1 **Hydrological and vegetation shifts in the Wallacean region of central Indonesia**
2 **since the Last Glacial Maximum**

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18

19 **Abstract**

20

21 Precipitation is the most important variable of Indonesian climate, yet there are
22 substantial uncertainties about past and future hydroclimate dynamics over the region.
23 This study explores vegetation and rainfall and associated changes in atmospheric
24 circulation during the past 26,000 years in Wallacea, a biogeographical area in central
25 Indonesia, wedged between the Sunda and Sahul shelves and known for its exceptionally
26 high rainforest biodiversity. We use terrestrial plant biomarkers from sediment cores
27 retrieved from Mandar Bay, off west Sulawesi, to reconstruct changes in Wallacean
28 vegetation and climate since the Last Glacial Maximum (LGM). Enriched leaf wax
29 carbon isotope ($\delta^{13}\text{C}_{\text{wax}}$) values recorded in Mandar Bay during the LGM, together with
30 other regional vegetation records, document grassland expansion, implying a regionally
31 dry, and possibly more seasonal, glacial climate. Depleted leaf wax deuterium isotope
32 ($\delta\text{D}_{\text{wax}}$) values in Mandar Bay during the LGM, and low reconstructed precipitation
33 isotope compositions from nearby sites, reveal an intensified Austral-Asian summer
34 monsoon circulation and a southward shift of the mean position of the Intertropical
35 Convergence Zone, likely due to strong southern hemisphere summer insolation and the
36 presence of large northern hemisphere ice sheets. Mandar Bay $\delta^{13}\text{C}_{\text{wax}}$ was anti-correlated
37 with $\delta\text{D}_{\text{wax}}$ during the LGM and the last deglaciation, but was positively correlated during
38 most of the Holocene, indicating time-varying controls on the isotopic composition of
39 rainfall in this region. The inundation event of the Sunda Shelf and in particular the
40 opening of the Java Sea and Karimata Strait between 9.4 and 11.1 thousand years ago
41 might have provided new moisture sources for regional convection and/or influenced

42 moisture source trajectories, providing the trigger for shifts in atmospheric circulation
43 and the controls on precipitation isotope compositions from the LGM to the Holocene.

44

45 **Keywords**

46 Precipitation; Vegetation; Biomarker; Isotope; Indonesia; Last Glacial Maximum

47

48 **1. Introduction**

49

50 Alterations in modern-day rainfall patterns over the Indonesian archipelago
51 influence the stability of its rich and diverse tropical rainforests (Underwood et al. 2014)
52 and impact the livelihoods of more than a quarter billion people living in the region
53 (Naylor et al. 2007, Oktaviani et al. 2011). However, there is relatively weak agreement
54 among Intergovernmental Panel on Climate Change climate models on the trend and
55 magnitude of future precipitation changes over Maritime Southeast Asia (Knutti and
56 Sedláček 2013), reflecting substantial uncertainty about hydroclimate dynamics in the
57 archipelago under future climate boundary conditions. Similarly, there is a long-running
58 debate on changes in Maritime Southeast Asian hydroclimate during the Last Glacial
59 Maximum (LGM), when atmospheric CO₂ concentrations were approximately 80 ppm
60 lower than pre-industrial values and Sundaland and Sahul were two, separate contiguous
61 landmasses (Figure 1). Nearly all terrestrial- and marine-based climate records indicate
62 cooler temperatures in the Indonesian archipelago during the LGM, but regional
63 precipitation patterns are still poorly understood (Reeves et al. 2013 and references
64 therein) and vegetation responses to regional climate change are even less clear (Wurster

65 and Bird 2014). Given the importance of LGM climate to climate model validation and
66 testing (e.g. diNezio and Tierney 2013), new reconstructions of precipitation and
67 vegetation response during the LGM are needed to close these knowledge gaps.

68 At the heart of the Indonesian archipelago lies Wallacea, a biogeographical region
69 comprised of Sulawesi, the Maluku Islands, and the Lesser Sunda Islands (Figure 1,
70 Whitten et al. 1987, Monk et al. 1997). Although the biogeographical lines delineating
71 Wallacea have been the subject of much academic debate (e.g. Mayr 1944 and Simpson
72 1977), the widely accepted delineation of Wallacea at present defines it as the region
73 separating Sundaland (the Malay Peninsula, Sumatra, Borneo, Java, and Bali) to the west
74 and Sahul (Papua and Australia) to the east, while excluding the Philippines. Wallacea is
75 separated from Sundaland and Sahul by deep seaways, and thus had a different
76 environmental history from the neighboring islands connected to the Sunda and Sahul
77 shelves, which were exposed during the LGM. Wallacea is a global biodiversity hotspot
78 due to its high species diversity, yet is also at high risk of habitat loss due to changes in
79 land use and climate (Myers et al. 2000). Here we present new reconstructions to assess
80 Wallacean climate and vegetation within the larger context of climate changes in
81 Maritime Southeast Asia since 26 kyr BP to improve our understanding of regional
82 vegetation and climate history, with a particular emphasis on the LGM.

83 The LGM climate of Maritime Southeast Asia has proved challenging to
84 understand, due to significant disagreement among reconstructions in the direction and
85 magnitude of regional precipitation change. This uncertainty could be partly due to the
86 use of different precipitation proxies in different archives, which are sensitive to different
87 elements of paleohydrologic change. For example, records of precipitation isotopes such

88 as speleothem oxygen isotopic compositions (e.g. Griffiths et al. 2009, Ayliffe et al.
89 2013) and seawater oxygen isotope data derived from foraminifera (Mohtadi et al. 2011)
90 have been interpreted to reflect changes in the amount of annual precipitation. The
91 interpretation of foraminiferal $\delta^{18}\text{O}$ is underpinned by the dilution of isotopically-
92 enriched seawater with isotopically-depleted rainfall, which is itself influenced by the so-
93 called “amount effect”, in which local rainfall amount is considered as the primary
94 control of rainfall isotopes (e.g. Dansgaard 1964 and Rozanski et al. 1992). However,
95 recent modeling studies and observations of modern precipitation isotope data indicate
96 that precipitation isotopes can record a range of atmospheric circulation processes
97 including moisture sources and trajectories (e.g. LeGrande and Schmidt. 2009, Kurita et
98 al. 2009, Lewis et al. 2010, Moerman et al. 2013). Better constraints on the interpretation
99 of rainfall isotope data through comparisons of multiple climate indicators will therefore
100 help to elucidate not only the regional hydroclimate sensitivities to global climate
101 forcings, including atmospheric greenhouse gas levels (e.g. Tachikawa et al. 2013),
102 orbital variations (e.g. Holbourn et al. 2005, Kershaw et al. 2011, Tierney et al. 2012,
103 Carolin et al. 2013), ice volume changes (e.g. Lee et al. 2014, Russell et al. 2014,
104 Wicaksono et al. 2015), and aerial exposure of the Sunda and Sahul Shelves (Griffiths et
105 al. 2009, Partin et al. 2007, diNezio and Tierney 2013, Dubois et al. 2014, Niedermeyer
106 et al. 2014), but also the atmospheric circulation changes that govern the regional climate
107 response.

108 Here we employ stable isotope analyses of terrestrial plant biomarkers to
109 reconstruct paleoclimate and paleovegetation in Wallacea. Carbon and hydrogen isotope
110 ratios of terrestrial plant leaf wax compounds (e.g. *n*-alkanes, *n*-acids, and others) can

111 record changes in vegetation type and the isotopic composition of precipitation, providing
112 valuable paleoclimate data (Sachse et al. 2012). We analyzed the carbon and hydrogen
113 isotopic composition of long-chain *n*-alkanoic acids extracted from a new marine
114 sediment core collected from Mandar Bay, offshore central Sulawesi, and compare these
115 data with several published hydroclimate and vegetation records from Wallacea and other
116 parts of Maritime Southeast Asia. Our record builds upon and extends previous
117 reconstructions from Mandar Bay (Tierney et al. 2012) and helps to: 1) determine the
118 patterns of Wallacean hydroclimate change during and since the LGM, 2) understand the
119 response of vegetation to long-term hydroclimate variations in this high biodiversity
120 region, and 3) elucidate the climate processes controlling rainfall isotope variations
121 through time.

122

123 **2. Regional Setting**

124

125 Wallacea lies in the center of Maritime Southeast Asia and extends from 5 °N to
126 12 °S (Figure 1). Precipitation in the northern part of Wallacea (the majority of Sulawesi
127 and the Maluku Islands, from 5 °N to 6 °S) averages more than 2000 mm/year, with all
128 months experiencing rainfall of more than 50 mm/month. There is a distinct wet season
129 from November through May, when strong northerly trade winds associated with the
130 Austral-Asian Monsoon flow over the region, carrying moisture originating in the
131 northern hemisphere (Aldrian and Susanto 2003). Over northern Wallacea, this boreal
132 winter monsoon flows from the northeast direction, whereas south of the equator this
133 monsoon is manifested as the northwest Austral-Asian winter monsoon. The southern

134 part of Wallacea (the Lesser Sunda Islands from 7 °S to 12 °S) is notably drier than
135 regions to the north, with average annual precipitation between 500-1500 mm but
136 experiences a similar wet season from November through April. The region experiences
137 a drier season between June and November, when the southeast trade winds bring dry air
138 from the Australian land mass (Monk et al. 1997). Both regions, but in particular in
139 central Sulawesi, exhibit strong orographic controls on precipitation, with higher
140 elevation areas having up to twice the rainfall of the coastal zones (Whitten et al. 1987,
141 Monk et al. 1997).

142 Vegetation in Wallacea tracks variations in precipitation, seasonality, and
143 associated climate variables (Figure 2). Wetter, and less seasonal northern Wallacea is
144 dominated by evergreen rainforests, including lowland, lower montane, upper montane,
145 and subalpine rainforests that vary as a function of altitude (Whitten et al. 1987).
146 Conversely, to the south a significant portion of the Lesser Sunda Islands is covered in
147 tropical deciduous forests and savanna woodlands. Ever-wet rainforests in the Lesser
148 Sunda Islands are found only at high elevations and in areas facing the rain-bearing winds
149 (Monk et al. 1997).

150 Our leaf wax-based hydroclimate and vegetation records were reconstructed from
151 a marine sediment core retrieved from Mandar Bay during the 2011 Sonne-217 “MAJA”
152 expedition. Mandar Bay is west of central Sulawesi adjacent to the Makassar Strait
153 (Figures 1 and 2). Core SO18515 (3.63 °S, 119.36 °E, water depth: 688 m, core length:
154 12.175 m) was recovered 12 km off the Saddang (or Sa’dan) river mouth from a
155 submarine topographic high, therefore protecting its depositional environment from
156 submarine sediment redeposition (Kuhnt et al. 2011). The Saddang River catchment has a

mean elevation of 1,173 m, a maximum altitude of 3,428 m, a length of 182 km, and a total watershed area of approximately 6,380 km², covering much of the interior of west central Sulawesi including Tana Toraja. The Saddang River lies at a vegetation divide with evergreen C₃ rainforests in the catchment to the north (Whitten et al. 1987), and tropical deciduous forests with substantial C₄ grasses to the south (Figure 2). These more open vegetation types are typical of the natural vegetation in the southern arm of Sulawesi, which has a transitional climate between the wetter central Sulawesi and drier Lesser Sunda Islands (Whitten et al. 1987).

3. Methods

We analyzed the carbon and hydrogen isotopic composition of long-chain *n*-alkanoic acids from terrestrial leaf waxes ($\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{wax}}$) in sediment samples from core SO18515. We extracted lipids from homogenized, freeze-dried sediment samples using a Dionex Accelerated Solvent Extractor and purified acids from the lipid extracts using aminopropyl column chromatography. The fatty acids were methylated with acidified methanol of a known isotopic composition, then further purified using silica gel chromatography. The resulting fatty acid methyl esters were analyzed via gas chromatography isotope ratio monitoring-mass spectrometry (GC-IR-MS) using Thermo Delta V Plus (for $\delta^{13}\text{C}_{\text{wax}}$) and Thermo Delta Plus XL (for $\delta\text{D}_{\text{wax}}$) at Brown University.

In total, we analyzed 100 samples for $\delta^{13}\text{C}_{\text{wax}}$ and 147 samples for $\delta\text{D}_{\text{wax}}$ in duplicate or greater. The average analytical error for $\delta^{13}\text{C}$ analysis of C₂₈ *n*-acid was 0.14 ‰, whereas for $\delta\text{D}_{\text{wax}}$ it was 1.96 ‰. Isotopic measurements of C₂₈ *n*-acid are

180 broadly representative of the isotopic composition of the long-chain waxes: the
181 correlations between $\delta^{13}\text{C}$ values of *n*-acid homologues were strong ($r^2 = 0.77$, $p < 0.001$
182 for $\delta^{13}\text{C}_{\text{wax}}$ of C_{28} and C_{26} , and $r^2 = 0.93$, $p < 0.001$ for $\delta^{13}\text{C}_{\text{wax}}$ of C_{28} and C_{30}). The
183 correlations were weaker for δD measurements ($r^2 = 0.47$ and 0.20 for C_{28} and C_{26} and for
184 C_{28} and C_{30} , respectively, but still significant at $p < 0.01$). We report $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{wax}}$
185 values of the C_{28} homologue, as it is the most abundant homologue in our sediment
186 samples and has the lowest analytical error. Except where otherwise noted, we interpret
187 the $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{wax}}$ values of C_{28} as representative isotope compositions of the waxes
188 preserved in our core.

189 Detailed discussion on the age model for core SO18515 (Figure S1) can be found
190 in Schroeder et al. (in press). Briefly, the age model was obtained by interpolating
191 twenty-two ^{14}C age control points (7 from wood samples and 15 from planktonic
192 foraminifera, Table S1), which covered the past 26 kyr BP and were calibrated with
193 IntCal 13 (Reimer et al. 2013) using Calib 7.0.1. (Stuiver et al. 2005). These age control
194 points are well-spread throughout the past 26 kyr BP (Figure 3).

195

196 **4. Ecological and climatic interpretations of leaf wax isotopes**

197

198 $\delta^{13}\text{C}_{\text{wax}}$ is a proxy for changes in vegetation type, and in particular the relative
199 abundance of C_3 versus C_4 plants. $\delta^{13}\text{C}_{\text{wax}}$ for C_4 plants range from -19‰ to -26‰ ,
200 whereas values for C_3 plants range from -30‰ to -45‰ (Chikaraishi and Naraoka 2007).
201 In the tropics, the relative abundance of C_3 versus C_4 plants is largely controlled by
202 precipitation and related hydroclimate variables, such as relative humidity and

203 evaporation, with dry environments favoring C₄ grasses, and wet environments favoring
204 C₃ trees (Scheffuss et al. 2005). Changes in the efficiency of carbon and water utilization
205 within C₃ and C₄ plants may also affect $\delta^{13}\text{C}_{\text{wax}}$, with more positive $\delta^{13}\text{C}$ values in drier
206 climates (Diefendorf et al. 2010). This process generally has a smaller effect on $\delta^{13}\text{C}_{\text{wax}}$
207 than the shifts between C₃ and C₄ plants; global C₃ biomass exhibits a ~6 ‰ enrichment
208 across precipitation gradients ranging from closed-canopy rainforest to desert ecosystems
209 (Diefendorf et al. 2010).

210 Within Indonesia, $\delta^{13}\text{C}_{\text{wax}}$ values observed in marine surface marine sediments are
211 consistent with climatically-driven vegetation gradients in the region, with strong
212 correlations between $\delta^{13}\text{C}_{\text{wax}}$ and average monthly precipitation and rainfall seasonality
213 (Dubois et al. 2014). $\delta^{13}\text{C}_{\text{wax}}$ values from surface marine sediments retrieved near the
214 Lesser Sunda Islands, where C₄ grasslands and savannas are common, are enriched
215 compared to those from wetter, less seasonal regions offshore from Sulawesi including
216 Mandar Bay. Thus, we assume that our $\delta^{13}\text{C}_{\text{wax}}$ record reflects the types of vegetation that
217 existed in west-central Sulawesi, and particularly the catchment of the Saddang River,
218 and, by inference, the region's hydroclimate.

219 $\delta\text{D}_{\text{wax}}$ has been shown to reliably track the δD of local precipitation ($\delta\text{D}_{\text{precip}}$) in a
220 range of terrestrial biomes and climatic conditions (Sachse et al. 2012 and references
221 therein). This includes many sites in the tropics, where $\delta\text{D}_{\text{precip}}$ is influenced by many
222 factors, including local/regional precipitation amount via “the amount effect” (Kurita et al.
223 2009), convective intensity/rain recycling (Worden et al. 2007), local SST interactions
224 (Moerman et al. 2013), downdrafts and large-scale subsidence (Brown et al. 2008), cloud
225 types and processes (Kurita et al. 2011), and horizontal moisture advection (Brown et al.

226 2008). Any spatial and temporal changes in the atmospheric circulation and precipitation
227 are thus likely to have at least some impact on δD_{wax} .

228 While δD_{wax} can thus provide information on a variety of hydroclimatic processes,
229 the isotopic fractionation between δD_{precip} and δD_{wax} (i.e. apparent fractionation) is
230 sensitive to plant physiology (Sachse et al. 2012 and references therein). Here we report
231 δD_{precip} after correcting δD_{wax} for changes in vegetation type following Konecky et al.
232 (2016, Supplementary Information). In brief, $\delta^{13}\text{C}_{\text{wax}}$ values from the same wax samples
233 were used to infer the relative abundance of C_3 and C_4 vegetation based upon $\delta^{13}\text{C}$ end-
234 member values from Chikaraishi et al. (2004). We then used D-H apparent fractionation
235 values for C_3 and C_4 end-members obtained from the literature (Sachse et al. 2012,
236 Chikaraishi and Naraoka 2007) to correct δD_{wax} . Due to differences in the number of
237 samples measured for carbon and hydrogen isotopes, we interpolated $\delta^{13}\text{C}_{\text{wax}}$ values to
238 match the δD_{wax} data points. To isolate atmospheric signals in our isotope data data we
239 applied a correction for global ice-volume changes to the δD_{precip} reconstruction based on
240 benthic foraminiferal oxygen isotope data (Waelbroeck et al. 2002) converted to δD
241 values using the global meteoric water line (Craig 1961).

242 To facilitate comparison to $\delta^{18}\text{O}$ records from regional speleothems, we also
243 converted ice volume-corrected $\delta^{18}\text{O}_{\text{speleothem}}$ from Liang Luar Cave, Flores and Gunung
244 Buda Cave, Borneo (Figure 2, Figure S3) to δD_{precip} . Because carbonate precipitation is
245 temperature-dependent, we use the carbonate-water paleotemperature equation (Fairchild
246 and Baker 2012) to calculate estimates of dripwater $\delta^{18}\text{O}$ from the $\delta^{18}\text{O}_{\text{speleothem}}$ values
247 measured on cave calcite. For Flores, assuming that Liang Luar cave temperatures have
248 always been ~3 degrees cooler than sea surface temperatures (Griffiths et al. 2010), we

249 subtracted 3 degrees from the closest available sea surface temperature records (Lombok
250 Basin, Levi et al. 2007) to estimate past temperatures of formation of cave calcite (Figure
251 S2). We then subtracted 1‰ from the computed dripwater $\delta^{18}\text{O}$, as Griffiths et al. (2010)
252 found a ~1-permil enrichment in $\delta^{18}\text{O}$ of the fluid inclusions within Liang Luar
253 speleothems relative to local rainfall. Finally, we converted Flores $\delta^{18}\text{O}_{\text{precip}}$ to $\delta\text{D}_{\text{precip}}$
254 using the modern local meteoric water line computed by Griffiths et al. (2009) (Figure
255 S2). Similar computations and corrections were done on Gunung Buda $\delta^{18}\text{O}_{\text{speleothem}}$
256 (Figure S2) based on sea surface temperature records from the Celebes Sea (Stott et al.
257 2007) and measurements of local meteorology and in-situ rainfall and dripwater isotopes
258 (Cobb et al. 2007, Moerman et al. 2013).

259

260 **5. Results**

261

262 During the past 26 kyr BP, Mandar Bay $\delta^{13}\text{C}_{\text{wax}}$ varied between -24.5 and -
263 29.6 ‰ (Figure 3). From ~26 to 19 kyr BP, $\delta^{13}\text{C}_{\text{wax}}$ was ^{13}C -enriched, averaging -25.0 ‰,
264 followed by a gradual transition to more negative values during deglaciation, until it
265 reached relatively stable values averaging -28.3 ‰ between 13 and 7 kyr BP. Afterwards,
266 $\delta^{13}\text{C}_{\text{wax}}$ became even more negative, reaching its most negative value (-29.6 ‰) ca. 5 kyr
267 BP, before rapidly shifting toward more positive values of -27.0 ‰ in the late Holocene.

268 Mandar Bay $\delta\text{D}_{\text{wax}}$ varied between -160 and -172 ‰ during the past 26 kyr BP
269 (Figure 4). Applying both ice-volume and vegetation corrections to estimate $\delta\text{D}_{\text{precip}}$, we
270 observe a slightly larger range of values (from -88 to -70 ‰), but very similar overall
271 trends as the uncorrected $\delta\text{D}_{\text{wax}}$ (Figure 4). The average difference of ~92 ‰ between

272 δD_{wax} and δD_{precip} represents the average apparent fractionation specific to C_{28} *n*-acid
273 (Konecky et al. 2016). More negative δD_{precip} values occur during the LGM, between 26
274 and ~17.5 kyr BP, with an average value of -82.5 ‰ and the most negative value (-88 ‰)
275 at ca. 20 kyr BP. Starting at 18 kyr BP, δD_{precip} becomes more positive, and averages -73
276 ‰ between 14 and 10 kyr BP. At the start of the Holocene, δD_{precip} shifts toward more
277 negative values toward the mid-Holocene, and averages -78 ‰ between 6 and 3 kyr BP.
278 Afterwards, δD_{precip} shifts to more positive values of approximately -73 ‰ in the late
279 Holocene.

280 The reconstructed late Holocene δD_{precip} value of -73 ‰ is slightly more negative
281 than δD_{precip} values of precipitation collected in recent years at two Sulawesi sites,
282 Timampu (Figure 5) and Makassar (Figure S3), where the amount-weighted annual
283 average δD_{precip} are ~-61‰ and ~-44 ‰, respectively (Konecky et al. 2016, Kurita et al.
284 2009). The excess depletion observed in Mandar Bay might be due to elevation effects,
285 given the 1170 m mean elevation of the Saddang River catchment, or due to inaccurate
286 estimates of the biosynthetic fractionation of deuterium vs. hydrogen in the formation of
287 leaf waxes by central Indonesian vegetation. Whatever the case, our reconstructed δD_{precip}
288 is within the seasonal range of isotopic values observed regionally (Konecky et al. 2016,
289 Tierney et al. 2012).

290

291 **6. Discussion**

292

293 **6.1 Wallacean vegetation and rainfall histories since the LGM**

294

295 Vegetation records covering the LGM to present from Wallacea are limited to
296 pollen records from the Banda Sea (van der Kaars et al. 2000), Lake Tondano on the
297 northeastern tip of Sulawesi (Dam et al. 2001), and the Wanda site near Lake Matano in
298 central Sulawesi (Hope 2001), and to plant biomarker records from Lake Towuti (Russell
299 et al. 2014), Lake Matano (Wicaksono et al. 2015), and Savu Sea (Dubois et al. 2014; see
300 Figure 2). These records span latitudes extending from 1 °N to 8 °S, and despite the
301 scarcity of data over such a large region, exhibit spatially and temporally coherent
302 changes that fit the vegetation and precipitation gradients observed in Wallacea today, as
303 discussed below.

304 In the present, Wallacea's southernmost island, Sumba, has the driest, most
305 seasonal climate in the region, resulting in the dominance of tropical deciduous forests in
306 northern Sumba and savannas in its southern lowlands (Peel et al. 2007).
307 Correspondingly, the core-top $\delta^{13}\text{C}_{\text{wax}}$ values at Savu Sea, adjacent to Sumba, are the
308 most positive in the region (-25 ‰) and suggest substantial inputs from C_4 vegetation
309 (Dubois et al. 2014). On the other hand, the modern $\delta^{13}\text{C}_{\text{wax}}$ values recorded from Lake
310 Towuti and Lake Matano (Figure 6) in the mountainous region of central Sulawesi,
311 average -39 ‰ and -37 ‰, respectively, reflecting closed-canopy C_3 rainforests and
312 montane rainforests with limited to no C_4 input (Russell et al. 2014, Wicaksono et al.
313 2015). Compared to these records, our Mandar Bay core-top $\delta^{13}\text{C}_{\text{wax}}$ values ($\sim$ -27 ‰) are
314 similar to, but more negative than, Savu Sea, and much more positive than the records
315 from Lakes Matano and Towuti (Figure 6), reflecting the mix of C_4 grassland and tropical
316 deciduous forest that exists today in southwestern Sulawesi. This vegetation type likely

317 stems from the area's transitional precipitation regime between the drier and seasonal
318 Lesser Sunda Islands and the ever-wet interior central Sulawesi (Whitten et al. 1987).

319 Variations in $\delta^{13}\text{C}_{\text{wax}}$ in Mandar Bay and other Wallacean biomarker records
320 (Figure 6) provide insight into the nature of vegetation changes during and since the
321 LGM. During the LGM, $\delta^{13}\text{C}_{\text{wax}}$ values from all four Wallacean sites were more positive
322 compared to their modern day/Late Holocene values (Figure 8), suggesting rainforest
323 opening and, at some sites, grassland expansion, during the LGM. This trend is in
324 keeping with pollen-based LGM climate reconstructions from the region. Savu Sea pollen
325 records indicate that grasses were abundant during the LGM, reaching up to 50 % of
326 Sumba's vegetation (Dubois et al. 2014), an inference supported by $\delta^{13}\text{C}_{\text{wax}}$ analyses
327 from the same core. Hope (2001) observed grass pollen concentrations of ~30% in late
328 glacial sediments compared to ~15% in the Holocene at Wanda Site, near Lakes Matano
329 and Towuti, and palynological analyses of cores from the Banda Sea, from eastern
330 Wallacea, also suggest expanded grasslands during the LGM (van der Kaars et al. 2000).

331 While vegetation can be controlled by numerous environmental and climatic
332 variables, temperature is unlikely to be the primary driver of the vegetation changes
333 observed in Wallacea, as changes in the relative abundance of trees and grasses are not
334 strongly temperature sensitive. Moreover, the longer $\delta^{13}\text{C}_{\text{wax}}$ data from Matano and
335 Towuti do not appear to track changes in Sulu Sea surface temperatures, in particular
336 during MIS 2 and 3 (Russell et al. 2014, Wicaksono et al. 2015). Similarly, although
337 atmospheric CO_2 concentrations can influence the relative abundances of C_3 vs. C_4
338 grasses (Ehleringer et al. 1997), the expansion of grasses relative to tree taxa observed in

339 $\delta^{13}\text{C}_{\text{wax}}$ and palynological data in Wallacea implies changes in rainfall and associated
340 hydroclimate variables, rather than CO_2 -driven shifts within grassland communities.

341 Thus, we interpret the enriched $\delta^{13}\text{C}_{\text{wax}}$ values observed across Wallacea to
342 indicate increased water stress during the LGM, likely due to a reduction in precipitation,
343 a longer dry season, and associated changes in other hydroclimate parameters (e.g. lower
344 humidity). A variety of data supports these interpretations of $\delta^{13}\text{C}_{\text{wax}}$ and other
345 paleovegetation records as an indicator for changes in Wallacean hydroclimate. Sediment
346 cores from Lake Tondano in North Sulawesi indicate lake level lowering during Marine
347 Isotope Stage (MIS) 2/LGM, attributed to reduced precipitation (Dam et al. 2001). The
348 concentrations of Fe and redox-sensitive trace elements in Lake Towuti increased during
349 the LGM, which have been linked to a longer or more severe dry season that induced lake
350 mixing (Costa et al. 2015). Despite evidence for more open vegetation, sedimentary Ti,
351 Al, and K concentrations in Lake Towuti decreased, reflecting reduced fluvial erosion
352 likely due to reductions in precipitation (Russell et al. 2014, Costa et al. 2015). Lake
353 levels were also lower in Towuti (Vogel et al. 2015), and regional wetlands appear to
354 have been desiccated reflecting sustained drying (e.g. Hope et al. 2001). The dry LGM
355 climate observed in Wallacea is also coherent with other climate records across Maritime
356 Southeast Asia (Reeves et al. 2013). Cave guano samples measured from Peninsula
357 Malaysia and Palawan Island in the Philippines display enrichment in $\delta^{13}\text{C}$ during the
358 LGM, indicating a dry environment (Wurster et al. 2010). In Java, grasses dominated the
359 local vegetation around Rawa Danau during the LGM, indicating a drier climate (van der
360 Kaars et al. 2001). However, LGM rainforest refugia have been detected in the high
361 mountains of Sulawesi as well as other highland areas within the Indonesian archipelago.

362 $\delta^{13}\text{C}_{\text{wax}}$ records from a marine core off west Sumatra's high mountain ranges
363 (Niedermeyer et al. 2014), along with pollen records from a high-altitude lake in west
364 Sumatra (Newsome and Flenley 1988) indicate the presence of rainforests and little
365 change in climate during the LGM. These higher elevation records demonstrate the long-
366 standing importance of orographic precipitation in the Indonesian archipelago amidst
367 large-scale changes in atmospheric regimes, similar to the patterns we observe today in
368 Wallacea.

369 Gradients in the $\delta^{13}\text{C}_{\text{wax}}$ values through time suggest that the present-day climate
370 gradient of Wallacea, with drier climates to the south, has remained largely in place over
371 the past 26 kyr. Savu Sea $\delta^{13}\text{C}_{\text{wax}}$ remained more positive than records from Sulawesi
372 throughout the last 26 kyr, likely indicating that the Lesser Sunda Islands were always
373 drier than Sulawesi. It is particularly striking that our new $\delta^{13}\text{C}_{\text{wax}}$ record from Mandar
374 Bay exhibits a nearly constant offset from that of the Savu Sea, suggesting that the two
375 regions experienced very similar relative changes in climate and vegetation through time.
376 This similarity occurs despite the significantly drier climate throughout the region during
377 the LGM inferred from these records and the possibility of southward migration of the
378 ITCZ during the LGM (e.g. Chiang et al. 2003, Chiang and Bitz 2005, Lee et al. 2014,
379 Russell et al. 2014, Wicaksono et al. 2015).

380 In contrast, variations in the amplitudes and absolute values of central Sulawesi
381 $\delta^{13}\text{C}_{\text{wax}}$ data indicate more localized and heterogeneous vegetation changes, though with
382 a similar trend from a dry LGM to a wetter climate in the Holocene reflecting the larger-
383 scale hydrologic changes. Lake Towuti $\delta^{13}\text{C}_{\text{wax}}$ registers the largest change, marked by a
384 12-permil depletion of $\delta^{13}\text{C}_{\text{wax}}$ from the LGM to the present. This change must have

involved major vegetation shifts, from open forests and C₄-dominated grasslands, such as in present-day Sumba, to the closed-canopy rainforests that dominate the region today. Indeed, the $\delta^{13}\text{C}_{\text{wax}}$ values of LGM sediment at Towuti are similar to the core-top Savu $\delta^{13}\text{C}_{\text{wax}}$ (Dubois et al. 2014), while Holocene $\delta^{13}\text{C}_{\text{wax}}$ exhibits much more negative values similar to those observed in nearby Lake Matano and in regional soils today (Wicaksono et al. 2015). In this light, the vegetation around Lake Towuti has oscillated between end-member vegetation communities in present-day Wallacea. $\delta^{13}\text{C}_{\text{wax}}$ in Lake Matano tracks the changes observed elsewhere, but with more depleted values, suggesting that the higher-altitude catchment of Lake Matano remained relatively wet, even during the LGM (Wicaksono et al. 2015), likely due to sustained orographically-induced precipitation at higher elevations. At Mandar Bay, $\delta^{13}\text{C}_{\text{wax}}$ data indicate much more modest changes than observed in the lake records, with open forest and grassland ecosystems similar to present-day Sumba at the LGM transitioning to a mix of closed-canopy forests, tropical deciduous forests, and grasslands in the Holocene, similar to the vegetation mosaic inhabiting southwest Sulawesi today.

Comparison of $\delta^{13}\text{C}_{\text{wax}}$ records from Wallacea may also provide insight into the application of the leaf wax proxy itself. Although these records are not from identical regions, the marine $\delta^{13}\text{C}_{\text{wax}}$ records from Savu Sea and Mandar Bay exhibit far fewer millennial- and centennial-scale variations compared to the lacustrine records (Figure 8), have different mean values, and have lower-amplitude glacial-interglacial signals. This occurs despite the fact that sedimentation rates are higher in the Mandar Bay core than in the lacustrine records, limiting effects of bioturbation. While these differences could reflect real differences in regional vegetation, they may also reflect the different sizes of

408 leaf wax source areas and transport mechanisms between offshore marine and inland lake
409 records. Both the Savu Sea and Mandar Bay records have much larger leaf wax source
410 areas compared to Lakes Matano and Towuti. For instance, the catchment of Saddang
411 River, associated with the Mandar Bay records, has an area of 6,380 km², compared to
412 the smaller catchment areas for Lakes Matano and Towuti (287 and 983 km², Wicaksono
413 et al. 2015). Palynological studies have shown that larger catchments often incorporate
414 different vegetation types, resulting in more averaging and smaller amplitude changes (i.e.
415 a “smoothing effect”) than in smaller catchments (Sugita 2007). Moreover, although most
416 leaf waxes found in Mandar Bay cores likely originate from the catchment, some waxes
417 might also come from northerly far away sources, brought into the site by oceanic
418 currents, in particular the strong currents of the Indonesian Throughflow (Gordon et al.
419 2003). This long distance oceanic transport is likely to carry waxes more enriched in ¹³C
420 given the drier climate of the regions north of Wallacea, possibly contributing to the
421 relatively enriched $\delta^{13}\text{C}_{\text{wax}}$ values in the Savu Sea and Mandar Bay samples. Our data
422 thus suggest that $\delta^{13}\text{C}_{\text{wax}}$ from marine records reflect a spatially broader view of
423 environmental changes compared to lacustrine records, at least in this region.
424 Observations of marine and terrestrial $\delta\text{D}_{\text{wax}}$ records, as discussed in the following
425 section, suggest similar patterns.

426

427 **6.2 Climatic controls on Wallacean rainfall isotopes since the LGM**

428

429 Our new record of $\delta\text{D}_{\text{precip}}$ from Mandar Bay builds upon and extends a
430 reconstruction of $\delta\text{D}_{\text{wax}}$ from a sediment core collected close to our study site from

431 Makassar Strait (Tierney et al. 2012). The two records, as well as δD_{wax} measurements of
432 different *n*-acids from our Mandar Bay core, exhibit generally coherent trends (Figure 7).
433 Tierney et al. (2012) do not report δD_{wax} values for C_{28} *n*-acid, but the Makassar Strait C_{30}
434 *n*-acid δD_{wax} are very similar to Mandar Bay C_{30} *n*-acid δD_{wax} , and both records exhibit
435 similar D-depletion from the late glacial to the Holocene (Figure 7).

436 There are, however, differences in the two records. Following the shift to more
437 enriched values during the last deglaciation, δD_{wax} becomes more negative in our core at
438 ~13 kyr BP. This depletion in Mandar is slightly later than in the Makassar Strait record,
439 which displays D-depletion since ~14 kyr BP. During the Holocene, our records indicate
440 a minimum in δD_{wax} at ~5 kyr BP, while the Makassar Strait δD_{wax} is relatively stable
441 through the Holocene. There could be many reasons for the differences between the
442 Mandar Bay and Makassar Strait datasets. Limitations in dating may have caused some of
443 the differences between the two records. For example, although the Makassar Strait age
444 model is generally well constrained, there is a relatively long interval between two ^{14}C
445 ages of 11,223 and 13,990 yr BP (Tierney et al. 2012), during which time the two records
446 display different trends. The two records employ slightly different marine reservoir age
447 corrections; the age model for Tierney et al. records used a constant 500-yr reservoir age
448 (Linsley et al. 2010), a variable reservoir age estimate (600-yr reservoir age in this
449 particular interval) based on estimates from paired ^{14}C ages on foraminifera and wood
450 particles (Table S1, Schroeder et al. in press). The reservoir age varies during the
451 deglaciation, such that time-varying reservoir effects could create uncertainty in both age
452 models. It is also possible that, despite the proximity of the two records, they experience
453 subtle differences in leaf wax source areas, though this hypothesis is very difficult to

454 evaluate. Lastly, we note that many of the differences in the records approach the
455 analytical uncertainty of the δD_{wax} measurements of $\sim 2\%$. While we cannot fully explain
456 these differences, the two records agree in the D-depletion observed during the late
457 deglacial, and our δD_{wax} record from Mandar Bay shows strong similarities to isotopic
458 records from other sites in Wallacea.

459 Indeed, long-term changes in δD_{precip} data from Mandar Bay, Lake Towuti, and
460 Flores Cave show striking similarities (Figure 8). During the LGM, these data sets
461 display more negative reconstructed δD_{precip} . This D-depletion is followed by D-
462 enrichment during the deglaciation, and then a trend toward D-depletion beginning ~ 11
463 kyr BP. The amplitude of δD_{precip} oscillations vary between the records, but the present-
464 day reconstructed values generally match the modern δD_{precip} gradients in the region
465 (Figure S3): both measured and reconstructed present-day δD_{precip} reveal more negative
466 values in central Sulawesi compared to northern Borneo and Bali (Figure S3). Though
467 inaccuracies in the conversion of leaf wax- and cave-based data sets to δD_{precip} could
468 contribute to the differences in the reconstructed absolute values (Figures 4 and S2), the
469 reasonable fit between reconstructed and observed δD_{precip} gradients therefore likely
470 reflects real regional variations in the factors controlling precipitation isotopes, including
471 rainfall amount (i.e. the ‘amount effect’), moisture sources, and trajectories.

472 Orbital-scale shifts in δD_{precip} in Towuti, Mandar Bay, and Flores are strikingly
473 similar, with D-depletion during the LGM, D-enrichment during the deglaciation, and a
474 return to D-depleted compositions in the early Holocene. This coherence suggests that
475 these sites, extending from $\sim 2^\circ$ to 10° South, experienced similar atmospheric circulation
476 changes since 26 kyr BP. In this context, the depletion of δD_{precip} in these records during

477 the LGM is challenging to explain in light of the widespread drying observed in these
478 same records and across the region. The ‘amount effect’, i.e. the negative relationship
479 between the isotopic composition of precipitation and precipitation amount (Bowen,
480 2008; Dansgaard, 1964; Rozanski et al., 1992), is often assumed to be the primary control
481 of past changes in δD_{precip} and $\delta^{18}O_{\text{precip}}$ in the tropics. In a strong convective system, the
482 removal of isotopically enriched water via condensation results in isotopically depleted
483 water vapor, such that both water vapor and rainwater become increasingly depleted as
484 convection continues (Vuille et al. 2003). Further, reduced evaporation of falling
485 raindrops in a wetter atmospheric column may reduce re-enrichment of the rainwater
486 (Dansgaard 1964, Lawrence et al. 2004). While these physical processes are relatively
487 well-understood, applying the amount effect to the interpretation of negative δD_{precip} data
488 observed in Wallacea would indicate a regionally wet climate during the LGM (e.g.
489 Griffiths et al. 2009, Ayliffe et al. 2013), in contrast to most paleoenvironmental
490 indicators from the region (e.g. Reeves et al. 2013 and references therein, Russell et al.
491 2014, Wicaksono et al. 2015).

492 What shifts in the atmospheric regime could have caused this LGM isotopic
493 depletion? The northerly monsoon winds are known to carry D-depleted air masses from
494 the Indian and Pacific Oceans relative to air masses originated locally or from the south
495 (Kurita et al. 2009, Griffiths et al. 2009, Konecky et al. 2016, Moerman et al. 2013).
496 Indeed, the effects of monsoon circulation and seasonal ITCZ movement on rainfall
497 isotopes are visible in central Sulawesi in the present. Data from Timampu suggest that
498 the δD_{precip} is much more depleted than predicted from the amount effect when the
499 northerly Austral-Asian Monsoon transports rain-bearing air masses from the northern

500 Pacific in November-January (Figure 6, Konecky et al. 2016). Meanwhile, convective
501 and stratiform rainfall within the ITCZ, which characterizes most rainfall in central
502 Sulawesi in March-May, have “limited depletion” (Konecky et al. 2016) compared to
503 precipitation associated with the northerly Austral-Asian Monsoon (Figure 6), likely due
504 to shorter transport paths at this time. The “extra depletion” in central Sulawesi might be
505 predominantly a function of moisture source and rainout along the long path taken by
506 northerly air masses into Wallacea, highlighting the importance of advection and
507 upstream atmospheric processes in controlling δD_{precip} along long transport paths (e.g.
508 Brown et al. 2008, Moerman et al. 2013, Konecky et al. 2016). Similarly, more negative
509 δD_{precip} values are observed between December and March in Denpasar, the closest
510 monitoring station to Flores, when the northwest trade winds are strong over the region
511 (Griffiths et al. 2009, Figure S3).

512 Thus, a stronger Austral Asian summer monsoon during the LGM could have
513 brought more depleted precipitation during the LGM to Wallacea, as registered by
514 depleted δD_{precip} observed at Mandar Bay, Lake Towuti, and Flores (e.g. Ayliffe et al.
515 2013, Konecky et al. 2016). In contrast, reconstructed δD_{precip} from northern Borneo
516 (Partin et al. 2007) indicates LGM precipitation was isotopically enriched relative to the
517 Holocene. The D-enrichment in Borneo is consistent with decreased regional convective
518 activity and evidence for drier conditions in northern Borneo during the LGM (Wurster et
519 al. 2010). In the present, precipitation amounts and isotopic composition at Mulu,
520 northern Borneo, exhibit relatively weak seasonality, with little D-depletion during boreal
521 winter when the northerly monsoon strengthens over southern and central Indonesia
522 (Moerman et al. 2013, Figure S3). The strong similarity of the reconstructions over

Wallacea, where the northerly monsoon plays a critical role during the modern-day wet season, relative to Borneo, where boreal winter precipitation is less important, strongly implicates an intensified northerly monsoon in the D-depletion observed across Wallacea during the LGM (Ayliffe et al. 2013, Konecky et al. 2016).

In the present, fluctuations in the tropical monsoons are linked to the seasonal migration of the Intertropical Convergence Zone. During the LGM, expanded ice sheets and northern hemisphere cooling is thought to have strengthened the East Asian Winter Monsoon (EAWM; Porter and An 1995, Shi et al. 2011), and given the modern-day relationship between the EAWM, the Austral-Asian northerly/boreal winter monsoon, and the ITCZ (Suppiah and Wu 1998, Hong and Li 2009), strengthening of the EAWM should have been associated with a stronger Austral-Asian northerly monsoon and a southward shift of the ITCZ over Wallacea. Modeling studies further indicate the large volume and high topography of the Laurentide ice sheets drove a net southward shift of the mean ITCZ position (e.g. Chiang et al. 2003, Chiang and Bitz 2005, Lee et al. 2014), consistent with a strengthened northerly monsoon. However, our vegetation reconstructions and other regional records indicate (e.g. Reeves et al. 2013) indicate that southward ITCZ migration and strengthened monsoon circulation did not lead to increased rainfall in Wallacea, indicating a reduction in precipitation within the ITCZ itself during the LGM relative to the Holocene. While the exact mechanism driving this reduction is unclear, possible causes include drying of the tropical atmosphere under the globally colder conditions of the LGM, regionally cooler SSTs, and changes in evaporation and evapotranspiration, partly related to increased exposure of land areas due to lowered sea level (e.g. DiNezio and Tierney 2013, Konecky et al. 2016). Alternatively,

one could speculate that the ITCZ migrated much further south, i.e. south of Wallacea; however such a large shift in the ITCZ over Maritime Southeast Asia during the LGM cannot be reconciled with the large changes in atmospheric heat transport that accompany ITCZ displacement (McGee et al. 2014).

Although global climate boundary conditions influenced the ITCZ during the LGM, a southward shift in the ITCZ, a strengthened northerly monsoon, and D-depleted rainfall may also be caused by orbital precession and seasonal insolation forcing. Indeed, Wallacean δD_{precip} records generally track southern hemisphere summer insolation throughout the past 26 kyr BP (Figure 8), suggesting a dynamical linkage between orbital precession, ITCZ displacement, and monsoon strength (e.g. Holbourn et al. 2005, Tachikawa et al. 2011, Kuhnt et al. 2015). Strong southern hemisphere summer insolation during the LGM might have pulled the caloric equator toward the southern tropics, shifting the seasonal and mean annual ITCZ positions southward. Moreover, weak southern hemisphere summer insolation ca. 12 kyr BP is synchronous with the maximum D-enrichment observed in the Wallacean δD_{precip} records, including Mandar Bay, Flores, and Towuti, when the ITCZ shifted to the north (Wang et al. 2001, Wang et al. 2008). As the ITCZ shifts northward, most of Wallacea was subject to increased influence of isotopically enriched precipitation brought by the southeasterlies, therefore increasing the overall δD_{precip} values. The strengthening of the northerly monsoon as the ITCZ shifted to the south after the Younger Dryas and during the Holocene is also recorded in the marine record from Timor Sea, to the south of the Lesser Sunda Islands (Kuhnt et al. 2015). Despite the correspondence between Wallacean δD_{precip} and southern hemisphere summer insolation, the D-depletion we observe at Mandar Bay and other Wallacean sites during

569 the LGM relative to the Holocene cannot be explained solely by insolation forcing,
570 implicating glacial boundary conditions in forcing the inferred changes in monsoon
571 strength and ITCZ position.

572 Our records may further explain the limited glacial-interglacial changes observed
573 in planktic foraminifera-derived $\delta^{18}\text{O}$ of seawater ($\delta^{18}\text{O}_{\text{sw}}$) records from offshore western
574 Sumatra (Mohtadi et al. 2014), off south Java (Mohtadi et al. 2011), and the Savu Basin
575 (Gibbons et al. 2014). The $\delta^{18}\text{O}_{\text{sw}}$ has been interpreted to primarily reflect mean annual
576 local sea surface salinity, likely controlled by shifts in regional precipitation/evaporation
577 and river runoff from nearby islands, and to reflect some influence from oceanic
578 advection and mixing of water with different isotopic compositions. In this context, the
579 authors explained the small LGM-present differences in $\delta^{18}\text{O}_{\text{sw}}$ as indicative of relative
580 similarities in precipitation amount during the LGM and present. This interpretation,
581 however, is in contrast to the substantial drying recorded by regional paleoenvironmental
582 records, as discussed above. Instead, it is possible that isotope depletion in rainwater
583 during the LGM due to shifts in glacial boundary conditions, orbital precession, and
584 seasonal insolation forcing has offset changes in precipitation amount and runoff in
585 regional $\delta^{18}\text{O}_{\text{sw}}$ records.

586

587 **6.3 Time-varying aspect of precipitation- $\delta\text{D}_{\text{precip}}$ relationship in Wallacea**

588

589 A prominent feature of our Mandar Bay data is a switch in the relationship
590 between $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ (Figure 3). $\delta\text{D}_{\text{precip}}$ is negatively correlated with $\delta^{13}\text{C}_{\text{wax}}$
591 during the LGM and the last deglaciation, but positively correlated during the Holocene.

592 Assuming $\delta^{13}\text{C}_{\text{wax}}$ records changes in precipitation amount, this switch could suggest that
593 seasonal changes in moisture source and atmospheric circulation dominate the $\delta\text{D}_{\text{precip}}$
594 signal when glacial boundary conditions prevailed, while $\delta\text{D}_{\text{precip}}$ records primarily
595 changes in precipitation amount during the Holocene. Although the precise changes in
596 atmospheric circulation remain unclear, this switch likely signifies a major shift in
597 atmospheric circulation and regional convection near the beginning of the Holocene.

598 By calculating a moving correlation window, i.e. searching for the time of the
599 switch that optimizes R-squared values for both the correlation and anti-correlation, we
600 found that the transition occurs between 11.1 and 9.4 kyr BP (Figure S4), suggesting that
601 rising sea level and flooding of the regional shelves exert an important control on
602 atmospheric circulation and precipitation isotopes (e.g. Griffiths et al., 2011). At 11.1 kyr
603 BP, the rapid inundation of the Sunda Shelf, associated with the Meltwater Pulse 1B, had
604 just been completed (Fairbanks 1989), resulting in the initial opening of the South China
605 Sea and the Java Sea (Sathiamurthy and Voris 2006, Hanebuth et al. 2009). Between 11.5
606 and 9.5 kyr BP, sea level rose from ~50 to ~20 meter below present level and flooded the
607 Karimata Strait and, more importantly, the rest of the Java Sea (Sathiamurthy and Voris
608 2006, Hanebuth et al. 2009), thereby creating a large moisture source region within the
609 Indonesian archipelago during this time window (Figure S4). Thus, by 9.4 kyr BP, the
610 channels and straits separating Java, Kalimantan, Sumatra, and the Malay Peninsular
611 were already becoming significant water bodies (Hanebuth et al. 2009).

612 While uncertainty remains regarding the effects of Sunda Shelf exposure on
613 regional atmospheric circulation (DiNezio and Tierney 2012), the presence of large,
614 warm seas in western Indonesia appears to have increased the relative influence of the

615 amount effect during the Holocene. Interestingly, the latter part of the time window of the
616 switch is coincident with an abrupt, non-linear switch in the monsoon system above
617 Flores at ~9.5 kyr BP observed in the correlations between $\delta^{18}\text{O}_{\text{speleothem}}$, Mg/Ca, and
618 Sr/Ca data from Flores cave and attributed to the rapid inundation of the Sunda shelf at
619 that time (Griffiths et al. 2011). Our data, however, suggest that the switch in the
620 atmospheric circulation or the monsoon system occurred gradually, in line with the
621 relatively gradual inundation of the Sunda Shelf. In particular, the flooding between 11.1
622 and 9.4 kyr BP likely represent a key interval for the establishment of the modern-day
623 relationship between precipitation and $\delta\text{D}_{\text{precip}}$ in the region.

624 During the Holocene, following the flooding of the Java Sea, records from
625 Wallacea exhibit much more heterogeneity. Borneo exhibits the most depleted $\delta\text{D}_{\text{precip}}$ at
626 ~4.5 kyr BP, Towuti shows more depleted $\delta\text{D}_{\text{precip}}$ at ~3.5 kyr BP, and Flores $\delta\text{D}_{\text{precip}}$ is
627 relatively stable (Figure 8). Our record from Mandar Bay shows more depleted $\delta\text{D}_{\text{precip}}$ at
628 ~4.5 kyr BP, similar to Borneo, though this finding contrasts with nearby $\delta\text{D}_{\text{precip}}$ records
629 from western Sulawesi (Tierney et al. 2012); thus, the trends in our Mandar Bay data
630 must be evaluated with caution. The heterogeneity in these records, in the absence of
631 strong glacial forcings, could reflect different sensitivities among the sites to insolation-
632 forced changes in the ITCZ and the monsoons (Partin et al. 2007, Griffiths et al. 2009)
633 and/or to coupled ocean-atmosphere processes operating in the Pacific and Indian Oceans
634 during the Holocene. The variations in Borneo rainfall isotopes during the Holocene have
635 been linked to insolation during boreal fall, which controls the intensity of precipitation
636 as the ITCZ passes above the site in the fall (Partin et al. 2007, Carolin et al. 2013). As
637 the mean position of the ITCZ shifts southwards over the course of the Holocene in

638 response to fall precessional forcing (Carolin et al. 2013), the peak of intense
639 precipitation and the isotopic minima also shift southward across the region (Partin et al.
640 2007). Combined with the depletion observed at Towuti ca. 3.5 kyr BP, these records
641 may jointly indicate the crossing of the ITCZ over the northern part of Indonesian
642 archipelago during the mid-Holocene. The relative stability of Holocene isotopic
643 compositions at Flores thus likely indicates that precipitation over the Lesser Sunda
644 Islands was less influenced by a shifting ITCZ during the Holocene, possibly because of
645 other dominant controlling factors such as sea surface temperatures.

646

647 **7. Conclusions**

648

649 $\delta^{13}\text{C}_{\text{wax}}$ measured from the Mandar Bay core indicate that regional vegetation is
650 highly sensitive to hydroclimatic changes, and that the largely meridional present-day
651 climate gradient of Wallacea has existed since at least 26 kyr BP. Wallacean
652 paleovegetation records document grassland expansion into many currently forested areas
653 in the region, implying a drier climate and increased seasonality during the LGM.
654 Wallacean $\delta\text{D}_{\text{precip}}$ records have similar trends during the past 26 kyr BP, with significant
655 isotopic depletion during a dry LGM implying a different rainfall isotope dynamic
656 compared to today. Our Mandar Bay records demonstrate that $\delta\text{D}_{\text{precip}}$ was anti-correlated
657 with $\delta^{13}\text{C}_{\text{wax}}$ during the LGM and the last deglaciation, but was positively correlated
658 during most of the Holocene. Based upon the similarity of the isotopic depletion in
659 records from equatorial and southern Wallacea, where the Austral-Asian northerly
660 monsoon dominates present-day precipitation patterns, we argue that the LGM depletion

661 was caused by a southward shift of the ITCZ mean position and an intensified, yet drier
662 northerly Austral-Asian monsoon. This circulation pattern, in turn, was likely forced by
663 stronger southern hemisphere summer insolation and the presence of large northern
664 hemisphere ice sheets. At Mandar Bay, $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ become positively correlated
665 during most of the Holocene. The switch in correlation likely occurred gradually between
666 11.1 and 9.4 kyr, when the Java Sea and Karimata Strait were flooded by the rising sea
667 level, suggesting that these new moisture sources and/or changes in moisture transport
668 pathways increased the influence of the amount effect in the region.

669 To further understand the LGM depletion of rainfall isotopes and better
670 characterize the extent and impact of ITCZ movements and seasonal variations on
671 Wallacean $\delta\text{D}_{\text{precip}}$ across multiple timescales, we will need to couple a larger spatial
672 coverage of precipitation isotope records from the region with a systematic study of high-
673 resolution isotope-enabled climate model simulations under different scenarios of
674 forcings and boundary conditions. Importantly, the time-dependent and varying controls
675 on the isotopic composition of rainfall suggest more caution is needed when interpreting
676 long-term rainfall isotope proxy data in this climatically complex region.

677

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686

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971 **Figure Captions**

972

973 **Figure 1** Map of Maritime Southeast Asia indicating locations discussed in the text.

974 During the Last Glacial Maximum, the exposed Sundaland was separated from Sahul by

975 deep and narrow straits, along with a group of islands termed Wallacea, comprised of

976 Sulawesi, Maluku islands, and Lesser Sunda Islands (from GeoMapApp, bathymetric

977 data from General Bathymetric Chart of the Ocean 2008). Locations of our study site

978 (Mandar Bay) and other paleoclimate records as well as rainfall isotope stations discussed

979 in the text are marked on the map.

980

981 **Figure 2** Map of Köppen–Geiger climate-vegetation classification for Wallacea (adapted

982 from Peel et al. 2007). Ever-wet rainforests dominate central Sulawesi, whereas monsoon

983 forests and savannas are common in southern Wallacea. Locations of our study site (core

984 SO18515, Mandar Bay) and other records discussed in the text are marked with white

985 stars. White dots indicate the locations of rainfall and/or rainfall isotope stations

986 discussed in the text.

987

988 **Figure 3** Mandar Bay $\delta^{13}\text{C}_{\text{wax}}$ (green) and $\delta\text{D}_{\text{precip}}$ (blue) since 26 kyr BP. Bold-colored

989 lines are a Stineman smoothing function, which applies a geometric weight to each data

990 point and ~10% of the data range. $\delta\text{D}_{\text{precip}}$ is largely anti-correlated with $\delta^{13}\text{C}_{\text{wax}}$ during

991 the LGM and the last deglaciation (the time period indicated by the purple arrow), but is

992 positively correlated during most of the Holocene (red arrow). The shaded area indicates

993 the transition period during which the switch in correlation occurs. Gray dots at the
994 bottom indicate the age control points for core SO18515.

995

996 **Figure 4** Comparison between the raw Mandar Bay δD_{wax} records (pink), ice volume-
997 corrected δD_{wax} (purple), and vegetation-corrected δD_{precip} (blue).

998

999 **Figure 5** Monthly mean precipitation for Timampu near Lake Towuti from 1988 to 2015
1000 (light blue bars; PT Vale Indonesia, unpublished), as well as for Saddang catchment close
1001 to Mandar Bay (pink bars) show that the two regions in central Sulawesi have a similar
1002 precipitation seasonality. The Saddang catchment data are the averages from three WMO
1003 stations within the catchment of Saddang River, namely Rantepao (34 year dataset),
1004 Mamasa (37 year dataset), and Enrekang (39 year dataset) obtained from the GHCN
1005 (2013). Green and red curves display δD_{precip} and $\delta^{18}\text{O}_{\text{precip}}$, respectively, measured from
1006 Timampu rainfall since 2013 (Konecky et al. 2016). At the bottom are back-trajectory
1007 analyses for Mandar Bay. The solid lines represent HYSPLIT (Draxler and Rolph 2013)-
1008 calculated 168-h back-trajectories of air parcels that generated rainfall during four
1009 trimesters (December-February, March-May, June-August, September-November) of
1010 July 2003–June 2004, a 1-year interval that did not experience El Niño Southern
1011 Oscillation and Indian Ocean Dipole-related anomalies. Different colors represent the
1012 height at which the air parcel originates. Back trajectory analyses for Timampu yield very
1013 similar patterns.

1014

1015 **Figure 6** Comparison of Mandar Bay $\delta^{13}\text{C}_{\text{wax}}$ record with other Wallacean $\delta^{13}\text{C}$ records.
1016 From top to bottom: 1) Lake Matano $\delta^{13}\text{C}_{\text{wax}}$ (Wicaksono et al. 2015) in red, 2) Lake
1017 Towuti $\delta^{13}\text{C}_{\text{wax}}$ (Russell et al. 2014) in purple, 3) Mandar Bay (this study) in green, and
1018 4) Savu Sea $\delta^{13}\text{C}_{\text{wax}}$ (Dubois et al. 2014) in brown.

1019

1020 **Figure 7** Comparison between C_{28} and C_{30} *n*-acid Mandar Bay $\delta\text{D}_{\text{wax}}$ records (dark blue
1021 and purple respectively) with Makassar Strait raw, C_{30} *n*-acid $\delta\text{D}_{\text{wax}}$ data (pink, Tierney et
1022 al. 2012).

1023

1024 **Figure 8** Comparison of Mandar Bay $\delta\text{D}_{\text{precip}}$ records with climate forcings and other
1025 Wallacean $\delta\text{D}_{\text{precip}}$ records. From top to bottom: 1) sea level fluctuations as a proxy for
1026 global ice volume, derived from benthic foraminiferal oxygen isotopes, as described in
1027 Waelbrock et al. (2002), 2) Mandar Bay $\delta\text{D}_{\text{precip}}$ (this study), 3) Lake Towuti $\delta\text{D}_{\text{precip}}$
1028 (Konecky et al. 2016), 4) Northern Borneo /Mulu Cave $\delta\text{D}_{\text{precip}}$ (computed from
1029 $\delta^{18}\text{O}_{\text{speleothem}}$ published in Partin et al. 2007), 5) Flores Cave $\delta\text{D}_{\text{precip}}$ (computed from
1030 $\delta^{18}\text{O}_{\text{speleothem}}$ published in Griffiths et al. 2009 and Ayliffe et al. 2013), 5) southern
1031 hemisphere summer insolation (calculated for 20°S , after Laskar et al. 2004).

1032

Legend

- Ever-wet rainforests
- Tropical deciduous forests
- Savanna

1. Northern Borneo cave
2. Lake Tondano
3. Lake Matano
4. Lake Towuti
5. Mandar Bay
6. Makassar Strait
7. Flores cave
8. Savu Sea

Yr BP

0 5000 10000 15000 20000 25000 30000

Sea Level
(m below present)

0
-20
-40
-60
-80
-100
-120

Mandar Bay

-90

Towuti

δD_{Precip} (‰)

Mulu/Borneo

Flores

Summer Insolation
(20 °S, W/m²)

490
480
470
460
450

Yr BP

0 5000 10000 15000 20000 25000 30000