The Influence of Orbital Forcing of Tropical Insolation on the Climate and Isotopic Composition of Precipitation in South America

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ABSTRACT

The $\delta^{18}O$ of calcite ($\delta^{18}O_c$) in speleothems from South America is fairly well correlated with austral summer [December–February (DJF)] insolation, indicating the role of orbitally paced changes in insolation in changing the climate of South America. Using an isotope-enabled atmospheric general circulation model (ECHAM4.6) coupled to a slab ocean model, the authors study how orbitally paced variations in insolation change climate and the isotopic composition of precipitation ($\delta^{18}O_p$) of South America. Compared with times of high summertime insolation, times of low insolation feature (i) a decrease in precipitation inland of tropical South America as a result of an anomalous cooling of the South American continent and hence a weakening of the South American summer monsoon and (ii) an increase in precipitation in eastern Brazil that is associated with the intensification and southward movement of the Atlantic intertropical convergence zone, which is caused by the strengthening of African winter monsoon that is induced by the anomalous cooling of northern Africa. Finally, reduced DJF insolation over southern Africa causes cooling and the generation of a tropically trapped Rossby wave that intensifies and shifts the South Atlantic convergence zone northward. In times of low insolation, $\delta^{18}O_p$ increases in the northern Andes and decreases in northeastern Brazil, consistent with the pattern of $\delta^{18}O_c$ changes seen in speleothems. Further analysis shows that the decrease in $\delta^{18}O_p$ in northeastern Brazil is due to change in the intensity of precipitation, while the increase in the northern Andes reflects a change in the seasonality of precipitation and in the isotopic composition of vapor that forms the condensates.

1. Introduction

In this study, we examine the impact of orbitally forced changes in insolation on the climate and isotopic composition of precipitation over South America. Changes in Earth’s orbit around the sun cause quasi-periodic changes in insolation reaching the top of the atmosphere. Particularly important for tropical precipitation is precession, with predominant periods at 19,000 and 23,000 years, and the modification to precessional forcing by changes in the eccentricity of Earth’s orbit—hereafter defined as the “precessationally forced” or “orbitally forced” changes in insolation. The long and well-dated oxygen isotope records stored in the calcite of stalagmites ($\delta^{18}O_c$) render the precessional cycle an ideal target for understanding the sensitivity of climate and, in particular, the monsoon circulations. Under the right conditions, the $\delta^{18}O_c$ of stalagmites reflects the isotopic composition of the drip water, which reflects the time-integrated oxygen isotope concentration in precipitation, that is, precipitation-weighted $\delta^{18}O (\delta^{18}O_p)$. Figure 1 shows the normalized $\delta^{18}O_c$ record of speleothems from South America (see also Table 1) that are sufficiently long to resolve the precessional cycle (see section 2 for details of the speleothem data used in this study). Superposed on each isotope record is the December–February (DJF) mean insolation at 30°S normalized to have the same standard deviation of each record but unitless. The isotope records line up well with the time series of insolation, implying a strong driving force from insolation.

Cruz et al. (2005) examined high-resolution oxygen isotope records in stalagmites from southeastern Brazil. They suggested that the decrease in $\delta^{18}O_c$ at times of high Southern Hemisphere summer insolation is due to an increase in the seasonality of precipitation (an increase in summer precipitation) and to an intensified...
convective activity in the South American summer monsoon (SASM), depleting the heavier isotope in the vapor transported to southern Brazil. Cheng et al. (2013) studied a suite of stalagmites from the eastern flank of the Andes in northern Peru, where in today’s climate 80% of the total vapor that forms the precipitation originates from upstream over the Amazon basin and from the tropical Atlantic. They also suggested that precessional changes in the $\delta^{18}O$ of stalagmites primarily reflect changes in the fractionation efficiency of water vapor upstream and hence changes in the strength of SASM. Cruz et al. (2009) studied a 25 000-yr-long speleothem record from Nordeste, Brazil; they concluded that the precessional signals in the speleothem record indicate changes in the rate of precipitation during the SASM (called the “amount effect” in the proxy literature).

Although the $\delta^{18}O$ of speleothems is a measure of the $\delta^{18}O$ of aggregated precipitation ($\delta^{18}O_p$) falling at the cave sites, we cannot directly infer changes in precipitation from the $\delta^{18}O$ because the latter is affected by many factors, including changes in the seasonality of precipitation, changes in the intensity of precipitation (the amount effect; Lee and Fung 2008), and changes in the isotopic composition of the vapor that is available to form precipitation. We cannot fully address all these factors and thus interpret the climatological significance of the $\delta^{18}O$ of speleothems without additional proxy.

FIG. 1. Normalized $\delta^{18}O$ records of speleothems from South America (colored lines), and DJF insolation at 30°S normalized to have the same but unitless standard deviation as in each record (black line). Different colors for $\delta^{18}O$ of a given site represent different speleothems from the same cave. Note that except for Rio Grande do Norte (RN), the insolation is multiplied by −1 for ease in comparison to $\delta^{18}O$. The vertical red lines denote the summer insolation used in the high-insolation (207 ka BP) and low-insolation (218 ka BP) experiments with ECHAM4.6. (Details on the speleothems used can be found in Fig. 3 and Table 1.)

TABLE 1. Speleothems used in this study and featured in Figs. 1, 3, and 4. For each record, we note the cave location, elevation above mean sea level (MSL), duration of the record, and references for the data. Also noted is the amplitude of the precessional cycle. The amplitude of the precessional cycle is found by a linear regression of the measured $\delta^{18}O$ against DJF 30°S insolation and then scaled by the difference in insolation 218 minus 207 ka BP (which is 90 W m$^{-2}$); the 95% confidence interval is noted in parentheses. The amplitude of the precessional cycle is noted only if the $\delta^{18}O$ is correlated with DJF Southern Hemisphere insolation at $p \leq 0.05$.

<table>
<thead>
<tr>
<th>Cave and location</th>
<th>Duration (kyr)</th>
<th>Amplitude ($%)</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rio Grande do Norte (10°10’S, 40°50’W; 500 m MSL)</td>
<td>24</td>
<td>—</td>
<td>Cruz et al. (2009)</td>
</tr>
<tr>
<td>Pacupahuan Cave (11°14’S, 75°49’W; 3800 m MSL)</td>
<td>32</td>
<td>—</td>
<td>Kanner et al. (2012)</td>
</tr>
<tr>
<td>Santiago Cave (3°1’S, 78°8’W; 980 m MSL)</td>
<td>86</td>
<td>—</td>
<td>Mosblech et al. (2012)</td>
</tr>
<tr>
<td>Cueva del Diamante (5°44’S, 77°30’W; 960 m MSL)</td>
<td>243</td>
<td>1.0 (0.2, 1.6)</td>
<td>Cheng et al. (2013)</td>
</tr>
<tr>
<td>El Condor (5°56’S, 77°18’W; 860 m MSL)</td>
<td>53</td>
<td>—</td>
<td>Cheng et al. (2013)</td>
</tr>
<tr>
<td>Botucara (27°13’S, 49°10’W; 230 m MSL)</td>
<td>116</td>
<td>3.3 (1.9, 4.2)</td>
<td>Cruz et al. (2005) and Wang et al. (2007)</td>
</tr>
</tbody>
</table>
data and the aid of a climate model that incorporates the water isotopes. And only with a climate model can we understand the dynamical processes responsible for the changes in $\delta^{18}O_p$ and climate. In turn, the large-amplitude, orbitally paced, and accurately dated $\delta^{18}O$ oscillations in the cave records provide a good constraint by which we can evaluate the efficacy of a climate model’s response to external forcing. Our confidence in the model is enhanced if the simulated $\delta^{18}O_p$ is very close to $\delta^{18}O$, of speleothems. In this study, we examine the impact of orbital forcing (again, defined as the combined effects of precession and eccentricity) on South American climate, with the aid of the isotope- enabled model.

There are three possible ways that orbital forcing can impact the climate of South America: 1) by changing the heating of the Amazon basin (i.e., changing the South American summer monsoon), 2) by changing the heating of Africa, and 3) by changing the heating of the Atlantic or/and Pacific Oceans [i.e., by changing the sea surface temperature (SST) distribution]. Kutzbach and Guetter (1986) proposed that the orbitally paced insolation strengthens (weakens) the summer monsoon by enhancing (weakening) the land heating. Given the large land area of tropical South America, it is quite likely that orbital forcing directly changes the strength of South American summer monsoon and thus changes the climate and $\delta^{18}O_p$ over tropical South America. On the other hand, Cook et al. (2004) proposed that Africa influences the climate of South America by way of an “intercontinental teleconnection.” They found that heating over Africa decreased the precipitation over northeastern Brazil through low-level moisture divergence and dry air advection. Paleoclimate data and archeological evidence have shown that changes in Earth’s orbit pace the climate of Africa (deMenocal and Tierney 2012). It is therefore possible that orbitally forced changes over Africa affect the climate of South America through this teleconnection. Finally, numerous studies have pointed out that changes in spatial distribution of SST in the tropical Atlantic and Pacific have a profound impact on the climate of tropical South America (Moura and Shukla 1981; Mechoso et al. 1990; Nobre and Shukla 1996; Dettinger et al. 2001). For example, Moura and Shukla (1981) proposed that anomalously warm water in the northern tropical Atlantic and/or cold water in the southern tropical Atlantic are responsible for severe droughts over northeastern Brazil by way of a thermally direct local Hadley circulation response. Hence, it is also possible that insolation-forced changes in the distribution of tropical SST impact the climate of tropical South America.

It is worth noting that these three processes are not mutually exclusive. For any given region, orbitally forced changes in local precipitation may be a result of the combined work of all these three processes. To illuminate the relative importance of these three processes, we will perform a unique set of experiments that are described in detail in section 5.

The paper is organized as follows. We describe in section 2 the speleothem data, the instrumental data, and the climate model that are used in this study, as well as the two core experiments. We also describe the decomposition method for determining the cause(s) of the changes in $\delta^{18}O_p$ simulated by the model. Section 3 presents the results from the experiments. Section 4 illuminates the causes of the changes in $\delta^{18}O_p$ in the experiments. In section 5, we identify the causes of the changes in precipitation. A discussion and conclusions are presented in sections 6 and 7, respectively.

2. Data, methods, and the core experiments

a. Experimental design

The ECHAM atmospheric general circulation model, version 4.6 (ECHAM4.6; Roeckner 1996), is used in the study. A water isotope module that accounts for the fractionation processes related to evaporation and condensation is included in ECHAM4.6 [more details can be found in Hoffmann et al. (1998)]. The model is run with spectral T42 resolution (approximately 2.8° horizontal resolution in latitude and longitude) and 19 vertical levels, and is coupled to a 50-m slab ocean. A cyclostationary heat flux ($q$ flux) is added to the slab ocean to account for ocean heat flux convergence by ocean currents and for biases in the surface heat flux because of biases in the atmospheric model. The $q$ flux is constructed in a three-step process: first, the observational climatological mean SST is used to force ECHAM4.6; then the surface flux output from the atmospheric component of ECHAM4.6 is used to force the uncoupled (offline) slab ocean; finally, the differences between the SST output from the offline slab ocean simulation and the observed SST are used to construct the cyclostationary heat flux that is added to the freely evolving coupled model.

We performed two core experiments with ECHAM4.6 to quantitatively study the effect of orbital forcing on climate and the isotopic composition of precipitation in South America. In the first experiment, called the high-insolation experiment, we used the insolation value of 207 000 years before present.
In the second experiment, called the low-insolation experiment, the insolation of 218 ka BP is used. These two model runs also formed the core experiments used by Battisti et al. (2014) to examine the Asian monsoon. Insolation of these two years represents respectively the minimum and maximum of Northern Hemisphere summer insolation over the past 950,000 years. These two years are also years of near-maximum and near-minimum summer insolation in the Southern Hemisphere, respectively.

Figure 2 shows the differences in the top-of-atmosphere insolation between low-insolation and high-insolation experiments. Compared to the high-insolation experiment, the low-insolation experiment has more insolation in June–August (JJA), but less insolation in DJF. This insolation deficit in DJF peaks in the high latitudes of Southern Hemisphere.

The same boundary conditions (360 ppm CO₂ and modern-day continental geometry, orography, and ice sheets) and the same climatological seasonal cycle of heat flux are used in both experiments: in effect, we do not take into account any changes in ocean heat flux convergence that would arise because of changes in the ocean advection or mixing. All experiments discussed in this study are run for 30 years, with output of the last 20 years used to construct climatologies and climatological differences; all differences discussed in this paper are statistically significant at p = 0.05 or better.

It is worth mentioning that these two years, 207 and 218 ka BP, are chosen to amplify the orbital signal and to avoid the influence of glaciations (both years fall into the interglacial periods). However, the response we have documented in the paper is not special to the particular time periods chosen for the experiments. A set of mid-Holocene (6 ka BP) and modern-day experiments was also performed (cf. section 6b); the differences in climate and isotopes between these two experiments scale nearly linearly with summer-averaged insolation (when compared to the difference between 218 and 207 ka BP). Furthermore, nearly the same climatic and isotopic responses are obtained in the high- and low-insolation experiments when the boundary conditions are replaced by those of the Last Glacial Maximum (200 ppm, Last Glacial Maximum ice sheets and continental geometry) (not shown).

b. Proxy data

The speleothem data we used to evaluate the model’s performance were obtained from the NOAA paleoclimatology website (http://www.ncdc.noaa.gov/paleo/speleothem.html). Figure 3 shows sites of the caves where the speleothems are from: Rio Grande do Norte in northeastern Brazil (Cruz et al. 2009), El Condor and Cueva del Diamante in northern Peru (Cheng et al. 2013), Santiago Cave in Ecuador (Mosblech et al. 2012), Pacupahuain Cave in Peru (Kanner et al. 2012), and Botuverá in southern Brazil (Cruz et al. 2005, Wang et al. 2007). These caves concentrate around three regions of South America: along the eastern flank of the Andes, in northeastern Brazil, and in southeastern Brazil.

Also shown in Fig. 3 is the modern-day climatological precipitation for DJF (these data are available at http://www.esrl.noaa.gov/psd/data/gridded/data.cmap.html). Three precipitation centers can be identified in Fig. 3: a narrow zonally oriented band over the tropical North Atlantic Ocean, a broad region over continental South America, and a broad northwest–southeast-oriented band over the subtropical Atlantic Ocean. The narrow tropical band over the ocean is called the Atlantic intertropical convergence zone (ITCZ) and denotes the convergence of the trade winds from both hemispheres. It migrates north–south with the seasonal cycle of insolation but stays north of the equator all year around (Waliser and Gautier 1993; Philander et al. 1996). The broad region of precipitation over
continental South America is known as the South American summer monsoon (Zhou and Lau 1998; Vera et al. 2006). It starts in early spring in northern South America, peaks to the south in austral summer, and retreats in austral autumn. The subtropical rainband, known as the South Atlantic convergence zone (SACZ), is due to the passage of extratropical transient frontal systems and to mean low-level convergence (Kodama 1993). It weakly migrates northward from austral winter to austral summer. A detailed discussion of these mean field features can be found in Garreaud et al. (2009).

From a climate perspective, caves in the northern Andes are just west and thus downwind of the SASM. Hence, speleothems from these caves are affected by changes in the SASM, including the seasonal migration and strengthening/weakening of the SASM. The cave site in northeastern Brazil is influenced in austral summer by the westernmost tail of the ITCZ and the easternmost edge of the SASM; it records changes in both the ITCZ and the SASM. The cave site in southern Brazil is on the border between the SASM and the SACZ, so it records climate changes in the SACZ as well as the SASM.

c. Decomposition method

The $\delta^{18}O_p$ is defined as $\sum \delta^{18}O_P / \sum P$, where $\delta^{18}O_P$ is monthly isotopic composition of precipitation, and $P$ is monthly precipitation.¹ Thus, the difference in $\delta^{18}O_p$ between the low-insolation and high-insolation experiment is

$$\delta^{18}O_{p,218} - \delta^{18}O_{p,207} = \frac{\sum \delta^{18}O_{P,218} P_{j,218}}{\sum P_{j,218}} - \frac{\sum \delta^{18}O_{P,207} P_{j,207}}{\sum P_{j,207}}.$$  (1)

Changes in $\delta^{18}O_p$ may result from changes in several local or nonlocal processes. To quantitatively address all possible factors, we decompose the total changes in $\delta^{18}O_p$ into two parts: 1) those resulting from changes in the amount of monthly precipitation and 2) those resulting from changes in the monthly isotopic composition of precipitation; the latter may be due to changes in precipitation intensity or to changes in the isotopic composition of the water vapor that is condensed to form the precipitate.

¹ The traditional definition of the climatological $\delta^{18}O$ features the ratio of the moles of $^{18}O$ molecules to the moles of $^{16}O$ molecules. The precipitation weighted formulation $\delta^{18}O_p$ is an excellent approximation of the climatological $\delta^{18}O$ [errors $O(0.2)$%; see Battisti et al. 2014] that provides additional insight on the processes and seasonality that contribute to the climatological $\delta^{18}O$. 

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The importance of changes in the seasonality of precipitation to the changes in $\delta^{18}O_p$ is indicated by

$$\frac{\sum_j \delta^{18}O_{j,207}P_{j,218}}{\sum_j P_{j,218}} - \frac{\sum_j \delta^{18}O_{j,207}P_{j,207}}{\sum_j P_{j,207}}. \quad (2)$$

The importance of changes in $\delta^{18}O$ of precipitation to the changes in total $\delta^{18}O_p$ is indicated by

$$\frac{\sum_j \delta^{18}O_{j,218}P_{j,207}}{\sum_j P_{j,207}} - \frac{\sum_j \delta^{18}O_{j,207}P_{j,207}}{\sum_j P_{j,207}}. \quad (3)$$

It is worth noting that the results from Eqs. (2) and (3) do not sum to the total changes in $\delta^{18}O_p$ [Eq. (1)] because of the inherent nonlinearity in the definition of $\delta^{18}O_p$.

3. Results

The differences in $\delta^{18}O_p$ between the low-insolation and high-insolation experiments (low insolation minus high insolation) are displayed in Fig. 4. Also plotted in Fig. 4 are the estimates of the differences in $\delta^{18}O_c$ from the speleothems, linearly scaled by the difference in summer insolation, 218 minus 207 ka BP. Our experiments were performed using the extreme values of summer insolation, last realized 218 000 and 207 000 years ago. The cave records do not extend this far back. Hence, we estimate the differences in $\delta^{18}O_c$ at the cave sites between 218 and 207 ka BP using linear fit of the $\delta^{18}O_c$ of speleothems to the 30°S DJF insolation, and then scale the regression value to reflect the changes in DJF insolation between 218 and 207 ka BP (90 W m$^{-2}$). The scaled differences in $\delta^{18}O_c$ between 218 and 207 ka BP are +3‰ at Botuverá in southeastern Brazil and +1‰ at Cueva del Diamante in the northern Andes. Records from the other three sites are too short to have a statistically significant (at $p = 0.05$) correlation with insolation. The model captures the gross dipole pattern indicated by the cave $\delta^{18}O_c$. It also gives the same quantitative estimates over the northern Andes. Over southeastern South America, the model underestimates the $\delta^{18}O_c$ change. The pattern of our simulated changes in $\delta^{18}O_p$ agrees with that of Cruz et al. (2009), who use the same GCM driven by modern sea surface temperature but set the insolation forcing to be 6 and 0 ka BP, respectively, which are also a minimum and a maximum of DJF insolation, but of smaller amplitudes.

Figure 5a shows the differences in precipitation; superposed in contours is the DJF mean precipitation of the high-insolation experiment. Compared to the high-insolation experiment, precipitation in low-insolation experiment decreases by over 1 mm day$^{-1}$ over the SASM region, indicating a weakening of summer monsoon. Precipitation increases greatly in an arc along the eastern coast of Brazil and offshore over the ocean, where the DJF mean precipitation increases from 5 mm day$^{-1}$ in the high-insolation experiment to over 13 mm day$^{-1}$, indicating a southward shift and intensification of the ITCZ and an intensification and northward shift of the SACZ. This northward shift of SACZ and southward shift of ITCZ constitutes a shrinking of the dry zone over the subtropical South Atlantic associated with the intensification of the subtropical high as a result of the anomalous cooling of southern Africa (see section 5). Curiously, this kind of dry zone shrinking is also seen in the Pacific when an El Niño event happens (Yuileava and Wallace 1994).

Figure 5b shows differences in temperature between the low-insolation and high-insolation experiments. The decrease in precipitation in the Amazon basin is accompanied by a slight warming, whereas the increase in precipitation in eastern Brazil is accompanied by a large

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2 Unless otherwise noted, “differences” discussed in the paper and plotted in the figures will always refer to low insolation minus high insolation.
cooling (by over 3 K). South of the SACZ, the decrease in precipitation again is accompanied by an increase in temperature (by over 1 K). These changes in temperature are consistent with the changes in clouds that are associated with changes in precipitation: the decrease in precipitation over monsoon region in the central Amazon basin is associated with fewer clouds and therefore more absorbed solar radiation, which more than offsets the cooling caused by decrease in local summer insolation. Similarly, more precipitation over eastern Brazil is associated with more clouds and thus less absorbed solar radiation, which adds to the orbitally induced cooling and makes the surface air temperature even lower.

In JJA, an increase in local heating (Fig. 2) causes precipitation to increase by over 3 mm day$^{-1}$ over northern Andes and western Amazon basin, but change little elsewhere over the South American continent (see Fig. 5c). Correspondingly, the surface air cools slightly over northern Andes and western Amazon basin, but warms up by over 3 K over the majority of South American continent (Fig. 5d). This once again implies the importance of clouds in changing the surface air temperature.

4. Analysis of isotopic responses

The decomposition method shown in Eqs. (2) and (3) is used to quantitatively illuminate the reason for changes in $\delta^{18}O_p$. Using the monthly $\delta^{18}O$ from high insolation and the monthly precipitation of low insolation in the calculation of $\delta^{18}O_p$ [Eq. (2)], we isolate the differences in $\delta^{18}O_p$ caused by changes in the seasonality of precipitation; the result is displayed in Fig. 6c. Similarly, using the monthly precipitation from high insolation and the monthly $\delta^{18}O$ from low insolation, we...
calculate the contribution of the changes in the isotopic composition of precipitation to the changes in $\delta^{18}O_p$ (Fig. 6d). Comparing Figs. 6c and 6d with Fig. 6a, we see that over the northern Andes changes in both $\delta^{18}O$ and the seasonality of precipitation contribute to the total change in $\delta^{18}O_p$. However, over northeastern Brazil, the change in $\delta^{18}O_p$ is largely due to the change in the $\delta^{18}O$ of precipitation.

We apply a similar decomposition method to examine the relative contribution of winter and summer changes in precipitation and $\delta^{18}O$ to the change in $\delta^{18}O_p$. Specifically, in the calculation of $\delta^{18}O_{p,\text{DJF}}$, we use the monthly precipitation and $\delta^{18}O$ values in DJF from the low-insolation experiment and use the values from the high-insolation experiment for all the other months. The differences between this partial $\delta^{18}O_p$ and $\delta^{18}O_{p,\text{DJF}}$ are displayed in Fig. 6b. Over the northern Andes, the change in $\delta^{18}O_p$ is apparently due to changes in both DJF and other seasons (cf. Figs. 6a and 6b) and involves changes in both the precipitation and the $\delta^{18}O$ of precipitation. Over northeastern Brazil, almost all the change in $\delta^{18}O_p$ is as a result of changes in the $\delta^{18}O$ of
The above results can be further understood by examining the seasonal cycle of precipitation and the $\delta^{18}O$ of precipitation in the low-insolation and high-insolation experiments. Figure 7 displays the seasonal cycles of the area-averaged precipitation and $\delta^{18}O$ of precipitation over northeastern Brazil and over the Andes. The regions used to calculate the area averages are shown in Fig. 4. We chose the boxes to encompass the regions of greatest change in simulated $\delta^{18}O_p$. Nonetheless, changes in the annual cycle of precipitation and $\delta^{18}O$ of precipitation of the cave sites are qualitatively similar to those in the boxes (not shown).

Before we present results for the impact of orbital changes, we first briefly discuss the ability of the model to simulate the modern climate. The seasonal cycle of precipitation over the Andes in the modern-day period is nearly identical to that observed (cf. black solid and dashed lines in Fig. 7b). Over northeastern Brazil, however, summer monsoon precipitation in the modern-day simulation is greater than that observed, and peaks one month too early (cf. black solid and dashed lines in Fig. 7a). This bias is mainly due to a bias in the marine portion of the box; the seasonal cycle of precipitation over land in northeastern Brazil is nearly identical to that observed (not shown). The $\delta^{18}O$ of precipitation during the rainy season in the modern-day simulation is

**Fig. 7.** (a) Area-averaged precipitation (mm day$^{-1}$) and (c) $\delta^{18}O$ (‰) in northeastern Brazil ($5^\circ$–$15^\circ$S, $40^\circ$–$30^\circ$W) in the high-insolation (blue) and low-insolation (red) experiments. (b),(d) As in (a),(c), but for the eastern flank of the Andes ($5^\circ$–$15^\circ$S, $75^\circ$–$65^\circ$W). Also plotted in (a),(b) are the area-averaged precipitation for the modern-day simulation (black dashed) and from observations (black solid). The observational data are from CMAP (Xie and Arkin 1997), from January 1979 to December 2010.
also heavier than that observed, which might be as a result of the inadequate rainout of the air mass when moving inland from the eastern Amazon basin (Vuille et al. 2003). The bias in the simulated seasonal cycles of precipitation and the \( \delta^{18}O \) of precipitation might cause the model to overestimate the contribution of insolation-driven changes in DJF precipitation to the changes in \( \delta^{18}O_p \) in northeastern Brazil.

**a. Change in \( \delta^{18}O_p \) over northeastern Brazil**

Compared with the high-insolation experiment, the low-insolation experiment has much more precipitation in DJF but almost the same amount in JJA. More precipitation in DJF in low-insolation experiment has little effect on the precipitation-weighted \( \delta^{18}O \), however, because there is little precipitation in winter compared to summer; hence, changes in the amount of summer precipitation do not affect the seasonality of precipitation. This is consistent with the decomposition analysis shown in Figs. 6a and 6c: changes in the seasonality of precipitation contribute little to the total change in \( \delta^{18}O_p \) over northeastern Brazil. Figure 7c shows that there is a large decrease in \( \delta^{18}O \) of precipitation in DJF of the low-insolation experiment; this accounts for the majority of the changes in \( \delta^{18}O_p \) because the bulk of the annual precipitation occurs in DJF in both the low-insolation and high-insolation simulations. Finally, to determine why the \( \delta^{18}O \) of summertime precipitation changes, we calculate the cumulative probability distribution function for daily precipitation over northeastern Brazil (Figs. 8a,c). The averaging region used in this calculation is the same as that shown in Fig. 4. The

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**FIG. 8.** Cumulative probability distribution of daily precipitation for (a),(c) northeastern Brazil and (b),(d) the northern Andes in the high-insolation (blue) and low-insolation (red) experiments in (top) austral winter (JJA) and (bottom) austral summer (DJF).
intensity of precipitation changes very little in JJA. In DJF, however, corresponding to the more negative $\delta^{18}O$, a greater fraction of precipitation is falling as heavy precipitation in low insolation compared to high insolation. This out-of-phase relationship between changes in the intensity of precipitation and changes in $\delta^{18}O$ of precipitation suggests that the amount effect (Lee and Fung 2008) is acting to deplete the $\delta^{18}O$ of precipitation in northeastern Brazil in DJF in the low-insolation experiment relative to the high-insolation experiment.

The more intense precipitation in DJF over northeastern Brazil is associated with an increase in near-surface moist static energy. Figure 9a shows the differences in moist static energy for DJF. Compared to the high-insolation experiment, the low-insolation experiment moist static energy increases south of the equator, including over northeastern Brazil, and decreases north of the equator over the tropical Atlantic, consistent with the southward movement of ITCZ (Fig. 5).

Therefore, we can conclude that the decrease in the $\delta^{18}O$ over northeastern Brazil for low insolation compared to high insolation is because of a decrease in the $\delta^{18}O$ of precipitation in summer, which appears to be as a result of an increase in the intensity of precipitation associated with a southward shift of ITCZ. The cause of this southward shift of ITCZ will be discussed in section 5.

b. Change in $\delta^{18}O_p$ along the Andes

In this section, we perform the same analysis as in section 4a, only for the northern Andes region. Figure 7b shows that low insolation has less precipitation than high insolation in the wet season (DJF), but more precipitation than high insolation in the dry season (JJA). Thus, a larger portion of annual precipitation for low insolation comes in the dry season. The weakening of seasonality of precipitation contributes to a higher $\delta^{18}O_p$ in the low-insolation experiment because the $\delta^{18}O$ of precipitation is more depleted in DJF compared to JJA in both the low-insolation and high-insolation experiments (Fig. 7d). Compared to high insolation, $\delta^{18}O$ of precipitation for low insolation also increases in DJF but changes very little in JJA (Fig. 7d). Since DJF is the wet season in both the low-insolation and high-insolation experiments, this change in DJF $\delta^{18}O$ also contributes significantly to the precessional changes in $\delta^{18}O_p$. Hence, unlike in northeastern Brazil, in the northern Andes changes in both the seasonality of precipitation and the $\delta^{18}O$ of precipitation contribute to the net change in $\delta^{18}O_p$. This is consistent with the decomposition analysis result shown in Figs. 6c and 6d.

Figures 8b and 8d show the cumulative probability distribution function for daily precipitation over the northern Andes. For JJA, daily precipitation is more intense in the low-insolation experiment than in the high-insolation experiment, as evident from the cumulative probability distribution of precipitation (Fig. 8b) and the changes in moist static energy (Fig. 9b). This suggests that the vapor condensed on site to form precipitation must be heavier to offset the decrease in $\delta^{18}O$ associated with more intense precipitation. In DJF, the intensity of daily precipitation changes very little, but the $\delta^{18}O$ of precipitation increases a lot (see Figs. 8d and 7d, respectively). Thus, the amount effect cannot
account for the simulated changes in $\delta^{18}$O in DJF. Less precipitation, coupled with very little change in the distribution of intensity, implies that the decrease in precipitation in DJF is as a result of a uniform reduction in the frequency of all types of precipitation; convection strength changes little. Thus, the heavier $\delta^{18}$O of the precipitation in DJF is due to heavier vapor imported into the region from the east, which is associated with the reduced intensity of the SASM (Fig. 5).

5. Analysis of the climate changes

a. The relative roles of insolation forcing in South America, Africa, and the oceans

To identify the essential processes that cause the orbitally induced changes in precipitation, we have performed a series of experiments that isolate the impact of the changes in insolation in selected regions: tropical South America, northern Africa, and southern Africa. Since changing local insolation would require significant recoding of the model, we approximate the changes in local insolation forcing by changing the local surface albedo. Starting from the boundary conditions of the high-insolation experiment, we mimic the changes in the local absorption of insolation in DJF in the low-insolation experiment by increasing the surface albedo by a factor of 2.5 (the a priori estimate of the appropriate scaling factor is presented in the appendix). These experiments are not perfect analogs to the local DJF forcing in the low-insolation experiment, however, because there are differences in the vertical distribution of forcing changes, and also because changing the surface albedo gives a year-round reduction in shortwave forcing whereas orbital forcing decreases the DJF insolation while increasing the springtime heating in the Southern Hemisphere. Nonetheless, so long as we focus on the SH summer season and changes in SST are not important [see section 5a(3) below and Fig. 11], the differences in the sign of forcing in JJA has little effect on the DJF response, and the response to the “increasing albedo” experiments serves as a reasonable proxy for the response to the low-insolation forcing in DJF because land has a sufficiently low heat capacity.

We first perform two fixed SST experiments to make sure that the fixed SST experiments reproduce closely the slab ocean experiments. In the first experiment, called highInsolation_highSST, we use insolation and SST output of the high-insolation simulation. In the second experiment, we use insolation and SST output of the low-insolation simulation. As expected, the difference in precipitation (and isotopes) between these two experiments (not shown) is nearly identical to the difference between the low-insolation and high-insolation experiments using the slab ocean model.

The highInsolation_highSST experiment thus serves as the reference experiment for all of the increasing albedo experiments discussed below. These three experiments involve increasing surface albedo by a factor of 2.5 (i) over South America only, (ii) over Africa only, and (iii) over both South America and Africa. These three experiments are called the “2.5x South America alone,” “2.5x Africa alone,” and “2.5x both South America and Africa” experiments, respectively. These experiments isolate the impact of DJF insolation changes over specific regions on the climate and isotopic composition of precipitation in South America.

1) ROLE OF SOUTH AMERICA

Figure 10b shows the differences in precipitation between 2.5x South America alone and highInsolation_highSST experiments. Keeping in mind that increasing albedo by a factor of 2.5 does not give exactly the same forcing as in the low-insolation experiment, we focus our discussion on the gross spatial pattern of the changes and not the detailed changes. Increasing albedo over South America decreases precipitation over all of the South American continent north of 40°S, as well as over the subtropical ocean south of the climatological mean SACZ. It also slightly increases precipitation over the tropical North Atlantic and to the north of SACZ. Compared with the differences between low-insolation and high-insolation experiments (Fig. 10a), increasing albedo over South America alone reproduces the decreases of precipitation over tropical and subtropical South America and the Andes (SASM) and weakly reproduces the northward shift of SACZ. However, it does not account for the increases in precipitation along and offshore of the eastern coast of Brazil. It also tends to move the ITCZ to the north, whereas the ITCZ is shifted southward in the low-insolation phase when compared to the high-insolation phase (low-insolation minus high-insolation experiments).

Therefore, the weakening of SASM in the low-insolation experiment is related to weakening of the heating of South American continent, which is caused by a reduction of insolation in the low-insolation experiment. These changes in the SASM are reflected in the changes in near-surface moist static energy (Fig. 9a). This weakening of land heating significantly cools the troposphere, lowers the moist static energy, and decreases the monsoon convection intensity.

2) ROLE OF AFRICA

Increasing albedo over Africa alone reproduces the precipitation increase along the eastern coast of Brazil
and south of the equator over the tropical Atlantic (cf. Figs. 10a and 10c). It also reproduces the northward shift and intensification of SACZ. However, it does not reproduce the precipitation decreases seen north of 40°S over the South American continent. Also worth noting is that the 2.5x Africa alone experiment reproduces the basinwide southward shift of the tropical Atlantic ITCZ. This implies that, independent of any changes in SST, cooling over Africa is an important driver for the orbitally driven shift of ITCZ in the Atlantic and the changes in the SACZ.

Increasing the surface albedo over both continents basically reproduces the precipitation changes over the whole South America, as well as the shifts in the ITCZ and SACZ (cf. Figs. 10a and 10d). The precipitation changes that result from increasing albedo over both Africa and South America are very similar to the sum of the precipitation changes from the experiment where the albedo is increased over each continent separately: the result in Fig. 10d is very close to the sum of Figs. 10b and 10c (not shown). In other words, the response is nearly linear.

3) ROLE OF OCEAN TEMPERATURE CHANGE

The only discrepancy between the orbitally forced DJF precipitation changes (Fig. 10a) and those resulting from the changes in albedo over both South America and Africa (Fig. 10d) is just north of the equator in the Atlantic Ocean. This is demonstrated by performing one additional experiment with ECHAM4.6. In this experiment, called highInsolation_lowSST, we use high insolation, but SST from the low-insolation experiment. The differences in precipitation between this experiment and the highInsolation_highSST experiment are due to differences in SST only, and are shown in Fig. 11. Note that the SST differences are due to both differences in in situ insolation and enhanced cold air advection by the mean northeasterly trade winds associated
with the anomalously cool Sahara in DJF (because of a deficit in insolation for the low-insolation experiment compared to high-insolation experiment). The differences in precipitation resulting from insolation-induced changes in SST are confined to a small region northeast of Brazil. Hence, drawing from the results of sections 5a(1) and 5a(2), we conclude that orbitally forced changes in precipitation over South America (and Africa) are caused solely by changes in the heating of land.

To sum up, we conclude that insolation impacts the summer precipitation over South America through two processes: 1) an insolation deficit that directly weakens the South American summer monsoon and decreases precipitation throughout inland South America by decreasing the near-surface moist static energy that drives convection in the SASM) and 2) an insolation deficit that weakens the heating of Africa, which indirectly increases the precipitation over the eastern edge of South America via teleconnected atmospheric changes (see below). Finally, the influence of ocean changes is very small and limited to a small oceanic region north of Brazil.

b. How does Africa influence South American climate?

The experiments in section 5a demonstrate that the changes in precipitation along the eastern coast and offshore of Brazil are mainly due to changes in the heating of Africa. Cook et al. (2004) added the entire African continent to an aquaplanet model and studied the influence of Africa on precipitation to the west of Africa. They concluded that Africa influences the climate of South America through a local Walker circulation response to the heating of southern Africa in January. In this section, we perform additional experiments to illuminate the relative importance of insolation changes over southern and northern Africa. We do this by proxy, first increasing the surface albedo of northern Africa by a factor of 2.5 and then increasing the albedo of southern Africa by a factor of 2.5; in both experiments, the SST is prescribed to be that from the high-insolation slab experiment. Results are then compared to those from the reference experiment, highInsolation_highSST. Results are shown in Fig. 12 and discussed below.
1) ROLE OF NORTHERN AFRICA

Figure 12a shows that increasing albedo over northern Africa alone intensifies and shifts the tropical Atlantic ITCZ southward; it has no influence on precipitation along the eastern coast of South America. This is associated with the low-level wind difference (Fig. 13b). In the control run (Fig. 13a), northeasterlies associated with the northern African winter monsoon dominate the circulation of northern Africa. These northeasterlies reach as far south as 5°N. Increasing albedo over northern Africa strengthens the northeasterlies, which now reach across the equator to 5°S (Fig. 13b). This strengthening of the winter monsoon shifts the ITCZ southward throughout the tropical Atlantic, increases the precipitation intensity in the western tropical Atlantic ITCZ, and hence decreases the $\delta^{18}O_p$.

2) ROLE OF SOUTHERN AFRICA

Increasing the albedo of southern Africa causes an increase in precipitation over eastern Brazil; it is also responsible for the intensification and northward shift in the SACZ that are seen in the differences between the low-insolation and high-insolation experiments (cf. Figs. 12b and 10a). It is also responsible for changes in precipitation over southern Africa, which reduce the local condensational heating of the troposphere and give rise to an anticyclonic Rossby wave to the west (Gill 1980). As shown in Fig. 13c, this anticyclone extends all the way to the eastern coast of South America. The low-level convergence associated with this Rossby wave is predominantly due to the zonal flow ($\partial u/\partial x$) near the equator and meridional flow ($\partial u/\partial y$) in the subtropics. Hence, the anomalous easterlies along the northern branch of the anticyclone converge near northeastern Brazil, contributing to the increase in precipitation just offshore. The northerlies along the western branch of the anticyclone give rise to anomalous convergence in the vicinity of the SACZ and thus enhance the mean moisture convergence (not shown) and increase the precipitation in the SACZ.

The increase in precipitation in the SACZ in the low-insolation experiment may also result from an increase in eddy activity aloft, which is in turn as a result of the changes in the upper-level flow (Fig. 14). The major difference in DJF zonal winds at 200 hPa between the low-insolation and high-insolation experiments is that the easterlies aloft over the equatorial Atlantic seen in the high-insolation experiment are replaced by westerlies (cf. Figs. 14a and 14b). Comparing the other panels in Fig. 14 with the panels for low insolation and high insolation, we see that the switch from easterlies to westerlies is uniquely associated with the cooling of southern Africa. The shift from easterlies to westerlies allows more upper-level eddies to enter the subtropical region and thus enhances the stochastic stirring of the air below, resulting in an increased likelihood of precipitation.

6. Discussion

In this section, we discuss the climatological significance of the $\delta^{18}O_c$ in the South American speleothems, and compare our model results with other proxy data and model results. We will also briefly discuss how the changes of the South American monsoon differ from the changes of the Asian monsoon on a precessional scale.

a. Climatological significance of $\delta^{18}O_c$ and comparison to other proxy records

Using the decomposition method outlined in section 2c, we find that the causes of the orbitally forced $\delta^{18}O_p$ excursions in the Andes and eastern Brazil support the interpretation of $\delta^{18}O_c$ of speleothems offered in the
literature. Specifically, our model results support the interpretation of Cruz et al. (2009) that the orbitally paced changes in $\delta^{18}O_p$ in northeastern Brazil are due to changes in the intensity of precipitation (i.e., the amount effect). Along the Andes, the precessional changes in $\delta^{18}O_p$ are due to changes in the $\delta^{18}O$ of the vapor arriving from the western Amazon basin and changes in the seasonality of precipitation, consistent with the original interpretation by Cheng et al. (2013). The degree of rainout in the SASM strongly impacts the isotopic composition of precipitation delivered downstream over the Andes, consistent with the results from previous studies (Salati et al. 1979; Grootes et al. 1989; Rozanski et al. 1993; Vuille and Werner 2005; Vuille et al. 2012).

The simulated orbitally forced changes in $\delta^{18}O_p$ are also qualitatively consistent with the several other proxy data from the Andes that show isotopic trends over the Holocene. These include trends in $\delta^{18}O_e$ in the speleothems from Huagapo, Peru (11°S; Kanner et al. 2013), and Cueva del Tigre Perdido, Peru (van Breukelen et al. 2008); trends in the $\delta^{18}O$ of authigenic calcite from Lake Junin, Peru (11°S; Seltzer et al. 2000), and Laguna Pumacocha (10°S, Bird et al. 2011) on the eastern flank of Peruvian Andes; and trends in the $\delta^{18}O$ of ice in the Huascaran glacier in the Peruvian Andes (9°S; Thompson et al. 1995). Each of these records suggests that precipitation becomes isotopically lighter in the late Holocene (high austral summer insolation) compared to the early Holocene (low insolation). Kanner (2012) presented an older segment of the speleothem record from Huagapo cave that extends from around 100 to 175 ka BP; it also features $\delta^{18}O_e$ excursions that are clearly paced by orbital forcing, with an amplitude and phase that is consistent with the excursions simulated by the model.

The simulated changes in precipitation as a result of orbitally forced insolation changes also agree with other types of proxy data that are sufficiently long to resolve the precessional-scale signal. These records include speleothem and travertine deposits from northeastern Brazil, which feature maxima in stalagmite and travertine growth during times of high austral autumn insolation at 10°S (Wang et al. 2004), which are also times of low austral summer insolation (Fig. 2) and thus times during which our models simulate high precipitation in northeastern Brazil. The natural $\gamma$-radiation records from sediment cores in Salar de Uyuni, Bolivia (20°S), feature alternating wet and dry periods over the past 50,000 years that align well with times of high and low January insolation of 15°S, respectively (Baker et al. 2001a; Fritz et al. 2004), also in agreement with our model results. A Holocene speleothem
record from Lapa Grande in east-central Brazil (14°S; Strikis et al. 2011) is located along the southern edge of the simulated negative $\delta^{18}O_p$ in northeastern Brazil (Fig. 4); as would be expected, the Lapa Grande record features an approximate $-0.5^{\circ}\text{o}$ decrease in $\delta^{18}O_p$ from 10 ka BP to the present. In addition to that, Haug et al. (2001) presented a record of titanium in sediments from the Cariaco basin and argued that the ITCZ has moved southward over the course of the Holocene; this is also consistent with our results (not shown). Finally, the marine sediment core from the continental slope off the Norte Chico, Chile (27.5°S; Lamy et al. 1998), shows that dry (wet) winters are associated with high (low) austral winter insolation, also consistent with our results (not shown).

There is a mismatch between the $\delta^{18}O_p$ record in the speleothems of Pacupahuain Cave at 11°S in the Andes (Kanner et al. 2012) and the $\delta^{18}O_p$ from our simulations; our model simulates a large ($3^{\circ}$) change in $\delta^{18}O_p$ while this speleothem record shows no orbital pacing. The likely cause for this misfit is that this record spans a limited duration when the changes in insolation are weak and when there are large changes in global ice volume occurring that affect the regional precipitation (see, e.g., Stansell et al. 2014). An analogous situation occurs in the speleothem of Sanbao in eastern China: the portion of the record that spans from present to 80 ka BP when insolation changes are relatively small reflects changes in global ice volume, whereas the portion of the record prior to 80 ka BP during a time of large insolation changes overwhelmingly reflects the changes in insolation (Cheng et al. 2009).

There are numerous lake level reconstructions from South America (e.g., Hodell et al. 1991; Curtis et al. 1999; Baker et al. 2001b; Polissar et al. 2013). However, it is difficult to compare results from our experiments with these records because lake level depends on precipitation, evaporation, and the geometry of the lake basin. Hence, only with a model of lake isotopes that is driven by the output from our climate model can we address these observed changes in lake level.

b. Dynamics of the precessional changes

Our model results support the hypothesis from earlier studies (Martin et al. 1997; Seltzer et al. 2000; Baker et al. 2001b; Cruz et al. 2009) that reduced insolation causes a weaker South American summer monsoon and thus less precipitation over the western Amazon basin. Cruz et al. (2009) further proposed that a weaker SASM increases the precipitation over northeastern Brazil by weakening the subsidence aloft. This is not supported by our experiments, however. Additionally, we found that the austral summer ITCZ moved southward in the low-insolation phase relative to the high-insolation phase, whereas Cruz et al. (2009) proposed the opposite.

Our results show that the orbitally driven changes in the heating of southern Africa influence the precipitation over and offshore of eastern South America. The mechanism appears to be a Rossby wave that is generated by the cooling of southern Africa that causes an increase in low-level time-mean (moisture) convergence in the western tropical Atlantic, enhancing the convection in the SACZ. This is unlike the impact of Africa on the basic climatology of South America (a Walker circulation impact; see Cook et al. 2004).

Participants in the Paleoclimate Modelling Intercomparison Project (PMIP) performed a series of experiments to examine the impact of changes in insolation between the present day and mid-Holocene (6 ka BP), which are mainly due to orbitally induced changes in insolation forcing (Braconnot et al. 2002). We performed the same pair of experiments with our model (Fig. 15b). The spatial pattern of changes in DJF precipitation in our model agrees fairly well with that from PMIP phase 2 (not shown) and PMIP phase 3 (PMIP3; Fig. 15a); precipitation decreases inland of South America and the ITCZ shifts southward in the low-insolation phase (modern day) relative to the high-insolation phase (Prado et al. 2013; Braconnot et al. 2007). The spatial pattern of the precipitation response of ECHAM4.6 to Holocene changes is also very similar to the difference between the low-insolation and high-insolation experiments (cf. Figs. 5a and 15b), implying that the response to insolation forcing is essentially linear. The only disagreement between our results and those from PMIP3 is in the SACZ, where precipitation does not change in PMIP3 but is intensified to the north of SACZ in our model. This disagreement may be as a result of the inclusion of ocean dynamics in the PMIP3 models because changes in SACZ are seen in both our mid-Holocene experiment and low-insolation experiment. That ocean dynamics are the root cause of the differences in the SACZ response is suggested by the results of Chameles (2014), who found a northward intensification and movement of SACZ in the Community Climate System Model, version 3 (CCSM3), when CCSM3 is coupled to a slab ocean model rather than a full ocean model.

Finally, Moura and Shukla (1981), Mechoso et al. (1990), and Nobre and Shukla (1996) showed that changes in SST play an important role for interannual changes in precipitation over eastern South America, especially the droughts of northeastern Brazil. Relative to the direct changes in land heating associated with insolation forcing, however, the impact of orbitally induced changes in SST on precipitation is small. On the precessional scale, changes in SST have a small influence on the precipitation over South America. This finding is corroborated by the similarity in precipitation response between our model and the PMIP3 models (cf. Figs. 15b
and 15a), in which a full ocean model was used. It is also consistent with Liu et al. (2003), who showed that the oceanic feedback has a modest influence on South American monsoon on a precessional scale, but it differs from the proxy studies by Baker et al. (2001a,b) and Polissar et al. (2013), who proposed that oceanic forcing plays a much larger role than previously suspected in modulating Holocene climate in South America.

Finally, our model results also show that in the low-insolation experiment there is a large increase in the JJA precipitation over the northern Andes (Fig. 7b), which contributes to the increase in $\delta^{18}O_p$ (Fig. 6c). With another set of experiments where the surface albedo is reduced that mimic the local increase in JJA net top-of-atmosphere radiation seen in the low-insolation experiment (not shown), we found that the increase in JJA precipitation amount, as well as the precipitation intensity, is due to the enhancement of local land heating of South America (see Figs. 2 and 9b).

c. Comparison to the response of the Asian monsoon to precessional forcing

The impact of the precessional cycle on the SASM is somewhat different from that of the Indian monsoon. For the SASM, the centroid of precipitation remains over the Amazon basin throughout the precessional cycle but waxes and wanes in intensity. In the case of the Indian monsoon, high Northern Hemisphere summer insolation features a centroid of rainfall that is over the land regions, whereas the low phase of summer insolation (such as today) features a centroid in monsoon precipitation that is over the ocean in the Bay of Bengal (Battisti et al. 2014).

7. Conclusions

The $\delta^{18}O$ of calcite ($\delta^{18}O_c$) in speleothems of South America shows a pervasive response to orbital forcing, indicating that the tropical insolation forcing plays an important role in changing the climate of South America. We examined the climatological significance of the $\delta^{18}O_c$ using an isotope-enabled general circulation model, ECHAM4.6. Two experiments were performed with the same modern-day boundary conditions, but insolation of 218,000 and 207,000 years ago, defined as the “low” and “high” Southern Hemisphere summer (DJF) insolation, respectively. Differences between these two experiments display as a dipole pattern in precipitation and precipitation-weighted $\delta^{18}O$ ($\delta^{18}O_p$). In the low-insolation experiment, precipitation increases over northeastern Brazil but decreases over the northern Andes and the majority of inland South America; $\delta^{18}O_p$ increases in the northern Andes but decreases in northeastern Brazil, consistent with what is seen in the $\delta^{18}O_c$ of speleothems found in these two regions.

Analysis of the changes in $\delta^{18}O_p$ reveals that the amount effect dominates the change in $\delta^{18}O_p$ in northeastern Brazil: compared with the high-insolation phase, austral summer precipitation shifts toward more frequent heavy precipitation in the low-insolation phase, which depletes the heavier isotope of precipitation and decreases the $\delta^{18}O_p$. In the northern Andes, changes in both the seasonality of precipitation and the intensity of precipitation (i.e., the amount effect) contribute to the change in $\delta^{18}O_p$. During the low-insolation phase, precipitation increases in austral winter (because of increased insolation) but decreases in austral summer, resulting in a higher ratio of winter (isotopically heavy) to summer (isotopically light) precipitation and thus higher $\delta^{18}O_p$. In
addition, the vapor that condenses and precipitates on the site in austral summer is heavier during the low-insolation phase (because of weaker rainout in the South American summer monsoon), which also contributes to the increase in δ¹⁸O<sub>p</sub> in the northern Andes. The causes of the δ¹⁸O<sub>p</sub> changes in the Andes and in northeastern Brazil are consistent with the proposed interpretation of the orbital signal in the δ¹⁸O of stalagmites from those regions (Cruz et al. 2009; Cheng et al. 2013).

We performed three additional experiments, with surface albedo increased respectively over South America, Africa, and both South America and Africa, to illuminate the processes responsible for the orbital changes in precipitation over South America. These experiments show that the decrease in summertime precipitation inland of South America in the low-insolation phase is caused by the weakening of South American summer monsoon, which is a result of the weaker local heating of land and the overlying atmosphere. The increases in precipitation in eastern Brazil and offshore along the east coast of Brazil are caused by the orbitally induced cooling of southern and northern Africa in DJF: the cooling of northern Africa in DJF shifts the Atlantic intertropical convergence zone (ITCZ) southward, increasing the intensity of convection over the South Atlantic convergence zone. Orbitally driven changes in SST contribute to the changes in precipitation in the tropical Atlantic north of Brazil; elsewhere changes in SST have a negligible impact on climate and the isotopic composition of precipitation.

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APPENDIX

Derivation of the Scaling Factor Used in the Increasing Albedo Experiments

The “increasing albedo” experiments in section 5 are designed to illuminate the changes in precipitation and circulation that are due to local- and regional-scale changes in the precessional forcing. Since the climate model does not allow localized changes in the downward top-of-the-atmosphere radiation, we approximated the regional changes in the insolation forcing associated with the low-insolation forcing by altering the surface albedo. The factor of 2.5 used in these increasing albedo experiments is derived as follows.

Following the method of Donohoe and Battisti (2011), we first decompose the planetary albedo $\alpha_p$ into the surface contribution $\alpha_{surf}$ and atmospheric contribution $\alpha_{atmos}$

$$\alpha_p = \alpha_{atmos} + \alpha_{surf}, \quad (A1)$$

where the surface contribution $\alpha_{surf}$ is

$$\alpha_{surf} = \alpha \frac{(1 - R - A)^2}{1 - aR}. \quad (A2)$$

In Eq. (A2), $A$ is the fraction of insolation that is absorbed within the atmosphere, $R$ is the fraction of the insolation that is reflected by the atmosphere, and $\alpha$ is the surface albedo.

Holding $\alpha_{atmos}$ constant, the change in surface contribution to planetary albedo $\Delta \alpha_{surf}$ required to obtain the same change in absorbed shortwave radiation as seen in the low- minus high-insolation experiment is

$$\Delta SW_{net} = S_0 \Delta \alpha_{surf} \quad \text{and} \quad \Delta \alpha_{surf} = \frac{\Delta SW_{net}}{S_0}, \quad (A3)$$

where $S_0$ is the incoming solar radiation at the top of the atmosphere and $\Delta SW_{net}$ is the net downward shortwave radiation at the top of the atmosphere that is seen in the response to precessional forcing:

$$\Delta SW_{net} = SW_{net}(218 \text{ ka BP}) - SW_{net}(207 \text{ ka BP}). \quad (A5)$$

A change in the surface contribution to planetary albedo $\Delta \alpha_{surf}$ induced by a change in the surface albedo $\Delta \alpha$ is then [see Eq. (A2)]

$$\Delta \alpha_{surf} = (\alpha + \Delta \alpha) \frac{(1 - R - A)^2}{1 - (\alpha + \Delta \alpha)R} - \alpha \frac{(1 - R - A)^2}{1 - aR}. \quad (A6)$$

Solving Eq. (A6) for $\Delta \alpha$ we find [using Eq. (A2)]

$$\Delta \alpha = \frac{(1 - aR)(\alpha_{surf} + \Delta \alpha_{surf}) - (1 - R - A)^2 \alpha}{(1 - R - A)^2 + R(\alpha_{surf} + \Delta \alpha_{surf})}. \quad (A7)$$

Values of $A$, $R$, $\alpha$, and $\alpha_{surf}$ are taken from the high-insolation (207 ka BP) experiment.
Plugging \( \Delta \alpha_{\text{surf}} \) from Eq. (A4) into Eq. (A7) and using the DJF values of \( A, R, \alpha, \) and \( S_0 \) from the high-insolation (207 ka BP) experiment, we estimate the surface albedo must be increased by a factor of 2.5 used in the high-insolation experiment to achieve the orbitally induced reduction in absorbed insolation. Figure A1 shows that the increase in surface albedo by a factor of 2.5 gives the similar net top-of-atmosphere (TOA) radiation change that is seen in the low-insolation minus high-insolation experiments (cf. Figs. A1e and A1f).
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