

The zonal patterns in late Quaternary tropical South American precipitation

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Key Points:

- The late Quaternary South American Precipitation Dipole drives opposing east-west precipitation anomalies in tropical South America.
- Dipole transitions can drive changes in rainfall greater than 1000 mm/yr.
- Spatial migration of the precipitation centroid can explain dipole transitions and reconcile proxy-model conflicts.

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Abstract

Speleothem oxygen isotope records ($\delta^{18}O$) of tropical South American rainfall in the late Quaternary show a zonal “South American Precipitation Dipole” (SAPD). The dipole is characterized by opposing east-west precipitation anomalies compared to the present—wetter in the east and drier in the west at the mid-Holocene (~ 7 ka), and drier in the east and wetter in the west at the Last Glacial Maximum (LGM; ~ 21 ka). However, the SAPD remains enigmatic because it is expressed differently in western versus eastern $\delta^{18}O$ records and isotope-enabled climate model simulations usually misrepresent the magnitude and/or spatial pattern of $\delta^{18}O$ change. Here, we address the SAPD enigma in two parts. First, we re-interpret the $\delta^{18}O$ data to account for upwind rainout effects that are known to be pervasive in tropical South America, but are not always considered in Quaternary paleoclimate studies. Our revised interpretation reconciles the $\delta^{18}O$ data with cave infiltration and other proxy records, and indicates that the centroid of tropical South American rainfall has migrated zonally over time. Second, using an energy balance model of tropical atmospheric circulation, we hypothesize that zonal migration of the precipitation centroid can be explained by regional energy budget shifts, such as changing Saharan albedo associated with the African Humid Period, that have not been modeled in previous SAPD studies. This hypothesis of a migrating precipitation centroid presents a new framework for interpreting $\delta^{18}O$ records from tropical South America and may help explain the zonal rainfall anomalies that predate the late Quaternary.

Plain Language Summary

Paleoclimate data suggest that, in the last ~ 25 thousand years, tropical South American precipitation has changed substantially, but in opposite directions between the east and west. This opposing east-west pattern in past rainfall is known as the “South American Precipitation Dipole”, and its end-member states approximately coincide with the Last Glacial Maximum (~ 21 thousand years ago) and mid-Holocene (~ 7 thousand years ago), respectively. However, the cause of the dipole is debated because different models produce different results, and the interpretations of data are in conflict. Central in this conflict are oxygen isotope tracers of past precipitation which show different trends over space. We present a new interpretation of these data, backed by model results, which suggests that the dipole is driven by the centroid, or focus, of tropical South American precipitation migrating from west-to-east (and back) across tropical South America. We test this precipitation centroid migration hypothesis with an energy balance climate model which reproduces the expected east-west differences for the Last Glacial Maximum and mid-Holocene. The precipitation centroid migration hypothesis is a possible solution to the precipitation dipole enigma, but it remains to be tested in more sophisticated climate models.

1 Introduction

Tropical South America spans about one-tenth of the Earth’s circumference from east to west (zonally). There is mounting evidence that rainfall across this stretch has varied in a zonal “dipole” fashion in the late Quaternary (here, the last ~ 25 kyr) with rainfall increasing in northeastern Brazil at the expense of drying in western Amazonia, and vice versa (Martin et al., 1997; Cruz et al., 2009; Cheng et al., 2013; M. C. Campos et al., 2022). This zonal rainfall pattern is called the “South American Precipitation Dipole” (SAPD), a term that describes the opposing east-west patterns of past rainfall anomalies (Fig. 1a,b), and is distinct from the precipitation dipole studied in the modern climate between southeastern South America and the South Atlantic Convergence Zone (Nogués-Paegle & Mo, 1997; Boers et al., 2014). The SAPD has been identified on precession (Martin et al., 1997; Wang et al., 2004; Cruz et al., 2009; Cheng et al., 2013) and glacial-interglacial timescales (Abouchami & Zabel, 2003; Mason et al., 2019), and

67 it corresponds with many high-amplitude signals in paleoclimate proxy data (P. A. Baker,
68 Seltzer, et al., 2001; P. A. Baker, Rigsby, et al., 2001; Tapia et al., 2003; Fritz et al., 2004;
69 Cruz et al., 2009; Wang et al., 2017). Still, conflicting model and proxy interpretations
70 cast doubt on what drives the SAPD (Cruz et al., 2009; Liu & Battisti, 2015; M. C. Cam-
71 pos et al., 2022), and even whether it exists at all (Wang et al., 2017).

72 On precession timescales, a primary challenge of the SAPD enigma is how to inter-
73 pret the speleothem oxygen isotope ($\delta^{18}O$) records that span the dipole region (Fig.
74 1c-e). These spatially and temporally complex $\delta^{18}O$ records are difficult to reconcile with
75 some independent proxy data. From the relatively high austral summer insolation phase
76 around the Last Glacial Maximum (LGM; ~ 20 ka) to the lower phase at the mid-Holocene
77 (~ 7 ka), the speleothem $\delta^{18}O$ records are zonally imbalanced—the $\delta^{18}O$ shifts to the east
78 are about twice as large as the opposing shifts in the west. East $\delta^{18}O$ is, if anything, less
79 sensitive to precipitation amount than west $\delta^{18}O$ today (Fig. S1), so it is speculated that
80 these data imply a zonally imbalanced SAPD with larger precipitation anomalies in the
81 east (Cheng et al., 2013). Yet, the implication of a more quiescent precipitation history
82 in the west is not consistent with previous evidence for substantial drying from the wetter-
83 than-present LGM to the mid-Holocene (P. A. Baker, Seltzer, et al., 2001; P. A. Baker,
84 Rigsby, et al., 2001; Fritz et al., 2004; Tapia et al., 2003). Further, evidence from stron-
85 tium isotopes in speleothems across tropical South America shows $\delta^{18}O$ is often decou-
86 pled from rainfall amount (Wortham et al., 2017; Ward et al., 2019). Thus, it is not clear
87 that zonally imbalanced $\delta^{18}O$ signals require a zonally imbalanced SAPD.

88 The speleothem $\delta^{18}O$ data also reveal important discrepancies with isotope-enabled
89 General Circulation Models (GCMs) forced with precession. In isotope-enabled GCMs,
90 opposing east-west precipitation and $\delta^{18}O$ anomalies have a similar magnitude—the SAPD
91 is zonally balanced (Cruz et al., 2009; Liu & Battisti, 2015). Under low summer inso-
92 lation, precipitation $\delta^{18}O$ decreases by 1-3‰ in the east, increases by the same amount
93 in the west, and shows no change in east-central Amazonia where there is a large $\sim 6\%$
94 shift in the speleothem data (Cruz et al., 2009; Liu & Battisti, 2015; Wang et al., 2017).
95 Thus, while the direction of change is reasonable, the magnitude and spatial pattern of
96 $\delta^{18}O$ is inconsistent with the speleothem data, suggesting factors other than precession
97 may contribute to the late Quaternary SAPD. Precession may also be insufficient to ex-
98 plain the apparent out-of-phase changes in speleothem $\delta^{18}O$ in the last ~ 15 kyr (Fig.
99 1c-e). Precession-driven insolation forcing is uniform east-to-west, but minimum and max-
100 imum $\delta^{18}O$ values occur at different times across tropical South America.

101 The goal of this manuscript is to develop a conceptual model for the late Quater-
102 nary SAPD that is consistent with the enigmatic features of the oxygen isotope records—
103 namely the zonally imbalanced $\delta^{18}O$ signals and their out-of-phase nature. We begin by
104 reinterpreting the $\delta^{18}O$ data to account for the effect of upwind rainout (where upwind
105 is east). Upwind rainout can decouple local $\delta^{18}O$ from local rainfall amount by gener-
106 ating low- $\delta^{18}O$ moisture that is transported downwind. The effect is widely known to
107 drive Amazon $\delta^{18}O$ in modeling, observational, and paleoclimate studies (Salati et al.,
108 1979; Grootes et al., 1989; Gat & Matsui, 1991; Vuille et al., 2003; Vimeux et al., 2005;
109 Vuille & Werner, 2005; Brienen et al., 2012; J. C. A. Baker et al., 2016; Ampuero et al.,
110 2020), yet has not been empirically constrained in previous interpretations of the SAPD
111 (van Breukelen et al., 2008; Cruz et al., 2009; Cheng et al., 2013). Accounting for up-
112 wind rainout yields two important results. First, it brings the $\delta^{18}O$ data in better agree-
113 ment with other proxy records, including strontium isotopes, and casts the SAPD as zon-
114 ally balanced—the magnitude of precipitation anomalies is similar in the east and west.
115 Second, the $\delta^{18}O$ data can be understood as recording zonal shifts in the location of max-
116 imum rainout, or the “precipitation centroid”, across tropical South America. A precip-
117 itation centroid that migrates east-west reconciles a zonally balanced SAPD with zon-
118 ally imbalanced $\delta^{18}O$ anomalies and it explains the out-of-phase $\delta^{18}O$ signals. Yet, the
119 mechanisms for a zonally migrating precipitation centroid are not immediately clear.

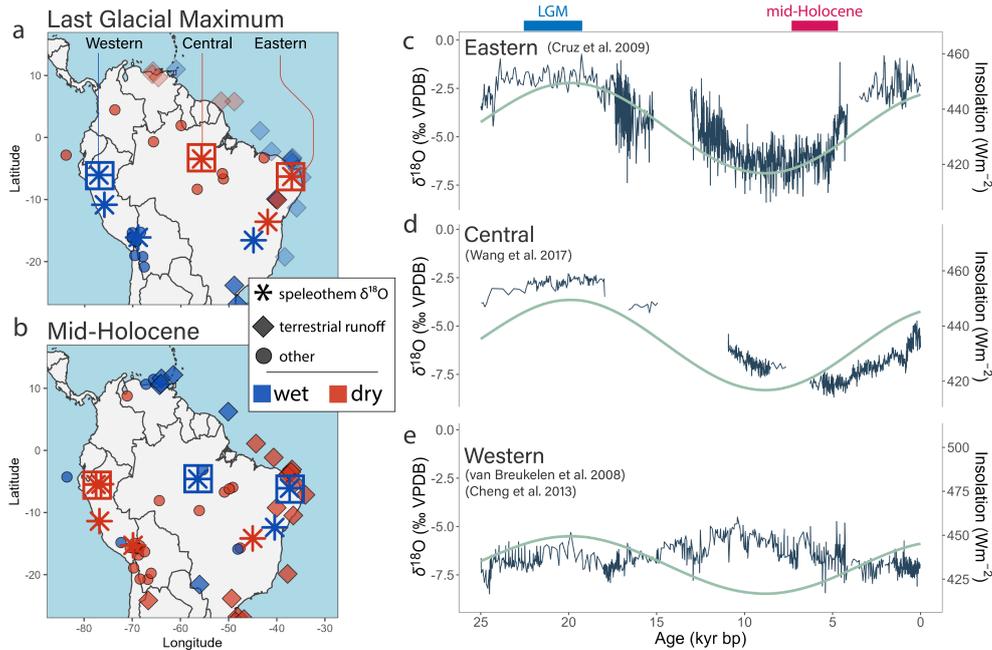


Figure 1. South America proxy map and isotope data. Proxy map for the LGM (~ 21 ka) (A) and mid-Holocene (~ 7 ka) (B). Data points in A and B are jittered to show instances of multiple proxy records from the same site. Offshore runoff proxies are lighter in panel A because they can record a sea-level signal at the Last Glacial Maximum. (C) Rio Grande do Norte (eastern) $\delta^{18}O$ record (Cruz et al., 2009). (D) Paraíso (central) $\delta^{18}O$ record (Wang et al., 2017). (E) Diamante and Tigre Perdido composite (western) $\delta^{18}O$ record (van Breukelen et al., 2008; Cheng et al., 2013). Teal lines (C-E) show February insolation at $10^{\circ}S$ following Cruz et al. (2009) (scales varied to match $\delta^{18}O$ magnitude).

120 In the second part of our analysis, we test whether precession forcing can explain
 121 a migrating precipitation centroid. Precession is considered the primary driver of the SAPD,
 122 and it was previously linked to east-west shifts in the pan-Asian Monsoon precipitation
 123 centroid (Battisti et al., 2014). These zonal pan-Asian Monsoon shifts caused large changes
 124 in $\delta^{18}O$ and precipitation, and we find similarly large changes in South America using
 125 an isotope-enabled reactive transport model (Kukla et al., 2019). However, the same GCM
 126 simulations presented in Battisti et al. (2014) showed no zonal migration in the tropi-
 127 cal South American precipitation centroid, and we also find no zonal shifts in the PMIP3/CMIP5
 128 models. Instead, we posit that land surface albedo change, in addition to precession, can
 129 explain the late Quaternary SAPD. We impose reasonable late Quaternary land albedo
 130 forcings in an energy balance model for tropical atmospheric circulation and find zonal
 131 shifts in the South American precipitation centroid that are consistent with the isotope
 132 data. We conclude that, while precession can drive a zonal precipitation dipole, addi-
 133 tional forcings such as land albedo are necessary to explain the zonal imbalance of $\delta^{18}O$
 134 signals, their magnitude, and their out-of-phase trends.

135 2 Late Quaternary speleothem $\delta^{18}O$ records and precipitation dynam- 136 ics

137 Our analysis leverages three existing speleothem $\delta^{18}O$ records that span tropical
 138 South America and have previously been used to identify the SAPD. We refer to these

139 as the eastern, central, and western records (Fig. 1). The eastern record is from the Rio
 140 Grande do Norte site of northeastern Brazil and shows a 5-7‰ decrease in $\delta^{18}O$ from
 141 the LGM to early-mid Holocene interpreted as evidence for a weakening South Amer-
 142 ican Monsoon (Cruz et al., 2009) (Fig. 1c). The central record comes from the Paraíso
 143 site in east-central Amazonia (Wang et al., 2017) and resembles the eastern record, but
 144 the $\delta^{18}O$ decrease lags behind by 1-2 kyr, and the records diverge in the late Holocene
 145 (Fig. 1d). Given its location near the monsoon’s deep convective region, data from the
 146 central site were interpreted as evidence for stronger convection in the mid-Holocene, in
 147 conflict with the eastern record interpretation (Wang et al., 2017). The western record
 148 is a composite of the Diamante (Cheng et al., 2013) and Tigre Perdido (van Breukelen
 149 et al., 2008) records (Fig. 1e). We adopt the cave temperature correction for these records
 150 following Wang et al. (2017) (see also Ampuero et al. (2020); Kukla et al. (2021)), in-
 151 creasing $\delta^{18}O$ by 1.4‰ to account for its relatively cooler cave temperatures. These records
 152 are interpreted to reflect Amazon or western Amazon rainfall amount, with a muted $\delta^{18}O$
 153 increase of ~ 2.5 ‰ from the LGM to early Holocene indicative of drying, then a grad-
 154 ual decrease to wetter, present conditions that starts when the eastern and then central
 155 $\delta^{18}O$ records initially decrease. The western record stands out from the central and east-
 156 ern records in that $\delta^{18}O$ increases, rather than decreases, from the LGM to the early-
 157 mid Holocene. This contrast defines the $\delta^{18}O$ expression of the SAPD, with the western-
 158 wet phase at the LGM and eastern-wet phase at the mid-Holocene representing end-member
 159 SAPD states.

160 The zonal SAPD is likely driven by multiple factors but, on precession timescales,
 161 it is agreed that changes in austral summer insolation are critical (Cruz et al., 2009; Cheng
 162 et al., 2013; Prado et al., 2013; Liu & Battisti, 2015; M. C. Campos et al., 2022). Pre-
 163 cession drives summer insolation with a ~ 21 kyr beat, and this forcing carries no zonal
 164 component. Low austral summer insolation (as during the mid-Holocene, ~ 7 ka) weak-
 165 ens the South American Monsoon and decreases rainfall in western tropical South Amer-
 166 ica (P. A. Baker, Seltzer, et al., 2001; Cruz et al., 2009; Liu & Battisti, 2015; M. C. Cam-
 167 pos et al., 2022). However, the opposing increase in eastern precipitation requires some
 168 zonal shift in atmospheric circulation, and the cause is debated. One theory posits that
 169 weaker subsidence over northeast Brazil must compensate for weaker convection to the
 170 west, increasing northeast Brazil rainfall (Cruz et al., 2009; Shimizu et al., 2020; M. C. Cam-
 171 pos et al., 2022). Another argues that northeast Brazil rainfall increases as south African
 172 summer cooling shifts the subtropical rain band, the South Atlantic Convergence Zone,
 173 northward, and north African cooling shifts the tropical rain band, the Inter-Tropical
 174 Convergence Zone (ITCZ), southward (Liu & Battisti, 2015), consistent with a broader
 175 seasonal ITCZ migration (Chiessi et al., 2021). In both cases models capture zonally op-
 176 posing $\delta^{18}O$ anomalies, but not their zonal imbalance, nor the magnitude of eastern and
 177 central $\delta^{18}O$ change (note that the central record was published after Cruz et al. (2009)
 178 and Liu and Battisti (2015)). Moreover, other simulations with precession forcing find
 179 no SAPD, or a zonal precipitation dipole in the austral summer that is offset by oppos-
 180 ing anomalies in the austral winter (Prado et al., 2013; Tigchelaar & Timmermann, 2016;
 181 Shimizu et al., 2020). One key limitation in the application of these models to the mid-
 182 Holocene is they do not account for the greening-induced decrease in Saharan land albedo—
 183 a major boundary condition change that has previously been linked to rainfall anom-
 184 alies in tropical South America (Lu et al., 2021). If such zonal forcings can impact trop-
 185 ical South American rainfall, they may be critical for explaining the zonal patterns of
 186 the SAPD.

187 Recent theoretical work demonstrates that South American rainfall, more so than
 188 other tropical regions, is energetically primed to shift east-west due to factors like non-
 189 local land surface albedo change (Boos & Korty, 2016). The precipitation centroid in trop-
 190 ical South America sits at the intersection of the energy flux equator (correlated with
 191 the ITCZ) and an energy flux prime meridian (Boos & Korty, 2016). These energy flux
 192 lines occur where column-integrated divergent atmospheric energy transport is zero in

193 the meridional (energy flux equator) and zonal (energy flux prime meridian) directions
 194 (Boos & Korty, 2016). The energy flux equator-prime meridian intersection conditions
 195 the precipitation centroid to migrate zonally because, just as the energy flux equator (and
 196 ITCZ) moves north and south following anomalous meridional energy sources (*e.g.*, changes
 197 in insolation and albedo), the energy flux prime meridian moves west and east in response
 198 to zonal energy anomalies. North-south shifts in the precipitation centroid, following the
 199 energy flux equator, are well documented in tropical South America and elsewhere (Haug,
 200 2001; Arbuszewski et al., 2013; Deplazes et al., 2013; Mulitza et al., 2017; J. L. P. S. Cam-
 201 pos et al., 2019; Chiessi et al., 2021), but east-west shifts are less thoroughly explored.
 202 The pan-Asian monsoon is also associated with an energy flux prime meridian and has
 203 been shown to migrate zonally with high-amplitude precession forcing, though preces-
 204 sion alone appears insufficient to shift the precipitation centroid east-west in South Amer-
 205 ica (Battisti et al., 2014; Liu & Battisti, 2015; Shimizu et al., 2020). If other factors drove
 206 the energy flux prime meridian over South America to shift zonally in the past, we ex-
 207 pect the precipitation centroid to shift with it (Boos & Korty, 2016), driving a zonal dipole
 208 in rainout expressed as the SAPD.

209 3 Methods

210 3.1 Paleo-isotope gradient justification

211 The isotope gradient is defined as downwind $\delta^{18}O$ minus upwind $\delta^{18}O$ along a given
 212 moisture trajectory and it is expressed in units of ‰ per thousand kilometers. This change
 213 in $\delta^{18}O$ is related to Rayleigh distillation interpretations of isotopic data as both $\delta^{18}O$
 214 and $\Delta\delta^{18}O$ decrease as net rainout (or distillation) increases (Salati et al., 1979; Gat &
 215 Matsui, 1991). Whereas one must assume the upwind $\delta^{18}O$ value to interpret a given
 216 $\delta^{18}O$ record in terms of net rainout, $\Delta\delta^{18}O$ explicitly accounts for these upwind vari-
 217 ations, theoretically isolating the $\delta^{18}O$ signal due to rainout alone (Hu et al., 2008; Win-
 218 nick et al., 2014; Kukla et al., 2019, 2021). This approach is particularly useful in tropi-
 219 cal South America because upwind effects are known to be a primary driver of $\delta^{18}O$ (Salati
 220 et al., 1979; Gat & Matsui, 1991; Vuille et al., 2003; Vuille & Werner, 2005; Lee et al.,
 221 2009; Liu & Battisti, 2015; J. C. A. Baker et al., 2016; Ampuero et al., 2020). Upwind
 222 and local rainout can be distinguished because upwind rainout will change the initial $\delta^{18}O$
 223 of a given domain but not $\Delta\delta^{18}O$ (Salati et al., 1979; Kukla et al., 2021). We note that
 224 $\Delta\delta^{18}O$ values are generally restricted to below zero (the “hydrostat”), since a zero iso-
 225 tope gradient reflects all precipitation being recycled between two sites or zero or neg-
 226 ligible precipitation (*i.e.* no net rainout) (Caves et al., 2015; Chamberlain et al., 2014;
 227 Kukla et al., 2019).

228 To validate our use of the isotope gradient approach, we analyze the connectivity
 229 of atmospheric moisture through transport and recycling among the eastern, central, and
 230 western sites in the modern climate. We use the 2-layer Water Accounting Model (WAM-
 231 2layers) (van der Ent et al., 2014) and the precipitation back-tracking scheme of Keys
 232 et al. (2012) (van der Ent & Savenije, 2013; van der Ent, 2016). The model simulates
 233 precipitation-sheds, the area where evaporation sources a site’s precipitation, and evaporation-
 234 sheds, the area where a site’s evaporation re-precipitates out. Contours enclosing the area
 235 where 70% of a site’s rainfall is sourced (for precipitation-sheds) or a site’s evaporation
 236 rains out (for evaporation-sheds) can be used to infer a meaningful dynamic connection
 237 between sites (Keys et al., 2012). We find that moisture recycling between our sites sur-
 238 passes this threshold (Fig. 2), demonstrating that these sites are sufficiently isotopically
 239 connected for $\Delta\delta^{18}O$ analysis (see Supplemental Text S1; Fig. S2, S3). We also find that
 240 the modern isotope gradient across tropical South America is negatively correlated with
 241 rainout and is negative throughout the year, consistent with theory for upwind signals
 242 propagating downwind with minimal attenuation (Fig. S4) (Kukla et al., 2019). Con-
 243 straining the isotopic connectivity between sites is challenging, in part because a strong
 244 connection today does not necessarily imply a strong connection in the past. Yet, as air-

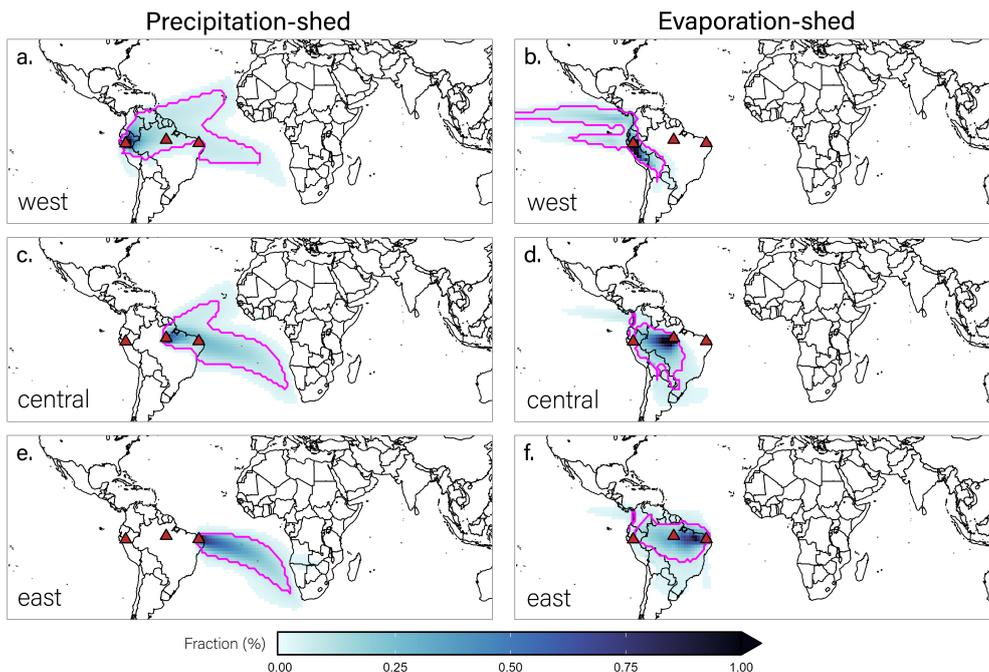


Figure 2. WAM-2layers hydrologic connectivity between speleothem sites. Annual mean precipitation-sheds (A, C, E) and evaporation-sheds (B, D, F) for the western (A, B), central (C, D), and eastern (E, F) sites. Magenta line denotes the spatial threshold where 70% of precipitation is sourced (precipitation-sheds) or where evaporation re-precipitates (evaporation-sheds), used to indicate a dynamically significant connection (Keys et al., 2012). Results show sites are significantly hydrologically connected within a single evaporation-precipitation cycle (See supplement for seasonal results).

245 masses shift in response to past forcings, moisture at one site can still be recycled to the
 246 next. Today, moisture recycling connects regions across tropical South America that re-
 247 ceive their peak rainfall at different times of the year (Staal et al., 2018). We discuss how
 248 weak isotopic connections would affect our conclusions in section 5.4.

249 3.2 Reconstructing paleo-precipitation rates from the central-to-western 250 sites

251 We focus exclusively on the isotope gradient between the central and western sites
 252 for our quantitative precipitation reconstruction because this trajectory aligns best with
 253 that of the prevailing winds (see Supplemental Text S1, S2). Oxygen isotope gradients
 254 along a dominant moisture trajectory depend on the balance of three fluxes: precipita-
 255 tion, evapotranspiration, and atmospheric transport (Salati et al., 1979; Winnick et al.,
 256 2014). We use a reactive transport model that simulates $\Delta\delta^{18}O$ as a function of these
 257 fluxes to quantify past precipitation rates from $\Delta\delta^{18}O$ data. To do so, we randomly sam-
 258 ple from uniform distributions of reactive transport model input parameters to estimate
 259 past precipitation from the simulations that agree with $\Delta\delta^{18}O$ data (Kukla et al., 2019,
 260 2021).

261 Our application of the reactive transport model to the central-to-western isotope
 262 gradient follows that of Kukla et al. (2021) with one key change. Kukla et al. (2021) used
 263 modern reanalysis data to analyze both the late Holocene and mid-Holocene isotope gra-

264 dients because PMIP3/CMIP5 results (Braconnot et al., 2012) show that reactive trans-
 265 port model inputs are similar for both time periods. However, modern reanalysis data
 266 cannot be reasonably applied to the LGM due to the $\sim 5^\circ\text{C}$ of tropical cooling. To ac-
 267 count for this cooling, we apply temperature-based scaling relationships to the reanal-
 268 ysis data to estimate LGM moisture content over the ocean (moisture source region) and
 269 potential evapotranspiration. Source region moisture content is calculated assuming rel-
 270 ative humidity remains constant over the ocean (Sherwood et al., 2010), and potential
 271 evapotranspiration is decreased following the scaling relationship defined by Scheff and
 272 Frierson (2014) and Siler et al. (2019). Decreasing source moisture content and poten-
 273 tial evapotranspiration both increase net rainout, all else equal. Therefore, these changes
 274 decrease the reconstructed LGM precipitation rates required to reproduce a given iso-
 275 tope gradient. Moisture content and humidity are allowed to change over land depend-
 276 ing on model-simulated rainout. We further account for unique LGM conditions by re-
 277 stricting the wind speed and transpiration fraction estimates. Proxy studies (McIntyre
 278 & Molino, 1996; Bradtmiller et al., 2016; Venancio et al., 2018) suggest that the north-
 279 easterlies were stronger at the LGM, so we restrict wind speeds to be equal to or greater
 280 than the late Holocene. Lower atmospheric $p\text{CO}_2$ implies lower plant water use efficiency
 281 suggesting that more transpiration may have been necessary to fix (approximately) the
 282 same amount of carbon. Since the rainforest largely remained intact at the LGM (*i.e.*
 283 similar biomass), we assume the transpired fraction of evapotranspiration is also equal
 284 to or greater than modern.

285 We find that our results are not sensitive to the shape of the distributions of model
 286 inputs, nor the sample size of the Monte Carlo routine (Fig. S5). We also test the im-
 287 portance of an additional input, rain re-evaporation, on model $\delta^{18}\text{O}$. Rain re-evaporation
 288 and its effect on $\delta^{18}\text{O}$ is heavily parameterized in models because it is difficult to directly
 289 measure (Worden et al., 2007; Dee et al., 2015; Konecky et al., 2019) (see Supplemen-
 290 tary text S2-S3). Using a parameterization fit to isotope data we find that it has a neg-
 291 ligible effect on $\delta^{18}\text{O}$ in the model (Fig. S6). Diagnostics of our late Holocene Monte Carlo
 292 results (essentially a modern analysis because late Holocene speleothem $\Delta\delta^{18}\text{O}$ is the
 293 same as modern rainfall) are provided in Fig. S7.

294 We further use the reactive transport model to calculate spatial $\delta^{18}\text{O}$ patterns for
 295 individual PMIP3/CMIP5 models (Braconnot et al., 2012). Using zonal profiles of at-
 296 mospheric moisture content, zonal winds, potential evapotranspiration, and temperature
 297 from the individual PMIP3/CMIP5 models, we run the reactive transport model to sim-
 298 ulate the isotope gradient for the LGM, mid-Holocene, and late Holocene (PMIP3/CMIP5
 299 pre-industrial). We then compare the predicted $\Delta\delta^{18}\text{O}$ derived from the PMIP3/CMIP5
 300 data to the speleothem data. If the predicted $\Delta\delta^{18}\text{O}$ is more negative than the observed
 301 $\Delta\delta^{18}\text{O}$, then the net rainout in that model is too high to reconcile the observed data in
 302 the reactive transport framework. We also analyze the precipitation rate necessary to
 303 match the paleo-isotope gradient if all other PMIP3/CMIP5 inputs to the reactive trans-
 304 port model are correct. This analysis effectively asks how much rainfall must increase
 305 or decrease relative to the PMIP3/CMIP5 prediction in order to reconcile the paleocli-
 306 mate $\Delta\delta^{18}\text{O}$ data.

307 **3.3 Application of a 2-dimensional atmosphere energy balance model**

308 We use a 2-dimensional energy balance model that is capable of tracking zonal shifts
 309 in the precipitation centroid (Boos & Korty, 2016) to accomplish two related goals. First,
 310 we identify the precipitation centroid to test whether it shifts zonally in the mid-Holocene
 311 or LGM simulations of the PMIP3/CMIP5 models. Second, we simulate the zonal pre-
 312 cipitation centroid response to conditions that likely characterize the LGM and mid-Holocene
 313 but are not accounted for in the PMIP3/CMIP5 experiments.

314 The energy balance model predicts how changes in energy input to the atmosphere
 315 would change atmospheric energy transport, thus altering atmospheric circulation and
 316 precipitation patterns. Here, we follow the methodology of Boos and Korty (2016) and
 317 consider how changes in continental albedo alter energy input to the atmosphere, and
 318 how atmospheric circulation would have to adjust in order to maintain the energy bal-
 319 ance. The anomalous energy flux generated by the energy balance model is then used
 320 to infer a shift in precipitation based on the assumption that the position of peak pre-
 321 cipitation migrates with the intersection of the energy flux equator and energy flux prime
 322 meridian (see equations 2-7 in Boos and Korty (2016)). We refrain from attributing the
 323 precipitation centroid anomalies to a specific atmospheric feature because the model is
 324 not designed to distinguish between the individual effects of, for example, the South Amer-
 325 ican Monsoon, the South Atlantic Convergence Zone, and the ITCZ.

3.3.1 Analysis of PMIP3/CMIP5 precipitation dynamics

326
 327 Using the energy balance model, we identify the PMIP3/CMIP5 ensemble mean
 328 location of the precipitation centroid, defined as the intersection of the energy flux equa-
 329 tor and prime meridian, for the LGM, mid-Holocene, and pre-industrial (or late Holocene).
 330 The LGM and mid-Holocene ensemble means are then used as the initial conditions for
 331 the perturbations discussed in the next section.

3.3.2 Simulating additional LGM and mid-Holocene constraints

332
 333 A critical step in determining whether the precipitation centroid migrated zonally
 334 in the past is quantifying the sensitivity of zonal shifts to energetic forcing. We impose
 335 anomalous moist static energy sources in the PMIP3/CMIP5 ensemble mean to quan-
 336 tify how the zonal location of the energy flux prime meridian (and thus the precipita-
 337 tion centroid (Boos & Korty, 2016)) changes with zonal forcing. The response of the South
 338 American precipitation centroid to anomalous energy forcing depends on (1) the mag-
 339 nitude and direction of energetic forcing; (2) the area over which the forcing is applied;
 340 and (3) the distance (especially zonally) of the anomalous forcing to the centroid.

341 During the mid-Holocene, lower land surface albedo likely increased the net col-
 342 umn energy over the grassy “green” Sahara by about $70 W/m^2$, accounting for the at-
 343 tenuation of the albedo anomaly at the top of the atmosphere (Boos & Korty, 2016). This
 344 forcing exceeds the magnitude of insolation change due to orbital variability ($\sim 10 W/m^2$
 345 in the mid-Holocene), but is applied over a smaller area (confined to the modern Sahara).
 346 Other modeling investigations of the late Quaternary SAPD (including PMIP3/CMIP5
 347 simulations) accounted for orbital forcing, but did not consider the Green Sahara (Cruz
 348 et al., 2009; Liu & Battisti, 2015). During the LGM there is evidence for forest dieback
 349 and grassland expansion in the African tropics, plus tundra expansion in the forests of
 350 modern Eurasia (Wu et al., 2007; Prentice et al., 2011; Binney et al., 2017). These veg-
 351 etation shifts would have brightened the regional land surface and, barring strong com-
 352 pensating feedbacks, the top of atmosphere. We note that our analysis does not account
 353 for other factors outside of moist static energy anomalies that can shift the precipita-
 354 tion centroid zonally. For example, there is evidence for stronger easterly winds across
 355 the tropical Atlantic at the LGM (McIntyre & Molino, 1996; Adkins et al., 2006; McGee
 356 et al., 2013; Bradtmiller et al., 2016; Zular et al., 2019) that could shift the maximum
 357 vector wind divergence, and thus precipitation centroid, westward, but stronger winds
 358 cannot be readily integrated to the energy balance model as an anomalous energy source.

359 Starting from the ensemble mean mid-Holocene and LGM climates, we simulate
 360 the effect of a darker Sahara (mid-Holocene) and a brighter African tropics and Eura-
 361 sia (LGM) as spatially uniform positive and negative moist static energy anomalies, re-
 362 spectively. This approach carries some important limitations and should be taken as a
 363 proof of concept for demonstrating how land surface albedo can modulate the zonal lo-

364 cation of the South American precipitation centroid. Our analysis implicitly assumes that
 365 the attenuation of the land surface anomaly to the top of atmosphere is spatially uni-
 366 form, which is unlikely when comparing tropical Africa and Eurasia. This analysis also
 367 ignores the role of an apparent shift to a less El Niño-dominant mean climate state af-
 368 ter the LGM (Koutavas & Joanides, 2012; Ford et al., 2018), which could affect the zonal
 369 energy balance (Boos & Korty, 2016) and was previously argued to contribute to the zon-
 370 ally imbalanced $\delta^{18}O$ signals by decreasing $\delta^{18}O$ everywhere, amplifying the eastern and
 371 dampening the western trend (Cheng et al., 2013). However, because this mechanism
 372 affects all sites similarly, so it cannot explain the $\Delta\delta^{18}O$ trends that we interpret as changes
 373 in the precipitation centroid’s location.

374 4 Results and interpretation

375 4.1 Isotope gradients and net rainout

376 The isotope gradients over space are distinct from any one $\delta^{18}O$ record, suggest-
 377 ing there is no single representative site that reflects basin-wide rainout. Figure 3a shows
 378 these gradients for the eastern-to-central sites (“eastern domain”; light green) and central-
 379 to-western sites (“western domain”; dark green), with the theoretical maximum $\Delta\delta^{18}O$
 380 value of zero labelled as the hydrostat (Chamberlain et al., 2014; Caves et al., 2015; Kukla
 381 et al., 2019). The hydrostat is the point where further drying has no affect on $\Delta\delta^{18}O$
 382 because nearly all precipitation is being recycled. The eastern domain gradient is near
 383 the hydrostat from the LGM to the early Holocene, then decreases to $\sim-2.5\%$ in the mid-
 384 late Holocene and increases by $<1\%$ to present. While at the hydrostat, $\Delta\delta^{18}O$ does
 385 not capture further drying that likely distinguishes the early-mid Holocene conditions
 386 from the mid-late Holocene (P. A. Baker, Seltzer, et al., 2001; Fritz et al., 2004; Cheng
 387 et al., 2013; Kukla et al., 2021). The western domain gradient shows a mostly oppos-
 388 ing trend, with $\Delta\delta^{18}O$ near $\sim-2\%$ at the LGM and increasing to zero, the hydrostat,
 389 by the mid-Holocene before decreasing to present. The late Holocene $\Delta\delta^{18}O$ value in this
 390 domain is similar to the rainfall $\Delta\delta^{18}O$ across the tropical South America today (Salati
 391 et al., 1979; Wang et al., 2017). Overall, despite $\delta^{18}O$ shifts that are zonally imbalanced
 392 (about twice as large in the eastern and central records compared to the west), the mag-
 393 nitude of $\Delta\delta^{18}O$ change is comparable in each domain, consistent with zonally balanced
 394 changes in rainout.

395 Following previous work using isotope gradients (Salati et al., 1979; Hu et al., 2008;
 396 Winnick et al., 2014), we interpret the $\Delta\delta^{18}O$ data as reflecting rainout between two sites
 397 and the $\delta^{18}O$ data as recording the net integrated upwind rainout signal. Figure 3b is
 398 an attempt to visualize both the local ($\Delta\delta^{18}O$) and upwind ($\delta^{18}O$) rainout signals. Here,
 399 the eastern, central, and western $\delta^{18}O$ records are smoothed and plotted together with
 400 the space between them colored to illustrate the magnitude of change in $\delta^{18}O$ between
 401 each site. This figure shows that the location where $\delta^{18}O$ decreases the most (indicative
 402 of the most rainout) shifts from the western domain (dark green) at the LGM to east
 403 of the eastern site, over the tropical Atlantic Ocean (blue), by the mid-Holocene, to some-
 404 where in between by the late Holocene.

405 This interpretive framework explains how the SAPD is zonally balanced despite
 406 zonally imbalanced $\delta^{18}O$ records. The western $\delta^{18}O$ shifts are small compared to the east-
 407 ern record because the focus of rainout is always upwind of the western site. In contrast,
 408 the focus of rainout is downwind of the eastern and central sites at the LGM, and up-
 409 wind of these sites at the mid-Holocene. Put otherwise, the focus of rainout shifts along
 410 the moisture trajectory relative to the eastern and central sites, but not the western site,
 411 driving larger amplitude $\delta^{18}O$ trends in the eastern and central sites. We note that ad-
 412 ditional complications at the eastern site, such as competing air-masses (Garreaud et al.,
 413 2009; Liu & Battisti, 2015) could modify the relationship between rainout and $\Delta\delta^{18}O$

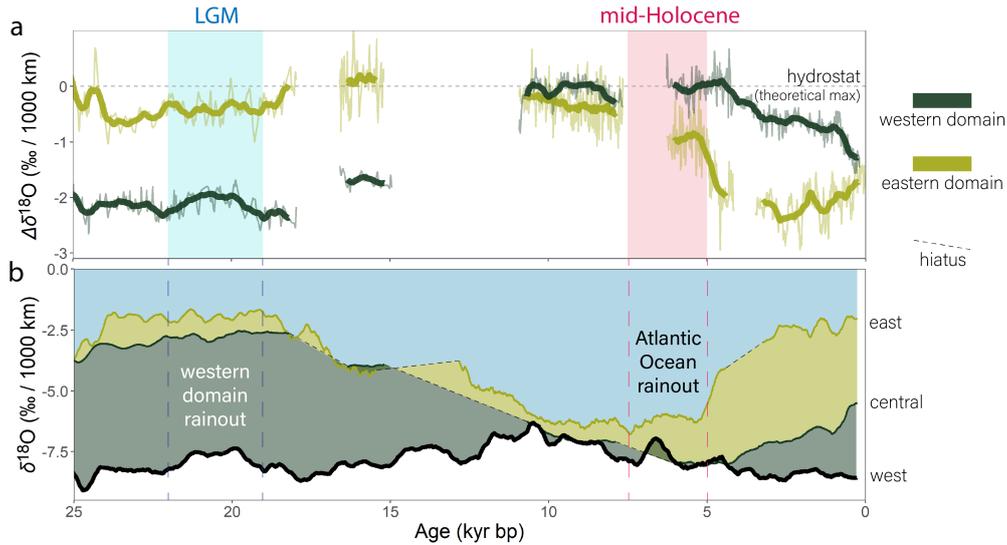


Figure 3. Isotope gradient and individual $\delta^{18}O$ records. (A) Eastern-to-central (light green) and central-to-western (dark green) isotopic gradients. More negative $\Delta\delta^{18}O$ is interpreted as more rainout between sites (wetter). (B) The three, smoothed $\delta^{18}O$ records (labels on right of panel) (van Breukelen et al., 2008; Cruz et al., 2009; Cheng et al., 2013; Wang et al., 2017). Y-axis color range is proportional to net moisture loss (rainout) within the western (dark green), eastern (light green), or ocean (turquoise) domains. The increase in blue toward the MH reflects a hypothesized increase in rainout over the ocean. Dashed lines with intervals of lighter shading are hiatus periods in the $\delta^{18}O$ records.

414 through time. We caution against interpreting eastern $\Delta\delta^{18}O$ as quantitative trends in
 415 rainout, and we expand on this point in section 5.4.

416 In addition to the zonally imbalanced $\delta^{18}O$ trends, another enigmatic feature of
 417 the $\delta^{18}O$ data is that the records are out-of-phase with one another. The out-of-phase
 418 nature of these $\delta^{18}O$ shifts can also be understood in the context of upwind effects. The
 419 western $\delta^{18}O$ record decreases from 10-5 ka (Fig. 3b) while $\Delta\delta^{18}O$ stays near the
 420 theoretical maximum value of zero (Fig. 3a), consistent with the $\delta^{18}O$ shift being driven
 421 by upwind rather than local rainout. Meanwhile, in the last 5 kyr, the focus of decreasing
 422 $\delta^{18}O$ shifts inland, first over the eastern domain and next over the western domain,
 423 revealing a time-transgressive trend that emerges from the central $\delta^{18}O$ data lagging the
 424 eastern record. Thus, the progressive inland migration of the focus of rainout provides
 425 a plausible mechanism for the enigmatic lag between these records.

426 4.2 Reconstructed annual precipitation rates

427 Our reactive transport results suggest that late Holocene precipitation rates were
 428 similar to modern, consistent with similar $\Delta\delta^{18}O$ values between the late Holocene speleothem
 429 data and modern rainfall. During the LGM, we find increased rainfall relative to the late
 430 Holocene (light blue distribution of Fig. 4a; 3000 ± 800 mm/yr). This result is consis-
 431 tent with extensive evidence for wetter conditions in western tropical South America (P. A. Baker,
 432 Seltzer, et al., 2001; P. A. Baker, Rigsby, et al., 2001; Fritz et al., 2004). When wind speed
 433 and transpiration are equal to or greater than modern values (see Methods), calculated
 434 rainfall increases to $\sim 3400 \pm 400$ mm/yr to compensate for increased moisture transport
 435 and decreased isotopic fractionation associated with transpiration (dark blue distribu-

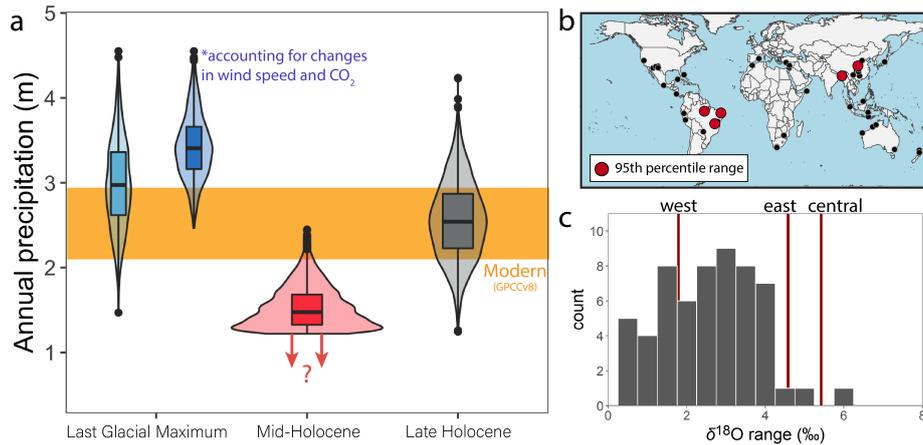


Figure 4. (A) Reconstructed precipitation for the LGM (blue), mid-Holocene (red) and late Holocene (gray). Mid-Holocene is restricted at the lower-bound because $\Delta\delta^{18}O$ is at the hydrostat. (B) Map of records with $\delta^{18}O$ ranges in the largest 5% (red) and all sites (black). (C) The distribution of $\delta^{18}O$ ranges. red, vertical lines show values for the west, central, and western sites discussed in the text. The central and eastern records are in the largest 5 percent of all similar sites (Supplemental text S4).

436 tions of Fig. 4a). We note that hydrogen isotope composition (δD) of leaf waxes from
 437 the Amazon River appears higher at the LGM than before or after, suggesting drier con-
 438 ditions than today (Häggi et al., 2017). However, the signal is small (about 1-2‰ in $\delta^{18}O$)
 439 and not inconsistent with our finding that the western domain $\Delta\delta^{18}O$ values are higher
 440 at the LGM compared to before and after.

441 During the mid-Holocene, the reactive transport model simulates rainfall decreas-
 442 ing to ~ 1200 mm/yr (about half of modern; red distribution of Fig. 4a). As discussed
 443 in Kukla et al. (2021), the $\Delta\delta^{18}O$ values in the mid-Holocene straddle zero—the theo-
 444 retical maximum value for a single moisture trajectory. At this point, further drying has
 445 a negligible effect on $\Delta\delta^{18}O$. The shape of the mid-Holocene distribution thus reflects
 446 the imposed lower-bound of annual precipitation, effectively restricting the solution to
 447 the wettest scenarios.

448 One limitation to our analysis is that we do not explicitly account for the possi-
 449 bility that changes in the seasonality of rainfall affect one site more than the other. Sea-
 450 sonality could be an issue in northeastern Brazil, where peak precipitation is offset from
 451 the central and western sites. However, seasonality, independent of rainout, is unlikely
 452 to drive the eastern (or central) $\delta^{18}O$ data because the amplitude of change is equal to
 453 or greater than the amplitude of $\delta^{18}O$ seasonality today (see Fig. S4). To formalize this
 454 point, we use high and low austral summer insolation results for northeastern Brazil from
 455 the isotope-enabled GCM experiments of Liu and Battisti (2015) to show that a 5-8‰
 456 decrease in wet-season precipitation $\delta^{18}O$ is required to explain the low mid-Holocene
 457 values (Fig. S8). Changes in monthly precipitation amount have a small effect on an-
 458 nual $\delta^{18}O$, as noted by Liu and Battisti (2015), indicating that a shift in the dominant
 459 air-mass cannot explain the speleothem signal. This required decrease in wet-season $\delta^{18}O$
 460 is about four times greater than that simulated by the isotope-enabled GCM (Fig. S8).
 461 Given the small influence of precipitation and air-mass changes, it is best explained by
 462 an increase in net upwind rainout. We expand on how changes in seasonality and atmo-
 463 spheric circulation affect our conclusions at other sites in the discussion section.

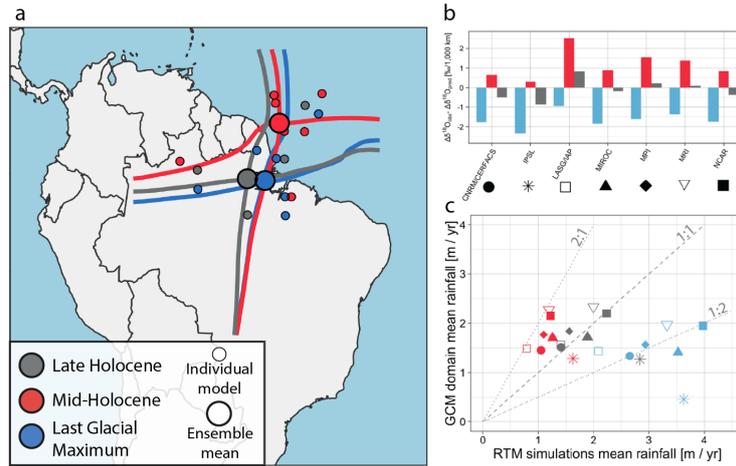


Figure 5. PMIP3/CMIP5 SAM centroid and isotope gradient analysis. (A) PMIP3 models show little zonal variation in the tropical South American precipitation centroid from the LGM, mid-Holocene, and late Holocene (pre-industrial) for months NDJFMAM. (B) When forced with PMIP3/CMIP5 output, the reactive transport model (Kukla et al., 2019) systematically predicts a steeper-than-observed $\delta^{18}O$ gradient at the mid-Holocene (red bars) and a shallower-than-observed gradient at the LGM (blue bars) with no systematic error in the late Holocene. This result is consistent with the $\delta^{18}O$ error found in the isotope-enabled simulations of Cruz et al. (2009) and Liu and Battisti (2015). (C) To match the observed oxygen isotope gradient, the reactive transport model requires similar rainfall amounts as predicted by the PMIP3/CMIP5 models at the late Holocene, but requires drier conditions than PMIP3/CMIP5 at the mid-Holocene and wetter conditions at the LGM.

464 The reactive transport model estimates of past precipitation show larger SAPD anomalies than predicted by GCMs (Cruz et al., 2009; Liu & Battisti, 2015; Shimizu et al., 2020).
 465 GCM simulations of the SAPD have accounted for precession and its impact on land surface heating, but not other “boundary condition” changes such as the land albedo response to vegetation change. If the speleothem $\delta^{18}O$ data reliably reflects past precipitation $\delta^{18}O$, we must consider the possibility that other factors, in addition to (and in response to) precession, shape the late Quaternary SAPD.
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471 Zonal migration of the precipitation centroid should drive larger precipitation anomalies than a zonally static precipitation centroid. Battisti et al. (2014), for example, argues that a zonal shift in the pan-Asian monsoon with changing northern hemisphere summer insolation could explain some of the largest documented speleothem $\delta^{18}O$ shifts. Using the SISALv2 database (Atsawawanunt et al., 2018; Comas-Bru et al., 2019, 2020) we find that magnitude of $\delta^{18}O$ shifts in the eastern and central records is in the top 5% of all comparable records (duration between 10^3 – 10^5 years and within 40° of the equator) (see Supplemental text S4) (Fig. 4b, c). The other records with large $\delta^{18}O$ ranges appear near the pan-Asian monsoon region, consistent with these two regions being among the most sensitive to zonal energy anomalies and precipitation shifts (Battisti et al., 2014; Boos & Korty, 2016).
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4.3 PMIP3/CMIP5 analysis with energy balance and reactive transport models

Our analyses with the PMIP3/CMIP5 data affirm previous isotope-enabled GCM results (Cruz et al., 2009; Liu & Battisti, 2015). The simulations do not capture zonal migration of the precipitation centroid and they under-estimate the magnitude of $\delta^{18}O$ variation. We find that, while the energy flux equator shifts northward in the mid-Holocene wet season (a result contested by other models; Liu and Battisti (2015); Chiessi et al. (2021)), the energy flux prime meridian does not show any systematic shift to the east or west (Fig. 5a). Meanwhile, when forced with PMIP3/CMIP5 output the reactive transport model correctly predicts late Holocene $\Delta\delta^{18}O$ data, demonstrating that the net rainout in the models is consistent with the isotope data despite broad precipitation biases. During the LGM, however, the PMIP3/CMIP5 output leads to an isotope gradient that is too shallow, consistent with the models being too dry (Fig. 5b,c). In contrast, the simulated isotope gradients are too steep at the mid-Holocene when driven by PMIP3/CMIP5 output, consistent with the models being too wet. Taken together, zonal shifts in the precipitation centroid are negligible in the PMIP3/CMIP5 models and their precipitation anomalies are smaller than suggested by the isotope data, despite good agreement in the late Holocene.

We therefore hypothesize that the the zonal migration of the precipitation centroid can resolve these discrepancies. This hypothesis is outlined in Figure 6, and we address its plausibility in the following subsection. We hypothesize that the precipitation centroid tracks the region of maximum net rainout (decreasing $\delta^{18}O$), located between the central and western records at the Last Glacial Maximum (Fig. 6c, f), upwind of the eastern record at the mid-Holocene (Fig. 6b, e), and somewhere in between in the late Holocene (Fig. 6a, d), consistent with its modern position (Boos & Korty, 2016). While zonal precipitation centroid migration aligns with the $\delta^{18}O$ data and its magnitude of change, it is unclear whether late Quaternary forcings could plausibly drive such zonal shifts.

4.4 Zonal migration of the precipitation centroid in an Energy Balance Model

We find that reasonable zonally asymmetric forcings for the mid-Holocene and LGM, not captured in the PMIP3/CMIP5 models, can cause the precipitation centroid to shift zonally relative to its initial PMIP3/CMIP5 ensemble mean state (Fig. 7). In our energy balance model simulations, the anomalous moist static energy source owed to a darker Sahara at the mid-Holocene is sufficient to pull the energy flux prime meridian east of the eastern speleothem record, consistent with its mid-Holocene $\delta^{18}O$ minimum (Fig. 7d-f). In contrast, a decrease in forest cover in tropical Africa and Eurasia pushes the energy flux prime meridian westward in the LGM (Fig. 7a-c). We note that the location of the energy flux equator-prime meridian intersection approximates, but may be offset from, the location of the precipitation centroid (Boos & Korty, 2016), although this offset should be constant as the precipitation centroid migrates (Adam et al., 2016; Boos & Korty, 2016). The simple energy balance model does not account for changes in the partitioning between latent and sensible heat, but any repartitioning does not alter the total energy flux from the land to the base of the atmospheric column (Laguë et al., 2019). Instead, repartitioning could affect the net energy imbalance via uncertain cloud feedbacks (Laguë et al., 2021) and possibly amplify the imbalance due to latent cooling-driven reductions in outgoing longwave radiation (Boos & Korty, 2016). A decrease in Saharan dustiness would also amplify the energy imbalance, though we do not account for it here. Our analysis shows that the precipitation centroid is sufficiently sensitive to remote forcing to explain the late Quaternary precipitation anomalies, although the exact location of rainout will depend on the initial state (here, from PMIP3/CMIP5) and the relative offset between the energy flux intersection and the precipitation centroid.

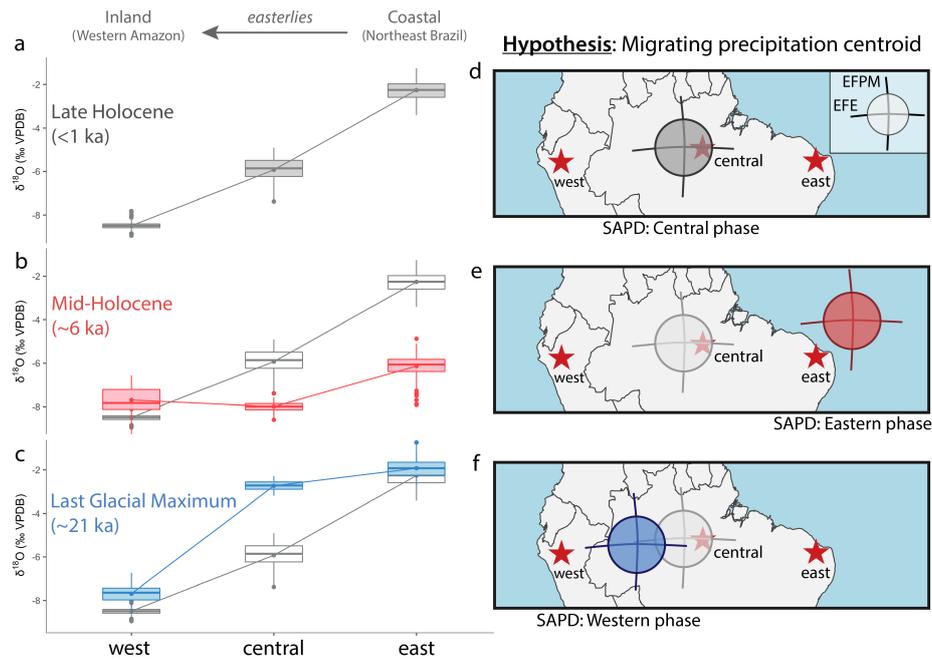


Figure 6. Precipitation centroid migration hypothesis and its isotopic expression. (A-C) Show the three distinct isotope profiles of the late Holocene (A), mid-Holocene (B), and LGM (C). Late Holocene is reproduced in panels B and C for comparison. Lines connect the mean of each site. Data from van Breukelen et al. (2008); Cruz et al. (2009); Cheng et al. (2013); Wang et al. (2017). (D-F) Illustrate hypothesized changes in the South American precipitation centroid (intersection of the energy flux equator (EFE) and energy flux prime meridian (EFPM)) based on where most of the $\delta^{18}O$ decrease (rainout) occurs (tropical Atlantic/northeast Brazil at mid-Holocene, and western Amazon at LGM).

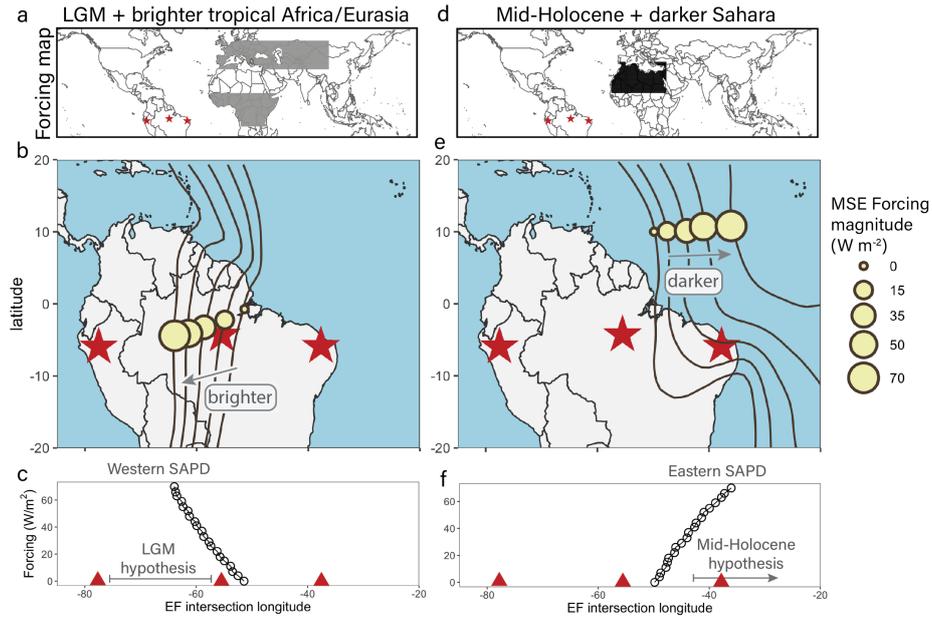


Figure 7. Sensitivity of the SAPD to zonal energy anomalies. Gray and black boxes in the maps of panels (A) and (D) show the locations of LGM and mid-Holocene moist static energy (MSE) forcings, respectively. Panels (B) and (E) show the response of the energy flux prime meridian (lines) and energy flux intersection (points; approximating the precipitation centroid) to selected forcing levels for the LGM and mid-Holocene. The energy flux intersection longitude versus the magnitude of forcing is shown in panels (C) and (F) for the LGM and mid-Holocene. We note that the points on the map panels are not a proposed path of the precipitation centroid in the late Quaternary, but rather the response to different forcing magnitudes starting from the PMP3/CMIP5 initial conditions.

5 Discussion

5.1 A zonally balanced SAPD

Previous work has argued that the distinct trends between the western and central/eastern $\delta^{18}O$ data reflect either (1) a zonally imbalanced precipitation dipole (Cheng et al., 2013); or (2) changes in the strength of convection, but not the location of peak rainout (Wang et al., 2017). Here, we address how our results support a third scenario—a zonally balanced SAPD that reconciles the zonally imbalanced amplitudes of $\delta^{18}O$ change. We also discuss how accounting for upwind rainout distinguishes our revised $\delta^{18}O$ interpretations from previous work.

Larger $\delta^{18}O$ signals in the eastern and central records do not require larger precipitation anomalies because a shift in the location of peak rainout has a small effect on sites that remain downwind. That is, whether the focus of rainout occurs near the speleothem site or a few hundred kilometers upwind, speleothem $\delta^{18}O$ will be approximately the same as long as the same magnitude of rainout occurs before the air mass reaches that site. We argue that this is why the western $\delta^{18}O$ trends are muted—the focus of rainout remains upwind of the western site for the entirety of the record. This may also explain discrepancies with basin-integrated precipitation isotope data. Häggi et al. (2017), for example, find basin-integrated δD trends that are small ($\sim 1\text{--}2\text{‰}$ in $\delta^{18}O$) and distinct from any one speleothem $\delta^{18}O$ record in the last 50 kyr, consistent with a decoupling of $\delta^{18}O$ and local precipitation amount. The magnitude of total rainout, and basin-integrated precipitation $\delta^{18}O$, appear relatively constant through time, regardless of whether that rainout occurs in the west or east. The western and eastern legs of the SAPD are approximately balanced.

Previous work has applied a different interpretive framework to the central and western $\delta^{18}O$ data to argue that the Amazon was wetter in the mid-Holocene and drier at the LGM, opposing our results (Wang et al., 2017). The key distinction with our work is that Wang et al. (2017) assume that upwind $\delta^{18}O$ is constant (with corrections for temperature and seawater $\delta^{18}O$ following P. A. Baker and Fritz (2015)) such that the central $\delta^{18}O$ record drives all variability in $\Delta\delta^{18}O$. We argue that the assumption of an effectively constant upwind $\delta^{18}O$ value is refuted by data (Cruz et al., 2009)—the strong correlation between central and eastern (upwind) $\delta^{18}O$ records is evidence that upwind $\delta^{18}O$ is propagating downwind without attenuation. Our approach avoids the assumption that upwind moisture loss is constant through time and, as we discuss in the next section, is consistent with evidence that $\delta^{18}O$ is not always strongly coupled with *local* precipitation amount (Wortham et al., 2017; Ward et al., 2019). Within our framework, the wettest time occurs when the isotope gradient is steepest, not when central $\delta^{18}O$ is lowest, consistent with our understanding of how $\Delta\delta^{18}O$ relates to precipitation today (Salati et al., 1979; Pattanayak et al., 2019; Ampuero et al., 2020).

5.2 Zonal and meridional components of precipitation centroid migration

Late Quaternary forcings, including precession and land surface change, should lead to zonal and meridional shifts in the precipitation centroid. The energy flux equator drives north-south migration that is well-documented in late Quaternary proxy records (Arbuszewski et al., 2013; Deplazes et al., 2013; Mulitza et al., 2017; Chiessi et al., 2021), and the energy flux prime meridian drives the east-west component (Boos & Korty, 2016) which we argue is evident in the speleothem $\delta^{18}O$ data. Because the location of the precipitation centroid varies near-linearly with anomalous forcing (rather than abruptly at some threshold; Fig. 7) its spatial migration should cause time-transgressive proxy trends as it reaches different locations at different times. Here, we compare the west-east (LGM to mid-Holocene) and east-west (mid-Holocene to present) SAPD transitions and dis-

583 cuss evidence for zonal and meridional structure of precipitation centroid migration with
 584 asynchronous proxy signals, consistent with our hypothesis.

585 We first focus on a broad comparison of the two SAPD transitions by comparing
 586 the eastern-to-central (eastern domain) and central-to-western (western domain) $\Delta\delta^{18}O$
 587 data, shown in Figure 8b. Here, lower values on the y-axis are interpreted as more west-
 588 ern domain rainout, and lower values on the x-axis as more eastern domain rainout. If
 589 the focus of rainout only migrates zonally, then a west-east trade-off in rainout will mark
 590 a diagonal line with slope -1 (as rainout in one domain increases at the expense of the
 591 other), and a shift in rainout further east over the ocean will trace a flat line with an in-
 592 tercept near zero (no rainout in the western domain, only moving through the eastern
 593 domain) (Fig. 8c, “expected if only zonal”). The $\Delta\delta^{18}O$ data, however, do not follow
 594 this trend. Instead, the LGM to early-mid Holocene marks a decrease in western domain
 595 rainout (increase in y-axis) with no compensating increase in the east (x-axis remains
 596 near zero) (Fig. 8b, points 1-2), followed by a mostly zonal progression into the eastern,
 597 then western domain (points 2-3 and 3-4, respectively). The precipitation centroid ap-
 598 pears to migrate eastward with a meridional component relative to the speleothem sites
 599 from the LGM to early-mid Holocene, and then westward following a zonal pattern through
 600 the speleothem sites to present.

601 This inferred pattern of migration from the $\Delta\delta^{18}O$ data is supported by indepen-
 602 dent proxy results. In the last eight thousand years, for example, a steeper isotope gra-
 603 dient reflecting more moisture distillation first appears in the eastern domain from ~ 8 -
 604 5 ka (points 2-3 of Fig. 8), and next in the western domain from ~ 5 - 0 ka (points 3-4)
 605 (see also Fig. 3). This result suggests that the precipitation centroid began passing over
 606 the central record ~ 5 thousand years ago, consistent with recent strontium isotope evi-
 607 dence from the same site pointing to high infiltration rates from 6 - 5 ka with less infil-
 608 tration before and after (Ward et al., 2019). The timing of migration is also consistent
 609 with a shift from dry to wet conditions in a nearby lake (Reis et al., 2017). After ~ 5 ka,
 610 rainfall begins increasing in the western domain and water infiltration rates at the central
 611 site temporarily decline (Ward et al., 2019). As discussed earlier, this gradual west-
 612 ward migration of rainout also explains the perplexing lag of central $\delta^{18}O$ behind the
 613 eastern record (Fig. 1c, d). The precipitation centroid first reaches the eastern site at
 614 ~ 8 - 7 ka when $\delta^{18}O$ values are lowest, and later the central site at ~ 5 ka, in tandem with
 615 records of the local water balance (Reis et al., 2017; Ward et al., 2019).

616 Unlike this east-west migration, the dipole transition from west to east spanning
 617 ~ 20 - 10 ka does not coincide with a decrease in eastern domain $\Delta\delta^{18}O$, and we suggest
 618 this reflects the precipitation centroid moving around, rather than through, the eastern
 619 domain (points 1-2 of Fig. 8). Movement around the domain would require a meridional
 620 component of precipitation centroid migration reflected by a change in $\Delta\delta^{18}O$ in one do-
 621 main that is not balanced by a corresponding change in the other (Fig. 8c, bottom panel).
 622 It is possible that the precipitation centroid moved southeast around the central $\delta^{18}O$
 623 site as there is evidence for wetter conditions to the southeast (Whitney et al., 2011; For-
 624 nace et al., 2016) and drier conditions to the north (Deplazes et al., 2013; Zular et al.,
 625 2019), as well as some evidence for a south-shift of the energy flux equator (Arbuszewski
 626 et al., 2013). The southeast appears to become drier around 12 ka, approximately when
 627 a nearby speleothem $\delta^{18}O$ shift occurs that is consistent with decreased Amazon and more
 628 Atlantic-derived moisture (Fig. S9) (Novello et al., 2017, 2018).

629 As the precipitation centroid migrates further east, after ~ 12 ka, pollen data from
 630 semi-arid northeastern Brazil (near the eastern $\delta^{18}O$ site) suggest humid conditions from
 631 ~ 10.9 - 6.7 ka (De Oliveira et al., 1999). Humidity peaks halfway through this interval
 632 (~ 8.9 ka) when eastern $\delta^{18}O$ reaches its lowest values (De Oliveira et al., 1999; Cruz et
 633 al., 2009), suggesting this marks the easternmost extent and turning point of the pre-
 634 cipitation centroid. As discussed earlier, this timing also corresponds with the onset of
 635 the time-transgressive westward shift in wet conditions that continues to the present. While

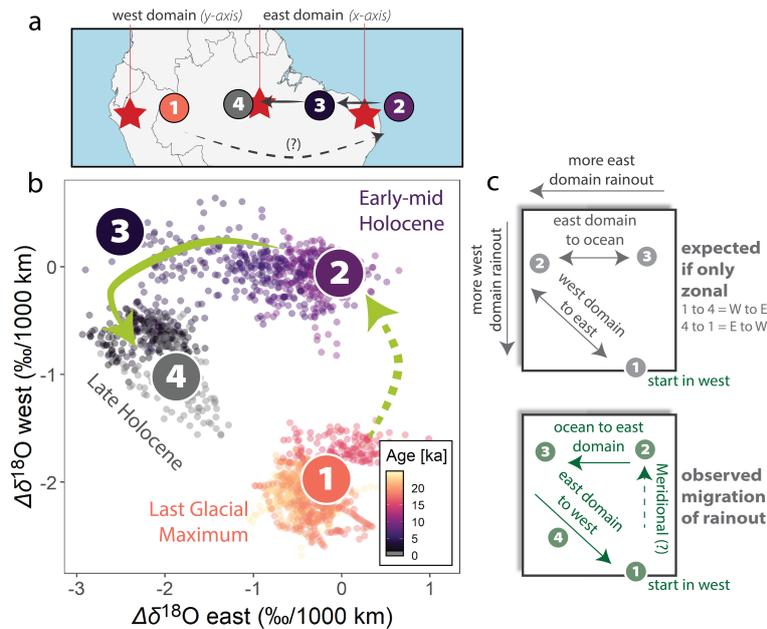


Figure 8. Isotope gradients reflect zonal and meridional shifts in the precipitation centroid. (A) Map of speleothem sites showing the east and west domains—the axes of panel B—and a schematic for the interpretation of panel B. (B) Crossplot of eastern and western domain data (derived from data in van Breukelen et al. (2008); Cruz et al. (2009); Cheng et al. (2013); Wang et al. (2017)). Numbered points in B correspond with numbers in panels A and C. (C) More negative $\Delta\delta^{18}O$ refers to more rainout in a given domain. Data should track a sideways “V” shape if the focus of rainout migrates only zonally (top panel, note different order of numbers). However, the LGM to early-mid Holocene does not follow this zonal trajectory, suggesting a meridional component (dashed arrow).

636 more work is needed to trace the past focus of rainfall, we suggest the progressive shifts
 637 in wet conditions across the continent (both east-west and west-east) provide empirical
 638 support and a testable framework for the pattern of precipitation centroid migration.

639 5.3 Mechanisms for zonal precipitation centroid migration

640 While climate model simulations are necessary to assess the dynamical drivers of
 641 precipitation centroid migration, our analysis allows us to present testable hypotheses.
 642 For example, the greening of the Sahara at the mid-Holocene (about $70 W/m^2$ anomalous
 643 heat source at top of atmosphere; Boos and Korty (2016)) is likely sufficient to drive
 644 the energy flux prime meridian eastward entirely over the tropical Atlantic (Fig. 7e, f).
 645 Comparison to proxy records from Africa generally support this remote influence on tropi-
 646 cal South American rainfall. Dust flux records of West African Monsoon behavior show
 647 pronounced precession-scale variability in the last 240 kyr with prominent exceptions at
 648 ~ 30 , ~ 70 , and ~ 150 ka when dust fluxes “skip” precession beats (Skonieczny et al., 2019).
 649 In South America, western Amazon $\delta^{18}O$ records lose sensitivity to precession at the same
 650 times (and $\Delta\delta^{18}O$ where there is data, in the ~ 30 ka case; Fig. S10) (Mosblech et al.,
 651 2012; Cheng et al., 2013; Wang et al., 2017). Further, there is a rapid increase in $\delta^{18}O$
 652 at the eastern site at ~ 5 ka, consistent with a westward (inland) shift of rainout, con-
 653 temporaneous with the termination of the African Humid Period in North Africa (Shanahan
 654 et al., 2015) where increasing land albedo would provide an anomalous energy sink. These

655 similarities are mostly preliminary and more data is needed to test if they hold over space
 656 and time, but they are consistent with expectations if the zonal location of the precip-
 657 itation centroid was sensitive to Saharan albedo.

658 At the LGM, vegetation change that increases land albedo in tropical Africa and
 659 Eurasia could push the energy flux prime meridian westward. However, it is not clear
 660 if the magnitude of forcing required for this shift could be accomplished by the LGM veg-
 661 etation change alone. For example, low $\Delta\delta^{18}O$ values (along with high runoff (Nace et
 662 al., 2014)) are a persistent feature in the western domain for at least ~ 20 kyr before the
 663 LGM (Fig. S10), suggesting the cause of a westward shift in rainout is not unique to this
 664 time interval. African dust fluxes were persistently high from 40-20 ka, consistent with
 665 a remote albedo forcing, but data for other possible drivers of precipitation centroid mi-
 666 gration, such as the strength of the easterlies, is sparse at this time.

667 Based on the zonal, meridional, and hysteresis-like migration of the South Amer-
 668 ican precipitation centroid, we suggest that multiple forcing mechanisms operate at dif-
 669 ferent times to drive these complex, precession-scale patterns. Remote land albedo change
 670 could play a particularly important role in driving zonal shifts in rainout, but more so-
 671 phisticated climate model simulations are needed to rigorously test these hypotheses. We
 672 note that Heinrich and Dansgaard/Oeschger events are also linked to remote forcing of
 673 tropical South American precipitation (Arz et al., 1998; Nace et al., 2014; Kanner et al.,
 674 2012), but these shorter, millennial-scale events are beyond the scope of this study.

675 5.4 Air-mass and seasonality complications

676 Up to now, our interpretation of a zonal shift in the precipitation centroid has hinged
 677 on the assumption that the three speleothem sites are isotopically connected through time.
 678 Here, we discuss how relaxing this assumption does not necessarily invalidate our con-
 679 clusions. This assumption warrants scrutiny because, at least in the eastern-to-central
 680 record domain, the timing of peak precipitation and the relevant air-mass can differ be-
 681 tween sites (Fig. S4; Garreaud et al. (2009); Liu and Battisti (2015)) suggesting the iso-
 682 topic connection may not be strong through time.

683 For simplicity, we consider two forms of isotopic connectivity between sites: (1) an
 684 ‘air-mass connection’, where the same air mass rains out at both sites; and (2) a ‘recy-
 685 cling connection’ where rainout at one site evaporates and re-precipitates at the other.
 686 An air-mass connection implies a recycling connection and allows upwind isotopic sig-
 687 nals to propagate fully downwind. When two sites share only a recycling connection, each
 688 site is dominated by a different air-mass and the upwind signal propagates downwind
 689 with some attenuation due to air-mass mixing. While changes in circulation might al-
 690 ter the strength of the air-mass connection across tropical South America through time,
 691 the recycling connection is likely more robust. For example, about half of western Ama-
 692 zon rainfall is derived from upwind recycling (Zemp et al., 2017; Staal et al., 2018) and
 693 upwind (northeastern Brazil) transpiration can travel over 1000 km before re-precipitating
 694 out, connecting the east and west (Staal et al., 2018).

695 First, we note that if two sites share an air-mass and recycling connection, then the
 696 mean isotope gradient is insensitive to differences in the timing of peak precipitation be-
 697 tween them. We demonstrate this point with a toy model that simulates the isotope gra-
 698 dient between two sites that have different phases of precipitation seasonality (Supple-
 699 mental Text S5). Differences in the seasonal phase lead to $\Delta\delta^{18}O$ errors (relative to the
 700 in-phase case), but these errors are negligible—less than 1% of the seasonal $\Delta\delta^{18}O$ am-
 701 plitude (Figs. S11 and S12). Thus, a difference in the timing of peak precipitation—as
 702 between the eastern and central sites—does not, itself, invalidate the $\Delta\delta^{18}O$ framework.
 703

704 Still, our conclusions could be impacted if the air-mass connection between sites
 705 is weak at some point in time. We evaluate two additional additional scenarios that ad-
 706 dress this possibility. First, we assume there is a weak air-mass connection between the
 707 eastern and central sites. In this case, the high- $\delta^{18}O$ values at these sites during the LGM
 708 indicate that their air-masses are delivering undistilled moisture. The precipitation centroid—
 709 where moisture distillation is strongest—is likely situated to the west. Toward the mid-
 710 Holocene, eastern and central $\delta^{18}O$ decrease in tandem, with no evidence for the down-
 711 wind attenuation that is expected if upwind recycling was mixing with an independent
 712 air-mass. Perhaps $\delta^{18}O$ of the central site’s air-mass also decreased from the LGM to
 713 mid-Holocene, hiding the attenuation. In this case, lest we invoke a third air-mass some-
 714 how, the central signal should propagate west. Wang et al. (2017) explain the lack of west-
 715 ern signal by invoking an increase in plant transpiration, but this mechanism has been
 716 discredited (Pattnayak et al., 2019; Ampuero et al., 2020).

717 In the alternate scenario, we assume the eastern and central sites share an air-mass
 718 connection with each other, but not the western site. In this case, the eastern and cen-
 719 tral $\delta^{18}O$ shift from the LGM to mid-Holocene makes sense, however it is not consistent
 720 with the lack of a signal in the west. The signal could be masked by a coincident decrease
 721 in the western air-mass’s rainout, but such a decrease may also be related to a zonal shift
 722 in the precipitation centroid. Still, western $\delta^{18}O$ shifts directions to track the central site
 723 as soon as $\Delta\delta^{18}O$ reaches the theoretical maximum value for two sites with a strong air-
 724 mass connection. Without a strong air-mass connection, this western $\delta^{18}O$ shift and the
 725 central and eastern $\delta^{18}O$ decrease after the LGM must be somewhat coincidental, driven
 726 by coeval changes in independent air-masses that happen to cancel out the attenuation
 727 of the upwind signal while obeying the theoretical maximum $\Delta\delta^{18}O$ of a single air-mass
 728 system. Arguments against a strong air-mass connection should address how these ap-
 729 parently unattenuated signals occur in the speleothem data. Overall, we argue that un-
 730 certainty in the strength of the air-mass connection makes our quantitative precipita-
 731 tion reconstruction less certain, but it does not conflict with our zonal precipitation cen-
 732 troid migration hypothesis.

733 6 Conclusion

734 Our analysis provides a path forward for resolving the enigmatic, non-uniform trends
 735 in tropical South American speleothem $\delta^{18}O$, but it rests on assumptions, many previ-
 736 ously discussed, that deserve further scrutiny. One critical assumption that is difficult
 737 to address is that speleothem $\delta^{18}O$ reliably tracks precipitation $\delta^{18}O$ at all sites. Kinetic
 738 fractionation and other confounding processes could decouple speleothem and precip-
 739 itation $\delta^{18}O$, challenging our model approach. Such effects have not been documented
 740 in these speleothems (van Breukelen et al., 2008; Cruz et al., 2009; Cheng et al., 2013;
 741 Wang et al., 2017), but additional proxy constraints (such as triple oxygen and mass-
 742 48 clumped isotopes) will provide more rigorous tests of local and kinetic effects (Huth
 743 et al., 2022). Another limitation lies in the simplified energy balance modeling approach.
 744 The goal of these model exercises is to present plausible drivers of zonal rainout shifts
 745 for further testing, while recognizing that our list of drivers is not exhaustive. Future stud-
 746 ies of the SAPD with more sophisticated models should analyze the zonal location of the
 747 energy flux prime meridian and its relation to zonal precipitation patterns to test whether
 748 this zonal precipitation centroid migration effect is present.

749 It is also fair to question whether the discrepancies between proxies and isotope-
 750 enabled GCMs are resolvable, as we posit. GCMs are known to struggle with tropical
 751 South American precipitation—there is substantial inter-model spread and dry-bias in
 752 seasonal and annual rainfall that complicate their application to exotic, paleoclimate states
 753 (Li et al., 2006; Ribas et al., 2022). However, these models show better agreement in their
 754 simulated precipitation change, and their precipitation biases do not appear to cause bi-
 755 ases in net rainout (see Fig. 5c). We also reiterate that, while we question whether pre-

756 cession without land albedo change sufficiently explains the late Quaternary SAPD, our
 757 work should not be taken to discredit the role of precession more generally. We expect
 758 that the spatial pattern and amplitude of $\delta^{18}O$ anomalies can vary from one precession
 759 cycle to the next, depending on how orbital forcing interacts with other forcings and feed-
 760 backs within the Earth system. A zonal shift in the precipitation centroid is not required
 761 to explain zonally opposing precipitation anomalies, but it helps explain certain features
 762 of the late Quaternary proxy data, including the zonally imbalanced amplitude of $\delta^{18}O$
 763 change and their notable phase-shifted trends. Our results build on previous work (Battisti
 764 et al., 2014) suggesting that zonal forcings may help explain some of the enigmatic proxy
 765 records found in places where tropical precipitation is energetically primed to migrate
 766 east-west.

767 Data Availability Statement

768 Code and data associated with this study can be found through Zenodo (Kukla et
 769 al., 2022) and Github (<https://github.com/tykukla/ZonalPrecipPatterns-Amazon>).
 770 The Zenodo/Github repository includes code and results for the energy balance and re-
 771 active transport model analysis, SISALv2 analysis, speleothem $\delta^{18}O$ data cleaning and
 772 smoothing, and the proxy compilation in Figure 1 of the main text. We note that the
 773 Tigre Perdido record (van Breukelen et al., 2008) from the western composite data was
 774 downloaded from the SISAL database (siteID: 25), while other speleothem records were
 775 provided by the original authors or taken from the supplementary materials of the rel-
 776 evant publication.

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