Insights into deglacial East Asian Monsoon seasonality and inter-regional teleconnections from Lake Suigetsu, Japan


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Abstract

The past evolution of the East Asian Monsoon (EAM) was affected by remote climate processes. However, there is uncertainty surrounding these teleconnections and their impact on each seasonal EAM mode during periods of rapid global change. Here we use the unique biannual precipitation patterns of central Japan to reconstruct winter and summer EAM evolution during the last glacial termination (22,000 – 10,000 cal BP). Oxygen isotope analysis of diatoms and hydrogen isotope analysis of n-alkanoic acids from the Lake Suigetsu sediment cores show that in Japan the winter EAM weakened and the summer EAM strengthened with deglaciation. Only the summer mode exhibited variations coeval with stade-interstade fluctuations. A relationship between the summer EAM and Antarctic temperature post-16,000 cal BP supports a remote link to the Southern Hemisphere. These trends were different to continental EAM records and highlight the value of broadening the geographical range of reconstructions to rationalise system heterogeneities.

Introduction

The East Asian Monsoon (EAM) region is located at a point of convergence of multiple, globally interconnected climatic systems, with the palaeo-EAM system demonstrably affected by remote surface and atmospheric processes on centennial to millennial timescales. This is most eminently exemplified by the teleconnection between summer EAM strength and North Atlantic temperature during the late Pleistocene; first evidenced by the relationship between the Hulu Cave stalagmite and Greenland ice core oxygen isotope (δ18O) records (Wang et al., 2001; Zhang et al., 2019) and subsequently corroborated by studies from across the region (Herzschuh, 2006; Liu et al., 2022). This relationship is thought to have been driven by the influence of Atlantic Meridional Overturning Circulation (AMOC) on the position of the Intertropical Convergence Zone (ITCZ) (Zhang et al., 2019). However, other reconstructions of EAM strength have noted an inverse relationship to Antarctic temperatures in this interval (Han et al., 2016; Zhang et al., 2016) and, as a result, the view...
of teleconnections acting on the EAM region has expanded to consider links to other climate systems, including Antarctica and the Pacific Ocean.

The behaviour of the EAM during the last glacial termination (TI; 22,000 to 10,000 cal BP) is an ideal focus for characterising the changing seasonality of this system and the persistence of these teleconnections. As a global transition, TI exemplifies a substantial, detectable change in many palaeoclimate archives with the climate at the geographic poles (representing two possible “end member” teleconnection origins) decoupled, with a Southern Hemisphere lead (Denton et al., 2010; WAIS Divide Project Members, 2013). In the North Atlantic, TI was characterised by colder conditions during Greenland Stade 2.1 (GS-2.1; ~22,900 to 14,700 cal BP) and Greenland Stade 1 (GS-1; ~12,800 to 11,700 cal BP), alternating with warmer conditions during Greenland Interstade 1 (GI-1; ~14,700 to 12,800 cal BP) and the Holocene (~11,700 cal BP onwards) (Rasmussen et al., 2014). GS-2.1, GI-1 and GS-1 were equivalent to the Last Glacial Maximum (LGM), Bølling–Allerød (BA) Interstade and Younger Dryas (YD) Stade of the wider North Atlantic region. In the far Southern Hemisphere, warming was more gradual and interrupted only by the Antarctic Cold Reversal (ACR; 14,500 to 12,900 cal BP), which bestrode GI-1 and GS-1 (EPICA Community Members, 2006; WAIS Divide Project Members, 2013).

Whilst current evidence supports a weakening East Asian Winter Monsoon (EAWM) and strengthening East Asian Summer Monsoon (EASM) during TI on an orbital scale (Wang et al., 1999; Gallagher et al., 2018), centennial- to millennial-scale fluctuations in EAM strength and the relative seasonal contributions to annual precipitation are poorly constrained (particularly in the case of the EAWM; Wen et al., 2016) and appear spatially heterogeneous (Zhang et al., 2019). It is unsurprising, therefore, that there is uncertainty surrounding deglacial EAM evolution on submillennial timescales (Yoshida and Takeuti, 2009; Hayashi et al., 2010; Liu et al., 2022), particularly in terms of conflicting evidence for teleconnections to other regions. A possible source of this uncertainty is the limited number of available high-resolution EAM reconstructions. Consideration of both EAWM and
EASM strength in isolation is vital for deconvolving EAM behaviour during TI because they do not always show a perfectly inverse relationship (Wang et al., 2012; Wen et al., 2016; Yan et al., 2020) and hence have the potential to exhibit unique behaviours (and teleconnection signals).

Due to its location, the EAM dominantly controls the climate of Japan; however, the relationship between the EAM and precipitation in Japan is unique. Japan is situated directly beneath the seasonally migrating monsoon front (Fig. 1), a critical geoclimatic boundary which moves northwards during summer and southwards during winter and separates the influence of the EAWM and EASM modes (Nakagawa et al., 2006, Yoshida and Takeuti, 2009). The Japanese archipelago lies between two water bodies (the Pacific Ocean and the Sea of Japan), hence both EAM modes bring heavy precipitation to the country; mixed rainfall and snowfall concentrated along the Sea of Japan coast during the EAWM, and rainfall across the entire country during the EASM (Chowdary et al., 2019; Amekawa et al., 2021). This is in contrast to Continental Asia, where the EASM brings heavy rainfall from the Pacific Ocean, but the EAWM is predominantly dry because it originates inland (Yancheva et al., 2007; Yan et al., 2020). As such, the annual distribution of rainfall in Japan uniquely reflects a balance between both seasonal modes of the EAM. In light of these distinctive climatic characteristics, EAM records from Japan are well suited to deconvolve the complexities of EAM evolution during TI. In this study we present new stable isotope-based proxy records of both EAWM and EASM evolution in central Japan from the Lake Suigetsu sediment cores and compare these to a series of global benchmark records of EAM variability and temperature changes (Fig. 1), to better understand the factors affecting both seasonal modes of the EAM in Japan during this interval.
Fig. 1 – Lake Suigetsu and other key sites. a) the key global benchmark record sites discussed in the main text. b) the sites within the EAM region, alongside the average positions of the modern monsoon front in January and July (adapted from Nakagawa et al., 2006). c) the location of Lake Suigetsu in Fukui Prefecture, central Honshu, relative to key cities and geographical features in Japan. Red boxes show the extent of subsequent panels in the sequence. Basemap is World Imagery (WGS84) from Esri (2023) (scale 1:37,314,123 (a), 1:7,851,046 (b), 1:530,430 (c)).

Lake Suigetsu is a tectonic lake located ~1 km from the Sea of Japan in Fukui Prefecture, Honshu Island (35° 35’ N, 135° 53’ E; Fig. 1; Nakagawa et al., 2012). Suigetsu was a freshwater lake for the last glacial-interglacial cycle, except during global sea level highstands in the Eemian and the mid-Holocene, and post-1664 CE when the lake was artificially connected to the Sea of Japan (Nakagawa et al., 2021). Four previous deep coring campaigns carried out in 1993 (SG93 core), 2006
(SG06), 2012 (SG12) and 2014 (SG14) recovered a series of overlapping sediment core sections from the present day to >150,000 yr BP (Nakagawa et al., 2012). The preserved sequence, covering >98 m of composite sediment depth, provides a palaeoenvironmental archive that captures two glacial and two full interglacial periods of continuous sedimentation. The Suigetsu cores contain annual laminations (varves) between ~70,000 yr BP and 10,000 cal BP, and between ~50,000 and 10,000 cal BP comprise the longest continuously varve-counted record from the Quaternary (Schlolaut et al., 2018). Thin-section microscopic analysis showed that the varves comprise a spring layer of Aulacoseira diatoms with some siderite, a detrital layer containing silt-sized quartz and feldspar, a summer layer of light amorphous organic material, an autumn layer of Enyonema diatoms within siderite, and a transitional autumn-to-winter clay layer (Schlolaut et al., 2012).

Previous studies of the Suigetsu sediments from TL substantiate that this was a period of significant climatic change at the catchment. Pollen-derived temperature reconstructions of this interval show a structure which resembles that of North Atlantic temperature: abrupt warming at the start of a “Late Glacial Interstade”, a more gradual decrease in temperature into a “Late Glacial Stade”, and an abrupt warming at the start of the Holocene (Nakagawa et al., 2021). However, whilst the onset of the Late Glacial Stade and the Holocene at Lake Suigetsu were synchronous with the Younger Dryas and Holocene periods in the North Atlantic, the Late Glacial Interstade at Lake Suigetsu commenced ~200 years prior to its North Atlantic counterpart (the Bølling–Allerød) (Nakagawa et al., 2021). This behaviour was rationalised by a sudden repositioning of the westerly jet to the north of the Himalayas (a topographical barrier which causes bimodality in the positioning of the jet) during boreal summer, which allowed warm air from the Pacific Ocean to propagate northwards and caused the step-change in summer temperatures in Japan. Conversely, it was an AMOC switch-on ~200 years later (due to a higher threshold response to insolation forcing), which resulted in a shift to interstadial conditions in the North Atlantic. On a subcentennial scale, multiproxy analysis of the Suigetsu cores showed a bipartite structure to the Late Glacial Stade with
a transition that was synchronous, albeit slower and more muted, to those observed at Lakes Meerfelder Maar (Germany) and Kråkenes (Norway) (Schlolaut et al., 2017). A shift from *Aulacoseira ambiguа* to *Aulacoseira subarctica* dominance during deglaciation was attributed to variations in the EAWM, affecting vegetation coverage and the water nutrient status (Kossler et al., 2011). Carbon and nitrogen isotope ratios from chlorins also suggest an increase in aquatic primary productivity from glacial to interglacial conditions in the lake (Tyler et al., 2010).

Here we present EAM precipitation reconstructions from Lake Suigetsu based on two proxy systems: the oxygen isotope composition of aquatic diatom silica frustules (δ¹⁸O_{diatom}) and the compound-specific hydrogen isotope composition of n-alkanoic acids (δ²H_{acid}) produced by aquatic algae (C₁₆ and C₁₈ homologues) and terrestrial plants (C₂₈ and C₃₀ homologues). Analysis was conducted using 100-year contiguous (continuous adjacent) samples. These proxies were selected based on target material abundances in the Lake Suigetsu cores and the ability to reconstruct both autochthonous (aquatic; δ¹⁸O_{diatom}, δ²H_{C₁₆acid} and δ²H_{C₁₈acid}) and allochthonous (terrestrial; δ²H_{C₂₈acid} and δ²H_{C₃₀acid}) changes during TI (Leng and Barker, 2006; Holtvoeth et al., 2019). Both isotope proxy systems are versatile tools in palaeolimnological reconstructions of climate and follow the principle that the isotope composition of palaeo-catchment water was incorporated (via biosynthetic processes) into target materials by diatoms, algae, and plants, and preserved in the lake (Leng and Barker, 2006; Holtvoeth et al., 2019). δ¹⁸O_{diatom} and δ²H_{acid} can be affected by fractionation processes during the life cycle of their respective source organisms but are dominantly controlled by the isotope composition of their water source (Leng and Marshall, 2004; Holtvoeth et al., 2019). Hence, broader-scale palaeohydrological inferences can be made, relating to regional-scale shifts in atmospheric circulation and water vapour transport (Leng and Marshall, 2004; García-Alix et al., 2021; Tierney et al., 2022). At Lake Suigetsu, the modern isotope composition of water within the catchment is closely related to that of precipitation (with minimal effects of evaporation in summer), which is dominantly controlled by both the EAWM and EASM (Rex et al., 2023, preprint), supporting...
the utilisation of δ\(^{18}\)O\(_{\text{diatom}}\) and δ\(^2\)H\(_{\text{acid}}\) for reconstructing past EAM behaviour. Other potential influencing factors are water temperature in the aquatic domain, and air temperature and vegetation changes in the terrestrial domain (Leng and Barker, 2006; Castañeda and Schouten, 2011; Sachse et al., 2012).

**Results**

**Observations at Lake Suigetsu**

The δ\(^{18}\)O\(_{\text{diatom}}\) values increased smoothly during TI with no abrupt shifts and the Holocene values were ~2.5 ‰ higher than in the LGM (Fig. 2a). Key features of this profile were high frequency centennial-scale fluctuations and higher values between 22,000 and 21,000 cal BP. The δ\(^2\)H\(_{\text{C16acid}}\) values followed a smooth trend to lower values through the record, except from 22,000 – 21,000 cal BP (when low values were observed), 19,000 – 17,000 cal BP (when values fluctuated drastically on a centennial scale), and sections of the Late Glacial Interstade and Late Glacial Stade (when this proxy exhibited periods of lower-than-average values; Fig. 2b). The difference between the δ\(^2\)H\(_{\text{C16acid}}\) values at 21,000 cal BP and 10,000 cal BP was ~50 ‰. The δ\(^2\)H\(_{\text{C18acid}}\) profile resembled that of δ\(^3\)H\(_{\text{C16acid}}\), however long-term trends were harder to identify due to limited data coverage (see Methods) and rapid high-amplitude fluctuations (Fig. 2c). Many of these coincided with equivalent, lower amplitude, variations in the δ\(^2\)H\(_{\text{C16acid}}\) values, suggesting that these were closely related. The δ\(^3\)H\(_{\text{C28acid}}\) and δ\(^3\)H\(_{\text{C30acid}}\) profiles also showed very similar behaviour to one another; decreasing through the LGM and into the Late Glacial Interstade before initially increasing smoothly for ~2000 years, and then decreasing from the early Holocene onwards (Fig. 2d, Fig. 2e). Like the other δ\(^2\)H\(_{\text{acid}}\) proxies, the δ\(^2\)H\(_{\text{C28acid}}\) and δ\(^2\)H\(_{\text{C30acid}}\) records showed a period of lower values between 22,000 and 21,000 cal BP before a marked shift at 21,000 cal BP to higher values. The difference between the δ\(^3\)H\(_{\text{C30acid}}\) values at 21,000 cal BP and 10,000 cal BP was ~35 ‰. The δ\(^2\)H\(_{\text{C28acid}}\) and δ\(^2\)H\(_{\text{C30acid}}\) profiles
were only decoupled between 19,000 and 17,000 cal BP; within this section both proxies showed rapid high amplitude changes which were offset by as little as one sample (100 years).

It appears that the shortest (C_{16}) and longest (C_{30}) chain lengths exhibited $\delta^2H_{\text{acid}}$ profiles with the most unique trends and the most complete data coverage, and hence we focus on these in subsequent discussions, whereas the $\delta^2H_{\text{C_{18}acid}}$ values were limited in quantity and the $\delta^2H_{\text{C_{28}acid}}$ values showed some features similar to the $\delta^2H_{\text{C_{18}acid}}$ profile, possibly due to C_{28} n-alkanoic acids originating from a combination of aquatic and terrestrial sources. Whilst C_{30} n-alkanoic acids originate from strictly terrestrial sources, C_{28} n-alkanoic acids have been known to originate from aquatic and terrestrial sources (van Bree et al., 2018; Tierney et al., 2022). None of the trends exhibited by the isotope proxies resembled those of the pollen-derived temperature reconstructions at Lake Suigetsu (mean temperature of the warmest month, MTWA or mean temperature of the coldest month, MTCO; Fig. 2f, Fig. 2g; Nakagawa et al., 2021). Proxies reflective of local catchment change (e.g., total organic carbon, n-alkanoic acid concentrations and diatom frustule concentration) exhibited smooth fluctuations during TI (Supplementary Fig. 1). Of these, only diatom frustule concentration showed any step-changes (at the onset of the Late Glacial Interstade), suggesting that the other proxies were driven by gradual changes, as for the $\delta^{18}O_{\text{diatom}}$ and $\delta^2H_{\text{acid}}$ values (Supplementary Discussion 1).
**Fig. 2** – Proxy changes at Lake Suigetsu during Termination I. a) $\delta^{18}O_{\text{diatom}}$ (this study), b) $\delta^{2}H_{\text{acid}}$ of C_{16} n-alkanoic acids (this study), c) $\delta^{2}H_{\text{acid}}$ of C_{18} n-alkanoic acids (this study), d) $\delta^{2}H_{\text{acid}}$ of C_{28} n-alkanoic acids (this study), e) $\delta^{2}H_{\text{acid}}$ of C_{30} n-alkanoic acids (this study), f) MTWA; mean temperature of the warmest month (pollen-derived; Nakagawa et al., 2021), g) MTCO; mean temperature of the coldest month (pollen-derived; Nakagawa et al., 2021). Panels A to E are displayed inverted, and show data derived from 100-year integrated contiguous samples. All panels are on the IntCal20 timescale. Vertical dashed lines indicate the phases of temperature change at Suigetsu as designated by the pollen-derived temperature reconstructions (Nakagawa et al., 2021).
Determination of the dominant isotope proxy drivers is critical for interpreting the results in the context of EAM behaviour. Whilst multiple factors can affect the proxy values presented here, valuable insights can be made by identifying the dominant driver (which can cause the largest amplitude changes) in each case. There was no identifiable relationship between our isotope proxy values and the pollen-derived temperature or the primary pollen taxa concentrations, which eliminates air temperature (and thus water temperature) and vegetative change as the main drivers, respectively. There is a temperature dependent fractionation on $\delta^{18}O_{\text{diatom}}$ in precipitation and a mineral-water fractionation, which have opposing effects, but the impact is likely to be small. Given the amount of variation, we suggest that changes in water source composition were most likely the dominant control on the $\delta^{18}O_{\text{diatom}}$ and $\delta^2H_{\text{acid}}$ values. The isotope composition of modern catchment waters is principally governed by the composition of EAM-borne precipitation, as shown by extended monitoring (Rex et al., 2023, preprint). EAM precipitation composition is determined by a combination of source composition and transport processes; in general, a “stronger” EAM, associated with strong winds and large quantities of precipitation integrated across the transport pathway, would act to lower $\delta_{\text{precipitation}}$, and the reverse qualities (for a “weaker” EAM) would provide higher $\delta_{\text{precipitation}}$ values (Yuan et al., 2004; Zhang et al., 2018). EAM strength is ultimately driven by an enhanced land-sea pressure gradient, moisture availability and conditions favourable for evaporation at source (Mohtadi et al., 2016). It follows that these were the critical drivers of the isotope proxy systems during TI. On shorter timescales, other influences (e.g., local summer evaporation) could have also affected the isotope proxies, although there is a limited influence of evaporation on catchment isotopes in the modern day (Rex et al., 2023, preprint), so we expect these to have caused small-amplitude effects, if any. The marked seasonality of EAM precipitation at Lake Suigetsu, combined with the short residence time of the lake (on the order of a few months; Rex et al., 2023, preprint) means that isotope proxy seasonality is crucial for attribution to one or
both seasonal EAM modes. A key element of this designation is related to the modern catchment transit lag of approximately three months (Rex et al., 2023, preprint), which signifies that surface lake water isotope composition ($\delta_{\text{lake}}$) reflects that of the precipitation ($\delta_{\text{precipitation}}$) delivered in the preceding season, i.e., spring $\delta_{\text{lake}}$ is equivalent to that of winter $\delta_{\text{precipitation}}$ and so on. Consistently, deep water $\delta_{\text{lake}}$ is homogeneous and reflects annually integrated $\delta_{\text{precipitation}}$.

In a palaeo context, different diatom taxa were associated with each of the predominant seasonal blooms in the lake (spring and autumn) and consequently, the $\delta^{18}$O$_{\text{diatom}}$ values reflect a combination of seasonal signals. Of the most abundant diatom taxa, mixed planktonic *Aulacoseira subarctica* and *Aulacoseira ambigua* diatoms bloomed in the spring and captured surficial spring $\delta^{18}$O$_{\text{lake}}$ (winter $\delta^{18}$O$_{\text{precipitation}}$), and benthic *Encyonema* diatoms bloomed during the autumn and captured annually integrated $\delta^{18}$O$_{\text{precipitation}}$ from the homogenous deep water (bloom season attributions were derived from the varve structure; Schlolaut et al., 2012). The signals captured by other taxa would have been dependent on their habitation depths and bloom seasons, however due to their low abundance relative to *Aulacoseira* and *Encyonema* diatoms, their seasonality was not ascribed as part of the varve structure. “Bulk” $\delta^{18}$O$_{\text{diatom}}$ therefore reflects mixed seasonality, with the relative weighting of each seasonal signal dependent on the mass contribution of each taxon (Leng and Barker, 2006). In the absence of quantified taxa-specific masses within each sample, the abundance (based on frustule counts) of each taxon can be used to provide an indication of the relative seasonal weighting of the $\delta^{18}$O$_{\text{diatom}}$ values; differences in taxon size can introduce uncertainty to this process, but we believe this to be a reasonable assessment because whilst other taxa are larger in size, the smallest genus of diatom, *Aulacoseira*, has thick valves and hence contributes significantly to the total diatom mass. Due to its relationship with winter $\delta^{18}$O$_{\text{precipitation}}$, *Aulacoseira* $\delta^{18}$O$_{\text{diatom}}$ offers the most robust links to EAWM behaviour. In theory, this connection may have been affected by alterations to the transit lag, however we believe that the relationship was maintained throughout TI, despite climatic variability, because the spring diatom bloom in Lake
Suigetsu is closely related to the delivery of the bulk of winter precipitation to the lake (i.e., both are caused by snow and ice melting). Additionally, winter is the wettest season of the year, minimising the effect of the preceding autumnal precipitation on the composition of surface $\delta_{\text{lake}}$ (an effect amplified by the short residence time of the lake). In light of this, it is reasonable to utilise the $\delta^{18}_{\text{O diatom}}$ values as an indicator of EAWM behaviour (i.e., strongly weighted towards winter $\delta^{18}_{\text{O precipitation}}$), when *Aulacoseira* was the most abundant genus. This was the case for earlier parts of the interval (22,000 – 19,800 cal BP) and from 15,000 cal BP onwards, when %*Aulacoseira* (*Aulacoseira* frustules as a percentage of total diatom frustules) was consistently greater than 60 % (and often greater than 70 %) on a centennial basis (Supplementary Fig. 2). This was likely to have been even greater in our $\delta^{18}_{\text{O diatom}}$ samples, which contained only diatoms which were smaller than 80 µm, favouring the (smaller) *Aulacoseira* over other taxa. However, the presence of other taxa may have contributed to the low amplitude centennial-scale variability in the $\delta^{18}_{\text{O diatom}}$ record throughout, and had greater influence between 19,800 and 15,000 cal BP. Maturation effects (i.e., post-mortem isotope exchange) remain an unconstrained influence on $\delta^{18}_{\text{O diatom}}$ (e.g., van Hardenbroek *et al.*, 2018), however (as with other studies) we do not expect this to be a dominant driver because the hydrous (most readily exchangeable) layer of each diatom was removed prior to measurement (see Methods) and only the more resistant structurally bound oxygen fixed during early sedimentation was analysed.

The $\delta^{2}_{\text{H acid}}$ values are associated with less complex seasonality than the $\delta^{18}_{\text{O diatom}}$ values, because the growth of terrestrial plants occurs predominantly during spring and summer (Sachse *et al.*, 2012) and lake-based organic matter production in Lake Suigetsu occurs almost exclusively in summer when the lake becomes stratified and aquatic productivity is high (Schlolaut *et al.*, 2012). Indeed, because the terrestrial $\delta^{2}_{\text{H acid}}$ values would have been unaffected by the catchment transit lag (the source water of terrestrial plants being soil pore water not lake water), we expect the $\delta^{2}_{\text{H C30acid}}$ values to have been heavily weighted towards summer (EASM) $\delta^{2}_{\text{H precipitation}}$. However, not
only would the aquatic $\delta^2$H$_{\text{acid}}$ values have been affected by the transit lag (and hence captured the isotope composition of spring $\delta^2$H$_{\text{precipitation}}$ = summer $\delta^2$H$_{\text{lake}}$), but during periods of cold winters (which TI typifies), we might expect extension of the transit lag time in the first half of the year due to greater longevity of snow on the surrounding hillsides and hence slower delivery of precipitation to the lake system (Rex et al., 2023, preprint). As a result, both winter and summer precipitation would enter the lake in quicker succession. Under these circumstances, summer $\delta^2$H$_{\text{lake}}$ would have become a greater seasonal mix of EAWM $\delta^2$H$_{\text{precipitation}}$ (entering the lake in late spring) and EASM $\delta^2$H$_{\text{precipitation}}$ (entering the lake in late summer). Consequently, the $\delta^2$H$_{\text{C16acid}}$ values reflect the weighted average of annual $\delta^2$H$_{\text{precipitation}}$. By calculating the numerical difference between the $\delta^2$H$_{\text{C16acid}}$ and $\delta^2$H$_{\text{C30acid}}$ values, it is possible to qualitatively infer the relative influence of the EAWM and EASM on the lake through time (Supplementary Discussion 2, Supplementary Fig. 3), although there is also an effect of reduced lag time with warming on this signal.

### EAM Behaviour in Japan During TI

By considering the high %$\text{Aulacoseira}$ intervals in isolation (Supplementary Fig. 2), variations in the $\delta^{18}$O$_{\text{diatom}}$ values strongly support a reduction in EAWM strength during TI in Japan (Fig. 3d). This aligns with Japanese stalagmite reconstructions of EAWM behaviour from the LGM and Holocene (Sone et al., 2013; Amekawa et al., 2021). It is possible that EAWM weakening caused higher $\delta^{18}$O$_{\text{diatom}}$ values both directly (as discussed above) and indirectly (by reducing EAWM surface runoff into the Sea of Japan, resulting in a transition from low- to high-salinity conditions in these EAWM source waters (Amekawa et al., 2021)). The $\delta^{18}$O$_{\text{diatom}}$ profile does not exhibit stade-interstade-like fluctuations as seen in the North Atlantic (typified by GRIP $\delta^{18}$O$_{\text{ice}}$; Fig. 3a; Rasmussen et al., 2014) or Antarctica (typified by WAIS Divide $\delta^{18}$O$_{\text{ice}}$; Fig. 3h; WAIS Divide Project Members, 2013), refuting the dominance of any such teleconnection between these regions and the EAWM during TI. Indeed, the $\delta^{18}$O$_{\text{diatom}}$ record does not show long term trends which parallel any of the
benchmark records shown in Fig. 3 suggesting that, instead, EAWM weakening during TI followed a more gradual trend akin to Northern Hemisphere summer insolation (albeit inversely; Fig. 3; Laskar et al., 2004). The trend in the δ18O_diatom values between 22,000 and 21,000 cal BP inversely aligns with sea surface temperature (SST) records from the Northwest Pacific (Fig. 3i; Oba and Murayama, 2004; Oba et al., 2006), suggesting that EAWM strengthening during this 1000-year interval was due to the intensification of glacial conditions in the region.

Variations in the δ2H_{C30acid} values support an overall strengthening of the EASM in Japan during TI (with lower values in the Holocene than the LGM), however this was non-linear and occurred in three phases: EASM strengthening from the LGM into the Late Glacial Interstade, EASM weakening from the Late Glacial Interstade to the early Holocene, and subsequent EASM strengthening (Fig. 3f). An overall increase of EASM strength in Japan is also sustained by grain size variations within the South China Sea core GIK17940-2 (<63 µm fraction; Fig. 3g; Wang et al., 1999); although the trend at Lake Suigetsu is not entirely equivalent to this because the South China Sea record shows a gradual, rather than tripartite, increase in EASM strength. Our reconstruction indicates that EASM behaviour in Japan was also characteristically different to EASM behaviour in Continental Asia (Chinese cave composite δ18O_speleothem; Fig. 3b; Cheng et al., 2016) and did not show a response to North Atlantic-style stade-interstade fluctuations (Fig. 3a; Rasmussen et al., 2014). However, a close inverse relationship is observed between EASM strength in Japan and Antarctic temperatures (illustrated by WAIS Divide δ18O_ice, Fig. 3h; WAIS Divide Project Members, 2013) between 16,000 and 10,000 cal BP, including coincident timing of inflections at ~14,500 cal BP, ~13,000 cal BP and ~12,000 cal BP (i.e., a response to the ACR). This provides compelling evidence for a teleconnection between the polar southern hemisphere and the EASM during this interval. Prior to this, the δ2H_{C30acid} values were decoupled from WAIS Divide δ18O_ice, suggesting that this teleconnection was only activated at this time. During the LGM, the δ2H_{C30acid} profile was distinctive and showed large amplitude fluctuations which did not occur synchronously with changes in the
global benchmark records, but many of these were also observed in the $\delta^{2}H_{C_{16}acid}$ values. Interpretation of this time period is complex and is presented in the Discussion section below. The $\delta^{2}H_{C_{16}acid}$ values display a composite trend (Fig. 3e) which encapsulates characteristics of both the $\delta^{18}O_{\text{diatom}}$ (Fig. 3d) and $\delta^{2}H_{C_{30}acid}$ (Fig. 3f) records. The influence of each of these unique behaviours appears to dominate the $\delta^{2}H_{C_{16}acid}$ values at different times during the interval: similarities between the $\delta^{2}H_{C_{16}acid}$ and $\delta^{2}H_{C_{30}acid}$ profiles are observed during the early part of the ACR and the Holocene; in the latter stages of the ACR the $\delta^{2}H_{C_{16}acid}$ profile resembles that of $\delta^{18}O_{\text{diatom}}$; and prior to 17,000 cal BP the $\delta^{2}H_{C_{16}acid}$ profile shows mixed behaviour. These observations are consistent with our assertions that the $\delta^{2}H_{C_{16}acid}$ values reflect a combined EAWM and EASM signal, weighted to the season with a greater quantity of precipitation, and these results support a changing balance between EAWM and EASM dominance in Japan during TI. The $\delta^{2}H_{C_{16}acid}$ values show more EAWM character during the later stages of the ACR, when the EAWM was strong and EASM strength was declining, but shows greater EASM affinity during the early ACR and the Holocene, when there was a strengthened EASM and a weakened EAWM. The numerical difference between the $\delta^{2}H_{C_{16}acid}$ and $\delta^{2}H_{C_{30}acid}$ values suggests an increasing EASM influence on Lake Suigetsu through time (Supplementary Fig. 3) which is consistent with our independent observations of EAWM and EASM strength from the $\delta^{18}O_{\text{diatom}}$ and $\delta^{2}H_{C_{30}acid}$ values.
Fig. 3 – EAM evolution at Lake Suigetsu during Termination I with contextual global benchmark records. a) GRIP δ¹⁸O (Johnsen et al., 1997; Rasmussen et al., 2014; remodelled onto U-Th timescale), b) Chinese Cave Composite (Hulu, Dongge, Sanbao) δ¹⁸O (speleothem (Cheng et al., 2016; U-Th timescale), c) Lake Suigetsu MTWA; mean temperature of the warmest month (pollen-derived; Nakagawa et al., 2021; IntCal20 timescale), d) Lake Suigetsu δ¹⁸O < sub>atom</sub>, loess smoothed (span = 0.1) with 1σ confidence bands (this study; IntCal20 timescale), e) δ²H for C₁₆ n-alkanoic acids, loess smoothed (span = 0.1) with 1σ confidence bands (this study; IntCal20 timescale), f) δ²H for C₃₀ n-alkanoic acids, loess smoothed (span = 0.1) with 1σ confidence bands (this study; IntCal20 timescale), g) South China Sea (GIK17940-2) grain size
Discussion

Our results provide novel insights into the evolution of the EAWM and EASM during TI, alongside their driving mechanisms and teleconnections to other regional climate systems. The gradual decrease in EAWM strength and overall increase in EASM strength in Japan strongly support increasing Northern Hemisphere insolation (driven by Milankovitch cycles) as the primary drivers of EAM evolution (Cheng et al., 2016; Mohtadi et al., 2016; Zhang et al., 2019). However, this common orbital driver breaks down as a result of different responses of each seasonal mode to submillennial temperature fluctuations in either hemisphere and it becomes necessary to evoke alternative mechanisms to rationalise these decoupled (rather than inversely related) seasonal behaviours.

EAWM evolution in Japan appears unaffected by stadials in either hemisphere, however other records from Continental Asia (e.g., Lake Huguang Maar (Yancheva et al., 2007)) exhibit significant North Atlantic character (i.e., a BA-YD structure). Despite high quality EAWM records being relatively sparse for this interval, we do not dispute the evidence for a North Atlantic-EAWM teleconnection during TI, and suggest that EAWM wind strength increased (decreased) during stadials (interstades) in the North Atlantic, driven by Northern Hemisphere cooling (warming) and an intensified (weakened) Siberian High as previously suggested (Nakagawa et al., 2006). Indeed, the pattern of winter temperatures at Lake Suigetsu provides evidence for the impact of North Atlantic temperatures on the Eurasian airmass (Nakagawa et al., 2021). However, as a consequence of the
interaction of the EAWM winds with the Sea of Japan, the decrease in EAWM strength here was gradual (Fig. 3d). Specifically, we suggest that any response to North Atlantic stade-interstade fluctuations was obscured from our reconstruction by a smoothed, low amplitude response of winter Sea of Japan SSTs to insolation (Wu et al., 2020) and greater inertia in the oceanic domain compared to the terrestrial. Furthermore, we are able to clarify that the previously evidenced bipartite Late Glacial Stade at Lake Suigetsu (Schlaut et al., 2017) was not associated with any shifts in EAWM precipitation; whilst there is a 300-year period of lower $\delta^{18}$O$_{diatom}$ values within this interval (between 12,284 and 12,093 cal BP; Fig. 3d), this fluctuation is not substantially different from other multi-centennial oscillations seen in the profile. However, it remains possible that there was a bipartite pattern in EAWM wind strength. An Antarctica-EAWM teleconnection was precluded by local EAWM moisture transport, the isolation of the Sea of Japan, and the blocking of Pacific Ocean signals by the monsoon front, all of which limited the influence of interhemispheric teleconnections.

Conversely, evidence from Lake Suigetsu supports a response of the EASM to stadial conditions in the southern hemisphere (specifically the ACR) post-16,000 cal BP (Fig. 3f). Previous studies which have noted a relationship between EASM strength and Antarctic temperatures have evoked the bipolar seesaw and migration of the ITCZ as a mechanism to explain this connection (Han et al., 2016; Zhang et al., 2016). Whilst this would facilitate an inverse correlation between EASM strength in Japan and the ACR, it would also suggest that equivalent shifts in the position of the ITCZ caused by North Atlantic behaviours could affect EASM strength, and we do not see evidence of such a relationship in our records. Instead, we suggest that the Antarctic-EASM teleconnection during TI resulted from Antarctic cooling (warming), which caused coupled atmospheric-ocean circulation changes that resulted in warming (cooling) in the Western Pacific Warm Pool, enhanced (decreased) evaporation and a strengthened (weakened) Western Pacific Subtropical High, which caused a stronger (weaker) EASM (Wang and Fan, 2005; Xu et al., 2020).
The coexistence of a North Atlantic teleconnection to summer temperature and an Antarctic teleconnection to summer (EASM) precipitation at Lake Suigetsu is reconcilable by noting that temperature is much more likely to exhibit patterns common to the same hemisphere, whereas the relationship of the EASM to Pacific Ocean behaviours facilitates the influence of the polar Southern Hemisphere on precipitation. Indeed, our observations of EASM strength at Lake Suigetsu, combined with the pollen-derived summer temperature record, support a “Southern Hemisphere lead” to stade/interstade fluctuations during TI; changes in EASM strength led changes in summer temperature because the former was driven by Antarctic processes and the latter by North Atlantic behaviours.

Our observations of EASM behaviour in Japan directly contradict those of the Chinese cave composite, which supports an EASM response to North Atlantic stade-interstade fluctuations and not Antarctic ones (Fig. 3b; Cheng et al., 2016). Because all of the caves included in the composite (Hulu, Sanbao and Dongge) and Lake Suigetsu are located south of the monsoon front during the summer (and hence under the influence of the Pacific air mass; Nakagawa et al., 2006), differences cannot be resolved by the blocking of teleconnections by the monsoon front, as previously suggested (Yoshida and Takeuti, 2009), and instead must be determined by a different geoclimatic boundary. One possible mechanism involves the overprinting of Antarctic signals in Continental Asia by the partitioning of precipitation between the EASM and post-EASM stages (jet-transition hypothesis; Zhang et al., 2018). This phenomenon is linked to the prevailing climatic conditions in the North Atlantic by the timing of the pre-summer northwards shift of the westerly jet and would have caused drier (wetter) conditions in central China during stades (interstades) in the North Atlantic For this to not impact on EASM evolution in Japan, the effect of this partitioning must have been spatially restricted to the continent, or limited to latitudes at, or lower than, the southernmost position of the westerly jet (Lake Suigetsu is located to the north of this position and Hulu, Dongge and Sanbao Cave to the south (Nakagawa et al., 2021)). An alternative explanation is that the
speleothem records contain an unresolved temperature-controlled component (demonstrated by their close relationship to the pollen-reconstructed temperature at Lake Suigetsu; both show a remote North Atlantic influence) which obscured the Antarctic signal in favour of a North Atlantic one. Regardless, we posit that the location of Japan at a higher latitude, adjacent to the Pacific Ocean and geographically separated from the rest of East Asia facilitated the preservation of Pacific (and hence Antarctic) signals in our record of EASM behaviour at Lake Suigetsu. Whilst this Antarctica-Japan teleconnection has not been observed in other EASM reconstructions (Wang et al., 1999; Ishiwatari et al., 2009; Hayashi et al., 2010), this is likely due to differences in proxy system and a direct relationship between the isotope composition of precipitation and EAM evolution.

However, prior to ~16,000 cal BP, EASM strength in Japan and Antarctic temperatures were decoupled. We suggest that the establishment of this teleconnection was as a consequence of the northwards repositioning of the westerly jet (relative to the Himalayas), a phenomenon previously used to explain the onset of the Late Glacial Interstade (BA-equivalent; 15,000 cal BP) at Lake Suigetsu (Nakagawa et al., 2021). Not only did this northwards shift of the westerly jet allow warm Pacific air to reach Japan, it also moved the EAM front to the north of Lake Suigetsu during boreal summer (because the position of the westerly jet also modulates the seasonal positioning of the EAM front), having been situated permanently (year-round) to the south during the LGM (Chiang et al., 2015; Nakagawa et al., 2021). This would allow the EASM to propagate to Japan and connect summer precipitation to the aforementioned components of the Pacific atmosphere-oceanic circulation (and hence the southern hemisphere). There were no abrupt shifts in the $\delta^{2}$H$_{C30}$acid values at the onset of the Late Glacial Interstade (15,000 cal BP) to reflect this expansion of EASM operation to include Japan; however, prior to 16,000 cal BP the $\delta^{2}$H$_{C30}$acid values were highly variable and structurally distinctive from WAIS Divide $\delta^{18}$O$_{Ice}$. Hence, we suggest that this Antarctic-Japan teleconnection was established at 16,000 cal BP, facilitated in the 16,000-15,000 cal BP period by intermittent northward migrations of the westerly jet and exposure of Japan to the Pacific airmass.
during summer (captured by our centennial sampling resolution), and then continued post-15,000 cal BP when the EAM front was situated permanently to the north of Lake Suigetsu during summer. It follows that prior to the establishment of this teleconnection, precipitation falling during summer in Japan would not technically be equivalent to the EASM (because it would originate and transit north of the EASM front), and instead would reflect local transport of precipitation under the influence of the continental airmass (possibly following a similar south-east to north-west trajectory due to pressure gradients, but with a significantly reduced transport distance). The highly variable behaviour of the $\delta^{2}H_{C30acid}$ values (and hence the $\delta^{2}H_{C16acid}$ values) between 22,000 and 16,000 cal BP may have been caused by the lack of constant, regional driver of summer precipitation (i.e., the EASM) in contrast to the $\delta^{18}O_{diatom}$ values, which were weighted towards the (enduring) EAWM. These fluctuations in the $\delta^{2}H_{acid}$ values could be driven by a number of factors, including precipitation amount, changing relative humidity, shifts in precipitation source and variable SSTs (particularly between 22,000 and 21,000 cal BP (Oba et al., 2006)). Another possible component of this signal is summer meltwater (which could have caused large negative excursions in the $\delta^{2}H_{C30acid}$ values, such as those observed between ~19,000 and 18,000 cal BP). The significant expansion of the EASM system at 16,000 cal BP has critical implications for other records from the region, particularly those derived from more northerly latitudes, where the influence of the westerly jet migration is more pronounced (Nakagawa et al., 2021). Careful consideration of the location of the monsoon front at this time is vital for robust interpretation of proxy changes and whether they indeed represent EASM evolution or, rather, non-monsoonal summer precipitation.

In conclusion, our findings provide centennial-scale reconstructions which highlight the distinctive seasonal behaviours of the EAM during T1, as well as novel insights into the spatially- and temporally-constrained teleconnections acting on the climate of the EAM region during this interval. Our records are unique in their reconstruction of both seasonal EAM modes in Japan from a single
archive. These results may begin to explain the observed heterogeneity between site-specific
reconstructions of EAM behaviour during deglaciation.

Methods

Core Materials

Materials for this study were subsampled from the SG12 core, which was taken from the
centre of Lake Suigetsu (35° 35' 08"N, 135° 52' 56"E) and spans a total composite depth of 41.8 m
(from the present day to 56,100 cal BP). The core consists of a series of overlapping ~90 cm sections
from four boreholes. SG12 is a direct stratigraphic copy of the SG06 core (Nakagawa et al., 2012)
due to the proximity of the SG12 and SG06 coring locations (the first borehole of each coring
campaign was made within 15 m of each other, and both at a water depth of 34 m). As with the
temporally equivalent sediments from the SG06 core, the SG12 sediment is varved throughout and
represents continuous sedimentation (Schlolaut et al., 2018). Chronological control for the SG12
cores is given by correlation to the SG06 cores via distinct visible marker layers. The SG06 chronology
(Bronk Ramsey et al., 2020; Staff et al., in review) is based on >800 radiocarbon dates for the past
~50 ka BP (Staff et al., 2011; Bronk Ramsey et al., 2020), thin section microscopic varve counts
(Schlolaut et al., 2018) and tephra tie points (McLean et al., 2018); all ages are presented on the
IntCal20 timescale (Reimer et al., 2020).

Subsamples of the SG12 core for this study were taken from 1.2 cm-wide LL-channels
(Nakagawa et al., 2012) from the longitudinally cut core across the composite depth (ver. 27
December 2022) range of 1274.8 – 2119.4 cm (9980 ± 30 cal BP to 22,040 ± 38 cal BP). This time
period was selected to encompass the maximum range of the existing Lake Suigetsu pollen-derived
temperature reconstruction for TI. The LL-channels were cut to produce contiguous (continuous
adjacent) subsamples each containing 100 integrated years of sedimentation (based on the SG06
2012 timescale). Sample ages were subsequently converted to the IntCal20 timescale. A contiguous
An integrated sampling approach was taken to avoid subsampling time periods associated with peaks or troughs of multidecadal cycles, superimposed on the multicentennial and multimillennial scale transitions of interest (Nakagawa et al., 2021). The sampling interval of ~100 years was selected to facilitate the detection of finer resolution variability within the key climatic transitions of the period. The (minimal) core expansion during storage was linearly modelled and accounted for by comparing the difference in event layer depth at the time of subsampling to high resolution photographs taken immediately after core extraction. Event layers (floods, turbidites and tephra) were removed during sampling to prevent skewing of the integrated signal by inclusion of a larger quantity of material from a single event. Particular care was taken to avoid contaminating the $\delta^2$H$_{acid}$ samples with modern organic compounds. Surfaces and instruments were regularly cleaned with ethanol.

Each subsample was split into two, with each half designated for $\delta^{18}$O$_{diatom}$ ($n = 120$) and $\delta^2$H$_{acid}$ ($n = 120$) analysis, respectively. The exposed surfaces of one half (for $\delta^2$H$_{acid}$ analysis) were scraped with a clean instrument prior to splitting to remove the outer sediments containing potential contamination from past handling. Wet weights ranged from 1.8 - 10.3 g (subsamples for $\delta^{18}$O$_{diatom}$ analysis) and 1.1 - 7.1 g (subsamples for $\delta^2$H$_{acid}$ analysis). The largest samples were from the younger end of the section, where the sedimentation rate (and thus depth range equating to ~100 years of integrated time) was high, and the smallest samples were from the older end of the section, where the sedimentation rate was relatively low. All subsamples were frozen and then freeze dried for 24 hours to obtain dry sediment masses before subsequent preparation steps.

$\delta^{18}$O$_{diatom}$ Analysis

Subsamples for diatom oxygen isotope analysis were prepared following standard procedures (Swann and Snelling, 2023). Specifically, subsamples were disaggregated overnight in 30 % H$_2$O$_2$ prior to three successive heavy liquid density separations using sodium polytungstate (SPT) at specific gravities of 2.25 g cm$^{-3}$, 2.20 g cm$^{-3}$ and 2.15 g cm$^{-3}$, removing the denser material.
each time. The subsamples were subsequently treated with 30 % H₂O₂ at 70 °C for one week to remove any remaining organic material, followed by 5 % HCl for 12 hours to remove carbonates. Subsamples were then filtered using an 80 µm nylon mesh to remove large sponge spicules and large diatoms and freeze dried ready for analysis. As with other studies, further taxa-specific separation was not performed beyond this sieve stage due to the similar size of many diatom taxa present in the samples, rendering physical separation techniques impossible. Qualitative visual purity checks were made using light microscopy at x800 magnification. The %Al₂O₃ for each subsample was quantified using a Malvern Panalytical Epsilon 3 X-Ray Fluorescence (XRF) spectrometer. Due to elevated %Al₂O₃ values in the majority of samples, a Zeiss Sigma 300 VP-FEG Scanning Electron Microscope (SEM) was used to visually identify the source of this aluminium in five samples of varying %Al₂O₃ (Fig. 4). Samples with %Al₂O₃ exceeding 4.2 % (Al/Si = 0.05) showed particulate Al-bearing contamination (most likely a detrital Al-rich silicate, but undetermined due to grain size) which was external to the diatom frustules. These samples (n = 3) were subjected to an additional SPT density separation at 2.15 g cm⁻³ to remove this contamination. Samples with %Al₂O₃ below 4.2 % showed only small amounts (on a sub-micron scale) of this Al-bearing contamination adhered to or trapped within the diatom frustules (Fig. 4). Additional density separations would not remove this contamination due to electrostatic charges adhering the contamination to the frustule, and air pockets within the cylindrical frustules preventing chemical treatment of internal contamination (Brewer et al., 2008), so a mass balancing approach was taken to correct for the effect of this contamination on the δ¹⁸O_diatom values (discussed below).
Fig. 4: Diatom SEM Imagery. SEM images for subsample 3D (left) and sample 65D (right) showing Al-bearing contamination adhered to frustules and inside cylindrical *Aulacoseira* spp., which are representative of the 120 analysed subsamples for $\delta^{18}O_{\text{diatom}}$ analysis. Scale bars are indicated on each image.

The purified diatom material (6 – 7 mg subsamples) was subjected to stepwise fluorination in nickel reaction vessels to liberate structural oxygen following Leng and Sloane (2008) at the British Geological Survey (UK). Briefly, this method involved outgassing samples at 250 °C to remove any water in the sample, “pre-fluorination” at 250 °C using a stoichiometric deficiency of BrF$_5$ to remove the exchangeable hydrous silica layer from the diatom frustules and minimise exchange effects, and then fluorination with BrF$_5$ overnight at 500 °C. The resulting oxygen gas was cryogenically purified and then converted to CO$_2$ by exposure to hot platinized graphite (Clayton and Mayeda, 1963). The average gas yield of this process was between 68 and 75 %. In each batch of diatom samples, an in-house laboratory diatomite standard, BFC, with a known value of $\delta^{18}$O (Chapligin et al., 2011) was also reacted and converted to CO$_2$ in the same manner following the principle of identical treatment. The oxygen isotope ratio of the diatom subsamples ($\delta^{18}O_{\text{diatom}}$) and BFC ($\delta^{18}O_{\text{BFC}}$) were measured
relative to a reference CO$_2$ gas on a Thermo Finnigan MAT 253 dual inlet isotope ratio mass spectrometer. Oxygen isotope ratios were calibrated using a single-point anchoring procedure by analysing CO$_2$ prepared from an in-house standard (MCS; $\delta^{13}$C$_{VPDB} = -0.7$‰ and $\delta^{18}$O$_{VPDB} = -9.2$‰) via reaction with H$_3$PO$_4$ (specific gravity = 1.92) at 25°C. All $\delta^{18}$O$_{VPDB}$ values were converted to the VSMOW scale using Coplen et al. (1983) and normalised such that the $\delta^{18}$O$_{VSMOW}$ of the within-run BFC = +28.9‰. The oxygen isotope composition of BFC has been calibrated via inter-laboratory comparison (Chapligin et al., 2011) using the certified international standard NBS 28 ($\delta^{18}$O$_{VSMOW} = +9.57$‰; Hut, 1987; Gröning et al., 2007). All oxygen isotope data are expressed in standard delta ($\delta$) notation in per mille (‰) deviations relative to VSMOW. External analytical error (1σ) was ±0.13‰ for BFC and ±0.33‰ for replicate analysis of the diatom samples.

The $\delta^{18}$O$_{diatom}$ values were corrected for the aforementioned Al-bearing silicate contamination using a geochemical mass balance approach (Mackay et al., 2011; Swann et al., 2018). End member contamination samples for Lake Suigetsu contained 19.45% Al$_2$O$_3$ (from XRF, n = 2), a $\delta^{18}$O composition of 10.36 ± 0.68‰ (1σ range, n = 8) and 48.93 wt% oxygen (from XRF, n = 2). These were prepared from raw materials (n = 2) and the residues of SPT separations from the $\delta^{18}$O$_{diatom}$ subsamples (n = 6), disaggregated overnight in 30% H$_2$O$_2$ and treated with 30% H$_2$O$_2$ at 70°C for one week to remove any remaining organic material, followed by 5% HCl for 12 hours to remove carbonates and 8% NaOH at 70°C for 48 hours to remove biogenic silica. Contamination-corrected $\delta^{18}$O$_{diatom}$ values were modelled using the %Al$_2$O$_3$ for each sample, the end member quantities above, and assuming a %Al$_2$O$_3$ of pure diatom material of 1.4% (Al/Si = 0.016), the minimum value measured across the 120 samples. The uncertainty associated with mass-balancing was calculated assuming a normal distribution for data uncertainty and Monte Carlo simulations (10,000 replicates). Errors presented on Fig. 2 and Fig. 3 show the 1σ range of these simulations (mean = ± 0.36‰). Of the 120 samples prepared, 113 datapoints are included in the final dataset. Five samples were excluded due to elevated %Al$_2$O$_3$ values (>4.2%; Al/Si = 0.05); these produced anomalously high
values during mass balancing, possibly due to atypically large quantities of diatom-bound Al which was unaccounted for by our methods. A further two samples were too small in mass for accurate $\delta^{18}$O$_{\text{diatom}}$ determination via the methods presented above.

**$\delta^2$H$_{\text{acid}}$ Analysis**

Subsamples for $\delta^2$H$_{\text{acid}}$ analysis were prepared following the Biomarkers for Environmental and Climate Science (BECs) group standard protocol at the University of Glasgow (UK). The total lipid extract (TLE) was extracted by a Dionex ASE 350 Accelerated Solvent Extractor using dichloromethane and methanol (9:1, v:v) and the TLE dried and weighed. The TLE was then separated into neutral and acid fractions using solid phase extraction through a LC-NH$_2$ silica gel column. The neutral fraction was eluted using dichloromethane:propan-2-ol solution (1:1, v:v) and the acid fraction using 4 % acetic acid in diethyl ether. The acid fraction was derivatised using 100 µL of 12 % boron trifluoride in methanol in sealed glass vials at 70 °C for 60 minutes. This converted n-alkanoic acids into fatty acid methyl esters (FAMEs) in preparation for analysis. These FAMEs were then cleaned through a silica gel column (35 – 75 µm particle size). Non-FAME material was eluted using hexane and the FAMEs eluted using dichloromethane. Subsamples were prepared in batches of 10 or 11, each with a procedural blank to check for contamination introduced by the stages described above.

Identification of the C$_{16}$, C$_{18}$, C$_{28}$ and C$_{30}$ n-alkanoic acids (as FAMEs) was performed on subsample 96B using an Agilent 7890B Gas Chromatograph connected to a 5977A mass spectrometer detector (MSD) with an electron impact (EI) ionisation source (GC-MS). A HP1-MS capillary column was used (60 m length, 0.25 mm internal diameter, 0.25 µm film thickness) with helium carrier gas at a constant flow rate (1.2 mL min$^{-1}$). The FAMEs were dissolved in DCM (200 µL) and a volume of 1 µL used in a splitless injection. The inlet was kept at 315 °C and the following oven programme used: the oven was held for two minutes at 60 °C and raised to 120 °C at 30 °C min$^{-1}$,
then increased to 310 °C at 5 °C min\(^{-1}\) and held for 33 minutes. The auxiliary temperature in the transfer line between the GC and MSD was kept constant at 315 °C. The MS source was set to 230 and the MS quad set to 160. The detector analysed all compounds between 12- and 65-minute retention time, except for the period between 34 and 39 minutes where flow was redirected to avoid a large phthalate compound peak from entering the detector. Data acquisition was performed in full scan mode at 2.4 scans s\(^{-1}\). Identification of contaminant peaks was also performed using this method. All large contaminant peaks (retention times: ~24.5, ~33 and ~37 minutes) were shown to be phthalate compounds. The largest of these (at ~33 minutes) was present in > 90 % of subsamples, however the others were only present in ~ 10 %. No contaminant peaks obscured the FAME peaks of interest.

The concentration of the C\(_{16}\), C\(_{18}\), C\(_{28}\) and C\(_{30}\) n-alkanoic acids (as FAMES) in each subsample were measured using an Agilent 7890B Gas Chromatograph fitted with a flame ionisation detector (GC-FID). A Restek Rtx-1 capillary column was used (60 m length, 0.25 mm internal diameter, 0.25 \(\mu\)m film thickness) with helium carrier gas at a constant flow rate (1.2 mL min\(^{-1}\)). FAMES were dissolved in DCM (200 \(\mu\)L - 1250 \(\mu\)L; depending on FAME yield) and a volume of 1 \(\mu\)L used in a splitless injection. The inlet was kept at 320 °C and the following oven programme used: the oven was held for two minutes at 60 °C and raised to 120 °C at 30 °C min\(^{-1}\), then increased to 330 °C at 5 °C min\(^{-1}\) and held for 15 minutes. The FID temperature was set to 250 °C. A chromatogram showing a representative sample with no phthalate peaks is shown in Supplementary Fig. 4. Peak areas were measured, and concentrations calculated from a set of external calibrations using a standard mix of eleven n-alkanes; ten straight-chained (C\(_{15}\), C\(_{18}\), C\(_{19}\), C\(_{23}\), C\(_{25}\), C\(_{26}\), C\(_{28}\), C\(_{30}\), C\(_{32}\) and C\(_{37}\) n-alkanes) and one branched (squalane). Calibration graphs were made using concentrations of 2.5 \(\mu\)g/mL, 5 \(\mu\)g/mL, and 10 \(\mu\)g/mL. A calibration for the 0 - 25 minute, 25 - 35 minute and 35 - 60 minute retention time intervals were made using the C\(_{16}\), C\(_{29}\) and C\(_{39}\) homologues in the standard, respectively. An R-squared value of >0.99 was achieved for each calibration. The n-alkanoic acid (as FAME)
concentrations were then normalised to dry sediment weight. Variations in n-alkanoic acid concentrations are shown in Supplementary Fig. 1.

The compound specific hydrogen isotopic composition (δ^2H) of the C_{16}, C_{18}, C_{28} and C_{30} n-alkanoic acids (as FAMEs) were measured using an Agilent 7890B gas chromatography (GC) system connected to an Elementar GC5 furnace and IRMS. The settings of the GC system were identical to those of the GC-FID outlined above with the difference being that the FAMEs were dissolved in between 30 and 120 µL of hexane and the injection volume set to either 1 or 2 µL depending on FAME yield. This ensured that FAMEs were measured as a similar response to the reference H_2 gas. Due to the presence of the largest phthalate compound (retention time ~33 mins) in >90% of the subsamples, flow was redirected between 1700 s and 1975 s to prevent entry into the IRMS. In 13 subsamples, this was extended to the interval 1400 s to 2100 s to remove the second largest phthalate peak (retention time ~37 minutes). The smallest phthalate peak (retention time ~24.5 minutes) was not larger than the FAME peaks and hence did not require redirection of flow. The furnace temperature was set to 1450 °C with an interface temperature of 350 °C. Subsamples were measured in duplicate and δ^2H values of each homologue were calculated relative to a calibrated reference H_2 gas. These values were then converted to the VSMOW scale by linear regression to an in-house standard (containing C_{16}, C_{19}, C_{23}, C_{25}, C_{28} and C_{32} straight chained n-alkanes in addition to squalane) calibrated to the standard Indiana n-alkane mixture B5 containing C_{16} to C_{30} n-alkanes. The in-house standard was measured four times prior to each batch, after every eight sample measurements, and in duplicate at the end of each batch. All hydrogen isotope data are expressed in standard delta (δ) notation in per mille (‰) deviations relative to VSMOW. Instrument precision was ±2.53 ‰ based on the standard deviation of all standard measurements. The H_3^+ correction factor ranged from 4.24 to 4.57.

Reliable δ^2H values for all chain lengths across all subsamples was not achieved due to some subsamples exhibiting raised baselines resulting in poorly defined peak areas and inaccurate
determination of $\delta^2$H for at least one homologue (affecting 26 subsamples). Raised baselines in some samples predominantly precluded accurate measurement of the aquatic chain lengths, though it was possible to accurately measure the C$_{28}$ and C$_{30}$ homologues in many of the impacted subsamples. Low concentrations affected the repeatability of the measurements of $\delta^2$H$_{\text{acid}}$ of at least one homologue in a further 9 subsamples (most commonly C$_{16}$ or C$_{18}$), hence these values were excluded from the final dataset. The mean precision of the $\delta^2$H$_{\text{acid}}$ measurements was ± 2.0 ‰ (1σ range). The resulting dataset has 95 datapoints for the C$_{16}$, 86 for the C$_{18}$, 112 for the C$_{28}$ and 108 for the C$_{30}$ n-alkanoic acids. Terrestrial and aquatic composites were not generated due to insufficiently strong statistical relationships between the $\delta^2$H$_{\text{C}{28}\text{acid}}$ and $\delta^2$H$_{\text{C}{30}\text{acid}}$ values, and the $\delta^2$H$_{\text{C}{16}\text{acid}}$ and $\delta^2$H$_{\text{C}{18}\text{acid}}$ values, although there was some visual correlation between the two pairs.

Values were corrected for the methylation process (which removes the exchangeable carboxylic acid hydrogen from each molecule and replaces it with a methyl group with three hydrogens) (Chivall et al., 2012). GC-IRMS measurements of (Z)-hexadec-9-enoic acid ($\delta^2$H = -154.02 ‰) and methyl (Z)-hexadec-9-enoate ($\delta^2$H = -143.13 ‰) were made to calculate the $\delta^2$H value of a single methanol-derived methyl hydrogen ($\delta^2$H = -37.86 ‰), and a mass balance scheme (Chivall et al., 2012) applied to correct the measured FAME $\delta^2$H values to n-alkanoic acid values (excluding the exchangeable hydrogen on the carboxylic acid group).

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**Data Availability**

Accompanying data can be found at [http://doi.org/10.5525/gla.researchdata.1439](http://doi.org/10.5525/gla.researchdata.1439).

**Supplementary Information**

**Supplementary Discussion 1**

The total lipid extract (TLE) exhibited low values in the early part of the section (~2 mg/g) before increasing gradually from ~15,500 cal BP (Supplementary Fig. 1). Values were then high (~6 mg/g) and fluctuating. This follows the pattern of sediment total organic carbon (TOC) during this period (Tyler *et al.*, 2010). Aquatic n-alkanoic acid concentrations did not show this long-term increasing trend, and the concentration profiles for the C<sub>16</sub> and C<sub>18</sub> n-alkanoic acids showed centennial-scale fluctuations throughout (Supplementary Fig. 1). Conversely, the terrestrial n-alkanoic acid concentrations were higher in the latter part of the section; the concentration of the C<sub>28</sub> n-alkanoic acid showed a greater increase than the C<sub>30</sub>, but also exhibited larger fluctuations (Supplementary Fig. 1). Throughout, the terrestrial n-alkanoic acid concentrations were generally higher than those of the aquatic n-alkanoic acids, aside from in a small number of samples. The close relationship between the terrestrial n-alkanoic acid concentrations and the TLE concentrations suggests that the terrestrial component was driving the observed TLE increase (allowing for the un-resolved non-acid fractions which also contribute but were not measured). The gradual increase in TLE, TOC and terrestrial n-alkanoic acid concentration was similar in character to the changes in EAWM and EASM strength and very different from the temperature reconstruction for the site (Main Text), suggesting that they were driven by precipitation changes rather than temperature changes. None of these variables show ACR-type inversions, however the seasonality of these variables suggests a relationship to the EASM is more likely than a link to the EAWM, despite the EAWM showing more gradual changes during this interval. Diatom frustule concentration also increased through the section (Supplementary Fig. 1), but the shift to higher values initiated ~500
cal BP later (~15,000 cal BP) and more abruptly than the TLE concentration. Hence, whilst an overall resemblance between frustule concentration and TOC can be argued, there was some decoupling of these variables between ~16,000 and ~15,000 cal BP. The sudden increase in diatom frustule concentration at ~15,000 cal BP was likely driven by temperature changes, in light of the abruptness of this shift and the coincident timing with the onset of the Late Glacial Interstadial (Main Text). *Aulacoseira* dominated the diatom assemblage counts, most notably at 21,500 cal BP, 20,000 cal BP and between 15,000 and 12,000 cal BP (Supplementary Fig. 1). Spikes of *Naviculoid* and benthic taxa were observed at 21,000 cal BP. Between 20,000 and 15,000 cal BP there were greater proportions of *Stephanodiscus*, *Naviculoid* and benthic taxa relative to *Aulacoseira*, and similarly a period of increased *Asterionella* between 12,000 and 10,000 cal BP.
Supplementary Figure 1

**Proxy weights, abundances and concentrations.** Variations in total organic carbon (TOC; Tyler et al., 2010), mass-normalised total lipid extract (TLE), mass normalised n-alkanoic acid concentrations (aquatic (C$_{16}$ and C$_{18}$) and terrestrial (C$_{28}$ and C$_{30}$)), diatom frustule concentration (Saito-Kato et al., unpub) and diatom taxa abundance (Saito-Kato et al., unpub) during TI in the Lake Suigetsu sediment cores.
Comparison of Lake Suigetsu δ\textsuperscript{18}O\textsubscript{diatom} and Lake Suigetsu diatom taxonomic counts. Diatom taxonomic counts show varying abundances of seasonally assigned genera (Aulacoseira and Encyonema, a benthic species) versus other taxa. Intervals associated with high %Aulacoseira are interpreted as a proxy of EAWM strength. The data presented here supports EAWM weakening during Termination I.

Supplementary Discussion 2

By considering the numerical difference between the δ\textsuperscript{2}H\textsubscript{C\textsubscript{30}acid} and δ\textsuperscript{2}H\textsubscript{C\textsubscript{16}acid} values (Supplementary Fig. 3), it is possible to semi-quantify the relative influence of the EAWM and EASM on the δ\textsuperscript{2}H\textsubscript{C\textsubscript{16}acid} values (which reflects a combination of the δ\textsuperscript{2}H\textsubscript{precipitation} of both seasons; Main Text). Whilst traditionally used to consider the effect of evaporation on the aquatic δ\textsuperscript{2}H signal (because the terrestrial δ\textsuperscript{2}H is not affected by lake evaporation, unlike aquatic δ\textsuperscript{2}H), under the circumstances presented here (and expanded upon in the main text), we interpret this measure to represent the balance between EAWM and EASM influence, based on the inference that the δ\textsuperscript{2}H\textsubscript{C\textsubscript{16}acid} values represent a seasonal mix, and the δ\textsuperscript{2}H\textsubscript{C\textsubscript{30}acid} values as seasonally constrained to summer. Following this logic, the numerical difference between these measures indicates increasing EASM influence on
aquatic $\delta^2$H during TI (Supplementary Fig. 3); larger values indicate greater EASM influence, and smaller values indicate greater EAWM influence (based on the assumption that EAWM $\delta^2$H_{precipitation} is higher than EASM $\delta^2$H_{precipitation}). However, it is inappropriate to apply this interpretation to the ~19,000 – 18,000 cal BP interval, where there was a significant amount of scatter in the $\delta^2$H_{C30acid} values (discussed in the main text). A reduction in lag time with warming temperatures remains a component of this signal.

**Supplementary Figure 3**

*Changing EASM influence on $\delta^2$H$_{acid}$ at Lake Suigetsu.* Comparison of Lake Suigetsu $\delta^2$H$_{C16acid}$, $\delta^2$H$_{C30acid}$, and the numerical difference between these quantities during TI in the Lake Suigetsu sediment cores. $\delta^2$H$_{C30acid} - \delta^2$H$_{C16acid}$ is interpreted as an indicator of EASM versus EAWM influence, with a component of a reduced lag time with warming temperatures.
Supplementary Figure 4

Chromatogram of the n-alkanoic acid fraction of Sample 6B. GC-FID Chromatogram of a representative uncontaminated sample showing clear, symmetrical peaks of interest with a flat chromatogram baseline, strong even-over-odd preference, and dominant terrestrial chain lengths. All n-alkanoic acids were measured as FAMEs.

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