Precipitation over South America during the Last Glacial Maximum: An analysis of the "amount effect" with a water isotope-enabled general circulation model

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[1] Low latitude paleoclimate records from speleothem δ^{18}O measurements are often considered to reflect variations in precipitation amounts. Here we test this interpretation with a water isotope enabled atmospheric general circulation model, comparing modern and Last Glacial Maximum (LGM) controls on the $\delta^{18}O_p$ in Brazil. We argue that the $\delta^{18}O_p$ is determined by the contributions from local evaporation (high δ^{18}O) versus water vapor convergence to the region (low δ^{18}O). Our analysis indicates that the amount effect, commonly used to infer precipitation amount from $\delta^{18}O_p$ in low latitude regions, works only where the isotopic composition of incoming vapor stays relatively constant such as in coastal regions of the subtropics. If the isotopic composition of incoming vapor has changed as a result of the variations in the upstream rainout, $\delta^{18}O_p$ cannot be used to estimate local precipitation. Our analysis supports the increase of precipitation over northeastern Brazil region during the LGM. **Citation:** Lee, J.-E., K. Johnson, and I. Fung (2009), Precipitation over South America during the Last Glacial Maximum: An analysis of the ''amount effect'' with a water isotope-enabled general circulation model, *Geophys. Res. Lett.*, *36*, L19701, doi:10.1029/2009GL039265.

1. Introduction

[2] One of the most widely used paleoproxies for reconstructing past hydrological variations in low latitude and monsoon regions is the oxygen isotopic composition of speleothem calcite [e.g., *Wang et al.*, 2001; *Cruz et al.*, 2005; *Partin et al.*, 2007]. Speleothem δ^{18}O has been shown to vary with $\delta^{18}O_p$, the δ^{18}O in precipitation [*Cobb et al.*, 2007], and has been used to infer changes in past precipitation patterns or amount [*Wang et al.*, 2001; *Fleitmann et al.*, 2003; *Johnson et al.*, 2006; *Cruz et al.*, 2005; *Partin et al.*, 2007; *Wang et al.*, 2008]. The basis of the relationship between $\delta^{18}O_p$ and precipitation rate is the so-called amount effect, wherein more depleted stable isotope values are observed where precipitation amount is large in tropical/subtropical coastal and island regions from modern data [*Dansgaard*, 1964; *Rozanski et al.*, 1993; *Vuille and Werner*, 2005; *Risi et al.*, 2008; *Lee and Fung*, 2008].

[3] Despite the wide use of speleothem δ^{18}O as a paleoproxy for precipitation amount over low latitude regions, a growing number of studies have shown that factors other than local precipitation amount may affect $\delta^{18}O_p$ [e.g., *Schmidt et al.*, 2007]. *Maher* [2008], for example, finds that the interpretation of speleothem $\delta^{18}O_p$ are inconsistent with rainfall records derived from loess/paleosol sequences and magnetic properties. *Hu et al.* [2008] hypothesize that differences between speleothem δ^{18}O from two sites (Heshang Cave and Dongge Cave) along an atmospheric transport pathway are directly related to the amount of rainout that occurred between the two sites, rather than the amount of precipitation at an individual site. From their modeling study, *Vuille et al.* [2003] show that the Asian speleothem-$\delta^{18}O_p$ is not directly related to the local precipitation amount but that it is linked to alterations in landward water vapor transport. *Yoshimura et al.* [2003] argue that the δ^{18}O variability is controlled by horizontal advection and not by local precipitation amount.

[4] In this study, we use an isotope-enabled atmospheric general circulation model (AGCM) to examine whether $\delta^{18}O_p$ differences in paleoproxies can be interpreted as precipitation amount differences over tropical/subtropical regions. We choose northeastern and southeastern Brazil during the LGM (21,000 years ago) and present-day from AGCM simulations as examples to show when and where $\delta^{18}O_p$ differences can or cannot be interpreted as precipitation amount differences. In Section 2, we present our model and the model configuration for the present-day and LGM conditions. Section 3 describes the changes in the LGM climatic conditions over tropical regions and tests the hypothesis explaining speleothem δ^{18}O, focusing on southeastern and northeastern Brazil. Section 4 discusses our findings.

2. Model Description

[5] We incorporated HDO and $H_2^{18}O$ into the NCAR CAM2; the global distribution of water isotopes in precipitation is reasonably simulated, and the limitations of the atmospheric model and isotopic scheme are described by *Lee et al.* [2007]. We ran the isotope-enabled NCAR CAM2 with fixed sea surface temperature (SST) and sea ice distributions for the present-day and the LGM. Present-day SST is the climatological monthly mean derived from observations from 1949 to 2001. In the LGM run, we used monthly SST and sea ice distribution simulated by the fully coupled atmosphere-land-ocean-ice Community Climate

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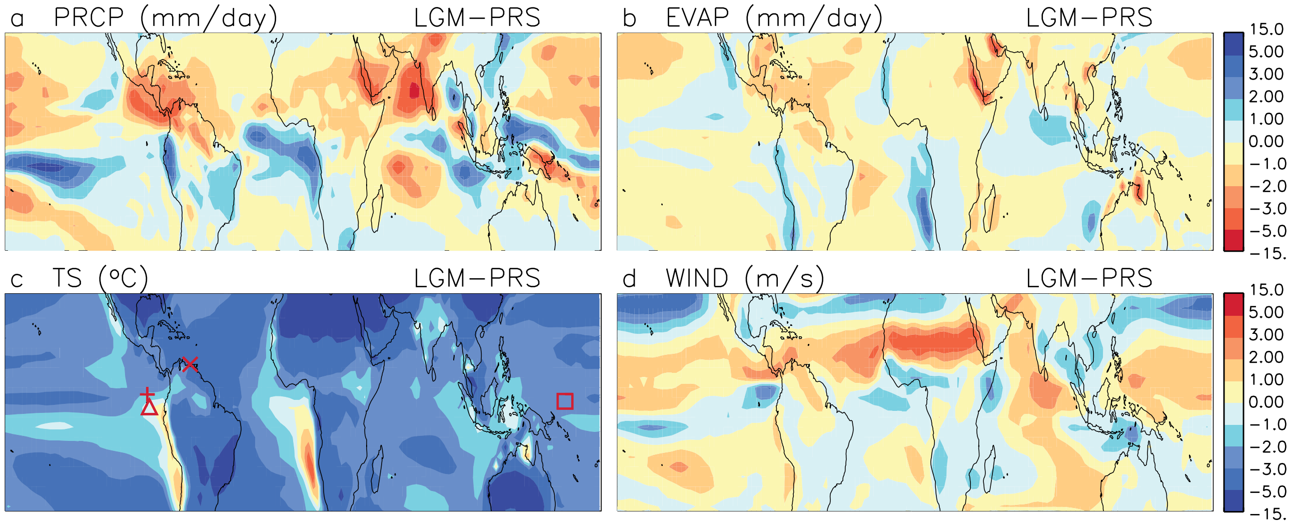

Figure 1. The LGM-present day differences in (a) precipitation, (b) evaporation, (c) surface temperature, and (d) wind speed in the lowest atmospheric layer of the GCM. Surface temperature (Ts) is SST for the ocean and ground temperature for the land in degrees Celsius. Locations of LGM SST estimated from paleoproxy data are marked in Figure 1c. Square and cross are for the northwest and northeast sites of *Lea et al.* [2000], triangle is for the site near Galapagos Islands [*Koutavas et al.*, 2002], and x is for the Cariaco Basin [*Peterson et al.*, 2000].

System Model [*Otto-Bliesner et al.*, 2006] with atmospheric carbon dioxide (CO_2), methane (CH_4), and nitrous oxide (N_2O) at 185 ppm, 350 ppb, and 200 ppb respectively, and with the continental ice sheet extent and topography prescribed. Our LGM and present-day runs have also been previously used to interpret ice core data [*Lee et al.*, 2008].

[6] Surface ocean δ^{18}O values for the present-day and LGM are prescribed as 0.5 [*Hoffmann et al.*, 1998] and 1.7‰ [*Schrag et al.*, 1996], respectively. The LGM simulation is reasonable and has been validated with the available proxy data [*Otto-Bliesner et al.*, 2006]. The LGM CCSM simulation has a global mean sea surface cooling of 6.3°C compared to the present-day. The isotope-CAM LGM simulation is initialized using the atmospheric state from the equilibrium simulation of the CCSM LGM run, and is integrated forward for 20 years using the CCSM SST's and glacial and sea ice extents as boundary conditions. The present-day simulation was integrated for 15 years. In both cases, averages of the last 10-year integrations were used for our analysis.

3. Results and Discussion

[7] As a result of the greater cooling simulated over the NH (6.54°C in NH versus 5.28°C in SH), the boreal summer ITCZ is located more equatorward during the LGM [*Chiang and Bitz*, 2005]. Figure 1 shows the LGM-PRS difference in mean annual precipitation, evaporation, surface temperature, and zonal wind over the tropical regions (30°S-30°N). In the LGM simulation, there is a substantial decrease in precipitation in NH, particularly over the Indian Ocean, Indian monsoon regions, and Caribbean regions. In the Southern Hemisphere (SH), a precipitation increase is shown over the eastern Pacific and Atlantic oceans. When the ITCZ shifts to the south during the LGM, the modeled strength of the trade winds decreases more near the equator of the Pacific and Atlantic Oceans. As a result, upwelling is smaller over the southeastern Pacific and Atlantic, and the temperature difference between the LGM and present-day is smaller (and even positive at some points) over the coastal regions of the southeastern Pacific and Atlantic oceans compared to the northeastern or western Pacific [*Fedorov and Philander*, 2000]. Due to the decrease in wind and/or temperature, the evaporation decrease is large over tropical and subtropical regions, particularly over the Atlantic.

[8] Paleoproxy data also indicate a larger LGM temperature decrease over the western and northeastern Pacific than over the southeastern tropical Pacific [*Koutavas et al.*, 2002]. Over the western tropical Pacific (2.26°N, 90.95°W, square in Figure 1) and northeastern tropical Pacific (0.32°N, 159.36°E, cross in Figure 1), *Lea et al.* [2000] estimates 2.6°C and 2.8°C of SST decrease respectively. The SST decrease near the Galapagos Islands (1.22°S, 89.68°W, triangle in Figure 1), on the other hand, is only 1.2°C [*Koutavas et al.*, 2002]. LGM SST's simulated by the CCSM agree with the paleoproxies for the western and northeastern tropical Pacific (2.5°C and 2.6°C respectively), and the southeastern tropical Pacific SST pattern captures the same decreasing trend from the west or north to the east, even if the magnitude of the modeled temperature decrease is too small (0.1°C).

[9] LGM paleoproxies from tropical and subtropical regions of South America also suggest equatorward shift of the precipitation band during boreal summer. The Cariaco basin (10.71°N, 65.27°W; x in Figure 1), currently located within the reach of the ITCZ, experienced relatively dry conditions at the LGM [*Peterson et al.*, 2000], while in the NE Brazil site (10.17°S, 40.83°W; triangle in Figure 1c), currently beyond the reach of the ITCZ, wetter conditions are inferred from the formation of travertine deposits, which do not form under modern dry conditions at this site [*Wang*

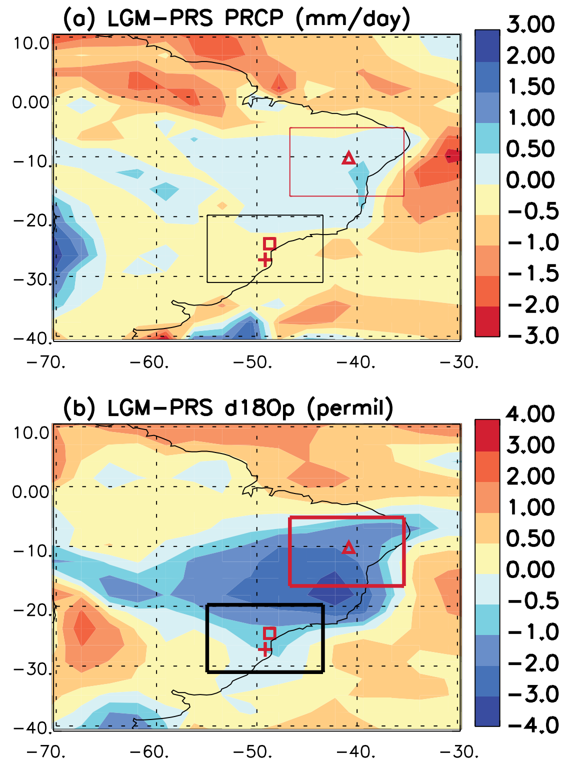

Figure 2. The LGM-present differences in (a) precipitation and (b) $\delta^{18}O_p$. The triangle and square denote the location of the NE Brazil speleothem record [*Wang et al.*, 2004] and the SE Brazil record [*Cruz et al.*, 2006] respectively. The boxes show the NE and SE Brazil regions used to calculate the model results shown in Table 1.

et al., 2004]. Figure 2a shows the modeled LGM-present day differences of precipitation over South America. Our model agrees well with this paleodata and captures the southward shift in the ITCZ location, the increase in precipitation over northeastern Brazil and the decrease over the Cariaco basin region.

[10] $\delta^{18}O$ differences between the LGM and present-day precipitation are plotted in Figure 2b. Speleothem $\delta^{18}O$ data from southeastern (SE) Brazil (24.53°S, 48.72°W, square in Figure 2b; and 27.22°S, 49.15°W, cross in Figure 2b) show a $\delta^{18}O_p$ decrease of ~0.5 to 1‰ [*Cruz et al.*, 2005, 2006; *Wang et al.*, 2007]. Our modeled $\delta^{18}O_p$ change from present-day to LGM is ~−0.86‰ for SE Brazil (31°S-20S, 55°W-44°W; black box in Figure 2b; Table 1), similar to the speleothem data. The present-day slope of the relationship between mean monthly precipitation amount and $\delta^{18}O_p$ at some tropical/subtropical island stations ranges from −0.48‰/(mm/day) and −0.66‰/(mm/day) [*Dansgaard*, 1964]. If we use this relationship, a change of −0.86‰ can be interpreted as a precipitation increase of 1.3 to 1.8 mm/day. The modeled precipitation change is only ~0.3 mm/day (Table 1). The $\delta^{18}O_p$ change in this region is, therefore, not fully explained by the changes in precipitation amount but rather by changes in the $\delta^{18}O$ of transported vapor. Most vapor in this region comes from the northwest land regions and underwent increased distillation in the LGM simulation, resulting in lower $\delta^{18}O$ values for transported vapor at this time (Figure 3). The incoming vapor from the Amazon basin has a $\delta^{18}O$ value of −25.2‰ during the LGM compared with the present-day value of −22.1‰. The decrease in Amazon $\delta^{18}O_v$ is due to the increased local and upstream rainfall and the resultant decrease of $\delta^{18}O_v$. Our result is consistent with the interpretation of *Cruz et al.* [2006] who assume that decreased speleothem $\delta^{18}O$ values are reflecting the contribution of more depleted Amazon moisture to SE Brazil during the LGM, not necessarily increased rainfall amount.

[11] Mean annual precipitation over NE Brazil (17°S-6°S, 46°S-35°S red box in Figure 2) increases about 50% during the LGM (2.4 mm/day versus 3.3 mm/day), and the isotopic composition of the precipitation within this region decreases by ~1.4‰. LGM local evaporation is similar to present-day, and so most of the increase in precipitation results from the increase in vapor convergence. The NE Brazil region receives onshore winds, and so most of the vapor transported into this region comes from the Atlantic Ocean for both the LGM and present-day (Figure 3). Water vapor over the Atlantic has similar isotopic composition for the LGM and present-day, because of the high evaporative recharge, and there is little distillation effect from the Atlantic Ocean to NE Brazil. The vapor transported to NE Brazil has a lower $\delta^{18}O_v$ (−13.0‰ for the present-day and −12.5‰ for the LGM) compared to that from local evaporation (−4.96‰ for the present-day and −4.92‰ for the LGM). The greater relative contribution from transported vapor than local evaporation in NE Brazil during the LGM thus leads to more precipitation and lower $\delta^{18}O_p$. Here, the present-day spatial relationship also predicts larger precipitation changes (2.1 to 2.9 mm/day compared with simulated difference of 0.95 mm/day) *Lee et al.* [2007] show that the $\delta^{18}O_p$ scales better with P/E than with P over oceanic regions. In the regions of our interest, evaporation is smaller (~3 mm/day) than the ocean values (4~5 mm/day), and this is why estimated precipitation is lower than the values expected from the modern relationship.

4. Conclusion

[12] The $\delta^{18}O_p$ over low latitude regions has been often interpreted as an indicator for the precipitation amount [*Wang et al.*, 2001; *Fleitmann et al.*, 2003]. We argue that the $\delta^{18}O_p$ is determined by the contributions from local

Table 1. Precipitation and Evaporation in Northeastern Brazil and Southeastern Brazil, Together With the Isotopic Composition of P, E, and Vapor Fluxes Into and out of the Region[a]

	NE Brazil			SE Brazil		
	PRS	LGM	LGM-PRS	PRS	LGM	LGM-PRS
P (mm/day)	2.35	3.30	0.95	3.18	3.51	0.33
E (mm/day	2.52	2.39	−0.13	2.85	2.68	−0.17
$\delta^{18}O_p$ (‰)	−3.15	−4.57	−1.42	−4.25	−5.11	−0.86
$\delta^{18}O_e$ (‰)	−4.96	−4.92	0.04	−5.17	−5.43	−0.26
$\delta^{18}O_{Fi}$ (‰)	−13.0	−12.5	0.05	−20.9	−23.2	−2.3
$\delta^{18}O_{Fo}$ (‰)	−13.4	−13.4	0	−22.1	−25.2	−3.1

[a]P, precipitation; E, evaporation. Northeastern Brazil, 17°S-6°S, 46°S-35°S, red box in Figure 2; southeastern Brazil, 31°S-20S, 55°W-44°W, black box in Figure 2. Fi, vapor fluxes into the region; Fo, vapor fluxes out of the region. The regions are denoted by black and red rectangles in Figure 2. Mean isotopic composition of each component i is represented as $\delta^{18}O_i$. Isotopic composition of incoming and out going vapor was computed using vertically integrated moisture transport of $H_2^{16}O$ and $H_2^{18}O$.

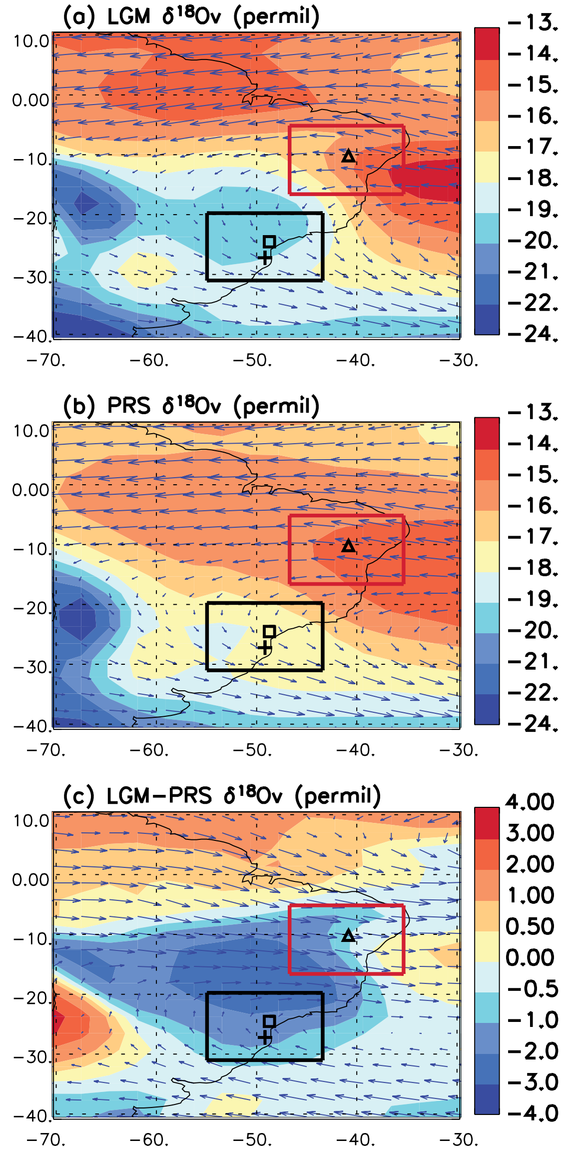

Figure 3. Isotopic composition of the annual mean column vapor for (a) LGM, (b) PRS and (c) the difference. Arrows are annual mean total column water vapor transport.

evaporation (high $\delta^{18}O$ in low latitude regions) versus transported vapor to the region (low $\delta^{18}O$). On an annual scale, $\delta^{18}O_p$ is determined by the contribution from local evaporation and vapor convergence since the amount of vapor in the atmosphere is very small compared with precipitation. For a shorter time scales, however, the interaction between water vapor and raindrops plays a significant role in determining how precipitation integrates the vertical distribution of water vapor [*Lee and Fung*, 2008; *Bony et al.*, 2008]. If the isotopic composition of both local evaporation and transported vapor remains the same, the changes in $\delta^{18}O_p$ should reflect the differences in the contribution. Since spatial variation of evaporation is smaller than precipitation variation [*Lee et al.*, 2007], lower $\delta^{18}O_p$ implies more transported vapor and hence more precipitation. Interpreting $\delta^{18}O_p$ change as a change in precipitation intensity requires minimal distillation of the incoming vapor. In practice, the LGM-present day $\delta^{18}O_p$ difference in a given region could arise from a varying contribution from transported vapor as well as variable $\delta^{18}O$ values of that transported vapor itself (due to upstream distillation effects), rather than changes in the amount of local precipitation. If the incoming vapor has a different isotopic composition between the present-day and the LGM, or if the atmospheric pathways have changed significantly, $\delta^{18}O_p$ cannot be used to estimate precipitation amount.

[13] Over SE Brazil where most vapor comes from the northwestern land regions, interpreting the modeled LGM changes in $\delta^{18}O_p$ by the amount effect would have yielded a precipitation rate difference of 1.3 $\sim$ 1.8 mm/day, contrary to the model simulation difference of 0.3 mm/day. Lower $\delta^{18}O_p$ in SE Brazil does not result from greater contribution of transported vapor, but lower $\delta^{18}O$ of transported vapor. NE Brazil region $\delta^{18}O_p$, however, can be better explained as the changes in precipitation amount because most of the vapor transported into this region comes directly from the Atlantic Ocean where large evaporation from the ocean and the short trajectory inhibits distillation. Our analysis supports the increase of precipitation over NE Brazil region.

[14] Water isotopes are a useful proxy for interpreting the hydrological cycle in low latitude regions, but caution is needed in interpreting these records, since $\delta^{18}O$ does not just depend on the precipitation intensity. Speleothem $\delta^{18}O$ records have great potential to provide information about past moisture transport and precipitation through investigating spatial patterns in past $\delta^{18}O_p$. In combination with isotope-enabled GCMs, such data can constrain interpretations of the paleo-hydrologic cycle.

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References

Bony, S., C. Risi, and F. Vimieux (2008), Influence of convective processes on the isotopic composition ($\delta^{18}O$ and HDO) of precipitation and water vapor in the tropics: 1. Radiative-convective equilibrium and Tropical Ocean-Global Atmosphere-Coupled Ocean-Atmosphere Response Experiments (TOGA-COARE) simulations, *J. Geophys. Res.*, *113*, D19305, doi:10.1029/2008JD009942.

Chiang, J. C. H., and C. M. Bitz (2005), Influence of high latitude ice cover on the marine Intertropical Convergence Zone, *Clim. Dyn.*, *25*, 477–496, doi:10.1007/s00382-005-0040-5.

Cobb, K. M., J. F. Adkins, J. W. Partin, and B. Clark (2007), Regional-scale climate influences on temporal variations of rainwater and cave dripwater oxygen isotopes in northern Borneo, *Earth Planet. Sci. Lett.*, *263*, 207–220.

Cruz, F. W., S. J. Burns, I. Karmann, W. D. Sharp, M. Vuille, A. O. Cardoso, J. A. Ferrari, P. L. S. Dias, and O. Viana (2005), Insolation-driven changes in atmospheric circulation over the past 116,000 years in subtropical Brazil, *Nature*, *434*, 63–66, doi:10.1038/nature03365.

Cruz, F. W., S. J. Burns, I. Karmann, W. D. Sharp, and M. Vuille (2006), Reconstruction of regional atmospheric circulation features during the late Pleistocene in subtropical Brazil from oxygen isotope composition of speleothems, *Earth Planet. Sci. Lett.*, *248*, 495–507.

Dansgaard, W. (1964), Stable isotopes in precipitation, *Tellus*, *16*, 436–468.

Fedorov, A., and S. G. Philander (2000), Is El Niño changing?, *Science*, *288*, 1997–2002, doi:10.1126/science.288.5473.1997.

Fleitmann, D., S. J. Burns, M. Mudelsee, U. Neff, J. Kramers, A. Mangini, and A. Matter (2003), Holocene forcing of the Indian monsoon recorded in a stalagmite from southern Oman, *Science*, *300*, 1737–1739, doi:10.1126/science.1083130.

Hoffmann, G., M. Werner, and M. Heimann (1998), Water isotope module of the ECHAM atmospheric general circulation model: A study on time-scales from days to several years, *J. Geophys. Res.*, *103*, 16,871–16,896, doi:10.1029/98JD00423.

Hu, C., G. M. Henderson, J. Huang, S. Xie, Y. Sun, and K. R. Johnson (2008), Quantification of Holocene Asian monsoon rainfall from spatially separated cave records, *Earth Planet. Sci. Lett.*, *266*, 221–232, doi:10.1016/j.epsl.2007.10.015.

Johnson, K. R., B. L. Ingram, W. D. Sharp, and P. Z. Zhang (2006), East Asian summer monsoon variability during Marine Isotope Stage 5 based on speleothem $\delta^{18}O$ records from Wanxiang Cave, central China, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *236*, 5–19, doi:10.1016/j.palaeo.2005.11.041.

Koutavas, A., J. Lynch-Stieglitz, T. M. Marchitto, and J. P. Sachs (2002), El Niño-like pattern in ice age tropical Pacific sea surface temperature, *Science*, *297*, 226–230, doi:10.1126/science.1072376.

Lea, D., D. Pak, and H. Spero (2000), Climate impact of late quaternary equatorial Pacific sea surface temperature variations, *Science*, *289*, 1719–1724, doi:10.1126/science.289.5485.1719.

Lee, J. E., and I. Fung (2008), "Amount effect" of water isotopes and quantitative analysis of post-condensation processes, *Hydrol. Processes*, *22*, 1–8, doi:10.1002/hyp.6637.

Lee, J. E., I. Fung, D. J. DePaolo, and C. C. Henning (2007), Analysis of the global distribution of water isotopes using the NCAR atmospheric general circulation model, *J. Geophys. Res.*, *112*, D16306, doi:10.1029/2006JD007657.

Lee, J. E., I. Fung, D. J. DePaolo, and B. Otto-Bliesner (2008), Water isotopes during the Last Glacial Maximum: New GCM calculations, *J. Geophys. Res.*, *113*, D19109, doi:10.1029/2008JD009859.

Maher, B. A. (2008), Holocene variability of the East Asian summer monsoon from Chinese cave records: A reassessment, *Holocene*, *18*, 861–866, doi:10.1177/0959683608095569.

Otto-Bliesner, B., E. Brady, G. Clauzet, R. Tomas, S. Levis, and Z. Kothavala (2006), Last Glacial Maximum and Holocene climate in CCSM3, *J. Clim.*, *19*, 2526–2544, doi:10.1175/JCLI3748.1.

Partin, J. W., K. M. Cobb, J. F. Adkins, B. Clark, and D. P. Fernandez (2007), Millennial-scale trends in west Pacific warm pool hydrology since the Last Glacial Maximum, *Nature*, *449*, 452–455, doi:10.1038/nature06164.

Peterson, L. C., G. H. Haug, K. A. Hughen, and U. Rohl (2000), Rapid changes in the hydrologic cycle of the tropical Atlantic during the Last Glacial, *Science*, *290*, 1947–1951, doi:10.1126/science.290.5498.1947.

Risi, C., S. Bony, and F. Vimeux (2008), Influence of convective processes on the isotopic composition ($\delta^{18}O$ and δD) of precipitation and water vapor in the tropics: 2. Physical interpretation of the amount effect, *J. Geophys. Res.*, *113*, D19306, doi:10.1029/2008JD009943.

Rozanski, K., L. Araguas-Araguas, and R. Giofiantini (1993), Isotopic patterns in modern global precipitation, in *Climate Change in Continental Isotopic Records*, *Geophys. Monogr. Ser.*, vol. 78, edited by P. K. Swart et al., pp. 1–36, AGU, Washington, D. C.

Schmidt, G. A., A. N. LeGrande, and G. Hoffmann (2007), Water isotope expressions of intrinsic and forced variability in a coupled ocean-atmosphere model, *J. Geophys. Res.*, *112*, D10103, doi:10.1029/2006JD007781.

Schrag, D. P., G. Hampt, and D. W. Murray (1996), Pore fluid constraints on the temperature and oxygen isotopic composition of the glacial ocean, *Science*, *272*, 1930–1932, doi:10.1126/science.272.5270.1930.

Vuille, M., and M. Werner (2005), Stable isotopes in precipitation recording South American summer monsoon and ENSO variability: Observations and model results, *Clim. Dyn.*, *25*, 401–413, doi:10.1007/s00382-005-0049-9.

Vuille, M., R. S. Bradley, M. Werner, R. Healy, and F. Keimig (2003), Modeling $\delta^{18}O$ in precipitation over the tropical Americas: 1. Interannual variability and climatic controls, *J. Geophys. Res.*, *108*(D6), 4174, doi:10.1029/2001JD002038.

Wang, X., A. S. Auler, R. L. Edwards, H. Cheng, P. S. Cristalli, P. L. Smart, D. A. Richards, and C. C. Shen (2004), Wet periods in northeastern Brazil over the past 210 kyr linked to distant climate anomalies, *Nature*, *432*, 740–743, doi:10.1038/nature03067.

Wang, X., A. S. Auler, R. L. Edwards, H. Cheng, E. Ito, Y. J. Wang, X. G. Kong, and M. Solheid (2007), Millennial-scale precipitation changes in southern Brazil over the past 90,000 years, *Geophys. Res. Lett.*, *34*, L23701, doi:10.1029/2007GL031149.

Wang, Y. J., H. Cheng, R. L. Edwards, Z. S. An, J. Y. Wu, C. C. Shen, and J. A. Dorale (2001), A High-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China, *Science*, *294*, 2345–2348, doi:10.1126/science.1064618.

Wang, Y. J., H. Cheng, R. L. Edwards, X. G. Kong, X. H. Shao, S. T. Chen, J. Y. Wu, X. Y. Jiang, X. F. Wang, and Z. S. An (2008), Millennial- and orbital-scale changes in the East Asian monsoon over the past 224,000 years, *Nature*, *451*, 1090–1093, doi:10.1038/nature06692.

Yoshimura, K., T. Oki, N. Ohte, and S. Kanae (2003), A quantitative analysis of short-term ^{18}O variability with a Rayleigh-type isotope circulation model, *J. Geophys. Res.*, *108*(D20), 4647, doi:10.1029/2003JD003477.

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