

1 Late-glacial to late-Holocene shifts in global precipitation $\delta^{18}\text{O}$

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22 **Abstract**

23 Reconstructions of Quaternary climate are often based on the isotopic content of paleo-
24 precipitation preserved in proxy records. While many paleo-precipitation isotope records are
25 available, few studies have synthesized these dispersed records to explore spatial patterns of
26 late-glacial precipitation $\delta^{18}\text{O}$. Here we present a synthesis of 86 globally-distributed
27 groundwater (n=59), cave calcite (n=15) and ice core (n=12) isotope records spanning the
28 late-glacial (defined as ~50,000 to ~20,000 years ago) to the late-Holocene (within the past

1 ~5,000 years). We show that precipitation $\delta^{18}\text{O}$ changes from the late-glacial to the
2 late-Holocene range from -7.1‰ ($\delta^{18}\text{O}_{\text{late-Holocene}} > \delta^{18}\text{O}_{\text{late-glacial}}$) to $+1.7\text{‰}$ ($\delta^{18}\text{O}_{\text{late-glacial}} >$
3 $\delta^{18}\text{O}_{\text{late-Holocene}}$), with the majority (77%) of records having lower late-glacial $\delta^{18}\text{O}$ than
4 late-Holocene $\delta^{18}\text{O}$ values. High-magnitude, negative precipitation $\delta^{18}\text{O}$ shifts are common at
5 high latitudes, high altitudes and continental interiors ($\delta^{18}\text{O}_{\text{late-Holocene}} > \delta^{18}\text{O}_{\text{late-glacial}}$ by more
6 than 3‰). Conversely, low-magnitude, positive precipitation $\delta^{18}\text{O}$ shifts are concentrated along
7 tropical and subtropical coasts ($\delta^{18}\text{O}_{\text{late-glacial}} > \delta^{18}\text{O}_{\text{late-Holocene}}$ by less than 2‰). Broad, global
8 patterns of late-glacial to late-Holocene precipitation $\delta^{18}\text{O}$ shifts suggest that stronger-than-
9 modern isotopic distillation of air masses prevailed during the late-glacial, likely impacted by
10 larger global temperature differences between the tropics and the poles. Further, to test how
11 well general circulation models reproduce global precipitation $\delta^{18}\text{O}$ shifts, we compiled
12 simulated precipitation $\delta^{18}\text{O}$ shifts from five isotope enabled general circulation models
13 simulated under recent and last glacial maximum climate states. Climate simulations generally
14 show better inter-model and model-measurement agreement in temperate regions than in the
15 tropics, highlighting a need for further research to better understand how inter-model spread in
16 convective rainout, seawater $\delta^{18}\text{O}$ and glacial topography parameterizations impact simulated
17 precipitation $\delta^{18}\text{O}$. Future research on paleo-precipitation $\delta^{18}\text{O}$ records can use the global maps
18 of measured and simulated late-glacial precipitation isotope compositions to target and
19 prioritize field sites.

20 **1 Introduction**

21 Isotopic compositions of late-glacial precipitation can be preserved in groundwaters, cave
22 calcite, glacial ice, ground ice and lake sediments. These records have been used to better
23 understand past climate changes for more than a half century (e.g., Münnich, 1957; Thatcher et
24 al., 1961; Münnich et al., 1967; Pearson and White, 1967; Tamers, 1967; Gat et al., 1969). Each
25 type of isotopic proxy record is distinguished by its temporal resolution, preservation of one or
26 both $^{18}\text{O}/^{16}\text{O}$ and $^2\text{H}/^1\text{H}$ ratios, and frequency on land surface. For example, groundwater
27 records contain both $^{18}\text{O}/^{16}\text{O}$ and $^2\text{H}/^1\text{H}$ ratios with widespread global occurrence, but have a
28 coarser temporal resolution than other paleoclimate proxies (Rozanski et al., 1985; Edmunds
29 and Milne, 2001; Edmunds, 2009; Corcho Alvarado et al., 2011; Jiráková et al., 2011).
30 Speleothem records, in contrast, have high temporal resolution but usually only report calcite
31 $^{18}\text{O}/^{16}\text{O}$ ratios (without fluid inclusion $^2\text{H}/^1\text{H}$ data) and are less common than groundwater
32 records (e.g., Harmon et al., 1978; 1979). Late-glacial ice core and ground ice records have

1 high temporal resolution, can be analysed for $^{18}\text{O}/^{16}\text{O}$ and $^2\text{H}/^1\text{H}$ ratios, but are rare on non-
2 polar lands (Dansgaard et al., 1982; Thompson et al., 1989; 1995; 1997; 1998). Lake sediment
3 records can have a high temporal resolution, can preserve $^{18}\text{O}/^{16}\text{O}$ and $^2\text{H}/^1\text{H}$ ratios and are
4 available for a multitude of globally-distributed locations (e.g., Edwards et al., 1989; Eawag et
5 al., 1992; Menking et al., 1997; Wolfe et al., 2000; Anderson et al., 2001; Beuning et al., 2002;
6 Sachse et al., 2004; Morley et al., 2005; Tierney et al., 2008). However, some lake water proxy
7 isotope records may be impacted by paleo-lake evaporative isotope effects that obscure the
8 primary meteoric water signal and mask paleo-precipitation isotope compositions (e.g., lake
9 sediment calcite, diatom silica; Leng and Marshall, 2004).

10 This study examines speleothem, ice core and groundwater isotope records, focusing primarily
11 on the groundwater isotope records due to their relative density in the published literature in
12 comparison to the more limited number of published speleothem and ice core records
13 (compilations by Pedro et al., 2011; Stenni et al., 2011; Clark et al., 2012; Shah et al., 2013;
14 Caley et al., 2014a). There exist roughly twice as many groundwater reconstructions of
15 late-glacial to late-Holocene precipitation $\delta^{18}\text{O}$ shifts ($n=59$) as the combined total of
16 speleothem and ice core records ($n=27$; where $\delta^{18}\text{O} = (^{18}\text{O}/^{16}\text{O}_{\text{sample}}) /$
17 $(^{18}\text{O}/^{16}\text{O}_{\text{standard mean ocean water}} - 1) \times 1000$). A recent global synthesis of paired precipitation-
18 groundwater isotopic data demonstrated that modern annual precipitation and modern
19 groundwater isotope compositions follow systematic relationships with some bias toward
20 winter and wet-season precipitation (Jasechko et al., 2014). Systematic rainfall-recharge
21 relationships shown by Jasechko et al. (2014) support our primary assumption in this study that
22 groundwater isotope compositions closely reflect meteoric water. Because groundwater records
23 can only identify climate change occurring over thousands of years due to hydrodynamic
24 dispersion during multi-millennial residence times (e.g., Davison and Airey, 1982; Stute and
25 Deak, 1989), we limit the focus of this study to meteoric water isotope composition changes
26 from the latter half of the last glacial time period to the late-Holocene. The latter half of the last
27 glacial period is defined as $\sim 20,000$ to $\sim 50,000$ years before present, using the end of the last
28 glacial maximum as the more recent age limit ($\sim 20,000$ years before present; Clark et al., 2009)
29 and the maximum age of groundwater that can be identified by ^{14}C dating as an approximate
30 upper age limit (i.e., groundwater ages more recent than $\sim 50,000$ years old).

31 For brevity, we refer herein to the time period representing the latter half of the last glacial
32 period ($\sim 20,000$ to $\sim 50,000$ years before present) as the *late-glacial* (e.g., $\delta^{18}\text{O}_{\text{late-glacial}}$). We

1 adopt a definition of the *late-Holocene* as occurring within the last 5,000 years following
2 Thompson et al. (2006). Other work proposes the late-Holocene be defined as within the last
3 4,200 years (Walker et al., 2012), which is consistent with the 5,000 years before present
4 definition (Thompson et al., 2006) within the practical uncertainty of ^{14}C -based groundwater
5 ages ($\pm \sim 10^3$ years). Further, although precipitation isotope compositions have varied over the
6 late-Holocene, groundwater mixing integrates this variability, prohibiting paleoclimate
7 interpretation at finer temporal resolutions.

8 Late-glacial to late-Holocene changes in precipitation isotope compositions provide important
9 insights into conditions and processes of the past. Perhaps the two best-constrained global-in-
10 scale differences between the late-glacial and the late-Holocene are changes to oceanic and
11 atmospheric temperatures (MARGO Members, 2009; Shakun and Carlson, 2010; Annan and
12 Hargreaves, 2013), and changes to seawater $\delta^{18}\text{O}$ (Emiliani, 1955; Dansgaard and Tauber,
13 1969; Schrag et al., 1996; 2002). Atmospheric temperatures have increased by a global average
14 of $\sim 4^\circ\text{C}$ since the last glacial maximum, with greatest warming at the poles and more modest
15 warming at lower latitudes (Figure 1; Shakun and Carlson, 2010; Annan and Hargreaves, 2013).
16 Seawater $\delta^{18}\text{O}$ during the last glacial maximum was 1.0 ± 0.1 ‰ higher than the modern ocean,
17 as constrained by paleo-ocean water samples collected from pore waters trapped within sea
18 floor sediments (Schrag et al., 2002).

19 Previous studies have proposed many different interpretations of past changes to precipitation
20 isotope compositions. Records of paleo-precipitation $\delta^{18}\text{O}$ have been used as a proxy for
21 regional land surface and atmospheric temperature (e.g., Rozanski, 1985; Nikolayev and
22 Mikhalev, 1995; Johnsen et al., 2001; Grasby and Chen, 2005; Akouvi et al., 2008; Bakari et
23 al., 2012); however, $\delta^{18}\text{O}$ -based paleotemperatures can be complicated by past changes to a
24 variety of other processes controlling precipitation $\delta^{18}\text{O}$, including moisture sources, upwind
25 rainout, transport pathways, moisture recycling and in-cloud processes (Ciais and Jouzel, 1994;
26 Masson-Delmotte et al., 2005; Sjoström and Welker, 2009). Process-based explanations for
27 observed meteoric water $\delta^{18}\text{O}$ variations in proxy records include changes to hurricane intensity
28 (e.g., Plummer et al., 1993), large-scale atmospheric circulation (e.g., Rozanski et al., 1985;
29 Weyhenmeyer et al., 2000; McDermott et al., 2001; Pausata et al., 2009; Asmerom et al., 2010;
30 Oster et al., 2015), aridity (e.g., Wagner et al., 2010), monsoon strength (e.g., Denniston et al.,
31 2000; Lachniet et al., 2004; Liu et al., 2007; Pausata et al., 2011a), local seawater $\delta^{18}\text{O}$ (Wood
32 et al., 2003; Feng et al., 2014), precipitation seasonality (e.g., Fawcett et al., 1997; Werner et

1 al., 2000; Cruz et al., 2005), moisture provenance (e.g., Sjoström and Welker 2009; Lewis et
2 al., 2010), storm tracks, climate oscillation modes (e.g., North Atlantic oscillation), moisture
3 recycling (e.g., Winnick et al., 2013; 2014; Liu et al., 2014a; 2014b) and groundwater flow path
4 architecture (Purdy et al., 1996; Stewart et al., 2004; Morrissey et al., 2010; Hagedorn, 2015).
5 While unravelling these mechanisms and delineating the primary and secondary processes can
6 be rather challenging, the use of climate models in combination with robust and extensive
7 precipitation isotope data can resolve many of these complexities with meaningful
8 interpretations and insight.

9 The objective of this study is to analyse spatial patterns of measured late-glacial to
10 late-Holocene precipitation $\delta^{18}\text{O}$ changes from published groundwater, ground ice, glacial ice
11 and cave calcite records, and to compare these measurements with output from five state-of-
12 the-art isotope-enabled general circulation model simulations of last glacial maximum and pre-
13 industrial or modern climate conditions. Synthesizing paleowater $\delta^{18}\text{O}$ records provides an
14 important constraint for isotope-enabled general circulation model simulations of atmospheric
15 and hydrologic conditions during glacial climate states (Jouzel et al., 2000). We combine a new
16 global compilation of late-glacial groundwater and ground ice isotope data ($n=59$) with existing
17 compilations for speleothems ($n=15$; Shah et al., 2013) and ice cores ($n=12$; Pedro et al., 2011;
18 Stenni et al., 2011; Clark et al., 2012; Caley et al., 2014a). This compilation of late-glacial
19 groundwater isotope compositions builds from earlier reviews of European and African
20 paleowater isotope compositions (Rozanski, 1985; Edmunds and Milne, 2001; Darling, 2004;
21 Edmunds, 2009; Négrel and Petelet-Giraud, 2011; Jiráková et al., 2011).

22 **2 Dataset and Methods**

23 In order to examine spatial patterns of change to meteoric water $\delta^{18}\text{O}$ values we compiled $\delta^{18}\text{O}$,
24 $\delta^2\text{H}$, $\delta^{13}\text{C}$ and ^{14}C data from 1713 groundwater samples collected from 59 aquifer systems
25 reported in 76 publications (data and primary references presented in the Supplement). $\delta^{13}\text{C}$,
26 ^3H and ^{14}C data were used to estimate groundwater age (details within Supplement). Changes
27 to precipitation $\delta^{18}\text{O}$ values over time were determined by comparing groundwater isotope
28 compositions of the late-Holocene ($\delta^{18}\text{O}_{\text{late-Holocene}}$ defined here as less than 5,000 years before
29 present; Thompson et al., 2006) and the latter half of the last glacial time period ($\delta^{18}\text{O}_{\text{late-glacial}}$:
30 20,000 to ~50,000 years before present). We acknowledge that these two relatively long time
31 intervals—necessarily long in order to examine groundwater isotope records—integrate
32 precipitation $\delta^{18}\text{O}$ variability over the course of each time interval. The late-Holocene time

1 interval integrates known precipitation $\delta^{18}\text{O}$ variability (e.g., Aichner et al., 2015), and the
2 late-glacial time interval likely incorporates groundwater preceding the last glacial maximum,
3 potentially during Marine Isotope Stage 3 or even older glacial time periods due to large
4 uncertainties in ^{14}C -based groundwater ages (Supplement).

5 Proxy-based meteoric water $\delta^{18}\text{O}$ changes from the latter half of the last glacial time period to
6 the late-Holocene are described herein as *measured* $\Delta^{18}\text{O}_{\text{late-glacial}}$, where measured $\Delta^{18}\text{O}_{\text{late-glacial}}$
7 = $\delta^{18}\text{O}_{\text{late-glacial}} - \delta^{18}\text{O}_{\text{late-Holocene}}$. A minimum groundwater age of 20,000 years before present
8 was used to define the late-glacial to remain consistent with the timing of the last glacial
9 maximum (~20,000 years before present; Clark et al., 2009). Samples having a deuterium
10 excess of less than zero (deuterium excess = $\delta^2\text{H} - 8 \times \delta^{18}\text{O}$; Dansgaard, 1964) and falling along
11 regionally-characteristic evaporation $\delta^2\text{H}/\delta^{18}\text{O}$ slopes (Gibson et al., 2008) were removed from
12 the analysis to avoid including groundwater samples impacted by partial evaporation. Further,
13 studies reporting saltwater intrusion were avoided on the basis of groundwater $\delta^{18}\text{O}$ and
14 salinities showing evidence of seawater mixing (e.g., Schiavo et al., 2009; Yechieli et al., 2009;
15 Hamouda et al., 2011; Han et al., 2011; Wang and Jiao, 2012; Currell et al., 2013). The 59
16 compiled groundwater measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are unevenly distributed among western
17 Europe (n=10), eastern Europe and the Middle-East (n=12), Africa (n=17), southeastern Asia
18 (n=6), Australia, Oceania and the Malay Archipelago (n=2), South America (n=2), temperate
19 and subtropical North America (n=8) and the High Arctic (n=2). Half of the compiled
20 groundwater records are located in the tropics or subtropics (that is, within 35° of the equator;
21 n=29) and half are located in the extra-tropics (n=30).

22 Speleothem and ice core isotope proxy records were also compiled. Lacustrine sediment $\delta^{18}\text{O}$
23 records are not considered in this study because these records may preserve meteoric waters
24 impacted by evaporative isotope effects (Leng and Marshall, 2004). Speleothem and ice core
25 measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values were calculated by subtracting average $\delta^{18}\text{O}$ values for each of
26 the two time intervals defined for the groundwater records: the late-Holocene (<5,000 years
27 before present) and latter half of the last glacial time period (20,000 to 50,000 years before
28 present). This step effectively lowered the temporal resolution of speleothem and ice core
29 precipitation isotope records to be consistent with the temporal resolution of the groundwater
30 records. A correction factor was applied to speleothem $\delta^{18}\text{O}$ values to account for different
31 $\text{H}_2\text{O}-\text{CaCO}_3$ isotopic fractionation factors during the late-glacial and the late-Holocene because

1 of differing land surface temperatures during each time period (details presented within
2 Supplement).

3 Simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values were compiled from five isotope-enabled general circulation
4 models (simulated $\Delta^{18}\text{O}_{\text{late-glacial}} = \delta^{18}\text{O}_{\text{last glacial maximum}} - \delta^{18}\text{O}_{\text{pre-industrial}}$): CAM3iso (e.g., Noone
5 and Sturm, 2010; Pausata et al., 2011a), ECHAM5-wiso (e.g., Werner et al., 2011), GISSE2-R
6 (e.g., Schmidt et al., 2014; LeGrande and Schmidt, 2008; 2009), IsoGSM (e.g., Yoshimura et
7 al., 2003) and LMDZ4 (e.g., Risi et al., 2010a). ECHAM5-wiso and IsoGSM outputs are for
8 modern climate rather than pre-industrial conditions; however, the difference between the
9 isotopic composition of pre-industrial and modern climate are expectedly small compared to
10 late-glacial to late-Holocene $\delta^{18}\text{O}$ shifts. An offset factor was applied to simulated mean
11 seawater $\delta^{18}\text{O}$ in all five models (Table S1) to account for known glacial-interglacial changes
12 to seawater $\delta^{18}\text{O}$ (Emiliani, 1955; Dansgaard and Tauber, 1969; Schrag et al., 1996; 2002).
13 Possible spatial differences in seawater $\delta^{18}\text{O}$ changes from the last glacial maximum to the pre-
14 industrial time period are not incorporated into simulations with prescribed sea surface
15 temperatures (CAM3iso, ECHAM5-wiso, IsoGSM, LMDZ4) but are simulated by the coupled
16 ocean-atmosphere simulation of GISSE2-R (Supplement Table S1). GISSE2-R was submitted
17 to the CMIP5 archive and participated in PMIP3. LMDZ4 was submitted to the CMIP3 archive.
18 ECHAM5 and CAM3iso did not participate in CMIP5, while IsoGSM uses different boundary
19 conditions than proposed for CMIP5 (Yoshimura et al., 2008). The five models span a range of
20 spatio-temporal resolutions and isotopic/atmospheric parameterizations described in detail in
21 the above references. A selection of the inter-model similarities and differences are summarized
22 in Table S1 (Supplement).

23 For clarity, empirical $\Delta^{18}\text{O}_{\text{late-glacial}}$ values that are based on measured isotope contents of
24 groundwater, speleothem, ground ice or ice core records are referred to herein as *measured*
25 $\Delta^{18}\text{O}_{\text{late-glacial}}$; simulated precipitation isotope compositions obtained from general circulation
26 model results are referred to as *simulated* $\Delta^{18}\text{O}_{\text{late-glacial}}$. We acknowledge that the general
27 circulation models explicitly analyse the last glacial maximum and the pre-industrial climate
28 conditions (i.e., simulated $\Delta^{18}\text{O}_{\text{late-glacial}} = \delta^{18}\text{O}_{\text{last glacial maximum}} - \delta^{18}\text{O}_{\text{pre-industrial}}$), whereas proxy
29 record reconstructions of $\Delta^{18}\text{O}_{\text{late-glacial}}$ integrate hydroclimatology over multi-millennial time
30 scales that are different from the model simulations.

1 3 Results and Discussion

2 3.1 Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values

3 Measured groundwater (n=59), speleothem (n=15) and ice core (n=12) $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are
4 presented in Figure 2 (references presented in the Supplement). Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values
5 range from -7.1‰ (i.e., $\delta^{18}\text{O}_{\text{late-glacial}} < \delta^{18}\text{O}_{\text{late-Holocene}}$) to $+1.7\text{‰}$ (i.e., $\delta^{18}\text{O}_{\text{late-glacial}} >$
6 $\delta^{18}\text{O}_{\text{late-Holocene}}$). Three-quarters of the compiled records have negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$
7 values and one-quarter of compiled records have positive measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values. More
8 than 80% of records with positive measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are located within 35° of the
9 equator and within 400 km of the nearest coastline (e.g., Bangladesh $\Delta^{18}\text{O}_{\text{late-glacial}}$ of $+1.5\text{‰}$,
10 less than 300 km from the coast; Figures 2-4). In comparison, negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$
11 values are found in both coastal regions and farther inland. Negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$
12 values of the greatest magnitude are located at high latitudes (e.g., northwestern Canada,
13 latitude 64°N : $\Delta^{18}\text{O}_{\text{late-glacial}}$ of -5.5‰ ; northern Russia latitude 72°N : -5.4‰) and far from
14 coastlines (e.g., Hungary: -3.7‰ , ~ 500 km from Atlantic Ocean; Peru: -6.3‰ , ~ 2000 km
15 from Atlantic Ocean, the modern moisture source to Peru; Garreaud et al., 2009). Greenland
16 and Antarctic ice cores have negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values that are of greater
17 magnitude than non-polar measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values (Antarctic and Greenland $\Delta^{18}\text{O}_{\text{late-glacial}}$
18 values range from -3.6‰ to -7.1‰ ; Figure 3).

19 Our synthesis shows that measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in the tropics are closer to 0‰ (i.e., no
20 change) than $\Delta^{18}\text{O}_{\text{late-glacial}}$ values at high latitudes and continental interiors that generally have
21 high magnitude, negative $\Delta^{18}\text{O}_{\text{late-glacial}}$ values. High magnitude, negative measured
22 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are most common where present day precipitation $\delta^{18}\text{O}$ values are at a
23 minimum (e.g., Bowen and Wilkinson, 2002). This broad spatial pattern is consistent with the
24 non-linear isotopic distillation of air masses undergoing progressive rainout (i.e., Rayleigh
25 distillation). Because seawater $\delta^{18}\text{O}$ values were $\sim 1\text{‰}$ higher-than-modern during the last
26 glacial maximum (Schrag et al., 1996; 2002), our finding that the majority of measured
27 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are negative suggests that isotopic distillation of air masses was greater
28 during the late-glacial than under present climate. This finding is consistent with land surface
29 temperature reconstructions that show larger glacial-to-modern changes to land temperatures at
30 high latitude and continental settings (Figure 1; Annan and Hargreaves, 2013). Tropical versus
31 extratropical patterns of late-glacial/late-Holocene temperature change (Figure 1a) are broadly
32 similar to measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values (Figure 3), where both temperature and isotope shifts

1 are greater at high latitudes relative to the equator. Therefore, it is possible that the larger
2 late-glacial to late-Holocene temperature shifts at the poles relative to the equator may have
3 served to amplify the non-linear, Rayleigh relationship describing the heavy isotope depletion
4 of air masses undergoing progressive rainout during transport from lower to higher latitudes.
5 Further, the late-glacial was characterized by: (i) lower-than-modern atmospheric temperatures
6 with larger coastal-inland gradients, and (ii) lower-than-modern eustatic sea level leading to
7 longer overland atmospheric transport distances. Each of these late-glacial/late-Holocene
8 changes favours stronger-than-modern isotopic distillation of air masses transported inland
9 from the coast during the late-glacial (Dansgaard, 1964; Rozanski et al., 1993; Winnick et al.,
10 2014), potentially contributing to the broad, global observation that most (77%) $\delta^{18}\text{O}_{\text{late-Holocene}}$
11 values exceed $\delta^{18}\text{O}_{\text{late-glacial}}$ values on continents.

12 Pairings of groundwater and speleothem records are available within ~500 km of one another
13 in the southwestern USA, central China and Israel. Southwestern USA speleothem and
14 groundwater records ~400 km apart show similar $\Delta^{18}\text{O}_{\text{late-glacial}}$ values, with San Juan Basin
15 groundwaters having a measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ value of -2.5 ± 1.0 ‰ (Phillips et al., 1986) and
16 speleothems ~400 km to the south having measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values of -3.0 ± 1.2 and
17 -3.4 ± 0.4 (Asmerom et al., 2010; Wagner et al., 2010). Central China speleothem and
18 groundwater records ~200 km apart overlap within uncertainty margins (i.e., $\Delta^{18}\text{O}_{\text{late-glacial}}$
19 values of -1.1 ± 1.7 ‰ and $+0.3 \pm 2.1$ ‰; Cai et al., 2010). Israeli speleothem and groundwater
20 records ~100 km apart have different measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values. Two Israeli groundwater
21 $\Delta^{18}\text{O}_{\text{late-glacial}}$ records were compiled; the coastal Israeli aquifer has a $\Delta^{18}\text{O}_{\text{late-glacial}}$ value of
22 $+0.3 \pm 0.4$ ‰ (Yechieli et al., 2009), whereas groundwater of the Dead Sea Rift Valley has a
23 $\Delta^{18}\text{O}_{\text{late-glacial}}$ value of -1.8 ± 0.6 ‰ (Burg et al., 2013). Speleothem records have $\Delta^{18}\text{O}_{\text{late-glacial}}$
24 values close to $+1$ ‰ (Frumkin et al., 1999; Bar-Matthews et al., 2003). In northern Turkey,
25 speleothem and groundwater separated by ~150 km have measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values that
26 differ by ~3 ‰ (speleothem $\Delta^{18}\text{O}_{\text{late-glacial}}$ -5.7 ± 0.4 ‰ versus groundwater $\Delta^{18}\text{O}_{\text{late-glacial}}$ of
27 -2.8 ± 1.0 ‰; Fleitmann et al., 2009; Arslan et al., 2013; 2015). While the locations of the
28 groundwater and speleothem records differ, the compiled data suggests that groundwater and
29 speleothem $\Delta^{18}\text{O}_{\text{late-glacial}}$ values may capture different $\Delta^{18}\text{O}_{\text{late-glacial}}$ values under similar climate
30 conditions.

31 A number of potential processes could bias the preservation of precipitation isotope
32 composition in ice core, speleothem or groundwater archives (Wang et al., 2001, Thompson et

1 al., 2006; Edmunds, 2009). For example, groundwater and speleothem archives preserve only
2 the isotope record of precipitation that traverses the vadose zone. Recent global analyses of
3 paired precipitation-groundwater isotope compositions show that winter (extratropics) and wet
4 season (tropics) precipitation contributes disproportionately to recharge (Jasechko et al., 2014),
5 meaning that paleoclimate records may be more sensitive to changes to winter and wet seasons
6 than summer or dry season (Vogel et al., 1963; Simpson et al., 1972; Grabczak et al., 1984;
7 Harrington et al., 2002; Jones et al., 2002; Darling, 2004; Partin et al., 2012). Similarly,
8 groundwater isotope records are unlikely to represent constant and continuous recharge fluxes
9 during the late-Holocene or the late-glacial (McIntosh et al., 2012). Modern groundwater
10 recharge fluxes are highest in humid climates (Wada et al., 2010). Groundwater $\delta^{18}\text{O}$ records
11 only represent precipitation that recharges aquifers, meaning that groundwater-based
12 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values could be biased to subintervals (e.g., interstadials, pluvial periods) within
13 the late-Holocene and late-glacial intervals when recharge fluxes were at local maxima.
14 Speleothem records may be further complicated by processes impacting the timing of calcite
15 precipitation. Recent modelling suggests that calcite precipitation in caves located outside of
16 the tropics is greatest during the cool season and reduced during summer months due to changes
17 in ventilation, meaning that higher latitude speleothems record oxygen isotope compositions
18 biased to cool season climate change (James et al., 2015). Other recent work suggests that
19 speleothem $\delta^{18}\text{O}$ data may be impacted by disequilibrium isotope effects (Asrat et al., 2008;
20 Daëron et al., 2011; Kluge and Affek, 2012; Kluge et al., 2013) or by partial evaporation of drip
21 waters resulting in ^{18}O -enrichment (e.g., Cuthbert et al., 2014a) and greater fractionation due
22 to evaporative cooling (Cuthbert et al., 2014b), potentially obscuring the preservation of
23 primary precipitation isotope contents in the speleothem record. Compiled ice core records may
24 have be influenced by post-depositional exchanges of ice with atmospheric vapour (Steen-
25 Larsen et al., 2014). The impact of atmospheric vapour exchanges on ice core isotope records
26 remains poorly understood. Potential biases in the preservation of precipitation $\delta^{18}\text{O}$ differ
27 among groundwater, glacial ice, and speleothem records, meaning that co-located records of
28 differing record-type may preserve different $\Delta^{18}\text{O}_{\text{late-glacial}}$ values under similar climate
29 conditions. Finally, all proxy records may be impacted by past changes in the seasonality of
30 precipitation, which can substantially impact annual precipitation $\delta^{18}\text{O}$ values (e.g., Werner et
31 al., 2000).

32 We cannot rule out the possibility that changes in seasonal biases of proxy record preservation
33 occurred between the late-glacial and the late-Holocene and have impacted measured

1 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values. Further, the chronologies of groundwaters and ice core records have
2 uncertainties on the order of thousands of years, meaning that the time intervals used to
3 calculate measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values may be inaccurate. However, the plateauing of isotope
4 content observed in most regional aquifers for 0-5,000 years before present and for >20,000
5 years before present supports our interpreting these data as records of late-glacial to
6 late-Holocene isotopic shifts (see figures in the Supplement). Notwithstanding potential $\delta^{18}\text{O}$
7 preservation biases and chronology uncertainties, the global data synthesized here show
8 patterns consistent with the enhanced distillation of advected air masses originating as
9 (sub)tropical ocean evaporate and undergoing progressive, poleward rainout under cooler-than-
10 modern late-glacial temperatures.

11 **3.2 Simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values**

12 Simulated precipitation $\Delta^{18}\text{O}_{\text{late-glacial}}$ values from five general circulation models are presented
13 in Figure 5. At least four of the five models agree on the sign of simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values—
14 that is values consistently above or consistently below zero—for 68.8 % of grid cells covering
15 Earth's surface (68.7 % of over-ocean areas and 68.9 % of over land areas; multi-model
16 calculation completed using 3 of 4 models as a threshold at high-latitudes where IsoGSM data
17 was unavailable). Simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are consistently negative over the North
18 Atlantic Ocean and the Fennoscandian and Laurentide ice sheets and consistently positive over
19 most of the tropical oceans, whereas poorer agreement is found over tropical land surfaces. The
20 negative simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values over the northern hemisphere ice sheets and North
21 Atlantic are likely driven by the difference in ice sheet topography and sea ice cover, between
22 the late-glacial and pre-industrial climate. The late-glacial to late-Holocene change in ice sheet
23 topography and sea ice cover impacted surface temperatures, which were more than $\sim 20^\circ\text{C}$
24 cooler over most of present-day Canada during the last glacial maximum (Figure 1). Cooler
25 temperatures in conjunction with ice sheet topography (>3000 m elevations; e.g., Peltier, 1994)
26 enhanced Rayleigh distillation for air masses transecting Northern Hemisphere ice sheets, as
27 evidenced by systematically low measured and simulated $\delta^{18}\text{O}_{\text{late-glacial}}$ values in these regions
28 (Figures 2, 3 and 5).

29 A comparison of simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values over tropical Africa, South America and
30 Oceania shows inter-model disagreement (Figure 5). Different tropical simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$
31 values among the models reflect the different isotopic parameterizations, inter-model spread in

1 simulated precipitation rates, and seawater $\delta^{18}\text{O}$ specifications used in each model
2 (Supplement). Inter-model spread in simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in some regions highlights
3 the importance of this global synthesis of measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values as a constraint for
4 isotope enabled climate simulations. Another potential source for the model disagreement is
5 introduced by the different ice-sheet topography used in each model. CAM3Iso, IsoGSM and
6 LMDZ4 used Ice 5G (Peltier 1994) as advised for PMIP2 (Braconnot et al., 2007), whereas the
7 GISSE2 replaces Ice 5G Laurentide ice with that of Licciardi et al. (1999) and ECHAM5-wiso
8 uses ice topography from PMIP3 (Braconnot et al., 2007; 2012; PMIP3 follows ice sheet
9 topography blended from multiple ice sheet reconstructions: Argus and Peltier, 2010; Toscano
10 et al., 2011). Ice sheet topography is an important driver of simulated temperature, precipitation
11 and atmospheric circulation during the last glacial maximum (e.g., Justino et al., 2005; Pausata
12 et al., 2011b, Ullman et al., 2014). Therefore, it is likely that inter-model differences in paleo-
13 ice sheet topographies impacts atmospheric circulation and thus high latitude simulated
14 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values reported in this study (Figure 5).

15 Differences in the specification of initial seawater $\delta^{18}\text{O}$ may also lead to inter-model differences
16 in simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values. Seawater $\delta^{18}\text{O}$ is set to be globally-homogenous in CAM3Iso,
17 IsoGSM and LMDZ4, and heterogeneous in ECHAM5-wiso (using modern gridded seawater
18 $\delta^{18}\text{O}$ heterogeneity of LeGrande and Schmidt, 2006) and GISSE2-R (coupled atmosphere-
19 ocean model; seawater $\delta^{18}\text{O}$ is calculated by the ocean model). Including surface ocean $\delta^{18}\text{O}$
20 heterogeneities in model simulations impacts land precipitation $\delta^{18}\text{O}$ by up to $\sim 1.5\%$ relative
21 to simulations with homogenous seawater $\delta^{18}\text{O}$ (LeGrande and Schmidt, 2006). However,
22 different seawater $\delta^{18}\text{O}$ specifications cannot account for all inter-model differences in
23 simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values.

24 The models also show deficiencies in simulating measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in the tropics,
25 particularly over tropical Africa. This finding could, in part, relate to the high sensitivity of
26 precipitation $\delta^{18}\text{O}$ to convective parameterizations (Lee et al., 2009, Field et al 2014), although
27 future research is required to test this. Another reason may be that the measured $\Delta^{18}\text{O}_{\text{late-glacial}}$
28 integrates the hydroclimatological signal over multi-millennial time scales, whereas the
29 simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values explicitly explore last glacial maximum and pre-
30 industrial/present-day climate conditions. The smeared temporal resolution of groundwater-
31 based measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values due to storage and mixing in the aquifer precludes an ideal
32 comparison of measured versus simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values. Further, as previously discussed

1 in section 3.1, the measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are susceptible to a number of potential biases
2 that may obscure the magnitude and direction of late-glacial to late-Holocene precipitation $\delta^{18}\text{O}$
3 changes. Notwithstanding, models correctly simulate the sign of measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values
4 (i.e., positive or negative) in the extratropics more frequently than in the tropics. Better
5 agreement in the sign of simulated versus measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in the extra-tropics
6 compared to the tropics is likely linked to the substantial changes to extra-tropical ice-sheet
7 topography and sea-ice cover between the two climate states in northern North America and
8 Europe. Substantial changes to northern hemisphere ice volumes between the late-glacial and
9 the late-Holocene likely enhanced upwind distillation of air masses leading to high-magnitude,
10 negative $\Delta^{18}\text{O}_{\text{late-glacial}}$ values that are well captured by the climate simulations. However,
11 simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values over Antarctica and Greenland show large inter-model spread,
12 suggesting that model-based interpretations of polar ice core records may vary widely among
13 different atmospheric models.

14 **3.3 Regional measured and simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values**

15 **3.3.1 Australia and Oceania**

16 Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values from Australia and Oceania fall between -1‰ and $+1\text{‰}$ (Figure
17 2). Australian climate during the last glacial time period was more arid (Nanson et al., 1992),
18 dustier (Chen et al., 1993) and cooler (Miller et al., 1997) than present day. Simulated
19 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values across Australia are variable among the five models. Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$
20 values across Oceania have been attributed to temporal changes in the strength of monsoons
21 and convective rains (Aggarwal et al., 2004; Partin et al., 2007; Williams et al., 2010)
22 potentially impacted by late-glacial to late-Holocene shifts in the position of the intertropical
23 convergence zone (Lewis et al., 2010; 2011).

24 **3.3.2 Southeast Asia**

25 Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values from southeastern Asia range from -2.3‰ to $+1.7\text{‰}$. The
26 highest regional measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are found in Bangladesh (measured $\Delta^{18}\text{O}_{\text{late-glacial}}$
27 of $+1.5\pm 1.3\text{‰}$; Aggarwal et al., 2000) and in central and southeastern China (measured
28 $\Delta^{18}\text{O}_{\text{late-glacial}}$ of $+0.3\text{‰}$ to $+1.7\text{‰}$; Wang et al., 2001; Yuan et al., 2004; Dykoski et al., 2005;
29 Cai et al., 2010; Yang et al., 2010). General circulation models have positive simulated
30 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values near to the Chinese coasts, but are more variable across western and

1 northern China (Figure 5). Chinese speleothem records show near-zero or positive measured
2 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values interpreted to reflect the reduced strength of the East Asian (Wang et al.,
3 2001; Dykoski et al., 2005; Cosford et al., 2008) or Indian monsoons (Pausata et al., 2011a).
4 Further research suggests that Chinese speleothem $\delta^{18}\text{O}$ variations reflect changes to regional
5 moisture sources and the intensity or provenance of atmospheric transport pathways (LeGrande
6 and Schmidt, 2009; Dayem et al., 2010; Lewis et al., 2010; Maher and Thompson, 2012; Caley
7 et al., 2014b; Tan, 2014).

8 North China Plain groundwaters have high-magnitude, negative $\Delta^{18}\text{O}_{\text{late-glacial}}$ values (measured
9 $\Delta^{18}\text{O}_{\text{late-glacial}}$ of -2.3 ± 0.6 ‰; Zongyu et al., 2003) compared to coastal, more southerly
10 counterparts. Combining the negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ in northern China (Zongyu et al.,
11 2003; Ma et al., 2008; Currell et al., 2012; Li et al., in press) with the positive measured
12 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in central and southeastern China (Wang et al., 2001; Yuan et al., 2004;
13 Dykoski et al., 2005; Cai et al., 2010; Yang et al., 2010) reveals a south-to-north decrease from
14 positive (south) to negative (north) measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values (Figures 2 and 6). Previous
15 studies of modern precipitation have identified increasing precipitation $\delta^{18}\text{O}$ values from the
16 coast to inland China during the wet season, sharply contrasting spatial patterns expected from
17 Rayleigh distillation (Aragúas-Aragúas et al., 1998). A more recent work suggests that low wet-
18 season precipitation $\delta^{18}\text{O}$ values over southern China are controlled by the deflection of
19 westerlies around the Tibetan Plateau, whereas precipitation $\delta^{18}\text{O}$ values over northern China
20 are controlled by local-scale rainfall and below-cloud raindrop evaporation (Lee et al., 2012).
21 Therefore, measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values from southern China may reflect changes to
22 atmospheric circulation at broader spatial scales, whereas measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values from
23 northern China may indicate changes to more localized atmospheric conditions impacting
24 processes such as raindrop evaporation in addition to meso- and synoptic-scale circulation
25 changes.

26 3.3.3 Africa

27 Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values from Africa range from -2.9 ‰ to $+0.1$ ‰ (Figures 2 and 6). 16
28 of 17 measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values from Africa are negative. Near-zero measured $\Delta^{18}\text{O}_{\text{late-glacial}}$
29 values are generally found near to coasts (e.g., Senegal $\Delta^{18}\text{O}_{\text{late-glacial}}$ of $+0.1\pm 0.8$ ‰; Madioune
30 et al., 2014), whereas higher magnitude, negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in Africa are
31 found farther inland (e.g., Niger $\Delta^{18}\text{O}_{\text{late-glacial}}$ values of -2.3 ± 2.0 ‰ and -2.9 ± 0.9 ‰: ~800 km
32 from the Atlantic coast). General circulation model $\Delta^{18}\text{O}_{\text{late-glacial}}$ values show poor agreement

1 with measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ over tropical Africa compared to model-measured comparisons for
2 Europe and North America (Figure 5), with positive simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values predicted
3 over large parts of Africa where negative $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are measured. Figure 5 shows that
4 Africa has the largest inter-model and model-measurement disagreements in the sign of
5 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values of the continents.

6 Northern African hydrological processes are influenced by interlinked controls such as
7 meridional shifts in the position of the intertropical convergence zone (Arbuszewski et al.,
8 2013) and the strength of Atlantic meridional overturning circulation (Mulitza et al., 2008).
9 Paleowater chemistry indicates that northern Africa was at least 2°C cooler than today
10 (Guendouz et al., 1998) and that westerly moisture transport was stronger than the present
11 during the late-glacial (Sultan et al., 1997; Abouelmagd et al., 2012).

12 Tropical Africa was 2°C to 4°C cooler and more arid than present day at the last glacial
13 maximum (Powers et al., 2005; Tierney et al., 2008). Early- and late-Holocene rainfall and
14 isotope compositions were highly variable across Africa (Tierney et al., 2008; Schefuß et al.,
15 2011; Tierney et al., 2013; Otto-Bliesner et al., 2014). Tropical African rainfall originates from
16 both Indian and Atlantic sources, with Atlantic-sourced moisture travelling across the Congo
17 rainforest (Levin et al., 2009). Lower-than-modern continental moisture recycling during the
18 late-glacial may partially explain negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values across some regions
19 of inland tropical Africa (e.g., Risi et al., 2013). Negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in
20 tropical Africa could also be interpreted to reflect higher-than-modern upwind rainout during
21 the late-glacial (see Risi et al., 2008; 2010b; Lee et al., 2009; Scholl et al., 2009; Lekshmy et
22 al., 2014; Samuels-Crow et al., 2014); however, this explanation necessitates stronger-than-
23 modern convection during the late-glacial, an explanation that would contradict the established
24 cooler-than-modern land surface temperatures. Therefore, changes to atmospheric transport
25 distances and vapour origins are more likely responsible for negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$
26 values across tropical Africa (Lewis et al., 2010).

27 3.3.4 Europe and the Mediterranean

28 Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values across Europe, the Middle-East and the eastern Mediterranean
29 range from -5.7 ‰ to +1.3 ‰. 80% of measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values across these regions are
30 negative. All five general circulation models agree on negative simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values
31 across Europe, consistent with the negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values across the majority

1 of Europe. Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are generally higher in western Europe (0.0 ‰ to
2 -1.0 ‰ in Portugal, the United Kingdom and France) than in eastern Europe (-1.0 ‰ to -5.7 ‰
3 in Poland, Hungary and Turkey; Stute and Deak, 1989; Le Gal La Salle et al., 1995; Darling et
4 al., 1997; Barbecot et al., 2000; Zuber et al., 2004; Galego Fernandes and Carreira, 2008; Celle-
5 Jeanton et al., 2009; Varsányi et al., 2011; Samborska et al., 2012; Arslan et al., 2013). This
6 spatial pattern of $\Delta^{18}\text{O}_{\text{late-glacial}}$ values is consistent with enhanced isotopic distillation of
7 westerlies during the late-glacial due to cooler-than-modern final condensation temperatures.

8 High magnitude, negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are located in Turkey and Georgia
9 south and east of the Black Sea (-2.8 ± 1.0 to -5.7 ± 0.4 ‰; Fleitmann et al., 2009; Arslan et al.,
10 2013; Melikadze et al., 2014). Westerly air mass trajectories distal to the Fennoscandian ice
11 sheet topography may not have changed considerably since the late-glacial over western and
12 central Europe (Rozanski, 1985; Loosli et al., 2001). Therefore, higher, near-zero measured
13 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in western Europe and lower, negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in
14 eastern Europe indicate enhanced distillation of advected air masses during the late-glacial
15 relative to the late-Holocene.

16 Changes to freeze-thaw conditions of the ground surface between the latter half of the last
17 glacial time period and the modern climates may have impacted the seasonality of the fraction
18 of precipitation recharging aquifers and thus $\Delta^{18}\text{O}_{\text{late-glacial}}$ (Darling, 2004; Darling, 2011;
19 Jasechko et al., 2014). Geomorphic evidence suggests permafrost covered portions of Hungary
20 at the last glacial maximum, suggesting that land temperatures may have been up to 15°C cooler
21 than present day (Fábián et al., 2014), a larger late-glacial to late-Holocene temperature shift
22 than earlier, noble gas based reconstructions ($5-7^{\circ}\text{C}$; Deák et al., 1987). European pollen
23 records indicate that northern Europe was tundra-like and that southern Europe was semi-arid
24 during the last glacial maximum (Harrison and Prentice, 2003; Clark et al., 2012). The European
25 late-glacial to late-Holocene transition from semi-arid deserts to temperate forests could have
26 lowered $\Delta^{18}\text{O}_{\text{late-glacial}}$ values as groundwater recharge ratios transitioned from more extreme
27 winter-biased (e.g., semi-arid lands during the late-glacial) to less extreme winter-biased
28 groundwater recharge ratios (e.g., forests during late-Holocene; Jasechko et al., 2014).

29 3.3.5 South America

30 Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values across South America range from -6.3 ‰ to $+0.6$ ‰ (Figures 2
31 and 6). The highest-magnitude, negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are found in Andean ice

1 cores ($\Delta^{18}\text{O}_{\text{late-glacial}}$ of -4.6 ± 1.0 and -6.3 ± 1.3 ; Thompson et al., 1995; 1998). Here the
2 importance of upstream convection upon modern Andean precipitation $\delta^{18}\text{O}$ has been
3 highlighted at inter-annual (Hoffmann et al., 2003; Vuille and Werner, 2005), seasonal (Vimeux
4 et al., 2005, Samuels-Crow et al., 2014) and daily time scales (Vimeux et al., 2011). It is
5 therefore possible that upstream convection controls past changes to Andean precipitation
6 isotope compositions recorded in ice cores.

7 The measured groundwater $\Delta^{18}\text{O}_{\text{late-glacial}}$ value located in eastern Brazil is -2.7 ± 1.3 ‰ (Salati
8 et al., 1974). Eastern Brazil was 5°C cooler than today during the latter half of the last glacial
9 period (Stute et al., 1995b). Four of the five general circulation models simulate positive
10 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values across eastern Brazil (Figure 5), highlighting a difference between
11 simulated and measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in parts of the tropics. The negative measured
12 $\Delta^{18}\text{O}_{\text{late-glacial}}$ value in eastern Brazil has been previously interpreted to reflect higher-than-
13 modern precipitation during the last glacial time period (Salati et al., 1974). Lewis et al. (2010)
14 show that localized rainfall governs precipitation $\delta^{18}\text{O}$ in eastern Brazil. Modern precipitation
15 $\delta^{18}\text{O}$ values are lowest in eastern Brazil when precipitation rates are at a maximum. Extending
16 Lewis et al.'s interpretation linking local precipitation amount to precipitation $\delta^{18}\text{O}$ would
17 suggest that the negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ value found in eastern Brazil may indeed
18 record wetter-than-modern conditions during the late-glacial as proposed by Salati et al. (1974).
19 Further, disagreement between measured and simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ in eastern Brazil
20 highlights the need to critically evaluate climate model performance in regions where the
21 precipitation amount is closely correlated with precipitation $\delta^{18}\text{O}$.

22 3.3.6 North America

23 Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ from North American proxy records range from -5.5 ‰ to $+1.0$ ‰.
24 Canadian records of groundwater recharge that took place beneath the Laurentide ice sheet are
25 not included in this synthesis ("subglacial recharge;" Grasby and Chen, 2005; Ferguson et al.,
26 2007; McIntosh et al., 2012; Ferguson and Jasechko, in press). These records were excluded
27 because the subglacial meltwaters that recharged aquifers likely reflect precipitation that fell
28 elsewhere on the paleo-ice sheet, potentially complicating the comparison of groundwater
29 isotope compositions for the late-Holocene and last glacial time period.

30 Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values along the USA east coast show the highest, positive values in
31 Georgia (latitude: 32°N ; measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ of $+1.0$ ‰; Clark et al., 1997), decreasing

1 northward to near-zero measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in coastal Maryland (latitude 39°N;
2 measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ of -0.1 ± 0.4 ‰; Aeschbach-Hertig et al., 2002). Decreasing $\Delta^{18}\text{O}_{\text{late-glacial}}$
3 values with increasing latitude along the USA east coast may be explained in part by the isotopic
4 distillation of air masses advected northward from the subtropics under cooler-than-modern
5 final atmospheric condensation temperatures. Indeed, paleoclimate records indicate that
6 Maryland was more arid and as much as 9-12°C cooler during the late-glacial relative to the
7 late-Holocene (Purdy et al., 1996; Aeschbach-Hertig et al., 2002; Plummer et al., 2012). In
8 addition to temperature change, late-glacial precipitation isotope compositions along eastern
9 USA coastline were likely impacted by the lower-than-modern late-glacial sea levels, which
10 changed overland atmospheric transport distances between the late-glacial and late-Holocene
11 (Clark et al., 1997; Aeschbach-Hertig et al., 2002; Tharammal et al., 2012).

12 Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in the central and southwestern USA have the highest magnitude,
13 negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values of temperate North America, ranging from -1.0 ‰ to
14 -3.4 ‰. Central and southwestern USA measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values contrast the positive
15 measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values found along the eastern USA coast at similar latitudes.
16 Consistently negative $\Delta^{18}\text{O}_{\text{late-glacial}}$ values in central and southwest USA suggest that advected
17 moisture to the region underwent greater upstream air mass distillation during the late-glacial
18 than under modern climate. Pollen, vadose zone and groundwater records show that late-glacial
19 southwestern USA was $\sim 4^\circ\text{C}$ cooler, had greater groundwater recharge fluxes, and had more
20 widespread forests than present day (Stute et al., 1992; 1995a; Scanlon et al., 2003; Williams,
21 2003). Negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values found in the southwest USA have been ascribed
22 to lower-than-modern summer precipitation (New Mexico, Phillips et al., 1986), latitudinal
23 shifts in the positions of the polar jet stream and the intertropical convergence zone (New
24 Mexico, Asmerom et al., 2010) and changes to over-ocean humidity, temperature or moisture
25 sources (Idaho, Schlegel et al., 2009). Wagner et al. (2010) interpret decreases to southwestern
26 precipitation $\delta^{18}\text{O}$ to reflect cooler and more-humid conditions. Extending this interpretation to
27 negative measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values found across the southwestern USA values supports
28 earlier conclusions that the region was cooler and more humid than today during the late-glacial,
29 possibly linked to changes in air mass trajectories and moisture sources (Asmerom et al., 2010;
30 Wagner et al., 2010). Simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values across North America closely match spatial
31 patterns of measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ synthesized in this study. Strong, multi-model agreement
32 with measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ patterns supports continued application of isotope enabled general
33 circulation models when interpreting North American precipitation isotope proxy records.

1 **4 Conclusions**

2 While changes to the isotope content of precipitation between the last glacial time period and
3 more recent times has been widely documented, few studies have synthesized these dispersed
4 data to explore the global patterns of $\delta^{18}\text{O}$ change driven by past shifts to regional climate. In
5 this study we compile groundwater, speleothem, ice core and ground ice records of $\delta^{18}\text{O}$ shifts
6 between the late-glacial (20 to ~50 thousand years ago) and the late-Holocene (within the past
7 5,000 years). Late-glacial to late-Holocene $\delta^{18}\text{O}$ shifts range from -7.1 ‰ (i.e., $\delta^{18}\text{O}_{\text{late-glacial}} <$
8 $\delta^{18}\text{O}_{\text{late-Holocene}}$) to $+1.7$ (i.e., $\delta^{18}\text{O}_{\text{late-glacial}} > \delta^{18}\text{O}_{\text{late-Holocene}}$). Aquifers with positive measured
9 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values (23% of records) are most common along the subtropical coasts. The
10 majority (77%) of measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values are negative, with the highest magnitude
11 differences between $\delta^{18}\text{O}_{\text{late-glacial}}$ and $\delta^{18}\text{O}_{\text{late-Holocene}}$ observed at high latitudes and far from
12 coasts. This spatial pattern suggests that isotopic distillation of advected air masses was greater
13 during the late-glacial than under present climate, likely due to the non-linear nature of Rayleigh
14 distillation, accentuated by larger glacial-interglacial atmospheric temperature changes at the
15 poles relative to lower latitudes. Regionally-divergent precipitation $\delta^{18}\text{O}$ responses to the $\sim 4^\circ\text{C}$
16 of global warming occurring between the late-glacial and the late-Holocene suggest that
17 continued monitoring of modern precipitation isotope contents may prove a useful for detecting
18 hydrologic changes due to ongoing, human-induced climate change. Future paleo-precipitation
19 proxy record $\delta^{18}\text{O}$ research can use these new global maps of $\Delta^{18}\text{O}_{\text{late-glacial}}$ records to target and
20 prioritize field sites. In the near term, a global compilation of large lake sediment isotope
21 records that accounts for paleo-evaporative isotope effects could enhance spatial coverage of
22 interglacial-glacial $\delta^{18}\text{O}$ shifts.

23 General circulation models agree on the sign and magnitude of terrestrial precipitation
24 $\Delta^{18}\text{O}_{\text{late-glacial}}$ values better in the extra-tropics than in the tropics. Differences in simulated
25 precipitation isotope composition changes amongst the models might be linked to different
26 parameterizations of seawater $\delta^{18}\text{O}$, glacial topography and convective rainfall, however, these
27 hypotheses require further testing. Future model research should focus on quantifying the
28 relative roles of inter-model spread in the simulated climate versus the isotopic response to
29 climate change on resulting simulated precipitation $\delta^{18}\text{O}$. This would provide guidelines to
30 interpret model-data isotopic differences and to identify what aspects climate models have
31 greatest difficulties capturing.

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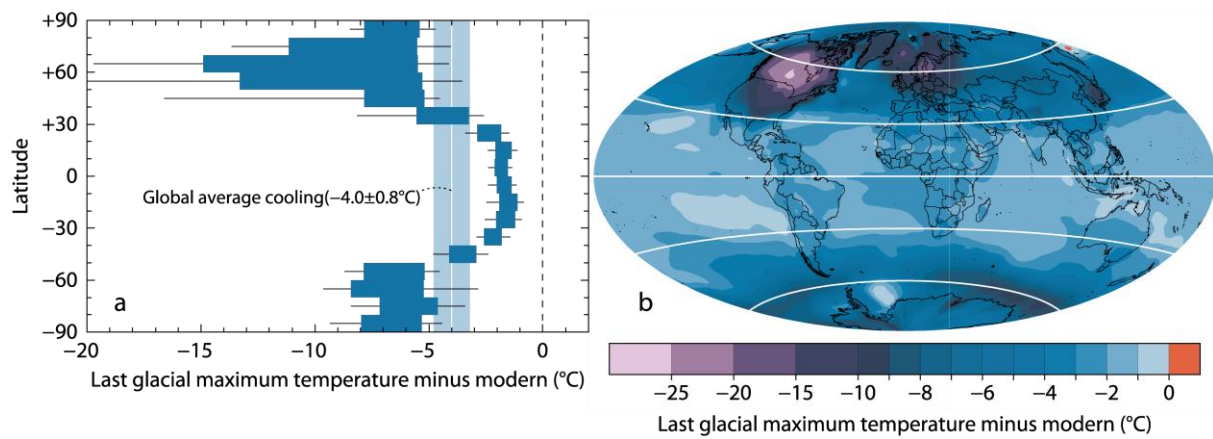
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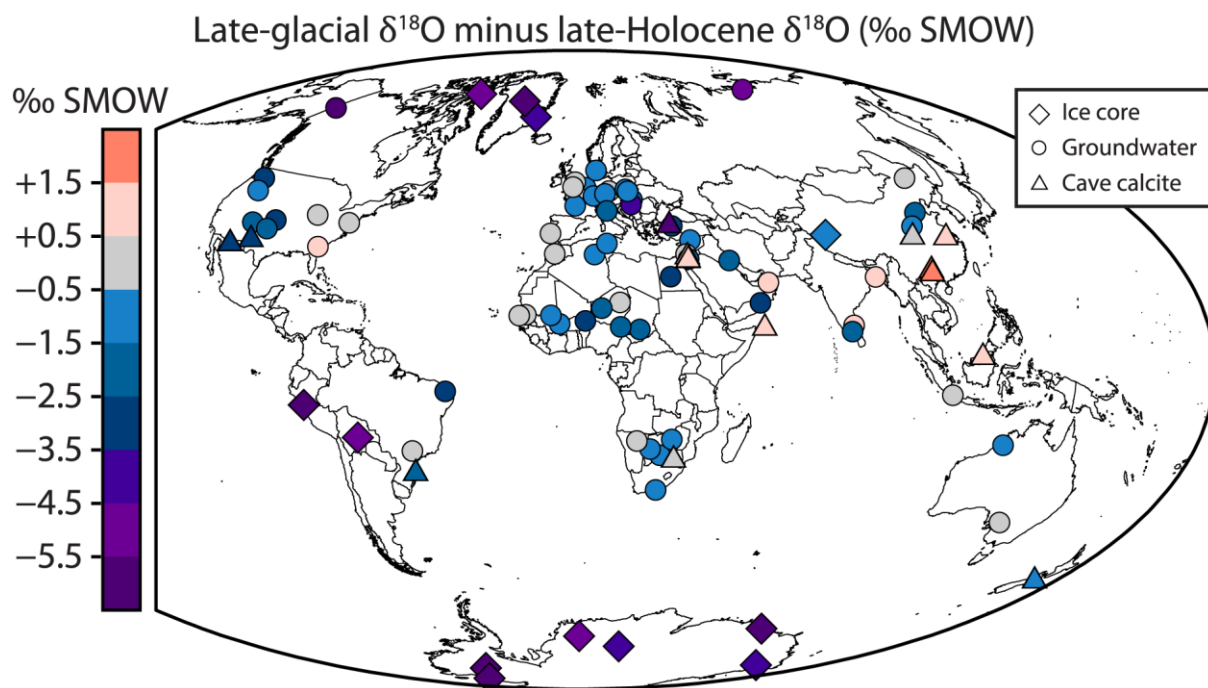
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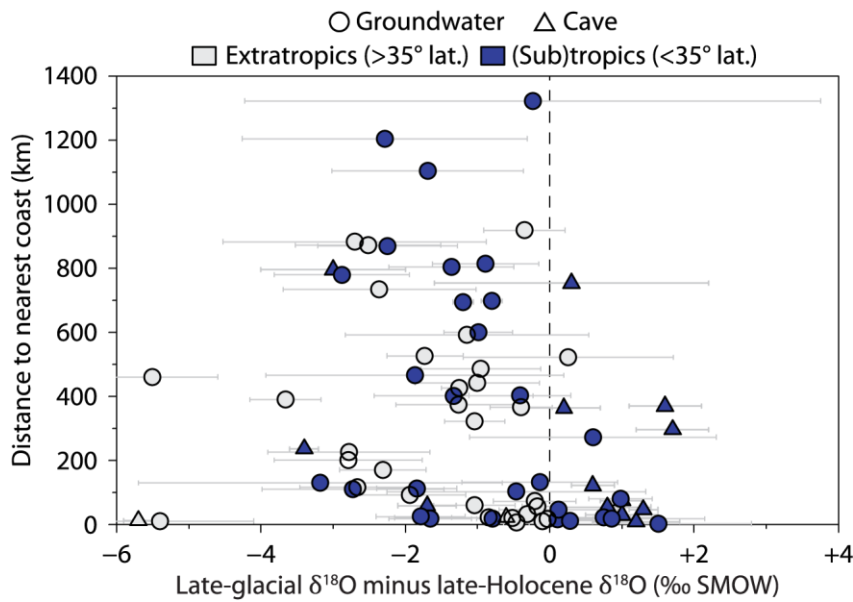


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 2 Figure 1. The change in surface air temperatures from the last glacial maximum to the
 3 preindustrial era (gridded data from Annan and Hargreaves, 2013). (a) Percentile ranges of
 4 temperature changes since the last glacial maximum for 10 degree latitudinal bands. Blue
 5 shading marks the 25th-75th percentile range; thin horizontal lines mark the 10th-90th percentile
 6 range. The grey band shows the globally-averaged estimate of temperature change since the
 7 last glacial maximum of -4.0 ± 0.8 °C. (b) Gridded surface air temperature anomaly from the
 8 last glacial maximum to the preindustrial era (data from Annan and Hargreaves, 2013).



1
2 Figure 2. Meteoric water $\delta^{18}\text{O}$ change from the late-glacial (20,000 to ~50,000 years ago) to the
3 late-Holocene (within past ~5,000 years; average $\Delta^{18}\text{O}_{\text{late-glacial}}$ values shown, where
4 $\Delta^{18}\text{O}_{\text{late-glacial}} = \delta^{18}\text{O}_{\text{late-glacial}} - \delta^{18}\text{O}_{\text{late-Holocene}}$). The low temporal resolution of groundwater
5 records means that $\delta^{18}\text{O}$ variations within each time period are smoothed and likely represent
6 unequal temporal weighting. References for measured meteoric water $\delta^{18}\text{O}$ changes for ice
7 cores, groundwater and cave calcite are presented in the Supplement.

8



1

2 Figure 4. Measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ value variability with distance to the nearest coast
 3 ($\Delta^{18}\text{O}_{\text{late-glacial}} = \delta^{18}\text{O}_{\text{late-glacial}} - \delta^{18}\text{O}_{\text{late-Holocene}}$). Tropical and subtropical locations are shown in
 4 deep blue (<35° absolute latitude), extra-tropical sites are shown in light grey (>35° absolute
 5 latitude). The shape of each point corresponds to groundwater and ground ice (circles) or cave
 6 calcite (i.e., speleothems; triangles). Error bars mark one standard deviation from the mean.

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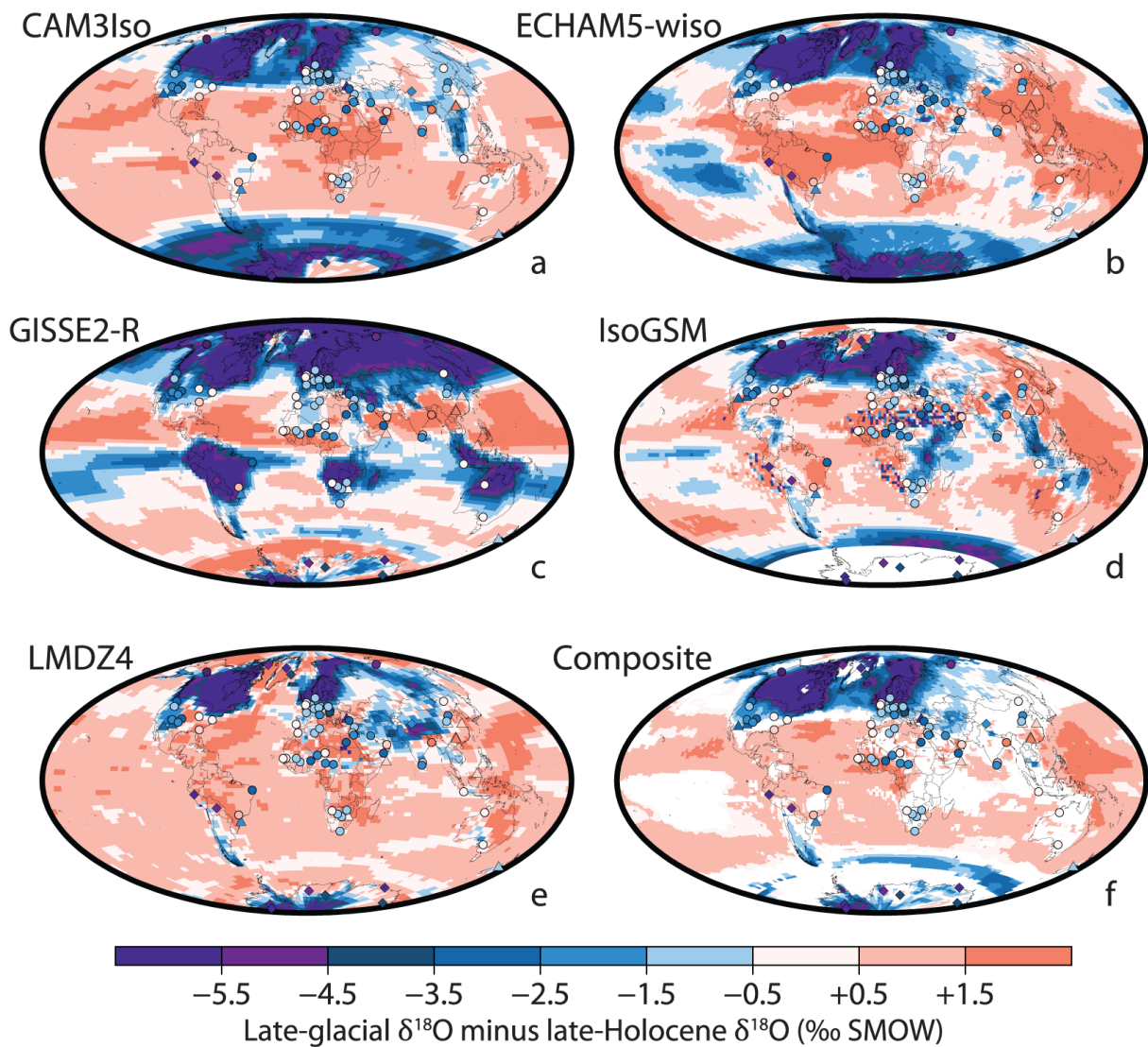
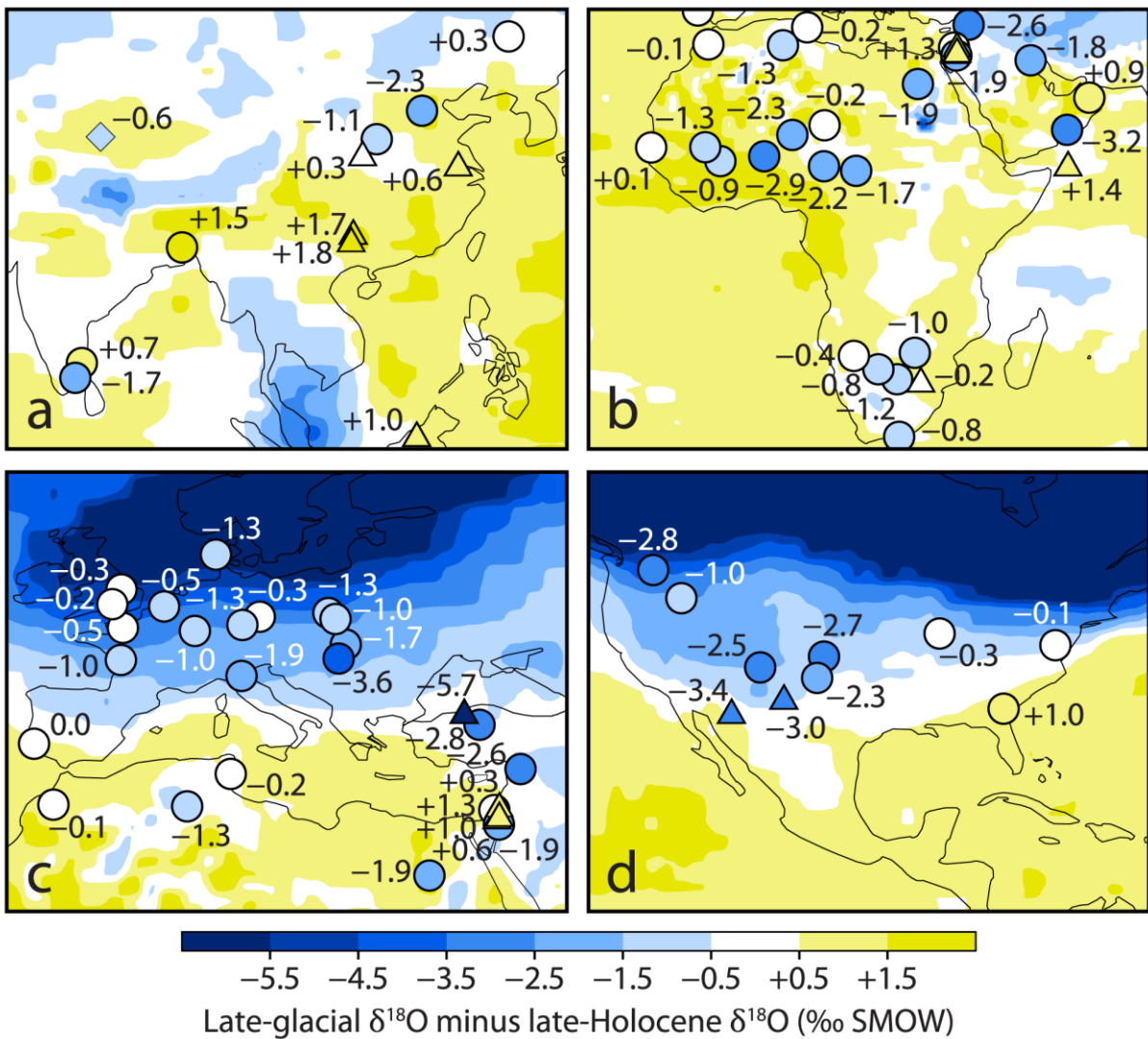


Figure 5. Simulated precipitation $\delta^{18}\text{O}$ differences between the last glacial maximum and pre-industrial time periods (i.e., $\delta^{18}\text{O}_{\text{last glacial maximum}} - \delta^{18}\text{O}_{\text{pre-industrial}}$) from five general circulation models: CAM3iso, ECHAM5-wiso, GISS2-R, IsoGSM and LMDZ4 (Yoshimura et al., 2003; Legrande and Schmidt, 2008; 2009; Risi et al., 2010a; Noone and Sturm, 2010; Pausata et al., 2011a; Werner et al., 2011). Circles (groundwater), triangles (speleothems) and diamonds (ice cores) show measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values from paleoclimate proxy records (Figure 1, original data presented in Tables S2–S5). The panel entitled “Composite” shows the multi-model ensemble median simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ value where at least four of the five models agree on the sign of simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ values (i.e., positive or negative; all five model simulations of $\delta^{18}\text{O}_{\text{last glacial maximum}} - \delta^{18}\text{O}_{\text{pre-industrial}}$ were used to calculate multi-model median shown in “Composite”).



1
 2 Figure 6. Regional proxy record $\Delta^{18}\text{O}_{\text{late-glacial}}$ values for (a) southeastern Asia, (b) Africa, (c)
 3 Europe, and (d) the contiguous United States of America (where $\Delta^{18}\text{O}_{\text{late-glacial}} = \delta^{18}\text{O}_{\text{late-glacial}} -$
 4 $\delta^{18}\text{O}_{\text{late-Holocene}}$). The multi-model ensemble median simulated $\Delta^{18}\text{O}_{\text{late-glacial}}$ value is shown as a
 5 grid (0.5 degree smoothing). Groundwater records are represented by circles, speleothems by
 6 triangles, and ice cores by diamonds, labels show measured $\Delta^{18}\text{O}_{\text{late-glacial}}$ values for each
 7 individual record.