Late-Pleistocene precipitation $\delta^{18}O$ interpolated across the global landmass

Scott Jasechko

1Department of Geography, University of Calgary, Calgary, Alberta, Canada

Abstract Global water cycles, ecosystem assemblages, and weathering rates were impacted by the $\sim 4\,^\circ C$ of global warming that took place over the course of the last glacial termination. Fossil groundwaters can be useful indicators of late-Pleistocene precipitation isotope compositions, which, in turn, can help to test hypotheses about the drivers and impacts of glacial-interglacial climate changes. Here, a global catalog of 126 fossil groundwater records is used to interpolate late-Pleistocene precipitation $\delta^{18}O$ across the global landmass. The interpolated data show that extratropical late-Pleistocene terrestrial precipitation was near uniformly depleted in $^{18}O$ relative to the late Holocene. By contrast, tropical $\delta^{18}O$ responses to deglacial warming diverged; late-Pleistocene $\delta^{18}O$ was higher-than-modern across India and South China but lower-than-modern throughout much of northern and southern Africa. Groundwaters that recharged beneath large northern hemisphere ice sheets have different Holocene-Pleistocene $\delta^{18}O$ relationships than paleowaters that recharged subaerially, potentially aiding reconstructions of englacial transport in paleo ice sheets. Global terrestrial late-Pleistocene precipitation $\delta^{18}O$ maps may help to determine 3-D groundwater age distributions, constrain Pleistocene mammal movements, and better understand glacial climate dynamics.

1. Introduction

Deglacial changes to atmospheric temperatures and water fluxes between the late-Pleistocene (~20000-50,000 years ago) and the late-Holocene (within the last ~5000 years) increased global sea levels, melted two large northern hemisphere ice sheets, and rearranged terrestrial biomes [Clark et al., 2009; Williams, 2009]. Late-Pleistocene precipitation isotope ($\delta^{18}O$) records are applied widely to understand the impacts of deglacial climate warming on earth systems and to better understand groundwater renewal rates (e.g., ice core, speleothem and groundwater records [Thompson et al., 1989; Bar-Matthews et al., 1997; Edmunds, 2009]; where $\delta^{18}O = ([^{18}O/^{16}O]_{\text{sample}}/[^{18}O/^{16}O]_{\text{Vienna standard mean ocean water}} - 1)\times 10^3$ ‰). A key limitation of some paleoclimate evaluations are regional absences of measurement-based constraints on glacial climate conditions [Harrison et al., 2015].

One type of Pleistocene precipitation $\delta^{18}O$ record that has been widely identified and explored are the fossil groundwaters that recharged during the Pleistocene, replenished either by the direct infiltration of precipitation, by focused recharge beneath rivers or lakes, or by the influx of pressurized subglacial meltwaters that existed beneath large northern hemisphere ice sheets during the last ice age [e.g., Edmunds and Wright, 1979; Phillips et al., 1986; Rozanski, 1985; Stute and Deák, 1989; Remenda et al., 1994; Aeschbach-Hertig et al., 2002; Edmunds, 2009; Taylor et al., 2013a]. Interpreting climate signals from fossil groundwater $\delta^{18}O$ records requires understanding processes that govern modern groundwater $\delta^{18}O$ values. Modern groundwater $\delta^{18}O$ values integrate a suite of processes that take place as the water vapor travels inland, condenses, falls, and flows to a location in the aquifer [Dansgaard, 1964; Gat, 1996]. Some of the processes impacting modern meteoric $\delta^{18}O$ values include physiochemical conditions at the ocean-atmosphere interface during evaporation at the moisture origin, upward precipitation-evapotranspiration balances and air mass mixing, seasonal variability in precipitation fluxes, seasonal or interannual variations in the ratio of groundwater recharge as a proportion of precipitation, postdepositional isotope effects, and mixing of waters in the aquifer or soil profile [Vogel et al., 1963; Simpson et al., 1972; Salati et al., 1979; Rozanski et al., 1993; Feng et al., 2009; Risi et al., 2013; Winnick et al., 2014]. It is therefore possible that deglacial $\delta^{18}O$ shifts recorded in fossil groundwaters capture information about the shifts to these and other hydrogeochemical processes that were brought about by the natural climate changes that took place between the late-Pleistocene and the late-Holocene.
Efforts to identify leading drivers of past precipitation $\delta^{18}O$ shifts often benefit from exploring multiple records in a region [Pausata et al., 2011; Mix and Chamberlain, 2014]. Further, several isotope-enabled general circulation models now exist and have been analyzed under recent and glacial climate states [e.g., Lee et al., 2009]. Combining global glacial climate simulations and measured late-Pleistocene precipitation $\delta^{18}O$ records could help to test ideas about deglacial climate changes. However, a measurement-driven map of late-Pleistocene precipitation $\delta^{18}O$ across the global landmass remains unavailable.

The objective of this study is to map changes to global terrestrial precipitation $\delta^{18}O$ between the late-Pleistocene and the late-Holocene across the global landmass. This analysis builds on two recent compilations of the isotopic compositions of fossil groundwaters recharged subaerially or subglacially [Jasechko et al., 2015; Ferguson and Jasechko, 2015]. This analysis differs from each of these two previous works in a number of ways, two of which include the analysis of a substantially greater number of isotope records than have been analyzed previously (~twice as many), and the development of the first measurement-driven estimate of late-Holocene/late-Pleistocene precipitation $\delta^{18}O$ shifts that spans across the global landmass. Critically, the much larger number of records analyzed here is sufficient to interpolate late-Pleistocene precipitation $\delta^{18}O$ values around much of the globe, which was not achievable using smaller data sets [Jasechko et al., 2015]. Herein, the difference between late-Pleistocene precipitation $\delta^{18}O$ and late-Holocene $\delta^{18}O$ is defined as: $\Delta^{18}O$late-Pleistocene $= \delta^{18}O$late-Pleistocene $- \delta^{18}O$late-Holocene.

2. Approach

2.1. Calculating Late-Pleistocene Precipitation $\delta^{18}O$

Deglacial precipitation $\delta^{18}O$ shifts were calculated for 154 proxy records compiled from the primary literature and specialist databases (supporting information; Figure 1). Fossil groundwaters constitute the great majority (82%) of the synthesized records; the other records explored here are speleothems and ice cores. This study interprets isotope compositions of groundwater samples that have radiocarbon dates exceeding 20,000 years as approximations of late-Pleistocene precipitation $\delta^{18}O$ (see section 2.3 entitled Calculation Limitations).

This study compares measured oxygen isotope compositions of paleoclimate records analyzed over two time intervals: the late-Holocene represented by the past 5000 calendar years ($\delta^{18}O$late-Holocene), and the late-Pleistocene represented by ~20,000 to ~50,000 calendar years before present ($\delta^{18}O$late-Pleistocene). The relatively coarse temporal resolutions of each time interval are necessary in order to interpret the ground-water records, which are characterized by uncertain $^{14}C$-based ages (~1000s of years). These boundaries set for the late-Holocene and late-Pleistocene are approximate and are defined by the uncertainty margins and practical limits of radiocarbon-based groundwater age dating.

Groundwater ages were determined using measured $^3$H and $^{14}C$ radioactivities, correcting the ages determined from the latter on the basis of measured $^{13}C/^{12}C$ ratios [Clark and Fritz, 1997] (see methods in Jasechko et al., 2015). Once an estimated age for each sample was calculated, $\delta^{18}O$ shifts between the late-Pleistocene and the late-Holocene ($\Delta^{18}O$late-Pleistocene) were determined by the difference between $\delta^{18}O$late-Pleistocene and $\delta^{18}O$late-Holocene (errors determined following Gaussian error propagation).

To complement the fossil ground records, a series of ice core and cave calcite records are also analyzed here. Cave calcite $\Delta^{18}O$late-Pleistocene records have been corrected for the different temperatures prevailing during the late-Holocene and late-Pleistocene time periods that likely impacted $H_2O-CaCO_3$ fractionation factors [Jasechko et al., 2015] (temperature-fractionation equation from O’Neil et al., 1969). Temperatures used to correct speleothem data for temperature changes were estimated using last glacial maximum [Annan and Hargreaves, 2013] and modern [New et al., 2002] temperature estimates. Therefore, the method used to correct speleothem records for temperature effects introduces additional sources of uncertainty, because the speleothem records were analyzed over a timeframe (~20–50 ka) that overlaps only in part with the Last Glacial Maximum (~19–23 ka), and because modern temperatures are imperfect approximations of late-Holocene temperatures [Marcott et al., 2013]. However, this additional uncertainty, which impacts only the speleothem records, is expected to be less than ~0.5‰. An arguably more important uncertainty lies in the averaging of speleothem data over the necessarily coarse late-Holocene and late-Pleistocene time intervals.
Reports of subglacial groundwater recharge from large, northern hemisphere ice sheets rarely include $^{14}$C measurements. In order to include these data in this synthesis, $\Delta^{18}$O$_{\text{late-Pleistocene}}$ values for subglacial recharge archives were calculated using the minimum measured groundwater $\delta^{18}$O value as an estimate of the subglacial recharge endmember [i.e., $\delta^{18}$O$_{\text{late-Pleistocene}}$] [Ferguson and Jasechko, 2015]. First, nearby (~100 km) subglacial recharge data sets were combined. Next, $\delta^{18}$O$_{\text{late-Holocene}}$ values for each subglacial recharge record were estimated using interpolated modern precipitation $\delta^{18}$O values [Bowen and Wilkinson, 2002]. This data analysis approach applied to subglacial groundwater records introduces several potential biases because modern precipitation $\delta^{18}$O likely differs from late-Holocene precipitation, because global precipitation $\delta^{18}$O often differs from modern groundwater $\delta^{18}$O [Vogel et al., 1963; Simpson et al., 1972], because the lowest $\delta^{18}$O value in a regional aquifer may not represent the absolute glacial end-member $\delta^{18}$O value due to mixing with more recent waters, and because paleo ice sheet isotope compositions likely varied throughout the last ice age and the timing of subglacial recharge remains unclear for most samples. Due to these potential biases, subglacial groundwater isotope records are discussed separately from the other groundwater records that capture direct, subaerial recharge; further, subglacial recharge records are excluded from the geospatial analysis described in the following section 2.2.

2.2. Mapping Late-Pleistocene $\delta^{18}$O

The spatial variability of late-Holocene and late-Pleistocene $\delta^{18}$O values was estimated following an approach developed for modern precipitation $\delta^{18}$O [Bowen and Wilkinson, 2002]. First, records located within 200 m of sea level were separated from records located at higher altitudes. These low elevation measurements. In order to include these data in this synthesis, combining the two regressions describing the empirical relationships of latitude, altitude, and $\delta^{18}$O values was estimated using interpolated modern precipitation $\delta^{18}$O values [Bowen and Wilkinson, 2002].

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Only groundwater and glacial ice records were used to develop equations (1) and (2); cave calcite records were not used because they record relative shifts in $\delta^{18}$O rather than absolute values. $\delta^{18}$O records located more than 200 m above sea level tend to be depleted in $^{18}$O relative to $\delta^{18}$O values predicted using equations (1) and (2). The magnitude of deviation between predicted versus measured $\delta^{18}$O values increases with altitude. An empirical “altitude effect” was determined by regression analysis of altitude versus the $\delta^{18}$O deviation (where $[\delta^{18}$O deviation] = [measured $\delta^{18}$O for the late-Holocene or late-Pleistocene] – [predicted $\delta^{18}$O from equation (1) or (2)]). Empirical “altitude effects” were estimated to be $-3.3\%_{\text{o}}$ km$^{-1}$ for the late-Holocene ($R^2 = 0.50$) and $-3.8\%_{\text{o}}$ km$^{-1}$ for the late-Pleistocene ($R^2 = 0.54$).

Predictions of groundwater $\delta^{18}$O for the late-Holocene and the late-Pleistocene were determined by combining the two regressions describing the empirical relationships of latitude, altitude, and $\delta^{18}$O [Bowen and Wilkinson, 2002]:

$$\Delta^{18}O_{\text{late-Holocene}} = -0.006566(|\text{Lat}|)^2 + 0.2897(|\text{Lat}|) - 7.5 \quad [R^2 = 0.66]$$

and

$$\Delta^{18}O_{\text{late-Pleistocene}} = -0.008467(|\text{Lat}|)^2 + 0.3828(|\text{Lat}|) - 8.8 \quad [R^2 = 0.65]$$

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$$\Delta^{18}O_{\text{late-Holocene}} = -0.006566(|\text{Lat}|)^2 + 0.2897(|\text{Lat}|) - 3.3(\text{Alt}) - 7.5 \quad (3)$$

and

$$\Delta^{18}O_{\text{late-Pleistocene}} = -0.008467(|\text{Lat}|)^2 + 0.3828(|\text{Lat}|) - 3.8(\text{Alt}) - 8.8 \quad (4)$$

where (Alt) represents the elevation of the isotopic record above sea level (in kilometers).

Equations (3) and (4) were used to estimate $\delta^{18}$O changes from the late-Pleistocene to the late-Holocene. Next, residuals were calculated for each isotope record by comparing predicted (i.e., results of equations (3) and (4)) and measured proxy record isotope compositions. Residuals were interpolated (inverse distance weighted) and applied as a correction factor to global grids developed from equations (3) and (4). Cave calcite records, while excluded from the development of equations (1–4), were applied in the residual correction step.

2.3. Calculation Limitations

This global and measurement-driven map of late-Pleistocene to late-Holocene precipitation $\delta^{18}$O shifts is complicated by a number of uncertainties.
First, calculated groundwater radiocarbon age dates implicitly assume that substantial mixing of young and old groundwaters does not take place in the aquifer or along the drilled well; neither assumption is likely valid for all collected samples [Torgersen et al., 2013]. This study uses a straightforward, widely applied, carbon-isotope-based approach to estimate the “mean” groundwater age of each individual sample [Clark and Fritz, 1997]. This method is vulnerable to aggregation errors when the groundwater sample contains a mixture of recent recharge and Pleistocene paleowater [Jasechko, 2016], potentially leading to differences between calculated and true $\Delta^{18}O_{\text{late-Pleistocene}}$ values. In an effort to reduce impacts of this known aggregation bias, the lower-limit of the late-Pleistocene time interval was set higher (~20,000 years ago) than the Pleistocene-Holocene boundary (11,700 years ago), which may be a more logical lower-limit [e.g., Turney et al., 2015]. For most sampled groundwater systems, the $\delta^{18}O$ values level off for samples with ages exceeding ~20,000 years (see supporting information Figures S1–S57 in Jasechko et al. [2015]). In addition, $^{14}C$-based groundwater dating techniques cannot distinguish groundwaters that recharged ~50,000 years ago from much older waters that may predate the Pleistocene [Torgersen et al., 2013]. The possibility that a fraction of a groundwater sample which has a calculated mean age falling within the ascribed late-Pleistocene time interval of ~20–50 thousand years is, in fact, considerably older than ~50 thousand years, cannot be ruled out [e.g., Patterson et al., 2005; Kulangoski et al. 2008].

Second, groundwater and speleothem $\delta^{18}O$ records are interpreted to represent precipitation $\delta^{18}O$ values; however, in some regions, the $\delta^{18}O$ of infiltration differs from the annual precipitation $\delta^{18}O$ value. Precipitation-groundwater comparisons show that recent recharge is often $^{18}O$-depleted relative to annual amount-weighted precipitation isotope compositions due to the seasonal bias in the groundwater recharge ratio to winter precipitation and intensive rainfall [e.g., Vogel et al., 1963; Simpson et al., 1972; Airey et al., 1980; Grabczak et al., 1984; Maué et al., 1994; Winograd et al., 1998; Abbott et al., 2000; Jones and Banner, 2003; O’driscoll et al., 2005; Florea, 2013; Jasechko et al., 2014; Jasechko and Taylor, 2015; Sánchez-Murillo and Birkel, 2016]. It is therefore possible that changes to the degree of seasonal bias in the groundwater recharge ratio between the late-Holocene and the late-Pleistocene influence $\Delta^{18}O_{\text{late-Pleistocene}}$ values recorded by groundwater and speleothem records.

Third, the time interval that has been ascribed to the late-Pleistocene (20–50 thousand years ago) includes numerous and substantial climate changes that are amalgamated, likely unevenly, in the $\Delta^{18}O_{\text{late-Pleistocene}}$ values calculated here. For example, this ascribed late-Pleistocene time interval spans 12 Dansgaard/Oeschger Events (2–13, inclusive), covers four Heinrich Events (H2–H5, inclusive), and poorly captures the timing of minima and maxima of well-established, millennial-scale $\delta^{18}O$ cycles found in some regions [Dansgaard et al., 1993; Grootes and Stuiver, 1997; Wang et al., 2007; Turney et al., 2015]. Further, it is unlikely that $\Delta^{18}O_{\text{late-Pleistocene}}$ records capture precipitation $\delta^{18}O$ shifts over the course of the ascribed late-Holocene and late-Pleistocene time periods evenly. For example, fossil groundwater records in the tropics are likely to disproportionately represent pluvial time periods when groundwater recharge ratios (recharge/pre-pitiation) are maximized [Taylor et al., 2013b; Jasechko and Taylor, 2015]. Further complications may arise near to river corridors where the possibility that groundwaters were recharged beneath riverbeds by precipitation that fell at higher altitudes cannot be dismissed [e.g., Datta et al., 1996]. The uneven sampling of amount-weighted, annual precipitation $\delta^{18}O$ by the proxy records likely impacts calculated $\Delta^{18}O_{\text{late-Pleistocene}}$ values presented here. However, by analyzing precipitation $\delta^{18}O$ records over the late-Holocene and late-Pleistocene time periods at coarser time-scales, an analysis of a large data set of fossil groundwater $\delta^{18}O$ records becomes possible, permitting an examination of how precipitation $\delta^{18}O$ values shifted as global climate shifted from cooler (~20–50 ka) to warmer (~0–5 ka) conditions.

A fourth assumption implicit in this analysis is that groundwater, speleothem, and ice core records preserve the primary $\delta^{18}O$ value of meteoric water. However, there are a number of hydrogeochemical processes that can impact $\delta^{18}O$ values during or following the preservation of these isotope records. Groundwater $\delta^{18}O$ records may be impacted by postrecharge processes such as partial evaporation, groundwater mixing, high-temperature water rock interactions, and other hydrogeochemical processes [Cartwright et al., 2012]. Efforts were made to minimize the possibility of substantial influence of these effects on the calculated $\Delta^{18}O_{\text{late-Pleistocene}}$ values by excluding samples reported to possibly be impacted by substantial geothermal activity or seawater intrusion [e.g., Morrissey et al., 2010]. Speleothem records may not capture annual precipitation isotope compositions due to evaporative cooling of cave drip waters [Cuthbert et al., 2014] or disequilibrium conditions imparted by seasonal differences in drip water fluxes [Mickler et al., 2006; Deininger
et al., 2012). Ice core records may be impacted by ice-vapor exchanges that can modify the $\delta^{18}O$ values that were ultimately preserved in the ice [Steen-Larsen et al., 2014].

Simultaneous compilation and analysis of groundwater, speleothem and ice core records introduces several key uncertainties, some of which are discussed above. Therefore, the maps created in this study constitute nondefinitive estimates of world $\Delta^{18}O_{late-Pleistocene}$ distributions. Nevertheless, no other type of paleoclimate record has been identified and measured with such high-frequency across the global landmass as fossil groundwaters. The strong spatial coherence of calculated $\Delta^{18}O_{late-Pleistocene}$ values presented in section 3 indicates strongly that fossil groundwaters do indeed capture and preserve useful information about late-Pleistocene to late-Holocene precipitation $\delta^{18}O$ shifts. Keeping in mind the processes described above, this global catalog of $\Delta^{18}O_{late-Pleistocene}$ data represents the first measurement-driven estimate of deglacial shifts to global land precipitation $\delta^{18}O$.

3. Results and Discussion

Geographic patterns of late-Pleistocene and late-Holocene precipitation $\delta^{18}O$ shifts are observed at both global and regional scales.
3.1. Global Deglacial $\delta^{18}$O Shifts

The global interpolation of $\Delta^{18}$O$_{late-Pleistocene}$ values demonstrates that late-Pleistocene precipitation $\delta^{18}$O was lower than late-Holocene precipitation $\delta^{18}$O over the great majority (~93%) of the global landmass (Figure 2). Spatial $\Delta^{18}$O$_{late-Pleistocene}$ patterns show that, broadly, tropical and subtropical $\Delta^{18}$O$_{late-Pleistocene}$ records have relatively low-magnitude $\Delta^{18}$O$_{late-Pleistocene}$ values that can be either positive or negative, whereas high-latitude records distal to coasts have negative $\Delta^{18}$O$_{late-Pleistocene}$ values (Figure 1). Positive $\Delta^{18}$O$_{late-Pleistocene}$ values are concentrated in warm humid climates influenced by the seasonal migration of the Intertropical Convergence Zone, including India, Bangladesh, south-central China, and Florida.

The 154 $\Delta^{18}$O$_{late-Pleistocene}$ records average $-2.3_{\text{mean}}$, have a median of $-1.3_{\text{median}}$, and a 10th-90th percentile range of $-6.6_{\text{lower}}$ to $+0.4_{\text{upper}}$ (Figure 1). Tropical versus extratropical $\Delta^{18}$O$_{late-Pleistocene}$ values differ significantly from one another ($p < 0.0001$; two-tailed heteroscedastic $t$-test). Tropical record $\Delta^{18}$O$_{late-Pleistocene}$ values (within 25° of equator; $n = 49$) have a median of $-0.9_{\text{median}}$, an upper-lower quartile range of $-2.7_{\text{lower}}$ to $+0.1_{\text{upper}}$, and a 10th–90th percentile range of $-4.7_{\text{lower}}$ to $+0.8_{\text{upper}}$. Extratropical record $\Delta^{18}$O$_{late-Pleistocene}$ values (more than 35° from the equator; $n = 70$) have a median of $-2.4_{\text{median}}$, an upper-lower quartile range of $-3.6_{\text{lower}}$ to $+1.0_{\text{upper}}$, and a 10th–90th percentile range of $-8.7_{\text{lower}}$ to $-1.0_{\text{upper}}$.

Negative $\Delta^{18}$O$_{late-Pleistocene}$ values are common (128 of 154 records (82%)) in spite of the fact that evaporating seawater—the ultimate source of atmospheric moisture—was $\sim 1_{\text{lower}}$ higher during the late-Pleistocene than during the late-Holocene [Schrag et al., 1996]. Records of late-Pleistocene precipitation isotope compositions are scarce throughout Latin America, eastern Europe, central Asia, and equatorial Africa, but relatively common in the contiguous USA, southeast Asia, Europe, and northern and southern Africa (Figure 2; supporting information Figures S1 and S2).

Globally, high magnitude, negative $\Delta^{18}$O$_{late-Pleistocene}$ values coincide with sites that have low late-Holocene precipitation $\delta^{18}$O values (Figure 3). Low modern precipitation $\delta^{18}$O values characterize air masses that have lost large fractions of their original moisture contents to upwind rainout [Dansgaard, 1964]. Therefore, a plausible, first-order explanation for the global $\Delta^{18}$O$_{late-Pleistocene}$ trends is that the late-Pleistocene climate was characterized by stronger-than-modern isotopic distillation of advected air masses—that is, a higher-than-modern late-Pleistocene precipitation $\delta^{18}$O gradient [Jasechko et al., 2015]. Stronger-than-modern isotopic distillation during the late-Pleistocene is consistent with a nonlinear response of Rayleigh distillation to cooler glacial climate conditions, intensified further by the nonuniform glacial-interglacial warming that was largest outside the tropics and away from the coasts [Annan and Hargreaves, 2013]. However, the global
The δ18O values of groundwater (circles), glacial ice (diamonds), and subglacial recharge that occurred beneath the northern hemisphere ice sheets that existed over northern Europe and Canada during the last ice age (squares). Sites that have lower late-Holocene δ18O values also have high magnitude, negative Δ18Olate-Pleistocene values. The highest magnitude, negative Δ18Olate-Pleistocene values are records of paleoice sheet meltwaters recharged beneath the Laurentide and Fennoscandian ice sheets ("subglacial recharge"). Error bars mark ±1 standard deviation from the mean.

Figure 3. Δ18Olate-Pleistocene and late-Holocene δ18O values of groundwater (circles), glacial ice (diamonds), and subglacial recharge that occurred beneath the northern hemisphere ice sheets that existed over northern Europe and Canada during the last ice age (squares). Sites that have lower late-Holocene δ18O values also have high magnitude, negative Δ18Olate-Pleistocene values. The highest magnitude, negative Δ18Olate-Pleistocene values are records of paleoice sheet meltwaters recharged beneath the Laurentide and Fennoscandian ice sheets ("subglacial recharge"). Error bars mark ±1 standard deviation from the mean.

Zonally, late-Pleistocene δ18O records located at latitudes of greater than 35° show near-uniform 18O-depletion relative to the late-Holocene (77 of 78 extratropical sites). Conversely, only two-thirds of late-Pleistocene δ18O records within 35° of the equator are 18O-depleted relative to the late-Holocene (51 of 76 subtropical and tropical sites), while the remaining sites have positive Δ18Olate-Pleistocene values (Figure 1a). Negative Δ18Olate-Pleistocene values linked to the cool late-Pleistocene climate corroborate precipitation δ18O estimates for the Eocene climate that simulate small modern-Eocene δ18O changes across the low latitudes juxtaposed by higher magnitude, positive (δ18Oocene > δ18Omodern) shifts closer to the poles [Winnick et al., 2015]. Global hot (Eocene, [Winnick et al., 2015]) and cold (late-Pleistocene, this study) house latitude-δ18O gradients suggest that detectable impacts of modern climate warming driven by anthropogenic greenhouse gas emissions on precipitation δ18O are likely to be maximized at the high latitudes rather than the equator.

Subglacial recharge records have the highest magnitude, negative Δ18Olate-Pleistocene values in this data-synthesis (Figures 1 and 3). Subglacial records appear to have a distinct δ18Olate-Holocene/Δ18Olate-Pleistocene isotopic relationship to that of direct, subaerial recharge (Figure 3). Specifically, some fossil groundwater systems replenished beneath the Laurentide and Fennoscandian ice sheets have high-magnitude, negative Δ18Olate-Pleistocene values (less than −7‰) and relatively high late-Holocene precipitation δ18O values (about −15‰ to −8‰; see points for subglacial recharge in Figure 3). Differing δ18Olate-Holocene/Δ18Olate-Pleistocene relationships of subaerially versus subglacially sourced paleowaters are plausibly due in part to the upstream englacial transport of ice that accumulated closer to topographic highs of the paleo ice sheets prior to subglacial recharge. δ18Olate-Holocene/Δ18Olate-Pleistocene relationships may be used to distinguish two distinct recharge processes: (i) subaerial recharge from rain and melting snow, versus (ii) glacier-sourced recharge that took place underneath and along the peripheries of the large northern hemisphere ice sheets that existed during the Pleistocene ice ages.

3.2. Regional Deglacial δ18O Shifts

High densities of Δ18Olate-Pleistocene records are found across Africa, southeast Asia, and North America (Figure 4).

In northern Africa, high-magnitude and negative Δ18Olate-Pleistocene values are found across western Niger, northern Nigeria, southern Chad, and southern Egypt (approximately −2 to −3‰). These continental Δ18Olate-Pleistocene minima form a zonal band across central Africa between the subhumid Sudanian Savanna to the south and hyperarid Sahara to the north, indicating a zonal shift in regional hydroclimate between the late-Pleistocene and the late-Holocene along the modern Sahelian corridor. Relict sand dunes suggest the southern margin of the Sahara Desert was ~5° farther south than its present position at ~19°N during the late-Pleistocene [Grove, 1958]. Presence of a more southerly Sahara during the late-Pleistocene is
Further evidenced by dust abundances in offshore sediment cores that suggest the Sahara expanded south during late-Pleistocene Heinrich stadials [Collins et al., 2013]. Therefore, it is possible that the $\Delta ^{18}O_{\text{late-Pleistocene}}$ minima along the modern Sahelian corridor are in part due to a deglacial shift in the Intertropical Convergence Zone from south (late-Pleistocene) to north (late-Holocene). However, the occurrence of the African Humid Period during the early Holocene—critically omitted from this data analysis and discussion because of the prescribed time intervals—provides strong evidence that nonlinear and complex hydroecological feedbacks linked to ocean circulation patterns and orbital precession are key drivers of African hydroclimate conditions [e.g., Pokras and Mix, 1985; deMenocal et al., 2000; Gasse, 2000; Tjallingii et al., 2008; Tierney et al., 2011]. Therefore, substantial fluctuations in African hydroclimate are not straightforwardly and solely linked to deglacial global warming, highlighting that detailed information about expansion-contraction cycles of the Sahara-Sahalian boundary are unlikely captured by the two-stage Pleistocene-to-Holocene precipitation $\delta ^{18}O$ relationship calculated here.

In other regions of Africa, most $\Delta ^{18}O_{\text{late-Pleistocene}}$ values are negative (Figure 4a; see also Edmunds [2009]). Near-zero $\Delta ^{18}O_{\text{late-Pleistocene}}$ values are confined to coastlines such as west Morocco, Senegal, and Namibia. In southern Africa, western records have near-zero $\Delta ^{18}O_{\text{late-Pleistocene}}$ values, whereas continental and southeastern records have negative $\Delta ^{18}O_{\text{late-Pleistocene}}$ of moderate magnitudes (approximately $-1^{\circ}$). In equatorial Africa, few if any fossil groundwater $\Delta ^{18}O_{\text{late-Pleistocene}}$ records are available, possibly highlighting a data gap that may be met by future field testing of groundwater systems located within $\sim 10^{\circ}$ of the equator. However, leaf wax $\delta ^{2}H$ records for east African rift lakes show large isotopic variations between the late-Pleistocene and the late-Holocene [Tierney et al., 2008]. These fluctuations occur at time scales of less than $\sim 5000$ years and are therefore unlikely captured in full by fossil groundwater isotope records, highlighting the importance of other isotope proxy records that capture shifts at a relatively high temporal resolution (e.g., speleothems, ice cores, lake, and marine sediment cores).

In southeast Asia, $\Delta ^{18}O_{\text{late-Pleistocene}}$ values are positive or near-zero throughout southcentral China, the Malay Archipelago, Bangladesh, and most of India (Figure 4b). Conversely, $\Delta ^{18}O_{\text{late-Pleistocene}}$ values are consistently negative across northern China and along the Malay Peninsula. Near-zero and negative $\Delta ^{18}O_{\text{late-Pleistocene}}$ values along the Malay Peninsula are located where two regional monsoon troughs converge [Aggarwal et al., 2004] and have lower $\Delta ^{18}O_{\text{late-Pleistocene}}$ values than those observed to the west (Bangladesh), north (southern China), and east (Borneo), potentially impacted by different glacial-interglacial land-atmosphere feedbacks (e.g., moisture recycling by evaporation or plant transpiration) due to Pleistocene subaerial exposure of the Sunda Shelf [Partin et al., 2007].

In the contiguous USA, $\Delta ^{18}O_{\text{late-Pleistocene}}$ values near to the southeastern coastline have near-zero and positive $\Delta ^{18}O_{\text{late-Pleistocene}}$ values, whereas modern-day southwest deserts and central plains have consistently high magnitude, negative $\Delta ^{18}O_{\text{late-Pleistocene}}$ values close to $-2.5^{\circ}$ [Phillips et al. 1986] (Figure 4c). Positive $\Delta ^{18}O_{\text{late-Pleistocene}}$ values characterize hot humid climates along the southeast coast (e.g., Georgia and Florida) [Plummer, 1993; Morrissey et al., 2010], whereas $\Delta ^{18}O_{\text{late-Pleistocene}}$ values along the southwest Pacific...
coast are near-to-zero or negative (e.g., near Oakland, Los Angeles, San Diego in California). Continental-scale patterns of $\delta^{18}O$ are broadly similar for the late-Pleistocene and late-Holocene time periods, implying that the architecture of the largest, synoptic-scale moisture supplies to the contiguous USA may have remained broadly similar between the two climate states. However, $\Delta^{18}O_{\text{late-Pleistocene}}$ patterns show a steeper continental gradient in precipitation $\delta^{18}O$ during the late-Pleistocene relative to the late-Holocene, consistent with a nonlinear response of isotopic distillation that was plausibly intensified by larger glacial-interglacial temperature shifts distal to the coasts.

4. Concluding Remarks

The primary result of this study is a first estimate of the spatial distribution of deglacial precipitation $\delta^{18}O$ shifts across the global landmass (Figure 2). The global map of deglacial precipitation $\delta^{18}O$ shifts suggests (i) that extratropical precipitation was near-uniformly $^{18}O$-depleted relative to the late-Holocene, (ii) that subtropical and tropical precipitation $\delta^{18}O$ responses to deglacial climate warming diverge between the two monsoonal climates of Africa and southeast Asia, (iii) that groundwaters recharged by meltwaters of the Fennoscandian and Laurentide ice sheets might be distinguished from subaerial-sourced paleo-groundwaters on the basis of their $\delta^{18}O$ values, (iv) that late-Pleistocene precipitation at the great majority of high latitudes and high altitudes was strongly $^{18}O$-depleted relative to the late-Holocene, and (v) that polar air masses likely underwent stronger-than-modern isotopic distillation as they advected poleward during the late-Pleistocene.

Broader implications of late-Pleistocene precipitation $\delta^{18}O$ maps could extend to a number of different fields, as evidenced by the array of applications of modern precipitation $\delta^{18}O$ maps to mammal migration and ecosystem water use studies. It is therefore possible that late-Pleistocene precipitation $\delta^{18}O$ distributions, while of poor temporal resolution, could provide some constraints on certain late-Pleistocene earth and biologic systems processes. For example, late-Pleistocene paleo-groundwater $\delta^{18}O$ values have in past been used to distinguish paleo-groundwaters from younger Holocene-aged groundwaters, thereby better constraining groundwater velocity and mixing [Samborska et al., 2012]. The maps developed here show that the usefulness of $\delta^{18}O$ measurements to partition groundwater age distributions (i.e., fractions of Holocene and paleo-groundwater in a sample) is likely to be greatest in regions with high magnitude $\Delta^{18}O_{\text{late-Pleistocene}}$ values (e.g., north Africa, continental North America, east Europe, central Asia). Further, the interpolated $\Delta^{18}O_{\text{late-Pleistocene}}$ data could help provide data-driven constraints on Pleistocene/Holocene spatial precipitation $\delta^{18}O$ gradients that have been used in past to reconstruct and constrain Pleistocene mammal migration extents [e.g., Koch et al., 1998]. Finally, the $\Delta^{18}O_{\text{late-Pleistocene}}$ maps developed here may help to evaluate the numerous isoype-enabled climate models that are applied frequently to evaluate hypotheses about glacial climate dynamics and the impacts of deglacial climate warming on water resources.

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