehumbert, 1999].

Recycling in this paper is approximated using regional water budgets, i.e. the proportion of moisture derived from within the region versus the proportion of advected moisture, rather than oceanic versus terrestrial moisture sources. It therefore seemed appropriate to compare our findings with recycling values that were calculated the same way (Brubaker et al., 1993; Eltahir and Bras, 1994). However, we will add a sentence adding that recycling estimates derived from water tagging (Risi et al., 2013; Yoshimura et al., 2004) are higher still.

From l431:

“However, this is a region where changes in precipitation recycling also appear to be important. Based on regional water budget estimates, recycling presently contributes ca 25-35% of the precipitation over the Amazon (Brubaker et al., 1993; Eltahir and Bras, 1994) while these figures increase up to ca 40-60% based on moisture tagging studies (Risi et al., 2013; Yoshimura et al., 2004).”

• l 466: could the greater water-calcite fractionation at colder temperature also contribute to the observed change in speleothem δ 18O? Could you do at least a simple back-of-the-envelope calculation to check this?

Using equations for drip water to calcite (Tremaine et al., 2011) or aragonite (Grossman and Ku, 1986) oxygen isotope fractionation, the ~2°C cooler LGM temperatures for the ISM and EAM would cause $\delta^{18}O_{spel}$ to be ~0.4‰ more depleted and ~3°C cooling for the IAM results in $\delta^{18}O$ values ~0.6‰ more depleted. This will partially counteract the relatively more enriched $\delta^{18}O_{spel}$ values of the LGM, although the effect of cave temperature on $\delta^{18}O_{spel}$ is within the uncertainty of regional signals. We will amend the text to include these back-of-the-envelope calculations.

From l437:

“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). Cooler air temperatures will also result in a depletion of $\delta^{18}O_{spel}$ values during the LGM of ca 0.4 ‰ and 0.6‰ for the ISM/EAM and IAM respectively, as a result of water-calcite/aragonite fractionation (Grossman and Ku, 1986; Tremaine et al., 2011). This has the effect of slightly reducing the regional LGM $\delta^{18}O_{spel}$ signals, although the change is small and within the uncertainty of the regional signals. The 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.”

• Fig 3: I still find it very inconvenient to have a different axis for the model and observations. The figure allows us to see the sign of changes, but not the amplitudes. If the model capture the sign but not the amplitude, this is a very important information. So can you please use the same axis?

We reiterate that the key point of this figure is to compare the relative shifts of $\delta^{18}O$ and climate variables between the MH, LGM and LIG and we therefore fix the axes. However, we agree that comparing the amplitude and shifts of change between model and observation is interesting of itself. Therefore, we will include a figure of $\delta^{18}O_{precip}$ and $\delta^{18}O_{spel}$ (converted to $\delta^{18}O_{drip\ water}$), on the same axis, in the supplementary materials. It is worth noting the caveat that this necessary conversion of $\delta^{18}O_{spel}$ to $\delta^{18}O_{drip\ water}$ requires temperature values. We use modelled temperatures here, which adds a further source of uncertainty to $\delta^{18}O$ values.

We will add the following figure and caption to the supplement:

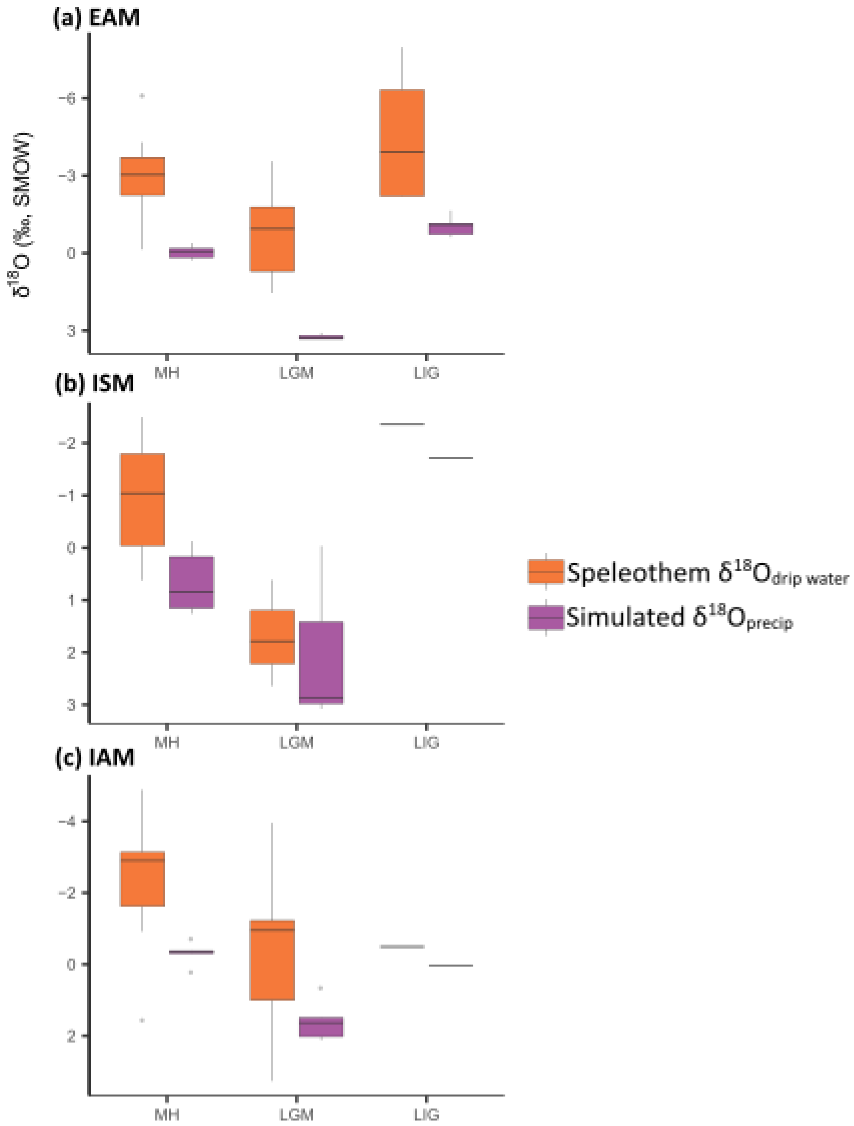


Figure S8: Speleothem $\delta^{18}O$ anomalies, converted to their drip water equivalent, compared to anomalies of $\delta^{18}O_{precip}$ from the ECHAM simulations for the (a) East Asian (EAM), (b) Indian (ISM) and (c) Indonesian-Australian (IAM) monsoons. $\delta^{18}O_{spel}$ (PBD) is converted to $\delta^{18}O_{drip\ water}$ (SMOW) following the methodology in Comas-Bru et al. (2019), using simulated temperature. 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. The difference in amplitude of $\delta^{18}O$ signals is small (<0.5 ‰) between simulated and observed values for the ISM and IAM. In the EAM, simulated $\delta^{18}O$ values have a ~1.4 ‰ higher amplitude than $\delta^{18}O_{drip\ water}$ observations.

We will add the following text, from l313:

"The simulated $\delta^{18}O_{precip}$ anomalies show a similar pattern, with more positive anomalies during the LGM than during the MH or the LIG. The amplitude of this pattern is also similar between $\delta^{18}O_{precip}$ and $\delta^{18}O_{spel}$, when the observations are converted to their drip water equivalent (Fig. S8)."

• Fig 3 caption: ”shown” -> “show”

We will amend the text accordingly.

Anonymous referee #2

The revision has considerably improved the manuscript. I only have some minor comments/suggestions.

(1) Lines 62-63: “a change in the absolute amount of precipitation (Cai et al., 2012; Cheng et al., 2006)” is clearly a misinterpretation of the cited papers.
We apologise for the confusion here. We amended the text but failed to amend the citations appropriately. We will correct this mistake. We also acknowledge that the description of interpretations of the Chinese speleothem record in the introduction was over-simplified. We will amend the text to include the more complex interpretations by several papers:
From l61:
“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, often interpreted predominantly as a change in the absolute amount of precipitation (Cheng et al., 2013; Fleitmann et al., 2003) or a change in the ratio of more negative $\delta^{18}O$ summer precipitation to less negative $\delta^{18}O$ winter precipitation (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 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 (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 $\delta^{18}O_{spel}$ have included rainfall source changes (Tan 2009, 2011, 2014) or local rainfall changes (Tan et al., 2015). 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 changes in precipitation (either local or upstream) or rainfall seasonality required to reproduce $\delta^{18}O_{spel}$ trends would be unreasonably large. 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. Overall, there are several plausible climate mechanisms that could contribute to $\delta^{18}O_{spel}$ on multi-millennial timescales. East Asian monsoon speleothem records are often interpreted as a combination of several of these processes (Cheng et al., 2016; Dykoski et al., 2005) which overall represent monsoon intensity (Cheng et al., 2019).”

(2) In order to further demonstrate the various interpretations of Chinese speleothem δ18O records, it would be better to cite the recent review paper (Cheng et al., 2019. Chinese

stalagmite paleoclimate researches: A review and perspective. Science China: Earth Sciences 62 (10), 1489-1513), as suggested previously.

We will expand our discussion of East Asian monsoon $\delta^{18}O_{spel}$ in the introduction, to elaborate that interpretations often include multiple mechanisms, with $\delta^{18}O_{spel}$ representing an integrated result of processes, representing monsoon intensity, citing the Cheng et al. (2019) paper. We will amend the text as follows (from l76):

"East Asian monsoon speleothem records are often interpreted as a combination of several of these processes (Cheng et al., 2016; Dykoski et al., 2005) which overall represent monsoon intensity (Cheng et al., 2019)."

(3) It would be proper to move the description of the δ18Ospel in the Indian summer monsoon region (in lines 84-88) to line 77 (after the description about the East Asian monsoon), because both Indian summer monsoon and East Asian summer monsoon are a subsystem of the Asian summer monsoon that show strong interplays.

We will reorder this part of text to make the order more logical. From l76:

"There are also multiple interpretations of the causes of $\delta^{18}O_{spel}$ variability in other monsoon regions. 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). In the Indonesian-Australian monsoon region, $\delta^{18}O_{spel}$ 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 $\delta^{18}O_{precip}$ 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., 2005; 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., 2006; Wang et al., 2006)."

(4) The authors' response to the previous comment 9 is oversimply. Regarding the Asian monsoon, the LGM (~21 ±1ka) is under the condition of a Northern Hemisphere insolation minimum, whereas the MH and LIG are at the condition of an insolation maxima. This may at least partially explain the δ18Ospel records. More broadly, the heavy δ18Ospel excursions also occurred, for example, during the MIS 5b and 5d in the interglacial period (MIS 5) with similar amplitudes (values) to that of the LGM (e.g., Cheng et al., 2016. Nature). This is also the case in the South American monsoon regime, where the LGM δ18Ospel value is much lighter than those in MIS 5c (e.g., Cruz et al., 2005. Nature), and apparently these observations contradict in principle, to the current interpretation.

In this study, we focus on MIS 5e, i.e. the peak of the Last Interglacial, for comparison with the mid-Holocene and because this is the time period for which we have simulations. An examination of the variability within the stage 5 interglacial is beyond the scope of the paper. Similarly, we do not examine the entirety of the last Glacial period, but focus of the

LGM, when global ice volumes were at a maximum. We will amend the text to better clarify that we are focusing on the peak of the Last Interglacial (Stage 5e).

From l16-l18 (abstract):
"We focus on differences in $\delta^{18}O$ signals between the mid-Holocene, the peak of the Last Interglacial (Stage 5e) and Last Glacial Maximum, and on $\delta^{18}O$ evolution through the Holocene."

From l103-l105 (introduction):
"...to investigate the plausible mechanisms driving changes in $\delta^{18}O$ in monsoon regions through the Holocene period (last 11,700 years) and between the mid-Holocene, the peak of the Last Interglacial (Stage 5e) and Last Glacial Maximum."

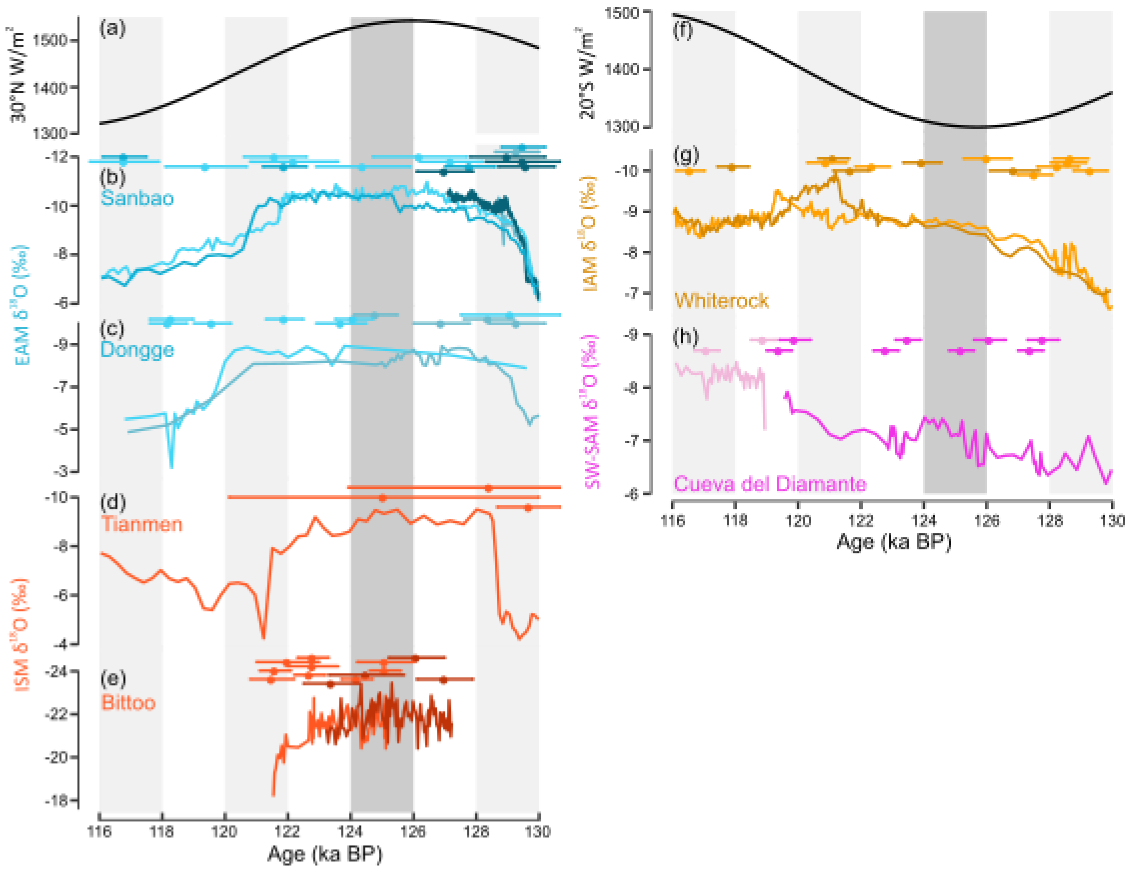
From l121 (methods):
"For the analysis of mid-Holocene (MH), Last Glacial Maximum (LGM) and Stage 5e during the Last Interglacial (LIG) $\delta^{18}O$ signals, the records contain samples within at least one of these time periods, defined as 6,000±500 years for the MH, 21,000±1,000 years BP for the LGM and 125,000±1,000 years BP for the LIG, where BP (before present) is 1950 CE;

From l194 (methods):
"2.4. Glacial-interglacial changes in $\delta^{18}O$
We examined 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, between the MH, LGM and LIG. Values are given as anomalies with respect to the present day for speleothems and the control simulation for model outputs."

From l305 (results):
"3.2 Regional interglacial-glacial differences
To investigate the causes of shifts in $\delta^{18}O$ between the MH, LGM and LIG, we compare simulated and observed regional $\delta^{18}O$ signals during these periods with shifts in climate variables (precipitation and temperature)."

From l477 (conclusions):
"LGM $\delta^{18}O_{spel}$ signals are best explained by a large decrease in precipitation, as a consequence of lower atmospheric moisture content driven by global cooling."

We will also amend the caption of figure 5 to clarify that we are only examining variability within stage 5e, not the entire interglacial period. We will annotate figure 5 to mark the LIG (stage 5e) time slice used in the section 2.4 analysis:



"Figure 5: Comparison of changes in summer insolation and $\delta^{18}O_{spel}$ through the peak of the Last Interglacial (Stage 5e) 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. The LIG (Stage 5e) time slice used in the analysis in section 2.4 is shown by the dark grey bar."

Steven Clemens

The manuscript has improved considerably with the revisions in response to the two previous reviewers. It now reads more easily for those not familiar with the analytical approaches used. The authors utilize SISAL proxy data to define monsoon regions with similar speleothem d18O responses and then interrogate isotope enabled models to explain the patterns in the context of climate variables. The findings lend strong support to the evolving consensus that there is significant variability in the speleothem d18O response of the various monsoon systems and variability in the components of the climate system that drive the regional speleothem d18O signals (source area variability, dynamics along the moisture path, precipitation…).

Most of the comments below are largely along the lines of clarifications. Two can be considered more important, including the discussion of glacial-interglacial variability and the definition of source areas, both of which, in my opinion, can be addressed at the discretion of the authors; I recommend publication with minor revision.

Title:
I recommend dropping '….on orbital timescales' from the title. The three short-duration time slices MH, LIG, and LGM explored here don't address the range of dynamics of orbital-scale variability expressed in paleoclimate records. More importantly, dropping these three words makes the title broader, better reflecting the idea that this approach can be applied to any time-scale of variability.
We agree that this paper does not fully explore orbital timescales. We will therefore change the title of the paper to:
"A data-model approach to interpreting speleothem oxygen isotope records from monsoon regions"

Abstract:
"Differences in speleothem δ18O 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."

Might this be due to the fact that CO2 is more similar in the mid-Holocene and present, implicating an internal radiative forcing versus an external insolation forcing? CO2 as a driver for East Asian monsoon variability is supported by most (all?) non-speleothem-based East Asian summer monsoon proxies examined at orbital time scales. See, for example, the Pleistocene loess proxies.
The difference in CO_2 concentrations is small between these two periods (LIG is 11 ppm higher; Otto-Bliesner et al., 2017). Whilst loess proxies at the margin of the East Asian monsoon show a close relationship between high northern latitude CO_2 concentrations and ice volumes (Lu et al., 2013; Sun et al., 2015), it is likely that the core monsoon region (where most speleothem sites are located) is less sensitive to these changes (Sun et al., 2015). Regardless, loess records suggest the East Asian monsoon is stronger when CO_2 concentrations are higher and ice volumes are lower. On this basis, higher CO_2 concentrations during the LIG would reinforce the enhancement of the monsoon by summer insolation and therefore cannot easily explain the similarity of $\delta^{18}O_{spel}$ signals between these two periods. We will amend the text to discuss CO_2 and ice volume difference.
From l l411:
"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). The LIG (125ka) period was defined by higher summer insolation, higher CO_2 concentrations (Otto-Bliesner et al., 2017) and lower ice volumes (Dutton and Lambeck, 2012) than the MH, suggesting that the LIG ISM and EAM monsoons should be stronger than the MH monsoons. The lack of a clear differentiation in the isotope signals between the LIG and MH suggests that other factors play a role in modulating the monsoon response to these forcings and may reflect the importance of global constraints on the externally-forced expansion of the tropical circulation (Biasutti et al., 2018)."

1. Introduction:

Paragraph on isotope-enabled modeling (lines 93-105).

Jalihal et al., 2019 is a useful reference here as well (https://doi.org/10.5194/cp-15-449-2019).7

The Jalihal et al. (2019) is an interesting study of the complexity of influences on monsoon precipitation changes. However, in this paragraph we are focusing explicitly on results from isotope-enabled modelling - largely to provide the background to our use of isotope-enabled models in our analyses. Summarising the literature on the various influences on monsoon precipitation changes would be a much bigger task and not strictly relevant to our study.

***

2.6 Multiple regression analysis

"We investigate the drivers of regional δ18Oprecip, and by extension δ18Ospel, through the Holocene using multiple linear regression (MLR) of annual precipitation-weighted mean δ18Oprecip 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 δ18O 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 three climate variables chosen are useful (and if recycling had not been presented, reviewers would have requested it). Nevertheless, I wonder if Including some variable that monitors changes in tropical deep convective precipitation (e.g. ISM) versus subtropical precipitation that is frontal in nature (e.g. EAM) could be useful in the MRL as well.

We agree that it would be worthwhile to investigate the role of deep convection. Unfortunately, the only variable we have that provides a measure of convection is convective precipitation. We calculated the relative proportion of convective precipitation (as the ratio of convective: total precipitation). However, both this ratio and convective precipitation itself are highly and significantly ($P < 0.001$) correlated with precipitation (with correlation coefficients of 0.87 and 0.97 respectively) and therefore cannot be included in the model. Thus, including convective precipitation in the model in addition would not provide any additional useful information.

***

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).

This appears to be in conflict with the assumption (line 231-232) that "(i) precipitation-weighted mean annual δ18Oprecip is equivalent to mean annual drip-water δ18O (Yonge et al., 1985)". Please clarify; if the proxy reflects mean annual conditions, why compare to summer – season climate variables?

Speleothem $\delta^{18}O$ records are biased to the months when rainfall is highest, hence we use precipitation-weighted mean $\delta^{18}O_{precip}$ values. For most sites used in this study, there is a clear summer precipitation maximum; most are located within monsoon regions (defined by Wang and Ding, 2008) and hence are dominated by summer precipitation. We therefore use the summer mean for meteorological values.

We will add the following figure and caption to the supplement:

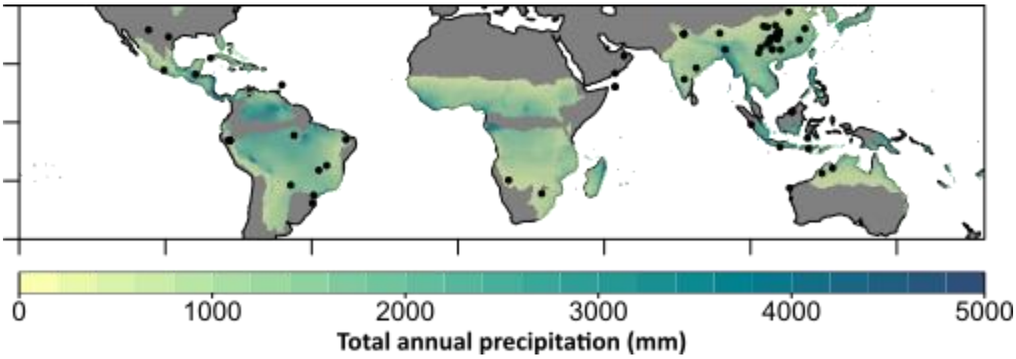


Figure S7: Spatial distribution of speleothem sites used in this study, shown with monsoon regions, defined by the precipitation-based criteria (Wang and Ding, 2008): the annual precipitation range (summer minus winter) exceeds 300 mm and 50 % of the annual mean. Summer is defined as May to September for the northern hemisphere and November to March for the southern hemisphere, vice versa for winter. Precipitation data is from the WFDEI dataset (Weedon et al., 2014).

We will amend the text from l272:
"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), on the basis that these regions are dominated by summer season precipitation (Fig. S7)."

***
3.2 Regional interglacial-glacial differences
"To investigate the causes of glacial-interglacial shifts in δ18O, we compare simulated and observed regional δ18O signals during the LIG, LGM and MH with shifts in climate variables (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 δ18Ospel and simulated δ18Oprecip. T. ….. The most positive δ18Ospel anomalies in all three regions occur at the LGM, with more negative anomalies for the MH and LIG."

A hallmark of the EAM orbital-scale composite speleothem record from the Yangtze River Valley is that it has no 100-kyr glacial-interglacial variance in the spectrum; it's virtually all precession. Discussion of termination occurrences in Cheng et al., is in the context of one every 4 or 5 precession cycles. Hence, this section (and associated figure 3 and discussion), is interpreting EAM precession-scale variance as glacial-interglacial (100-kyr) variance. In contrast, the ISM record from Xiaobialong (XBL) contains "real (100-ky)" glacial-intgerglacial variability and the associated publication (Cai et al., 15) discusses it in the context of global ice volume and sea level change. In this context, the proxy-model comparison result that "The glacial-interglacial changes in δ18Oprecip 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'' may be valid for the ISM but not the EAM. The EAM differences between values at the time of the LGM and the times of the MH and LIG may be better interpreted in the context of precession-band differences in model results (precession maxima vs precession minima) instead of LGM vs MH or LIG. I guess the point is that it's odd to discuss glacial-interglacial differences for records that have no 100-kyr variance.

For this analysis, we focus on the MH and LIG to represent interglacial peaks, whilst the LGM represents a period of maximum ice extent during the Last Glacial Period. Hence, this analysis provides snapshots of glacial-interglacial variability, but not a fuller picture capturing multiple terminations. We will clarify that the LIG here represents Stage 5e. Furthermore, we will further clarify the aim of this analysis in the text.
From l434:
"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. The MH and LIG (Stage 5e) periods represent peaks in the present and last interglacial periods, whilst the LGM represents maximum ice extent during the Last Glacial Period. Hence, comparison of these time periods provides a snap-shot view of glacial-interglacial variability. 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)."

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Figure 1.
The ISM source region seems too small if it includes only the red box encompassing the Arabian Sea; Indian summer monsoon moisture is sourced on the order of 50% from monsoon lows and depressions originating in the Bay of Bengal and tracking NW into India. Please clarify if the Bay of Bengal is or is not included as possible source region for the ISM and if not, explain why.
We defined monsoon regions based on modern moisture tracking studies (described from l270-272) and modelled surface winds. We also kept all source regions broadly a similar size. We originally tested several different source areas for each region, finding that our extracted values were not sensitive to the exact choice of source region limits.
We agree that Bay of Bengal is also an important moisture source for the ISM. However, expanding the ISM source region to include the Bay of Bengal as well as the Arabian Sea has a negligible effect on the results of the multiple linear regression model. We will include the figure and table for the MLR analysis for when this expanded source region is used in the supplement, and amend the text.

From l357:
"The exact choice of source region has a negligible impact on the model, for example expanding the ISM source region to include the Bay of Bengal does not change the outcome of this analysis (Fig S9, Table S1)."

In the supplement:

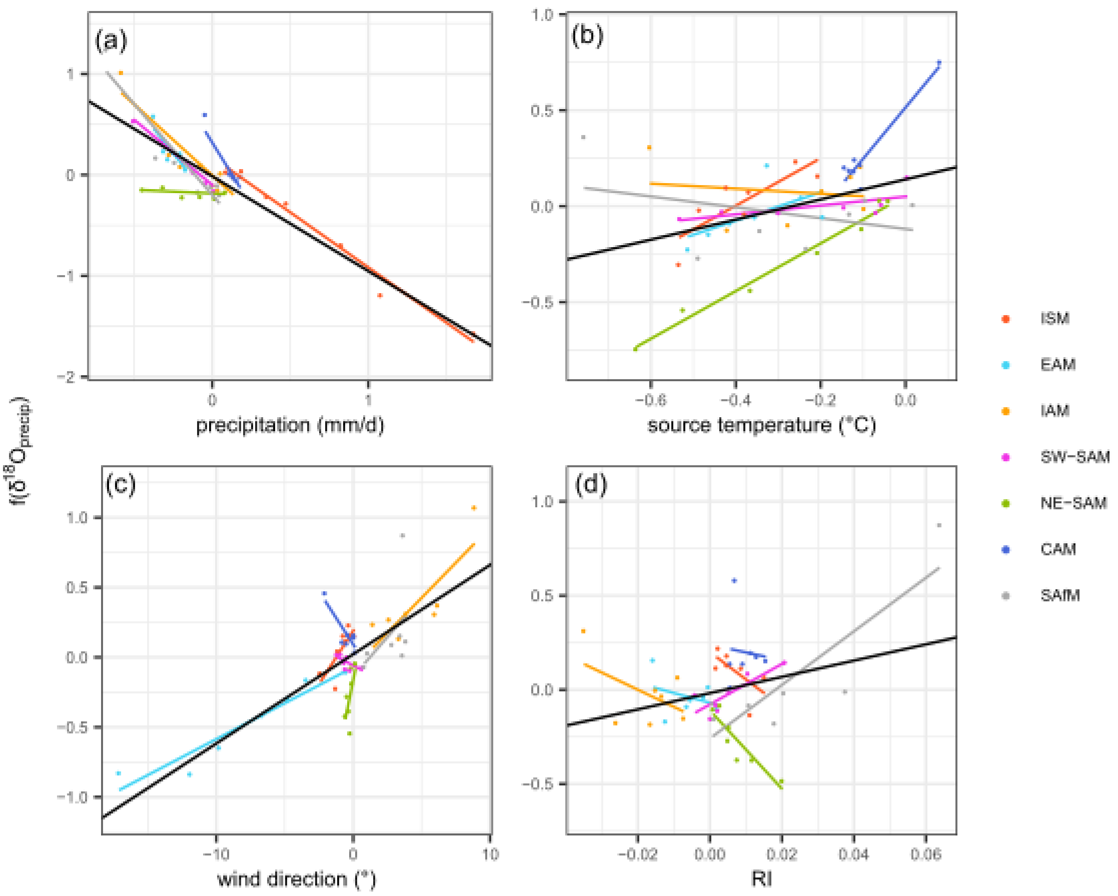

Figure S9: Same as figure 6, except Indian Summer Monsoon (ISM) source region has been expanded to include the Bay of Bengal.

	Regression coefficient	T value
Regional precipitation	**-0.94**	-11.22
Source area temperature	**0.52**	2.95
Wind direction	**0.06**	8.30
Precipitation recycling	4.32	1.95

Table S1: Results of the multiple linear regression analysis when the Indian monsoon source region is expanded to include the Bay of Bengal. Significant relationships (P > 0.01) are shown in bold.

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3.4 Multiple regression analysis of Holocene δ18Oprecip

"The global model for the Holocene (1 to 9ka) δ18Oprecip trends has a pseudo-R2 of 0.80 and shows statistically significant relationships between the anomalies in δ18Oprecip and anomalies in regional precipitation, temperature and surface wind direction (Table 3)."

The term 'global model' is not defined anywhere in the paper and, hence, confusing. Please clarify what model is used and if the results apply to the entire globe (pole to pole) or all monsoon regions combined…

The "global model" refers to the incorporation of all monsoon regions. We understand that this wording is potentially misleading and will therefore amend the text to clarify how the model is constructed.

From l262 (section 2.6):

“We investigate the underlying relationships between regional $\delta^{18}O_{precip}$ (and by extension $\delta^{18}O_{spel}$) and monsoon climate through the Holocene using multiple linear regression (MLR)”
From l286:
“We incorporate mean meteorological variables and $\delta^{18}O_{precip}$ for all Holocene time slices (1ka to 9ka) and all monsoon regions (CAM, ISM, EAM, SW-SAM, NE-SAM, SAfM, IAM) into the MLR model. Thus, the relationships constrained by the overall (global) MLR model represent the combined response across all monsoon regions.”
From l349 (section 3.4):
“The global MLR model includes the Holocene (1 to 9ka) $\delta^{18}O_{precip}$ trends combined across all monsoon regions. This global monsoon MLR model has a pseudo-R^2 of 0.80…”
From l358:
“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 MLR model (representing the combined response across all monsoon regions). Data points from the ISM, SW-SAM, IAM and SAfM are well aligned with the overall relationship with regional precipitation (Fig. 6a), indicating that precipitation is an important control on changes in $\delta^{18}O_{precip}$ in these regions. The NE-SAM, EAM and CAM values deviate somewhat from the overall relationship and, although there are relatively few points, this suggests that changes in precipitation are a less important influence on $\delta^{18}O_{precip}$ changes in these regions. The impact of temperature changes (Fig. 6b) in the ISM, EAM and SW-SAM is broadly consistent with the overall relationship. 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 overall relationship between anomalies in temperature and $\delta^{18}O_{precip}$ 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 overall relationship between changes in $\delta^{18}O_{precip}$ and changes in wind direction (Fig. 6c), indicating that changes in source area or moisture pathway are an important control on changes in $\delta^{18}O_{precip}$ in these regions. However, values for CAM, SW-SAM, NE-SAM and SAfM deviate strongly from the overall relationship.”

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Figure 6

Why are the predictor variables summer mean values whereas the d18Oprecip anomalies are precipitation weighted annual average values? Please clarify why summer mean d18Oprecip is not used.

The speleothem records used in this study are broadly-speaking in the monsoon domain and hence the signals will be dominated by the summer season rainfall. We will clarify this in the figure 6 caption:

“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, representing the summer monsoon, where summer is defined as May to September for northern hemisphere monsoons and November to March for southern hemisphere monsoons."

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Discussion

"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)."

What about greenhouse gasses as a possible explanation? Is the radiative forcing LIG to MH more similar compared to that of insolation forcing?

CO_2 concentrations are similar between the MH and LIG, based on ice core values (and as used as boundary conditions in the PMIP protocol: Otto-Bliesner et al., 2017). The MH and LIG have CO_2 concentrations of 264 and 275 ppm, respectively. Ice volumes may have also been slightly different between the two time periods: whilst MH ice volumes were similar to PI, higher sea-levels during the LIG (>6 m relative to today: Dutton and Lambeck, 2012) suggest lower ice volumes in Greenland and Antarctica at this time. The role of high-latitude ice volumes and CO_2 concentrations on East Asian monsoon variability has been widely discussed in loess record studies. Several studies show a strong correlation between EASM variability (as recorded by loess) and ice volumes/CO_2, with higher ice volumes and lower CO_2 driving a weaker monsoon (and vice versa), either due to a shift in the mean position of the ITCZ (Lu et al., 2013), or due to a strengthening and southward shift of the westerlies (Sun et al., 2015). On this basis, the slightly higher CO_2 and lower ice volumes of the LIG relative to the MH would drive a stronger monsoon, reinforcing the summer insolation driven strengthening of the monsoon during the LIG. Thus, neither CO_2 or ice volume explain the similarity of the MH and LIG $\delta^{18}O_{spel}$ signals. It is, however, also worth noting that the monsoon front (where the EASM loess records are located) are likely more sensitive to high latitude ice-volume/CO_2 conditions, than the core of the EASM (Sun et al., 2015), where all the LIG speleothem records are located.
We will amend the text to include a brief discussion of CO_2/ice volume conditions of the MH vs. LIG signals.
From l411:
"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). The LIG (125ka) period was defined by higher summer insolation, higher CO_2 concentrations (Otto-Bliesner et al., 2017) and lower ice volumes (Dutton and Lambeck, 2012) than the MH, suggesting that the LIG ISM and EAM monsoons should be stronger than the MH monsoons. The lack of a clear differentiation in the isotope signals between the LIG and MH suggests that other factors play a role in modulating the monsoon response to these forcings and may reflect the importance of global constraints on the externally-forced expansion of the tropical circulation (Biasutti et al., 2018)."

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References:

Dutton, A. and Lambeck, K.: Ice volume and sea level during the last interglacial, Science, 337(6091), 216-219, 2012.

Fleitmann, D., Burns, S.J., Mudelsee, M., Neff, U., Kramers, J., Mangini, A., and Matter, A.: Holocene Forcing of the Indian Monsoon Recorded in a Stalagmite from Southern Oman, Science, 300(5626), 1737-1739, https://doi.org/ 10.1126/science.1083130, 2003.

Lu, H., Yi, S., Liu, Z., Mason, J.A., Jiang, D., Cheng, J., Stevens, T., Xu, Z., Zhang, E., Jin, L. and Zhang, Z.: Variation of East Asian monsoon precipitation during the past 21 ky and potential CO2 forcing, Geology, 41(9), 1023-1026, 2013.

Otto-Bliesner, B.L., Braconnot, P., Harrison, S.P., Lunt, D.J., Abe-Ouchi, A., Albani, S., Bartlein, P.J., Capron, E., Carlson, A.E., Dutton, A. and Fischer, H.: The PMIP4 contribution to CMIP6–Part 2: Two interglacials, scientific objective and experimental design for Holocene and Last Interglacial simulations, Geosci. Model Dev., 10(11), 3979-4003, 2017.

Sun, Y., Kutzbach, J., An, Z., Clemens, S., Liu, Z., Liu, W., Liu, X., Shi, Z., Zheng, W., Liang, L. and Yan, Y.: Astronomical and glacial forcing of East Asian summer monsoon variability, Quat. Sci. Rev., 115, 132-142, 2015.

Weedon, G. P., Balsamo, G., Bellouin, N., Gomes, S., Best, M. J., and Viterbo, P.: The WFDEI meteorological forcing data set: WATCH Forcing Data methodology applied to ERA-Interim reanalysis data, Water Resources Research, 50, 7505–7514, https://doi.org/10.1002/2014WR015638, 2014.