TITLE: Glacial aridity in central Indonesia coeval with intensified monsoon circulation

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ABSTRACT

The Last Glacial Maximum was cool and dry over the Indo-Pacific Warm Pool (IPWP), a key region driving global oceanic-atmospheric circulation. Both low- and high-latitude teleconnections with insolation, ice sheets, and sea level have been suggested to explain the pervasive aridity observed in paleoecological and geomorphic data. However, proxies tracking the H- and O-isotopic composition of rainfall (e.g., speleothems, sedimentary biomarkers) suggest muted aridity or even wetter conditions than the present, complicating interpretations of glacial IPWP climate. Here we use multiproxy reconstructions from lake sediments and modern rainfall isotopic measurements from central Indonesia to show that, contrary to the classical “amount effect,” intensified Australian-Indonesian monsoon circulation drove lighter H- and O-isotopic composition of IPWP rainfall during the LGM, while at the same time, dry conditions prevailed. Precipitation isotopes are particularly sensitive to the apparent increase in monsoon circulation and perhaps also decreased moisture residence time implied by our data, explaining contrasts among proxy records while illuminating glacial IPWP atmospheric circulation, a key target for climate models.

1. INTRODUCTION

The Indo-Pacific Warm Pool (IPWP) underwent enormous changes during the late Pleistocene, when lowered sea levels and altered climate patterns reorganized Southeast Asian biogeography and facilitated early human migration (Wurster et al., 2010). During the Last Glacial Maximum (LGM; ~21 kyr before present [B.P.]), pollen, charcoal, stable carbon isotope, sediment geochemistry, and geomorphic evidence point to cool and dry conditions, contracted rainforests, and expanded savannas across Sundaland and northern Australia (DiNezio and Tierney, 2013; Reeves et al., 2013; Russell et al., 2014). These regional climate changes were driven by both orbital and glacial forcings (Carolin et al., 2013; Russell et al., 2014), and teleconnections within the tropics (Tierney et al., 2012).

The longest, most continuous records of terrestrial Pleistocene hydroclimate from the IPWP are based on the oxygen and hydrogen isotopic composition of rainfall ($\delta^{18}\text{O}_p$, δD_p) recorded in cave calcite and terrestrial biomarkers preserved in marine and lacustrine sediments. Tropical $\delta^{18}\text{O}_p$ and δD_p variations are often attributed to the “amount effect,” the empirically observed inverse correlation between monthly precipitation amount (P) and $\delta^{18}\text{O}_p$ and δD_p (Dansgaard, 1964; Rozanski et al., 1993), which on longer timescales should lead to ^{18}O - and D-enriched precipitation during dry periods such as the LGM. Despite widespread evidence for IPWP aridity during the LGM, $\delta^{18}\text{O}_p$ and δD_p proxy records indicate a muted response or even ^{18}O - and D-depleted conditions (Ayliffe et al., 2013; Carolin et al., 2013; Griffiths et al., 2009; Niedermeyer et al., 2014; Partin et al., 2007). If interpreted in light of the amount effect, these isotopic ratios imply only moderate drying in Borneo (Carolin et al., 2013) and wetter conditions in southern Indonesia (Ayliffe et al., 2013), in contrast to other geological proxies. These disagreements could reflect the confounding influences of multiple climate variables on proxies

such as fossil pollen, and/or real spatial heterogeneities in terrestrial moisture balance, and hence regional to local sensitivity to climate forcings. Alternatively, the disagreement between paleoecological/geochemical and O- and H-isotopic proxies could reflect the need to further develop the interpretation of $\delta^{18}\text{O}_\text{P}$ and δD_P proxy archives in the IPWP, as recent studies of isotope ratios in tropical precipitation and water vapor have revealed diverse physical processes that can either contribute to (Moore et al., 2014; Risi et al., 2012) or oppose (Brown et al., 2008; Moerman et al., 2013) the amount effect. Therefore, multiproxy reconstructions of both $\delta^{18}\text{O}_\text{P}/\delta\text{D}_\text{P}$ and geochemistry/paleoecology from the same site are critical for understanding of the response of IPWP hydrology to climate change.

Here, we reconstruct δD_P using terrestrial leaf wax compounds ($\delta\text{D}_\text{wax}$) preserved in the sediments of Lake Towuti (2.9°S, 121.5°E), central Sulawesi (Figure 1), since 60 kyr B.P. The δD of long-chain ($> n\text{-C}_{25}$) lipid *n*-alkanoic acids, components of waxy coatings on terrestrial plant leaves that are transported by water and wind into lakes and preserved in sediments, has been successfully been employed to reconstruct hydroclimate worldwide (Sachse et al., 2012), including in Indonesia (Konecky et al., 2013; Niedermeyer et al., 2014; Tierney et al., 2010a; 2012). $\delta\text{D}_\text{wax}$ reflects the δD of plant source waters, and hence δD_P , after a series of offsets during soil evaporation and biosynthesis (Sachse et al., 2012). Plant type can influence the apparent D/H fractionation between meteoric water and wax (Sachse et al., 2012), so we use the carbon isotopic composition of waxes ($\delta^{13}\text{C}_\text{wax}$) to constrain vegetation type and correct for potential influences on reconstructed δD_P ($\delta\text{D}_\text{P-R}$; Section 2). Both $\delta\text{D}_\text{wax}$ and $\delta\text{D}_\text{P-R}$ values are given throughout the study.

We constrain our interpretation of $\delta\text{D}_\text{wax}$ using geochemical and vegetation proxies from the same sediment cores, which reflect rainfall variations during the past 60 kyr (Costa et al.,

2014; Russell et al., 2014; Vogel et al., 2015), allowing moisture balance (e.g., precipitation, P-E,) and circulation (e.g., convection, moisture transport, moisture convergence) to be interpreted independently. We interpret δD_{wax} in light of modern δD_P with the aid of a new daily precipitation isotopic dataset from our study site. Our multiproxy approach explicitly tests the influence of the “amount effect” on glacial δD_P and $\delta^{18}O_P$, and highlights important changes to glacial monsoon circulation over the IPWP.

2. METHODS

2.1 Leaf wax purification and vegetation correction

Lipids were extracted from ~2-10 g of dry sediment using a DIONEX Accelerated Solvent Extractor 350, and long-chain fatty acids were purified using previously described methods (Konecky et al., 2011). The δD of the C_{28} *n*-alkanoic acid was measured in 180 samples using Gas Chromatography-Pyrolysis Isotope Ratio Mass Spectrometry at Brown University. Samples were run in duplicate, with the exception of 24 samples (~13%) that were run in triplicate. The pooled standard deviation of triplicate samples was 1.9‰, and the standard deviation of a standard run every 6 injections was 2.9‰. Samples have been corrected for the methyl group added during fatty acid derivitization using the known isotopic composition of the acidified methanol used in methylation. δD values are reported relative to Vienna Standard Mean Ocean Water (VSMOW), and have been corrected for the effect of global ice volume on the ocean isotopic reservoir by assuming a 1‰ ^{18}O -enrichment in the LGM oceans (Schrag et al., 1996) scaled to changes in global ice volume recorded by the LR04 benthic oxygen isotope stack (Lisiecki, 2005).

We test and then apply a vegetation correction to δD_{wax} that best accounts for the contrast in the apparent fractionation ($\epsilon_{\text{wax-P}}$) of C_4 grasses vs. C_3 rainforest vegetation, the biggest vegetation change at our site. Different plant types have been shown to exhibit different net fractionation between precipitation δD and wax δD ($\epsilon_{\text{wax-P}}$), potentially due to differences in evapotranspiration and/or physiology (Sachse et al., 2012). Major changes in the types of vegetation on the landscape in a given region can therefore influence the δD_P reconstructed using δD_{wax} . In some cases these effects are considered negligible because paleo-vegetation varied within groups that exhibit similar $\epsilon_{\text{wax-P}}$ values (e.g. Tierney et al., 2010b) and/or because paleo-vegetation changes were too small to significantly alter the average $\epsilon_{\text{wax/precip}}$ fractionation on a landscape (e.g. Konecky et al., 2013). In other cases, paleo-vegetation changes were substantial enough to require significant corrections to δD_{wax} based on pollen and/or the carbon isotopic composition of waxes (Feakins, 2013; Shanahan et al., 2015; Y. V. Wang et al., 2012). At Lake Towuti, the carbon isotopic composition of plant waxes ($\delta^{13}C_{\text{wax}}$) reveals vegetation changes that are too large to be negligible for $\epsilon_{\text{wax-P}}$ (Russell et al., 2014). Therefore, we employ a correction that constrains, to first order, the role of changing vegetation on δD_{wax} .

We follow a method that has been successfully employed in other tropical studies (e.g., Feakins, 2013; Konecky et al., 2014; Shanahan et al., 2015). Namely, we calculated the vegetation-corrected $\epsilon_{\text{wax-P}}$ ($\epsilon_{\text{wax-P}_{\text{veg}}}$) based on the following equation:

$$\epsilon_{\text{wax-P}_{\text{veg}}} = (fC_3 * \epsilon_{C27.alk-P_{C3}}) + (fC_4 * \epsilon_{C27.alk-P_{C4}}) + \epsilon_{\text{alkane-acid}} \quad (1)$$

where fC_3 and fC_4 are the fraction of C_3 and C_4 plants, respectively. $\epsilon_{C27.alk-P_{C3}}$ and $\epsilon_{C27.alk-P_{C4}}$ refer to the average $\epsilon_{\text{wax-P}}$ for the C_{27} *n*-alkane, and $\epsilon_{\text{alkane-acid}}$ is the average alkane-acid offset

(25‰) that occurs between C₂₈ *n*-acid and C₂₇ *n*-alkane during *n*-alkane biosynthesis (Chikaraishi and Naraoka, 2007). In a large global survey of published values, Sachse et al. (2012) determined $\epsilon_{C27.alk-P_C3}$ and $\epsilon_{C27.alk-P_C4}$ to be $-112.8 \pm 34.7\text{‰}$ and $-124.5 \pm 28.2\text{‰}$, respectively. As the global ϵ_{wax-P} compilation is strongly biased towards Northern Hemisphere mid-latitude ecosystems (Sachse et al., 2012), we repeated this calculation using a subset of observations from tropical locations only (equatorwards of 30°N/30°S), and determined the endmembers for $\epsilon_{C27.alk-P_C3}$ and $\epsilon_{C27.alk-P_C4}$ are $-129.0 \pm 29.8\text{‰}$ and $-125.4 \pm 34.0\text{‰}$, respectively.

We estimated the fraction of C₃ and C₄ plants in each of our 177 samples based on $\delta^{13}C_{wax}$ measured on the C₂₈ *n*-acid in that sample, or, for 32 samples that lacked $\delta^{13}C_{wax}$ measurements, from linearly interpolated values from neighboring samples. We used a two-member linear mixing model and C₂₈ *n*-acid endmember values of -36.9‰ and -22.0‰ for C₃ and C₄ plants (Chikaraishi et al., 2004). Values more negative than -36.9‰ , which likely reflect high carbon use efficiency or carbon recycling under closed-canopy rainforest conditions (Russell et al., 2014), were treated as 100% C₃ vegetation.

We then reconstructed the vegetation-corrected δD_P (δD_{P-R}) for each sample based on the following equation:

$$\delta D_{P-R} = [(\delta D_{wax} + 1000) / (\epsilon_{wax-P_veg}/1000 + 1)] - 1000 \quad (2)$$

where δD_{wax} is the measured δD value of the C₂₈ *n*-acid.

As the endmember $\epsilon_{C27.alk-P_C3}$ and $\epsilon_{C27.alk-P_C4}$ have significant uncertainty, we performed a Monte Carlo sensitivity analysis to quantify how the large spread ϵ_{wax-P} values ($1\sigma = 34.7\text{‰}$

and 28.2‰ for C₃ and C₄ plants, respectively) may affect our vegetation-corrected δD_P reconstructions. We randomly selected endmember $\epsilon_{C27.alk-P_C3}$ and $\epsilon_{C27.alk-P_C4}$ values from the full distribution of values in the global compilation, and used these to calculate ϵ_{wax-P_veg} and δD_{P-R} following Equations 1-2. We repeated this procedure 1000 times, resulting in 1000 possible timeseries, then used the full range of these timeseries to estimate the 25th and 75th percentiles of each point in the δD_{P-R} timeseries. We then used this ensemble of 1000 timeseries to calculate the full distribution of possible δD_{P-R} values at the LGM and early Holocene, two key periods in our timeseries, and to perform statistical hypothesis testing of the differences between the means of these distributions via two-tailed, right-tailed, and left-tailed t-tests. For comparison, we repeated the Monte Carlo error analysis using the full distribution of $\epsilon_{C27.alk-P_C3}$ and $\epsilon_{C27.alk-P_C4}$ from tropical locations (equatorwards of 30°N/30°S), rather than all values in the global compilation.

2.2 Sediment core chronology

Lithology, geochemistry, and chronology for core IDLE-TOW10-9B are described in Russell et al., 2014. Briefly, our age model for TOW10-9B is based on 23 Accelerated Mass Spectrometry ¹⁴C ages, including 20 ages on bulk organic carbon and 3 ages on terrestrial macrofossils, with a ¹⁴C reservoir age of 500-1000 ¹⁴C years determined from paired macrofossil/bulk sediment ¹⁴C ages and ²¹⁰Pb ages. An average sedimentation rate was used to extrapolate the ages of sediments older than 50 kyr. The error on the age model, calculated using mixed-effect regression (Heegaard et al., 2005), is ~100 years near the core-top and ~900 years at the base of the ¹⁴C-dated interval.

2.3 Local precipitation isotopic measurements

Precipitation samples were collected daily starting in March 2013 from the town of Wawondula, approximately ~8km from Lake Towuti. Samples were sealed, refrigerated, and later analyzed on a Picarro L1102-I Isotopic Water Liquid Analyzer at Brown University. Precipitation amount was measured every day at the collection site via a Stratus RG202 Professional rain gauge mounted next to the precipitation collector. We present data from March 2013 through March 2015. We supplement these data with rainfall measurements made from 2009-2014 by a nearby automated weather station operated by local mining company PT Vale Indonesia. We also compare our data with a back-trajectory analysis of rain-bearing air masses during March 2013-2015 using the Hybrid Single Particle Lagrangian Integrated Trajectory (HYSPLIT4) model (Draxler, 1999) with NCEP/NCAR Reanalysis V2 gridded dataset (available at <http://www.esrl.noaa.gov/psd/>).

3. RESULTS AND DISCUSSION

3.1 Controls on modern δD_p at Lake Towuti and relationship with precipitation amount

Like most of Indonesia (Hendon, 2003), the majority of Towuti's 2,700 mm of annual rainfall occurs during the AISM wet season, from Nov-May, with a drier season occurring from Jun-Oct (Figure 1). Peak rainy conditions typically occur in April, when the ITCZ is directly overhead and local convective activity is enhanced (Figure 1, Figure 2). Rainfall throughout the Nov-May wet season is primarily of monsoonal origin, borne by northwesterly air masses (Figure 2), even though peak wet and dry months (MAM and ASO, respectively) occur slightly later in the season than average monsoonal climates in Indonesia (DJF and JAS) (Aldrian and Susanto, 2003). Seven-day back trajectories of rain-bearing air masses show a strong seasonal

contrast between AISM moisture, which is associated with westerly, northwesterly, and northeasterly air masses, and dry season rainfall, which is associated with southeasterly sources (Figure 2). Even northwesterly monsoonal air masses are connected with the large-scale northeasterly flow that is present throughout DJFMAM (Figure S1). This marked seasonal contrast has also been noted at Flores, southern Indonesia (Griffiths et al., 2009).

Despite its slightly delayed (MAM) rainfall maximum, the precipitation isotope climatology at Lake Towuti is broadly similar to the classic monsoonal regime of western Sulawesi (where rainfall peaks during DJF), suggesting a dominant role for the AISM on precipitation isotope seasonality (Figure 3). By comparison, the precipitation isotope climatology of Borneo, which does not fall within the core AISM region (Aldrian and Susanto, 2003), exhibits weak precipitation and isotopic seasonality, with no clear monsoonal influence (Moerman et al., 2013) (Figure 3). Towuti precipitation anomalies are strongly correlated to precipitation amount across Sulawesi (Figure S2), but Towuti precipitation isotopic composition is negatively correlated with rainfall across a broad region of monsoonal central and southern Indonesia (Figure S2). This predominantly regional, rather than local, nature of precipitation isotope ratios has been previously noted throughout the IPWP (Kurita et al., 2009; Moerman et al., 2013).

The AISM has a strong impact on monthly-scale δD_P at Towuti that extends beyond a local precipitation amount effect. During our observational period from March 2013-2015, daily δD_P and precipitation amount were uncorrelated ($r^2 = 0.03$, $p < 0.001$). Monthly δD_P and precipitation amount were more strongly correlated ($r^2 = 0.40$, $p < 0.001$), but with strong seasonal variability (Figure 4c): δD_P was strongly correlated with precipitation amount during the dry season (JJASO; $r^2 = 0.81$) but not during the monsoonal wet season δD_P (NDJFMAM; $r^2 = 0.003$).

The five-month periods with the strongest correlation between δD_P and precipitation amount occurred during the dry and transition seasons, from June-October 2013, August-December 2013 ($r^2 = 0.84$), and June-October 2014 ($r^2 = 0.93$), and the weakest correlation occurred during the AISM season from Nov-Mar 2013-2014 and 2014-2015 ($r^2 = 0.02$).

The beginning of the 2013-14 monsoon season was abnormally dry at Towuti, with monthly average rainfall in NDJF 45% lower than the 2009-2014 average. Rather than reflecting locally dry conditions, Towuti δD_P exhibited D-depleted values around -75‰ in December/January 2013-14, nearly identical to δD_P during the wettest months of April/May 2013. δD_P in the following December/January (2014-2015) was around -60‰, more D-enriched than the previous December/January despite slightly more rainfall. Therefore, the amount effect is variable and inconsistent at Lake Towuti, with a stronger presence during the dry season but a weak presence during the AISM season.

Recent studies have highlighted that the IPWP's complex circulation and cloud processes can either enhance or interfere with the amount effect caused by local rainout. Horizontal moisture transport and convergence (Moore et al., 2014), which is strong in the IPWP, has been shown to cause "upstream" isotopic fractionation from mesoscale convective systems and remote rainfall to be imprinted on local precipitation isotopic compositions (Conroy et al., n.d.; Moerman et al., 2013). The frequent presence of mesoscale convective systems in the IPWP, which are associated with deep, stratiform cloud systems that adjoin deep convective clouds (Schumacher and Houze, 2003), also produce isotopic depletion independent of local rainout by enhancing stratiform rainfall and injecting D-depleted vapor into the subcloud layer (Kurita, 2013; Kurita et al., 2011).

Our 2013-2015 observations confirm that these remote signals are extremely important for precipitation δD and $\delta^{18}O$, especially during monsoon season. “Extra” (i.e., non-amount effect) isotopic depletion occurs during NDJF (Figure 4), when AISM circulation dominates local rainfall (Figure 2). In addition to local rainout, this isotopic depletion is associated with the region’s large-scale circulation, specifically the northeasterly and westerly winds that drive convergence and rainfall over the IPWP (Figure 1, Figure S2). Therefore, isotopic depletion at Towuti is affected by, and likely records, the strong transport of moisture from the tropical western Pacific and Indian Oceans. These sources may be strongly isotopically depleted because they contain significant amounts of stratiform rainfall (Kurita et al., 2011), which is more prevalent over the open ocean than over land (Schumacher and Houze, 2003). These sources may also carry an imprint of prior moisture convergence, (Moore et al., 2014), for example along the oceanic ITCZ, or the imprint of rain re-evaporation that characterizes tropospheric water vapor in the western tropical Pacific (Noone, 2012; Sutanto et al., 2015; Worden et al., 2007). Therefore, in modern rainfall, and likely in past rainfall, isotopic depletion need not correspond to increased precipitation amount in central and southern Indonesia.

3.2 Sensitivity of reconstructed δD_P to changes in vegetation

Results of the vegetation correction are shown in Figure 5. When all $\epsilon_{C27,alk-P}$ values in the global compilation (Sachse et al., 2012) are used (Figure 5a), accounting for changes in the relative distribution of C_3 and C_4 vegetation reduces the amplitude of the δD_{P-R} signal by up to ~8‰ from 18-30 kyr B.P., with maximal offset during times of C_4 grass expansion (Figure 6). However, when only tropical $\epsilon_{C27,alk-P}$ values are used (Figure 5b), accounting for changes in the relative distribution of C_3 and C_4 results in a δD_{P-R} that is only slightly different from

uncorrected δD_{P-R} (Figure 5b). The vegetation correction is minimal in the tropics-only case because the difference between $\epsilon_{C27.alk-P_C3}$ and $\epsilon_{C27.alk-P_C4}$, while more than 10‰ globally, is minor (<4‰) when the analysis is restricted to tropical plants. While the spread of $\epsilon_{C27.alk-P}$ values is still large ($1\sigma \geq 30\%$), the negligible difference in the means of C_3 vs. C_4 tropical plants suggests that the variations in apparent hydrogen isotope fractionation among different plant types is not inherent to photosynthetic pathway. Rather, the differences are strongly tied to climatic effects on soil-water and leaf-water evaporative enrichment, and to physiological and biosynthetic differences, as previously indicated (Sachse et al., 2012). Soil-water and leaf-water evaporative enrichment are expected to be large in the mid-latitudes, where evaporation exceeds precipitation; this may explain why the offset between $\epsilon_{C27.alk-P_C3}$ and $\epsilon_{C27.alk-P_C4}$ is larger in the global compilation, which is strongly biased towards mid-latitude field studies, than when restricted to the tropics. To err on the conservative side, we base our interpretations on the more extreme vegetation-corrected δD_{P-R} (Figure 5a), even though this correction may be an overestimate.

Our Monte Carlo error analysis indicates that the δD_{P-R} values are strongly impacted by the choice of ϵ_{wax-P} endmembers (Figure 5). However, the structure of the time series is generally maintained, and amplitudes of major changes remain similar, if not enhanced. This follows logically from the distributions of errors on the C_3 and C_4 ϵ_{wax-P} mean values in the global compilation, which are approximately normal. Examining the absolute δD_{P-R} values, our correction using the full global dataset produces δD_{P-R} values that are approximately ~18‰ more negative at the core top than observed modern δD_P (-55‰, Mar-Mar, weighted). This discrepancy is minor given the age of the core top (~1.2 kyr BP) and the ~30‰ range of δD_{P-R} values observed during the Holocene. However, we note that the tropics-only vegetation-

corrected δD_{P-R} yields a core top value that closely matches (within 3‰) the modern observed δD_P . This provides more support that our correction using the full global dataset is likely an overcorrection. Overall, while the absolute values of reconstructed precipitation δD may be too D-depleted, the relative magnitude of variations should be robust.

3.3 Co-variation in δD_P , rainfall and vegetation at Lake Towuti during the past 60kyr

Lake Towuti δD_{P-R} exhibits a ~50‰ range over the past 60 kyr, with the most D-enriched values during mid-Marine Isotope Stage (MIS) 3 (-59‰, 33.0-50.0 kyr BP) and during the early Holocene (-53‰, 10.5-11.7 kyr BP). By contrast, precipitation was highly D-depleted during early MIS3, the majority of MIS2, and the late Holocene, with average values of -80‰, -75‰, and -71‰, respectively (Figure 6).

On multi-millennial to orbital timescales, these variations in Lake Towuti δD_{wax} and δD_{P-R} bear strong resemblance to proxy reconstructions of terrigenous runoff (sedimentary Ti content) and terrestrial vegetation ($\delta^{13}C_{wax}$) from the same sediment core (Figure 6), both indicators of precipitation and moisture balance (Russell et al., 2014). Highly coherent variations in these three proxies suggest that the atmospheric circulation processes traced by δD_{P-R} are related to long-term changes in rainfall. However, contrary to the δD_{P-R} signal predicted from the local amount effect (Dansgaard, 1964) or a regional amount effect (Figure S2), δD_{P-R} is positively related to wetter climate states (Figure 6). For instance, during mid-MIS3 and the early Holocene, when δD_{P-R} is D-enriched, a high concentration of Ti in Towuti's sediments indicates a large influx of surface and river waters into the lake under wetter conditions. At the same time, $\delta^{13}C_{wax}$ is strongly negative, suggesting that these sustained wet conditions supported closed-canopy rainforest vegetation in the catchment (Russell et al., 2014). Conversely, Ti is low, and

$\delta^{13}\text{C}_{\text{wax}}$ is more positive, during early MIS3, the majority of MIS2, and the late Holocene, implying drier conditions when $\delta\text{D}_{\text{P-R}}$ is D-depleted.

The MIS3/2 transition ca. 30 kyr BP was particularly pronounced, corresponding to a 16‰ D-depletion in $\delta\text{D}_{\text{P-R}}$ (22‰ without the vegetation correction to $\delta\text{D}_{\text{wax}}$; see Section 2). Depleted $\delta\text{D}_{\text{P-R}}$ values occurred at the peak of the Last Glacial Maximum (~21 kyr B.P.), when arid conditions characterized this region. Elsewhere in Sulawesi, lake levels dropped substantially (Dam et al., 2001) and swamps became entirely desiccated (Hope, 2001). Although previous work has indicated a severe dry season during the LGM (Dubois et al., 2014), our data show that LGM aridity was a year-round phenomenon at Towuti. Decreased Ti reflects decreased surface runoff during peak precipitation season, as a result of lower erosion rates and sediment yield during these months (Milliman and Syvitski, 1992). At the same time, expansion of C_4 grasses (Russell et al., 2014) and increased deposition of redox-sensitive trace elements (Costa et al., 2014) indicate a longer or more severe dry season (Dubois et al., 2014). A reduction in both wet season and dry season rainfall drove a large reduction in mean annual P-E, which caused lake levels to fall substantially, such that Towuti became a hydrologically closed basin (Vogel et al., 2015).

The contrast between D-depleted precipitation at the LGM and more D-enriched precipitation during MIS3 and the early Holocene may be dampened slightly when considering uncertainties in the vegetation correction (Figure 5). However, results from our modeling and hypothesis testing indicate that D-depletion during the LGM in $\delta\text{D}_{\text{P-R}}$ between could also be enhanced, depending on the selection of endmembers. The probability of eliminating the LGM-to-early-Holocene contrast is extremely low (Figure 5a). Two-tailed, right-tailed, and left-tailed t-tests of the distributions of 1000 possible LGM and early Holocene $\delta\text{D}_{\text{P-R}}$ values show that, at

the 99% confidence level, the means are distinct from one another ($p < 0.0001$) and the LGM is more D-depleted ($p < 0.0001$).

3.4 Regional aridity and isotopically light rainfall during the LGM

The evolution of Lake Towuti δD_{P-R} since the LGM is similar to other long isotopic proxy records from the IPWP (Figure 7). Precipitation δD and $\delta^{18}O$ were moderately light during the LGM, became maximally heavy during the deglaciation (~ 16 kyr BP at Borneo, ~ 11.5 kyr BP at Towuti and Flores), and became lighter again during the Holocene. A mid-Holocene peak in isotopically light precipitation occurred at Borneo, whereas the lightest values at Towuti occur during the late Holocene, and Holocene values are relatively stable at Flores and in the Makassar Strait. A pronounced enrichment occurred at both Towuti and NE Kimberley around 2 kyr BP. Differences in the timings of isotopic minima and maxima likely reflect real regional differences in atmospheric processes, and perhaps precipitation seasonality, due to variable regional sensitivities to insolation forcing, teleconnections to the Indian Ocean, and changes in local sea level (Griffiths et al., 2009; Partin et al., 2007; Russell et al., 2014; Tierney et al., 2012). Differences in timing, particularly at the sub-millennial level, could also be due to filtering and smoothing inherent to each proxy system, such as the residence time of leaf waxes on the landscape, or residence time of groundwater within soils and the karst system before forming cave dripwater, which in modern studies can smooth $\delta^{18}O$ signals by ~ 3 -10 months (Moerman et al., 2014). However, this smoothing likely has minimal impact on geologic timescales (Moerman et al., 2014). If a seasonal bias exists in the proxies, both speleothem $\delta^{18}O$ and leaf wax δD would be biased towards the wet season. Plants grow year-round in Sulawesi, so δD_{wax} represents weighted mean annual δD_P . Weighted mean annual δD_P , and therefore δD_{wax} , is likely

biased towards the monsoon season, as ~70% of annual rainfall falls during these months. Similar arguments apply to speleothem records, since most rainfall occurs during the wet season, and groundwater recharge is biased toward heavy rainfall events (Moerman et al., 2014; Taylor et al., 2012). Overall, the similarity in the evolution of δD and $\delta^{18}O$ across the IPWP suggests that these records trace common atmospheric circulation processes occurring during the LGM, the deglaciation, and the early Holocene. Therefore, depleted δD_{P-R} values during the LGM largely reflect D-depleted monsoonal precipitation.

In particular, the Lake Towuti δD_{P-R} record is strikingly similar to the speleothem $\delta^{18}O$ record from Flores, southern Indonesia (Figure 7). In both Sulawesi and Flores, the LGM was isotopically depleted relative to the early Holocene, with δD_P comparable to modern conditions. The LGM and deglacial similarities between these two records suggest regionally coherent changes in the AISM, which drives modern rainfall and isotope climatology at both sites. Previous studies have attributed these depleted LGM values to a southward shift of the AISM system, and hence the ITCZ, bringing increased and more depleted rainfall to southern Indonesia (Ayliffe et al., 2013; Griffiths et al., 2009). Depleted δD_{P-R} values at Towuti during the LGM support these inferences of strengthened AISM circulation (Ayliffe et al., 2013). However, geochemical and vegetation proxies from Lake Towuti indicate that the intensified glacial monsoon accompanied dry, rather than wet conditions. We note that precipitation isotopic composition during the LGM versus the Holocene cannot be easily compared in the AISM region of northern Australia, as these two speleothems were taken from caves ~500 km apart.

Our finding of depleted δD_{P-R} during arid intervals in Sulawesi stands in apparent contrast to other terrestrial IPWP records interpreting isotopically depleted precipitation to reflect wetter conditions, directly or indirectly, during the LGM and the early deglacial period

(Carolin et al., 2013; Denniston et al., 2013b; Griffiths et al., 2009; Partin et al., 2007). Two possible explanations could account for this apparent paradox. First, Lake Towuti δD_{P-R} could reflect wetter conditions across the region, while aridity prevailed locally. In this scenario, arid conditions would have been geographically restricted to central Sulawesi while increased rainfall and moisture convergence across southern Indonesia under a southward-shifted AISM front drove rainfall in the southern IPWP to become isotopically lighter than at sites to the north (Ayliffe et al., 2013; Griffiths et al., 2009). However, although some parts of NW Indonesia appear to have remained wet at the LGM (Williams et al., 2009), the vast majority of multiproxy evidence from both marine and terrestrial environments indicates widespread LGM aridity across most of the IPWP (Dam et al., 2001; DiNezio and Tierney, 2013; Hope, 2001; Reeves et al., 2013; Williams et al., 2009; Wurster et al., 2010). Reduced precipitation and forest contraction is inferred from pollen, charcoal, guano, and lithostratigraphy in northern Sundaland (Anshari et al., 2001; Caratini and Tissot, 1988; Wurster et al., 2010) as well as in the monsoonal regions of central/southern Indonesia and northern Australia (Hope, 2001; X. Wang et al., 1999; Williams et al., 2009). Even northern Java and northeastern Sulawesi, which are perennially wet today, were occupied by open herbaceous swampland, grass, and seasonally-dry forest vegetation during MIS2 (Dam et al., 2001; Hope, 2001; Kershaw et al., 2007; van der Kaars et al., 2000; Williams et al., 2009). Maximum aridity occurred between ~29-17 ka, similar to the timing of maximum D-depletion at Lake Towuti (Williams et al., 2009). Therefore, advection of a depleted δD_{P-R} signal from wetter source areas within the Maritime Continent, while only central Sulawesi remained anomalously dry, seems unlikely.

We argue, instead, for a second possible explanation: widespread aridity during the LGM occurred concomitant with an intensified monsoon circulation over the IPWP, driving isotopic

depletion of rainfall. Proxy data indicate that both aridity and D- and ^{18}O -depleted precipitation were regionally widespread, with evidence for isotopically light precipitation in isotopic proxy records and evidence for aridity in geochemical and paleoecological records. Despite widespread paleoecological and lithostratigraphic evidence for reduced precipitation throughout the IPWP, fossil groundwaters in Thailand, Vietnam, and elsewhere in southeast Asia were isotopically similar or light during the LGM relative to modern (Aggarwal et al., 2004). Speleothem $\delta^{18}\text{O}$ during the LGM was comparable to the early Holocene at Flores (Griffiths et al., 2009; Partin et al., 2007) despite nearby evidence for LGM aridity (Bird et al., 2005; Dubois et al., 2014), similar to our findings at Towuti (Figure 7). By contrast, speleothem $\delta^{18}\text{O}$ at Borneo, where climate is rainy year-round (Cobb et al., 2007; Moerman et al., 2013) rather than predominantly monsoonal like Sulawesi and Flores (Figure 3), is more enriched during the LGM relative to the Holocene (Partin et al., 2007). Therefore, “extra” isotopic depletion during the LGM — i.e., not explained by the amount effect — is a feature common to records of continental AISM precipitation isotopes spanning the LGM to present. The “extra” depletion may be further enhanced at Towuti and Flores because these records are likely biased towards the monsoonal wet season.

Support for the inference of strongly depleted monsoonal rainfall derives from present-day precipitation samples collected from Lake Towuti, which show isotopic depletion during the AISM season even when precipitation amounts are relatively low (Figure 4) (Section 3.1). Though our 2013-2015 precipitation dataset is too short to evaluate interannual variations in δD_p , let alone glacial/interglacial controls, these data can at least provide clues into the mechanisms driving isotopic depletion during the LGM. Modern AISM precipitation is D-depleted at Lake Towuti owing to the strong moisture convergence over the IPWP, which fuels the transport of

isotopically depleted moisture from the tropical western Pacific and Indian Oceans, and to the strong influence of isotopically depleted rainfall from mesoscale convective systems (Section 3.1). These processes remain important during the AISM season even when local rainfall amounts are reduced, weakening the $\delta D_P/P$ correlation. Our multiproxy reconstructions from Lake Towuti indicate similar processes may operate on glacial-interglacial timescales, particularly during the LGM, when rainfall amounts were low. During the LGM, the advection of depleted moisture sources, which were isotopically depleted due to intensified circulation, had a stronger influence on δD_P than rainfall amount.

4. Implications for glacial monsoon circulation in the IPWP

During the LGM, isotopic depletion occurred over central and southern Indonesia despite arid conditions, indicating that the character of IPWP circulation — convection, moisture transport, and moisture convergence — was substantially different from today's. This modified glacial circulation over the IPWP may have arisen in part from interactions between an intensified AISM and other regional monsoons. Previous work on the paleo-AISM has suggested a strong relationship with the East Asian Winter Monsoon (EAWM)(Miller et al., 2005), which enhances the modern AISM (Godfred Spinning and Reason, 2002; Liu et al., 2003) and may have helped to produce wet conditions in Australia during the Eemian (~115-130 kyr BP) and early Holocene despite low summer insolation in the AISM's core Southern Hemisphere region (Magee et al., 2004; Miller et al., 2005; L. Wang et al., 2012). However, this mechanism cannot fully explain Australian and IPWP aridity under glacial boundary conditions (Johnson et al., 1999; Magee et al., 2004). While a strengthened EAWM could have invigorated the AISM circulation during the LGM, it did not result in increased precipitation: Both Australia and Lake

Towuti were dry during MIS2, when the EAWM was notably strong (Sun et al., 2011), and both experienced wet intervals during MIS3 when the EAWM was weak (Magee et al., 2004; Miller et al., 2005). High obliquity during MIS3 could have contributed to wet conditions in Australia by increasing the equator-pole heat gradient and thus the transfer of moisture onto land; however this mechanism cannot explain concurrent wet conditions at Lake Towuti, which should have become dry by this mechanism (Wyrwoll et al., 2007).

Instead, our data indicate that glacial boundary conditions modified the EAWM's relationship with Indonesian and Australian rainfall. While the exact mechanism for this "push/pull" relationship is not clear, during MIS2, expansion of Northern Hemisphere ice sheets and cooling of the Asian land mass strengthened the Tibetan high, enhancing EAWM circulation (Sun et al., 2011). This could have contributed to the intensified AISM circulation and isotopic depletion that we observe over Indonesia. Stronger EAWM outflow and enhanced northerly winds over the IPWP did not lead to increased rainfall, however, perhaps because cooler SST, lowered sea level, and increased exposure of land surfaces reduced low-level moisture entrainment and cross-equatorial transport by monsoonal winds. At the same time, converged moisture over the IPWP may have been drier and more D-depleted because of cooler, more arid conditions in the subtropics, especially in the western Pacific. Together, enhanced EAWM and glacial boundary conditions lead to a strengthened, isotopically depleted, and drier zone of convergence over the Maritime Continent and the exposed Sunda Shelf.

Reconstructions of IPWP δD_p and $\delta^{18}O_p$ allow a qualitative assessment of atmospheric moisture residence time, i.e. the total precipitable water relative to precipitation rate, which tracks circulation intensity (Aggarwal et al., 2012). Because isotopically enriched precipitation is positively correlated with longer residence time (RT), the depleted $\delta^{18}O$ and δD that we observe

during the LGM necessitates a shorter RT in the IPWP, while enriched δD and wet conditions during MIS3 implied a longer RT during that time. In order to maintain both isotopic depletion and short RT during arid LGM conditions, total precipitable water would have had to decrease more dramatically than precipitation rates. With precipitable water in a given atmospheric column depending on global surface temperatures and their spatial distribution (Soden and Held, 2006; Vecchi and Soden, 2007), LGM cooling could have resulted in greatly reduced humidity while precipitation was reduced less — i.e., an apparent increase in circulation but a decrease in rainfall.

In particular, our results suggest that pervasive aridity and reduced rainfall during the LGM coincided with an apparent increase in circulation intensity in the IPWP, manifested locally by intensified convergence of depleted moisture sources during the AISM. Conversely, in simulations of 21st century climate under global warming scenarios, rising temperatures are projected to lead to increased atmospheric water vapor content, a weakened monsoon circulation, and increased monsoonal rainfall in some areas of the tropics (Christensen et al., 2013). General Circulation Models have had trouble reproducing observed LGM aridity in the IPWP, potentially due to their treatment of convection and/or the response of exposed land surfaces of the Sunda and Sahul shelves to LGM cooling (Dinezio et al., 2011; DiNezio and Tierney, 2013). Shelf exposure and subsequent flooding are considered important drivers of IPWP hydroclimate during the LGM, the deglaciation, and the early Holocene (DiNezio and Tierney, 2013; Griffiths et al., 2009; 2013; Tierney et al., 2012), and it is possible that δD_{P-R} became more D-depleted when exposed shelf areas increased the distance over land that monsoonal air masses traveled, exposing them to additional rainout via the continental effect (Aggarwal et al., 2012; Winnick et al., 2014). However, areal shelf exposure was similar during the LGM and late MIS3 (Russell et

al., 2014), yet we observe a $\sim 30\text{‰}$ difference in δD_{P-R} between the two periods (Figure 6). Therefore, δD_{P-R} depletion during the LGM likely arose from a combination of factors that intensified regional circulation, including interactions between regional monsoons and ice sheets, insolation, global sea level, and local land configuration. Water may have had a shorter residence time in the atmosphere as a result of cooler temperatures and decreased atmospheric humidity. Isotope-enabled GCM simulations of LGM climate that include both glacial and orbital forcings are necessary to test these hypotheses and their impact on local atmospheric circulation, and to understand how LGM climate dynamics in the IPWP differed from modern climate dynamics.

FIGURE CAPTIONS

Figure 1

Average seasonal wind and rainfall climatology for Indo-Pacific Warm Pool region (1979-2010). Shading, precipitation (mm/day) from the GPCP v 2.2 Combined Precipitation Dataset. Arrows, vector winds (m/s) at 850 mb. Location of Lake Towuti (yellow star) and other proxy site locations discussed in the text (red circles) are shown. Inset, average rainfall climatology for Lake Towuti, derived from 9 rain gauge stations in Central Sulawesi and inversely weighted by their distance from the study site.

Figure 2

Origins of rain-bearing air masses at Lake Towuti. a) Map of Central Indonesia (inset) showing location of Lake Towuti (star). Lines depict the 7-day back-trajectories of rain-bearing air masses during our isotopic observation period (March 2013-2015) using HYSPLIT4. Only trajectories that produced rain within 12 hours of reaching Lake Towuti were plotted. Trajectories are grouped and color-coded by month (see panel b). b) Precipitation isotope climatology (squares, δD ; circles, $\delta^{18}O$) from Lake Towuti during March 2013-2015. Each month is color-coded to match the trajectories in panel a. Shading on the x-axis indicates the months corresponding to the wet northwest monsoon/higher rainfall amounts at Towuti (purple) vs. months corresponding to the dry southeast monsoon/lower rainfall amounts at Towuti (yellow).

Figure 3

Regional precipitation isotope climatology. δD (triangles) and $\delta^{18}O$ (circles) climatology from Lake Towuti (this study; as in Figure 2; observed from 3/2013-3/2015); Makassar, western Sulawesi (Kurita et al., 2009) (observed from 11/2002-12/2006); and Borneo (Moerman et al., 2013) (observed from 7/2006-12/2012). Purple shading indicates AISM season (Nov-May) as in Figure 2.

Figure 4

Lake Towuti precipitation isotope ratios, March 2013-2015. a) Precipitation, δD_p , and $\delta^{18}O_p$ at Wawondula, Sulawesi, from March 2013 through March 2015. Shaded columns, rainfall amount measured onsite via a rain gauge (blue) and measured at a nearby meteorological station (purple). Lines, average monthly δD_p (red line, triangles) and $\delta^{18}O_p$ (black line, x's). Dashed

lines, monthly average δD_P and $\delta^{18}O_P$ weighted by daily precipitation amount. b) Black, local meteoric water line (LMWL) derived from all daily samples (n=430) at Lake Towuti. LMWLs from Borneo (red) (Moerman et al., 2013) and Flores (green) (Griffiths et al., 2009). c) Monthly precipitation amount effect at Lake Towuti for months occurring during the dry season (JJASO; red circles) and wet season (NDJFMAM; blue circles) during 2013-2015.

Figure 5

Vegetation-corrected δD_{precip} and associated uncertainties. a) Grey line, calculated by applying a constant, average $\epsilon_{l/w}$ for a global compilation of C_3 and C_4 plants (Sachse et al., 2012)). Black line, calculated by applying a time-varying, weighted mean $\epsilon_{l/w}$ that accounts changes in the relative abundance of C_3 vs. C_4 plants and their respective $\epsilon_{l/w}$ values. Black line represents the median of 1000 Monte Carlo iterations of the vegetation correction, accounting for the full distribution of errors in published C_3 and C_4 $\epsilon_{l/w}$ values. Dashed lines represent the 25th and 75th percentile. b) As in a, but based only on $\epsilon_{l/w}$ values obtained from tropical locations (equatorwards of 30°N/30°S). Horizontal cyan line represents modern mean annual δD_{precip} measured at the study site.

Figure 6

Multiple proxies for hydroclimate measured in the same sediment core samples from Lake Towuti. Top, $\delta^{13}C$ of leaf waxes (Russell et al., 2014). Middle, Ti counts (grey) and 10-point running average (black) measured by scanning XRF; %TiO (red) measured via flux fusion in the laboratory. Bottom, δD_{wax} (grey) and δD_{P-R} corrected for effects of changing C_3/C_4 vegetation on

$\epsilon_{\text{wax-P}}$ (black). Note all δD_{wax} and $\delta D_{\text{P-R}}$ values have been corrected for changes in VSMOW due to global ice volume.

Figure 7

Regional reconstructions of continental δD_P and $\delta^{18}O_P$, 24 kyr B.P. to present. Lake Towuti $\delta D_{\text{P-R}}$ versus regional proxy reconstructions from the Indo-Pacific Warm Pool: Borneo (Carolin et al., 2013; Partin et al., 2007), Towuti (this study), Makassar Strait (Tierney et al., 2012), Flores (Ayliffe et al., 2013), northeastern Kimberly, western Australia (Denniston et al., 2013b) and southwestern Kimberley, western Australia (Denniston et al., 2013a). All records have been corrected for the effects of global ice volume on the δD of VSMOW. Note that all Y-axes are reversed, such that up is depleted.

ACKNOWLEDGMENTS

We wish to thank Sinyo Rio for onsite precipitation collection, Rafael Taroza, Angel Mojarro, Lily Cohen, and Kassandra Costa for laboratory assistance, and PT Vale Indonesia for local meteorological station data. This material is based upon work supported by the National Science Foundation under Grant Number NSF-EAR-0902845 to J. Russell, and an NSF Graduate Research Fellowship to B. Konecky.

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Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7