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TITLE PAGE

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Exact dating of the Meghalayan lower boundary based on high-latitude tree-ring isotope chronology

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Abstract

Subdivision of the Holocene Series/Epoch into the Greenlandian, Northgrippian and Meghalayan Stages/Ages has recently been ratified based on stable isotope records from ice-core and speleothem archives. The base of the most recent chronostratigraphic unit, corresponding to the Northgrippian−Meghalayan boundary, coincides with the '4.2 ka event' recognised as a low-latitude drought anomaly. The Global Stratotype Section and Point (GSSP) for the boundary, the oxygen isotope record from the Mawmluh Cave speleothem (India), demonstrates this event with markedly weakened Asian summer monsoon. Here we contribute to the assessments of the geological time scale and the global characteristics of this event by detailing an isotopic excursion in tree-ring carbon isotopes from high-latitude/subarctic Europe. The $\delta^{13}C$ chronology demonstrates extremely overcast (wet) conditions, especially between 2190 and 2100 BCE, with anomalous conditions sustaining until 1990 BCE. In addition to demonstrating its exact dating and duration, the $\delta^{13}C$ data also illustrate the two-stage nature of the event and highlight the greater magnitude of the earlier stage. This reinforces the characterisation of this 'Meghalayan anomaly' in the context of other proxy-sites around the world. The North Atlantic forcing, previously associated with weakened Asian summer monsoon, accords with the suggested roles of North Atlantic Oscillation and/or southwards shift of Inter-Tropical Convergence Zone in producing the hydroclimatic anomaly. In fact, it would be clarifying to separate the hydroclimatic 'Meghalayan anomaly' from the broadly cited and potentially longer lasting '4.2 ka event', to analyse their respective forcing mechanisms. We conclude by dating the Meghalayan lower boundary to 2190 BCE.

Keywords: Holocene; Paleoclimatology; Monsoon; Palaeoclimatology; Global; Scandinavia; Speleothems; Stable isotopes; Tree-rings

1. Introduction

Proposal for the subdivision of the Holocene Series/Epoch into the Greenlandian, Northgrippian and Meghalayan Stages/Ages, initiated by the Subcommission on Quaternary Stratigraphy and approved by the International Commission on Stratigraphy, has now been formally ratified by the Executive Committee of the International Union of Geological Sciences (Walker et al., 2018, 2019). The decision follows the recent ratifications of the Quaternary System/Period and the Pleistocene Series/Epoch, of which base is dated at 2.58 Ma (Gibbard et al., 2010), and in particular that of the Holocene Series/Epoch with its lower boundary formally dated at 11.7 ka as recorded in the Greenland ice (Walker et al., 2008, 2009). Being the most recent interval of Earth history, the Holocene extends to and includes the present day and there is an extensive amount of evidence, especially of fluctuating climate conditions, to characterise geologic-climatic units based on meticulously studied proxy indications. Rather than strictly lithological units, the new subdivision is, of necessity, indeed largely based on this (palaeoclimatic) evidence.

The tripartite division of the Holocene corresponds with the Lower (Early), Middle and Upper (Late) subepochs that have frequently been employed in the Quaternary palaeoclimate literature (Walker et al., 2012, 2018). Ratification of these stratigraphic units on the geologic time scale locates the boundary between the Greenlandian and Northgrippian at 8.2 ka and that between the Northgrippian and Meghalayan at 4.2 ka (Walker et al., 2018, 2019), each of these ages thus spanning successive roughly four thousand year intervals. With these regards, the boundaries are defined based on the palaeoclimate proxy evidence for the '8.2 ka event' (Alley et al., 1997; Alley and Ágústdóttir, 2005; Rohling and Pälike, 2005; Thomas et al., 2011) and '4.2 ka event' (Staubwasser et al., 2003; Booth et al., 2005; Liu and Feng, 2012; Zanchetta et al., 2016; Kathayat et al., 2018; Railsback et al., 2018; Bini et al., 2019). These events were generally governed by cold

climates but, importantly, they are also marked by prolonged droughts at low latitudes (Mayewski et al., 2004; Alley and Ágústdóttir, 2005; Wanner et al., 2011, 2015). This means that their signals are widespread across the sites sensitive not only to temperature but drought and moisture anomalies. Apart from palaeoclimatic evidence, the Meghalayan lower boundary, primarily indicated by the abrupt low-latitude aridification around 4.2 ka, is expected to resonate with archaeological and historical evidence of coinciding societal transformations from various low-latitude sites (Walker et al., 2012, 2018). Here we concentrate purely on the natural science evidence relevant to the Northgrippian–Meghalayan boundary. Recently, arguments against the criteria in support of the Meghalayan Stage/Age were indeed presented on palaeoclimatic grounds, with doubts on any safely dated global climate anomaly (Voosen, 2018). In fact, the Meghalayan lower boundary may not be hallmarked by such dramatic climatic extremes as was the transition from the last ice age to our current warm epoch, the Holocene (Walker et al., 2008, 2009). Climatic changes were more extreme also during the widespread cooling episode at 8.2 ka, when the catastrophic meltwater release from North American glacial lakes into the North Atlantic suppressed the deep-water formation (Alley et al., 1997; Clarke et al., 2004; Hoffman et al., 2012).

In any case, there is multiple overlapping palaeoclimate evidence to indicate megadroughts, especially at low latitudes around 4.2 to 3.9 ka. This evidence is so pronounced in the isotope data of the Mawmluh Cave speleothem (Berkelhammer et al., 2012), unearthed from the northeast Indian State of Meghalaya, that it was selected to serve as the GSSP for the new Meghalayan Stage/Age (Walker et al., 2018). The remaining worries about the Meghalayan lower boundary concern its accurate and precise dating together with its' putative global spread (Voosen, 2018). Additional well-dated evidence, reliably calibrated to a certain climatic variable from regions other than low latitudes, could disperse some of the issues still surfacing. Fortunately, such tree-ring stable carbon isotope ($\delta^{13}C$) data have been newly published from high-latitude continental North-West Europe, including Finnish Lapland (Helama et al., 2018a; Table S1) but any potential signature of the 4.2 ka

event has not yet been analysed from this $\delta^{13}C$ data. Here, we show a strong $\delta^{13}C$ excursion in this data that is correlative to the newly selected GSSP in the Mawmluh Cave speleothem. First, the exact dating of tree-ring materials, using dendrochronological methods (Fritts, 1976; Schweingruber, 1988; Speer, 2010), allows us to reliably date this excursion and to make inferences on the assessment of the Holocene chronostratigraphy. Second, the $\delta^{13}C$ record runs continuously and is consistently sampled using decadal blocks since 7.5 ka (Helama et al., 2018a), which makes it possible for us to analyse and characterise the particular isotopic event in this chronology similar to ice-core and speleothem records for which the Holocene event-stratigraphic approach has been applied (Walker et al., 1999). Third, the successful calibration and verification of this $\delta^{13}C$ dataset against the instrumental meteorological records have demonstrated its value for interpreting the climate variability in the North Atlantic sector (Helama et al., 2018a, 2018b). Guided by these previous findings, we use the $\delta^{13}C$ record as a basis for Holocene climatostratigraphy for regions sensitive to North Atlantic forcing mechanisms that are global in their reach and effect.

2. Characterising the Meghalayan event

The tree-ring isotope record runs from 5500 BCE until the present and thus covers almost entirely the Northgrippian and Meghalayan Stages/Ages (Fig. 1a). Transformed into estimates of past cloudiness, the stable carbon isotope data explains more than 80 percent of the variance in the instrumentally observed summertime (June to August) cloud cover between 1890 and 2010 CE (Helama et al., 2018a). We found that this $\delta^{13}C$-based data is unequivocally indicative of an hydroclimate anomaly between 2190 and 2100 BCE, with prolonged, but less anomalous, conditions sustained until 1990 BCE (Fig. 1b). This anomaly represents the cloudiest century-long period 2200-2100 BCE in the whole reconstruction and is consistent by its dating with the period of the markedly weakened monsoon recorded in the GSSP section in the Mawmluh Cave speleothem (Berkelhammer et al., 2012) showing a long-term $\delta^{18}O$ anomaly between 4.303 and 3.888 ka (before 1950 CE)

(Walker et al., 2018, 2019) (Fig. 1c). Compared to the tree-ring dated $\delta^{13}C$ proxy, the age control of the speleothem is based on the U-Th dating of the stalagmite, the most intense period of weakened monsoon occurring from 4.071 to 3.888 ka, with a potential uncertainty of U-Th dates of 30 and 20 years, respectively (Walker et al., 2018). Relative to the b2k convention, these dates translate into a range of 4151−4091 and 3968−3918 years before 2000 CE. The duration of approximately two centuries for both the $\delta^{18}O$ and $\delta^{13}C$ excursions provides additional consistency to our results. Moreover, both records are indicative of a strong isotopic excursion taking less than four to five decades to peak. In the tree-ring $\delta^{13}C$ record, this initiation phase occurred from 2210 to 2170 BCE. In its culmination phase, the reconstruction demonstrated increased cloud cover until 2100 BCE, with a tri-decadal phase of abruptly reduced cloud cover between 2100 and 2070 BCE. There appears yet another more short-lived phase of markedly increased cloudiness culminating at 2030-2020 BCE.

From a multi-site perspective, these characteristics closely mimic the bracketing dates of 4.15 to 3.93 ka (hence, 2200−1980 BCE), the bi-centennial duration, the two-stage nature and the greater magnitude of the earlier stage, previously suggested for the 4.2 ka event for various proxy types and localities around the world (Railsback et al., 2018) (Fig. 2a, b). In the context of the NW European climate, our tree-ring $\delta^{13}C$ results augment the multi-proxy findings of broadly coeval event of relatively wet conditions interpreted from considerably less well-dated, lower-resolution palaeoenvironmental records from the British Isles (Hughes et al., 2000; Barber et al., 2003; Jordan et al., 2017), the Jura Mountains (Magny, 2004) and south-central Sweden (Borgmark, 2005).

3. Chronostratigraphy *versus* climatostratigraphy

This analysis contributes to discussions on both the dating and spread of the hydroclimatic 4.2 ka event at the advent of the Meghalayan Stage. Moreover, the approach follows the procedure of

event stratigraphy (Björck et al., 1998; Walker et al., 1999) applied to recent geological record (here, subfossil tree rings) characterised by climatic shifts of considerable magnitude. Such events may be geologically short-lived occurrences (e.g. volcanic eruptions, earthquakes, sea-level changes or climatic events) which have left traceable signals in rock record and may be used for stratigraphic correlations (Whittaker et al., 1991). In this respect, the climate-based classification used here, i.e. the climatostratigraphic approach, is a type of event stratigraphy (Walker et al., 1999). Ideally, the climatostratigraphic boundaries should be placed at the climate change to equate them to a global scale; practically, the boundaries can be placed at midpoints of change i.e. between the maxima and minima as recorded in the data (Gibbard and Lewin, 2016; Gibbard, 2017). In fact, the midpoint between the dates of 4.3 ka and 4.1 ka, corresponding to the first and second shifts in the speleothem $\delta^{18}O$ data, is now formally used as the Meghalayan lower boundary at 4.2 ka (Walker et al., 2018). Fortunately, this date is in close agreement with the date of 2190 BCE at the exact timing of the observed change in hydroclimate interpreted from the high-latitude tree-ring carbon isotopes

In general, climatostratigraphic units should not be used interchangeably with chronostratigraphic units as the former may be time-transgressive in nature (Gibbard and Lewin, 2016; Gibbard, 2017): the climate changes themselves may exhibit diachroneity and/or the recorded events may be recognized only through the responses the climate changes initiate in depositional or biological systems, potentially with delay. While the diachroneity issue has been recorded for some other Holocene climate phases (e.g. Renssen et al., 2009), the coincidence between the isotopic excursions in $\delta^{18}O$ and $\delta^{13}C$ records from different proxy types, climate zones and widely separated sites is here well demonstrated. As for the delays in depositional or biological systems, such inertia were recently found for the first millennium CE proxy-records as the environmental indications (changes in glacial, aeolian, coastal and soil processes and in forested and aquatic ecosystems) appeared to lag approximately 100 years behind those recorded in temperature, precipitation and other climate proxy data (Helama et al., 2017a). No such lags are present in our tree-ring $\delta^{13}C$ data showing

immediate response to instrumentally observed climate variability over the 1890−2010 CE period (Helama et al., 2018a). This sensitivity, combined with accuracy and precision of tree-ring dating (Fritts, 1976; Schweingruber, 1988; Speer, 2010), emphasises the value of this $\delta^{13}C$ tree-ring proxy not only as climatostratigraphic, but also the chronostratigraphic record to express the hydroclimatic event and to pinpoint the Meghalayan Stage/Age lower boundary.

4. Global-scale climatic driving mechanisms

Locating downwind to the North Atlantic and proximal to Arctic suggests (Young et al., 2010, 2012; Loader et al., 2013; Helama et al., 2018a, 2018b) an interaction of large-scale atmospheric associations and respective climatic mechanisms to have played a role behind the recorded cloudiness/hydroclimatic event. The North Atlantic Oscillation (NAO) is the leading mode of atmospheric variability in the North Atlantic region modifying the flow of moist westerlies across the middle latitudes of the Atlantic onto Europe (Hurrell and Deser, 2010). During the summer season, the negative NAO phase generally corresponds to atmospheric configuration responsible for the known wet north−dry south anomaly pattern (Folland et al., 2009). Increases in cloud cover in northern Europe, and thus most likely in precipitation, as we observed between 2190 and 1990 BCE, are both consistent with conditions corresponding to this NAO phase. This configuration could also explain the indications of summer drought in the Mediterranean proxy records broadly over the same interval (Bini et al. 2019), but the seasonal influences of other circulation patterns affecting the spatiotemporal climate variability in the region (e.g. Krichak and Alpert, 2005; Comas-Bru and McDermott, 2014; Cook et al., 2016) may need to be critically evaluated. We note that the initial considerations of Meghalayan Stage (Walker et al., 2012, 2018) have mainly focused on the mechanisms tending to weaken the Asian summer monsoon, in combination with drought reflecting effects from the El Niño–Southern Oscillation (ENSO). In fact, more active El Niño events are known to inhibit the monsoon (Morrill et al., 2003) and lead to conditions prone to less negative $\delta^{18}O$ values

in the GSSP section, the Mawmluh Cave speleothem (Berkelhammer et al., 2012). They may well also have produced signals of coinciding droughts as shown also in other selected low-latitude records (Walker et al., 2012).

Moreover, the North Atlantic climate and monsoon variability have potentially been tied during the Holocene (Gupta et al., 2003; Wang et al., 2005; Dixit et al., 2014b; Kathayat et al., 2017) and even during the Younger Dryas Stadial and Heinrich Interstadials (Mohtadi et al., 2014). In these cases the intervals of cold North Atlantic sea surface temperatures (NASST) correlate with those of reduced monsoon rainfall (Mohtadi et al., 2016). Climate modelling suggests a link between the North Atlantic and monsoon through the NAO, in connection with the cold NASST decreasing meridional tropospheric temperature gradient over Eurasia, hence producing a decrease in monsoon rainfall (Goswami et al., 2006). This line of evidence could also explain why the intervals of released iceberg discharges into the North Atlantic, determined from millennial scale changes in sedimentary ice-rafted detritus (IRD) and indicating disruptions in the Atlantic meridional overturning circulation in connection to periods of low solar forcing (Bond et al., 1997, 2001) (Fig. 2c, d), appear to correlate with intervals of reduced monsoon rainfall. In fact, the IRD occurrences (Bond et al., 1997, 2001) and weak Asian monsoon (Dixit et al., 2014a, 2014b; Kathayat et al., 2017) recur to coincide at both the Greenlandian–Northgrippian and Northgrippian–Meghalayan Stage/Age boundaries. The North Atlantic forcing of cold climate could also have led to southward shift of the Northern Hemisphere summer position of the Inter-Tropical Convergence Zone resulting in weaker Asian monsoon system (Liu and Feng, 2012), in accordance with the “cool poles, dry tropics” configuration (Mayewski et al., 2004). There are also other ways to explain the ocean-atmosphere linkage between the NASST and Asian monsoon intensity, the robustness of the connections remaining however difficult to assess with the current set of high-resolution proxy records being far too still scarce to conclusively evaluate the suggested linkages (Mohtadi et al., 2016).

Apart from this hydroclimatic change, we highlight a shift indicative of summer cooling around the $\delta^{13}C$ sites in northern Finnish Lapland, where a sudden inability of high-altitude (here, over 150 m above modern sea level) trees to regenerate is indicated by decreased supply of pine subfossils (Helama et al., 2004) (Fig. 2e). The temporal relationship of this change to the near-synchronous evidence of glacier advances across a selection of Northern Hemispheric sites (Bakke et al. 2010), discussed also in the context of the 'Holocene Turnover' (Paasche et al., 2004; Paasche and Bakke, 2009), suggests that they are climatically linked. This finding, however, warrants further research and investigation.

5. Discussion

Previous proxy comparisons have implied that the anomalies observed at 4.2 ka may have occurred over a longer-term hydroclimatic and/or aeolian change (Drysdale et al., 2006; Axford et al., 2013). These findings would mimic those observed for the 8.2 ka event suggested to punctuate a longer and less extreme period of perturbed climate (Alley and Ágústdóttir, 2005; Rohling and Pälike, 2005). All these signals are likely superimposed within the palaeoclimatic archive but the slow and fast anomalies should not be mistakenly confused. The GSSP in the Mawmluh Cave speleothem stands primarily for a hydroclimatic anomaly and a general linkage of weakened monsoon to North Atlantic climate forcing has been proposed (Gupta et al., 2003; Wang et al., 2005; Dixit et al., 2014b; Mohtadi et al., 2014; Kathayat et al., 2017). It remains inconclusive as to whether this connection applies to correspondingly slow or fast anomalies, or both. For these reasons, it would also be more clarifying to cite the '4.2 ka event' and 'Meghalayan anomaly' separately, restricting the latter to the approx. bi-centennial two-stage hydroclimatic extreme (Railsback et al., 2018) reliably dated to last centuries of the 3rd millennium BCE (this study), the onset of this anomaly thus standing for the lower boundary of the Meghalayan Stage/Age. This separation could not only help to differentiate their climatic signals but also identify the respective forcing mechanisms.

Considering the need to focus on climate records based on well-dated chronologies when defining new climatic events (Helama et al., 2017a, 2017b), the isotope record from Finnish Lapland is an ideal time marker as it is reliably converted to calendric timeline using the established dendrochronological methods (Fritts, 1976; Schweingruber, 1988; Speer, 2010). Moreover, the reconstruction of cloud cover variability based on the tree-ring $\delta^{13}C$ data (Helama et al., 2018a) demonstrates the value of this isotope chronology to trace the past climatic events (*sensu* Whittaker et al., 1991) present in the recent geological record. Previously, tree-ring records have not served as GSSPs to subdivide formally the geological past. Traditionally most GSSPs utilise lithified and exposed rock outcrops; however, the GSSPs defining the lower boundary and subdivision of the Holocene Series/Epoch are now based on ice-core and speleothem isotope records (Walker et al., 2008, 2009, 2012, 2018). Interestingly, there have been suggestions of potential GSSP for the as-yet-undefined Anthropocene particularly in tree-ring samples (Lewis and Maslin, 2015; Turney et al., 2018). More generally, our results exemplify the potential of tree-ring archives to provide ideal time-stratigraphic marker horizons in terms of event stratigraphy.

We conclude by dating the Meghalayan lower boundary dendrochronologically to 2190 BCE. This dating is highly consistent with the ratified (Walker et al., 2018, 2019) lower boundary for the Meghalayan Stage/Age at 4.2 ka years ago.

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FIGURE CAPTIONS

Fig. 1. Tree-ring carbon isotope series from Lapland (a) and its record of high-latitude cloud cover changes (b) aligned to calendric timeline (CE/BCE) (Helama et al., 2018a; see also Table S2) compared with the oxygen isotope series from Mawmluh Cave speleothem (Berkelhammer et al., 2012), the Global Stratotype Section and Point (GSSP) for the Meghalayan Age (Walker et al., 2018) aligned to years BP (before 1950 CE) (c).

Fig. 2. Comparison of hydroclimatic, temperature, and the North Atlantic and solar proxy records. The $\delta^{18}O$ record from Stalagmite DP1, Namibia (Railsback et al., 2018) with horizontal dashed line denoting the mean $\delta^{18}O$ level (a), the tree-ring (TR) $\delta^{13}C$ record from Finnish Lapland (Helama et al., 2018a), the stacked North Atlantic multi-core record of percent hematite-stained grains (% HSG) indicative of ice-rafted debris (IRD) events, juxtaposed with the ^{14}C production rate (pr, atoms/cm^2/sec) indicative of variations in solar activity, both (% HSG and ^{14}C PR) as analysed and presented in Bond et al., (2001) for a period representing the 4.2 ka event (c) and over an interval corresponding to IRD-2, IRD-3 and IRD-4 occurrences (d), and the number of subfossil tree remains in the $\delta^{13}C$ sites in the northern Finnish Lapland per each century, with the mean levels of tree counts over the Northgrippian and Meghalayan ages (i.e. before and after 2190 CBE), indicative of long-term changes in summer temperature (Helama et al., 2004) and implying long-term cooling since the Meghalayan anomaly (e), aligned to years BP (before 1950 CE). The vertical dashed lines denote the bracketing dates of 4.15 to 3.93 ka (Railsback et al., 2018).