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Climate of the Past Discussions

Re-examining the 4.2 ka BP event in foraminifer isotope records from the Indus River delta in the Arabian Sea

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Abstract

The plains of northwest South Asia receive rainfall during both the Indian Summer (June-September) and Winter (December-March) Monsoon. Researchers have long attempted to deconstruct the influence of both precipitation regimes in paleoclimate records, in order to better understand regional climatic drivers and their potential impact on human populations. The Mid-Late Holocene transition between 5.3-3.3 ka BP is of particular interest in this region because it spans the period of the Indus Civilization from its early development to eventual transformation. The oxygen isotope record of the surface-dwelling planktonic foraminifer *Globigerinoides ruber* from the northeast Arabian Sea provided evidence for an abrupt decrease in rainfall and reduction in Indus River discharge at 4.2 ka BP, which the authors linked to the decline of the urban phase of the Indus Civilization (Staubwasser et al., 2003). Given the importance of this study, we used the same core (63KA) to replicate the oxygen isotope profiles of a larger size fraction of *G. ruber* than measured previously and, in addition, we measured two other foraminifer species at decadal resolution over the interval from 5.4 to 3.0 ka BP. By selecting both thermocline-dwelling (*Neogloboquadrina dutertrei*) and shallow-dwelling (*Globigerinoides sacculifer*) species, we provide enhanced detail of the climatic changes that occurred over this crucial time interval. We found evidence for a period of increased surface water mixing, which we suggest was related to a strengthened winter monsoon with a peak intensity over 200 years from 4.5 to 4.3 ka BP. The time of greatest change occurred at 4.1 ka BP when both the summer and winter monsoon weakened, resulting in a reduction in rainfall in the Indus region. The earliest phase of the Mature Harappan period coincided with the period of inferred stronger winter monsoon between 4.5-4.3 ka BP, whereas the end of the urbanized phase followed the decrease in both the summer and winter monsoon strength by 4.1 ka BP. Our findings provide evidence that the initial growth of large Indus urban centers was coincident with increased winter rainfall, whereas the contraction of urbanism and change in subsistence strategies followed a reduction in rainfall of both seasons.

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1. Introduction

The ~4.2 ka BP event is considered to be a defining event of the Mid-Late Holocene transition period (Mayewski et al., 2004), and is marked by intense aridity in much of western Asia, which has been linked to cultural transitions in Mesopotamia, Egypt, and the Indus Civilization (Staubwasser and Weiss, 2006; Weiss, 2016). Recently, a climate reconstruction from Mawmluh cave in northeastern India has been used to formally demarcate the post-4.2 ka BP time as the Meghalayan Age (Letter from the 44th International Union of Geological Sciences, 2018; Walker et al., 2012). However, defining the exact timing and extent of aridity at ~4.2 ka BP remains an open question (Finné et al., 2011; Wanner et al., 2008). In this special issue devoted to the "4.2 ka event", we provide new paleoclimate data from a marine core in the northern Arabian Sea over this critical time interval to better understand the changes that occurred in hydroclimate over the Indian Subcontinent.

The $\delta^{18}O$ record of *Globigerinoides ruber* from marine core 63KA, obtained from the Arabian Sea off the coast of Pakistan and produced by Staubwasser et al. (2003), was among the first well-resolved paleoclimate records to suggest a link between a decrease in Indus River discharge around 4.2 ka BP and the decline of the Indus Civilization. Since the publication of this record, several other terrestrial paleoclimate reconstructions from the region (Berkelhammer et al., 2012; Dixit et al., 2014, 2018; Giosan et al., 2012; Kathayat et al., 2017; Menzel et al., 2014; Nakamura et al., 2016; Prasad and Enzel, 2006), and a number of marine reconstructions (Giosan et al., 2018, in review; Gupta et al., 2003; Ponton et al., 2012) have added to our understanding of the complex relationship between the Indus Civilization and climate change. New questions have also emerged about the relative importance of winter rain from the Indian Winter Monsoon (IWM) system and summer rain from the Indian Summer Monsoon (ISM) during the critical time period from 5.4 to 3.0 ka BP, which spans the pre-urban, urban, and post-urban phases of the Indus Civilization (Giosan et al., 2018, in review; Petrie et al., 2017; Prasad and Enzel, 2006).

At its height, the Indus Civilization spanned a considerable geographical area with a greater extent than all the other ancient civilizations of its time (Agrawal, 2007; Possehl, 2003; Wheeler, 1968). Today, the region that was once occupied by Indus populations is marked by a heterogeneous rainfall pattern, and some sites in the central Thar desert receive as little as 100 mm yr^{-1}, which is only about 10% of the amount of direct annual rainfall seen in the northeastern region close to New Delhi. Scarce direct precipitation in the central regions around the Thar Desert is supplemented in some cases by fluvial or groundwater sources. In addition, the distribution of winter rain (increasing towards the northwest) is distinct from summer rain (increasing towards the east), making regions variably suitable for growing certain crops and grazing (Petrie et al., 2017; Petrie and Bates, 2017). While many paleoclimate studies from South Asia have theorized about the overall climatic impact of drought (and in most cases identified summer monsoon as the cause), it is important to identify changes in the relative contributions and timing of seasonal rainfall from both the winter and summer monsoons. Previously, it has not been possible to reliably differentiate winter and summer rain in reconstructions from the Indus region.

In this study, we re-examined the same marine core (63KA) used in the original Staubwasser et al. (2003) paper. We first assessed the reproducibility of the *Globigerinoides ruber* $\delta^{18}O$

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record using a larger size fraction of the same species for the time period 5.4-3.0 ka BP. We also measured the $\delta^{18}O$ of two additional foraminifer species, *G. sacculifer* (*Globigerinoides sacculifer*) and *N. dutertrei (Neogloboquadrina dutertrei)*, which live deeper than *G. ruber* in the water column. The different ecologies of these two species provide additional information with which to evaluate the multiple $\delta^{18}O$ records and assess seasonal changes in the paleoceanography of the northeastern Arabian Sea near the mouth of the Indus River.

The $\delta^{18}O$ of foraminifera has been widely applied as an indicator of temperature and salinity changes (Duplessy et al., 1992; Maslin et al., 1995; Wang et al., 1995; Rohling, 2000). Measuring the $\delta^{18}O$ of species calcifying at different depths can provide further information about upper ocean seasonal hydrography such as surface water mixing, depth of the thermocline, and upwelling (Ravelo and Shackleton, 1995). Such methods have been applied by several other studies (Billups et al., 1999; Cannariato and Ravelo, 1997; Norris, 1998; Steinke et al., 2010; Steph et al., 2009; among others), including a reconstruction of East Asian Winter Monsoon strength in the South China Sea (Tian et al., 2005). We applied a similar method to samples from core 63KA in the northeastern Arabian Sea because surface waters at this location are influenced by freshwater discharge from the Indus River and direct precipitation during the summer monsoon months, whereas enhanced upper ocean mixing and upwelling occur during the winter monsoon. By comparing the $\delta^{18}O$ of multiple species of foraminifera we seek to infer variations in the relative strengths of the summer and winter monsoons, and by comparing the 63KA record to other nearby marine and terrestrial records we evaluate the potential role that climate played in cultural transformation of the Indus Civilization.

2. Site Description

2.1 Monsoon – land-based processes

Today, most of the annual precipitation over northwest South Asia stems from the ISM, and occurs mainly between June and September. The pressure gradient between the low-pressure Tibetan Plateau and high-pressure Indian Ocean is accompanied by the ITCZ (Intertropical Convergence Zone) reaching its northward maximum in summer, which draws in moisture over the subcontinent via southwesterly winds from the Indian Ocean (Fleitmann et al., 2007; Gadgil, 2003). The summer rainfall gradient increases from the central Thar Desert (as little as 100 mm direct summer rainfall per year) to the Himalaya mountains in the north (>1000 mm) and the Aravalli range to the west (>500 mm) (Figure 1b).

The IWM rain falls between December through March, and is mainly the result of atmospheric Western Disturbances (Dimri and Dash, 2012; Yadav et al., 2012) originating over the Mediterranean and Black Sea (Hatwar et al., 2005) that allow for moisture incursion from the Arabian Sea (Rangachary and Bandyopadhyay, 1987). During the IWM, the pressure gradient is reversed from the summer condition, allowing the passage of Western Disturbances when the ITCZ moves southward. As winter transitions to spring, predominantly northeasterly winds shift to westerly winds (Sirocko, 1991) that result in peak winter rainfall over the plains of northwest India in February and March. The winter rainfall gradient increases from the southern Thar Desert (<10 mm per year) up to the Himalayas in the northwest (>400 mm)

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(Figure 1c). Overall, the IWM contributes between roughly 10 to 50% of the total annual rainfall of northwest South Asia today.

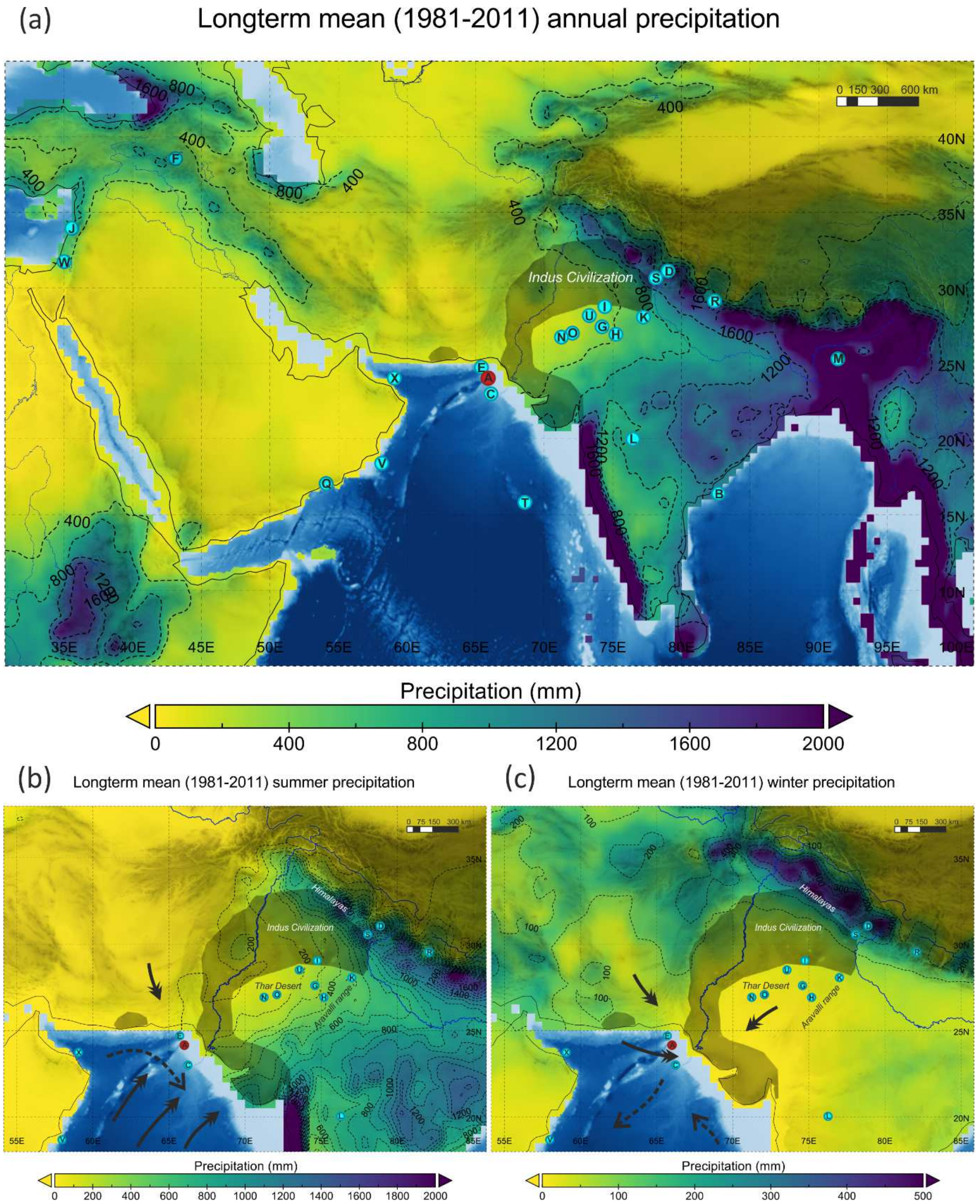

Figure 1. a. Annual **b.** ISM (JJAS) **c.** IWM (DJFM) mean precipitation (1981-2011) isohyets taken from the GPCC V7 global gridded dataset (0.5° x 0.5° resolution); note the difference in scale for summer and winter precipitation (0-2000 mm vs. 0-500 mm). Rainfall data overlain on GEBCO 2014 ocean bathymetry dataset, and shaded region shows extent of the Indus Civilization. Bold arrows show main wind directions, dashed arrows show ocean surface currents. Other studies discussed in this paper indicated by letters:

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A Core 63KA – (this study; Staubwasser et al., 2003)
B Core 16A – (Ponton et al., 2012)
C Core Indus 11C – (Giosan et al., 2018, in review)
D Din Gad peat record – (Phadtare, 2000)
E Core 39KG and 56KA – (Doose-Rolinkski et al., 2001)
F Lake Van record – (Wick et al., 2003; Lemcke and Sturm, 1997)
G Didwana playa lake – (Singh et al., 1990)
H Sambhar playa lake – (Sinha et al., 2006)
I Karsandi playa lake – (Dixit et al., 2018)
J Jeita cave speleothem – (Cheng et al., 2015)
K Kotla Dahar lake – (Dixit et al., 2014)
L Lonar lake – (Menzel et al., 2014)
M Mawmluh cave speleothem – (Berkelhammer et al., 2012)
N Kanod playa lake – (Deotare et al., 2004)
O Bap Malar playa lake – (Deotare et al., 2004)
Q Qunf cave speleothem – (Fleitmann et al., 2003)
R Rara lake – (Nakamura et al., 2016)
S Sahiya cave speleothem – (Kathayat et al., 2017)
T Foraminifer trap EAST – (Curry et al., 1992)
U Lunkaransar playa lake – (Enzel et al., 1999)
V Core 723A, RC27-14, RC27-23, RC27-28 – (Gupta et al., 2003), (Overpeck et al., 1996)
W Soreq cave speleothem – (Bar-Matthews et al., 2003; Bar-Matthews and Ayalon, 2011)
X Core M5-422 – (Cullen et al., 2000)

The Indus and the other rivers that make up Punjab are partly fed by winter snow and ice melt from their upper mountain catchment areas. Melting peaks during the summer months around July-August (Yu et al., 2013), which coincides with the peak of ISM rainfall, and Indus River discharge reaches its maximum during August (Karim et al., 2002). The proportion of winter to summer precipitation contributing to the Indus River is not entirely clear, although one study has estimated a 64-72% contribution of winter precipitation from the deuterium excess of Indus River water (Karim et al., 2002), whereas a previous study estimated a lower 15-44% contribution of snowmelt to Indus tributaries (Ramasastri, 1999). Since the 1960s, the Indus River has seen a more than a 50% reduction in discharge because of the construction of barrages as well as the diversion of water for agricultural uses (Ahmad et al., 2001).

2.2 Hydrography – ocean-based processes

Core 63KA was obtained by the PAKOMIN cruise in 1993 (von Rad et al., 1995). The laminated core from the northeastern Arabian Sea (24° 37′ N, 65° 59′ E) was taken at 316 m water depth on the continental shelf, ~100 km west of the Indus River delta. The core has high sedimentation rates (equivalent to a temporal resolution of around 18 years/cm in the period of interest, 5.4-3.0 ka BP), and all foraminifer proxies were produced from the same laminated core with no bioturbation. An important aspect of core 63KA is that different components of the monsoon system are co-registered in the same sediment core, thereby permitting an explicit evaluation of the relative timing of different parts of the climate system (e.g., ISM and IWM). Modern hydrographic conditions in the northeastern Arabian Sea are highly influenced by the seasonal monsoon. During summertime, highest sea surface temperatures (SSTs) are observed along with a shallow mixed layer depth <25 m (Schulz et al., 2002) (Figure 2a). A low salinity plume surrounds the Indus River delta and shoreline extending as far as the coring location (Supplemental Figure S1). The reverse occurs in winter when the lowest SSTs are accompanied by surface water mixing to >125 m, resulting in warming of the deeper waters (Schulz et al., 2002). During the transition from winter to spring, wind directions shift from northeasterly to westerly (Sirocko, 1991), promoting a period of upwelling in the northeastern Arabian Sea (Staubwasser et al., 2002; Rao, 1981).

The northern Arabian Sea is dominated by highly saline (up to 37 psu) surface waters of the Arabian Sea High Salinity Water Mass (ASHSW), which extend from the surface up to 100 m depth (Joseph and Freeland, 2005). This high salinity can be explained by the high evaporative

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rates over this region. Along coastal areas the ASHSW is starkly contrasted by the fresh water discharge of the Indus River, combined with direct precipitation. In contrast, surface waters in the Bay of Bengal on the eastern side of India have much lower surface water salinity, because of overall higher precipitation and stronger stratification from weaker winds (Shenoi et al., 2002). The heightened evaporative conditions and highly saline surface waters of the northeastern Arabian Sea make it a sensitive study location to observe changes in discharge of the entire Indus River catchment area – ultimately tracking changes in monsoon strength. Unlike individual terrestrial records, which may be affected by local climatic processes, the marine record from core 63KA is more likely to integrate regional changes of the large-scale ocean-atmosphere system.

Planktonic foraminifera complete their life cycle within a few weeks (Bé and Hutson, 1977). Peak abundances indicate the time of year when each species tends to calcify, thereby recording the $\delta^{18}O$ of the seawater in their $CaCO_3$ shells primarily during certain seasons. Foraminifer abundances in the eastern Arabian Sea have been studied by Curry et al. (1992) using sediment traps deployed at shallow (~1400 m) and deep (~2800 m) water depths ("T" in Figure 1a). Peak abundances for *G. ruber* and *G. sacculifer* occur during the summer months (June-September), whereas *N. dutertrei* peak mainly during the winter as well as with a secondary peak in summer months (Figure 2c). Preferred depth ranges for each species reflect their ecological niches, including requirements for nutrients and tolerance for ranges of temperature and salinity (Bé and Hutson, 1977; Hemleben et al., 2012). *G. ruber* lives in the upper surface waters (0-10 m), *G. sacculifer* is found in slightly deeper surface waters (10-40 m), and *N. dutertrei* inhabits the base of the mixed layer near the thermocline (40-140 m) (estimates based on ranges from Farmer et al. (2007) and the local CTD profile) (Figure 2d).

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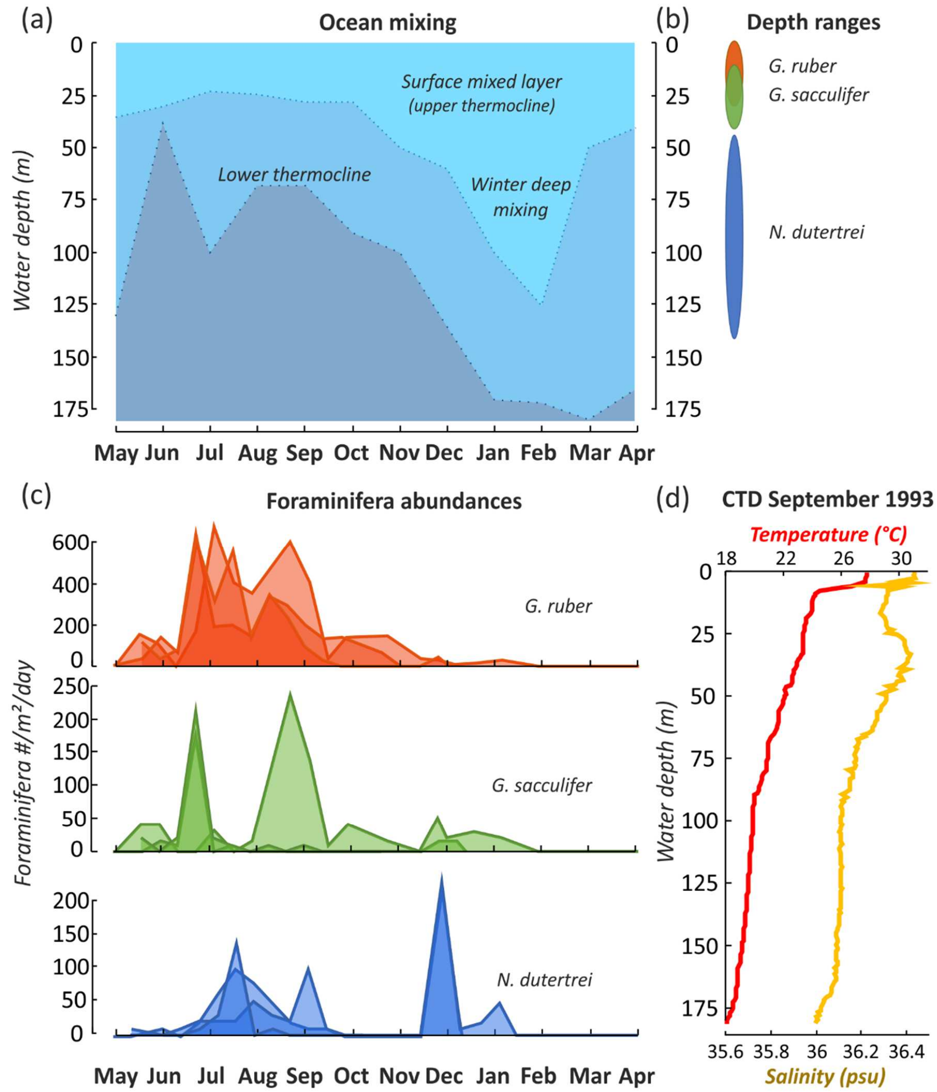

Figure 2. a. Seasonal surface water mixing depth based on station EPT-2 located nearby the coring site of 63KA (adapted from Schulz et al., 2002 who also used data from Hastenrath and Lamb, 1979) **b.** Foraminifer depth ranges based on CTD profile **c.** Foraminifer abundances from EAST trap: *G. ruber* (orange), *G. sacculifer* (green), and *N. dutertrei* (blue) (adapted from Curry et al., 1992 using Zaric, 2005) **d.** CTD temperature (red) and salinity (yellow) profile from station 11 at coring location, taken September 1993 (von Rad, 2013).

3. Materials and Methods

3.1 Age model

The radiocarbon dates from Staubwasser et al. (2002, 2003) were obtained from 80 samples of mainly the foraminifer *G. sacculifer* and three samples of *O. universa*. In the interval of

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interest (5.4-3.0 ka BP), there are 15 radiocarbon dates with a 95% confidence range of 30-130 years. Bayesian age modelling software, BACON v2.3.3 (Blaauw and Christen, 2011), was used as an R-package to update the age model of core 63KA (Supplemental Figure S5, Table S2). IntCal13 was used for radiocarbon calibration (Reimer et al., 2013) with marine reservoir ages provided by Staubwasser et al. (2002, 2003).

3.2 Stable isotope analysis

Oxygen and carbon isotopes were measured on three species of foraminifera selected from washed samples at 1-cm intervals throughout the core: *G. ruber* (white, *sensu stricto*), *G. sacculifer*, and *N. dutertrei*. For *G. ruber*, 12 ± 8 foraminifera were picked from the 400-500µm size fraction with an average weight of 21.4 ± 2.5µg. The 400-500µm size fraction was picked because too few specimens remained in the size fraction 315-400µm used by Staubwasser et al. (2003). For *G. sacculifer*, 34 ± 7 foraminifera were picked from the 315-400µm size fraction with an average weight of 21.9 ± 2.6µg. For *N. dutertrei*, 34 ± 4 foraminifera were picked from the 315-400µm size fraction with an average weight of 25.9 ± 2.2µg. At some depth levels in the core there were insufficient foraminifera for measurement, leaving 11 gaps in the *G. ruber* 400-500µm record, 3 gaps in the *G. sacculifer* record, and no gaps for *N. dutertrei*. The published *G. ruber* is from the 315-400µm size fraction and contains 17 gaps in the depth range examined (Staubwasser et al., 2003).

All foraminifera were weighed, crushed, and dried at 50° C. Samples were cleaned for 30 minutes with 3% H_2O_2, followed by a few drops of acetone, ultrasonication, and drying overnight. Where sample weights exceeded 80µg, oxygen and carbon isotopes were measured using a Micromass Multicarb Sample Preparation System attached to a VG SIRA Mass Spectrometer. In cases of smaller sample sizes, the Thermo Scientific Kiel device attached to a Thermo Scientific MAT253 Mass Spectrometer was used in dual inlet mode. This method adds 100% H_3PO_4 to the $CaCO_3$, water is removed cryogenically, and the dry CO_2 is analyzed isotopically by comparison with a laboratory reference gas. For both measurement methods, 10 reference carbonates and 2 control samples were included with every 30 samples. Results are reported relative to VPDB, and internal precision is better than ±0.08‰ for $\delta^{18}O$ and ±0.06‰ for $\delta^{13}C$. External precisions were estimated by five triplicate (three separately picked) measurements of *G. ruber* (400-500µm) that yielded one standard deviation of ±0.12‰ ($\delta^{18}O$) and ±0.10‰ ($\delta^{13}C$). For *G. sacculifer* (315-400µm) the standard deviation of eight triplicate measurements were ±0.07‰ ($\delta^{18}O$) and ±0.07‰ ($\delta^{13}C$), and for *N. dutertrei* (315-400µm) the standard deviation of nine triplicate measurements was ±0.06‰ ($\delta^{18}O$) and ±0.07‰ ($\delta^{13}C$).

To calculate equilibrium values of $\delta^{18}O_{calcite(PDB)}$, we used the CTD profile from station 11 (24.62° N, 66.07° E) (Figure 2d) taken during PAKOMIN *Sonne* cruise no. 90 (von Rad, 2013), which is nearly identical to the location of core 63KA (24.62° N 65.98° E). The $\delta^{18}O_{water(SMOW)}$ was calculated from salinity following Dahl and Oppo (2006), and $\delta^{18}O_{calcite(SMOW)}$ was further calculated using the calcite-water equation of Kim and O'Neil (1997). We also used the equation of Shackleton (1974) as a comparative method for calculating $\delta^{18}O_{calcite(PDB)}$.

3.3 Statistical treatment

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As in the original data of Staubwasser et al. (2003), the oxygen isotope results show great variability and distinguishing long-term trends in these data requires statistical smoothing. To reduce the variance in the data and identify trends, the $\delta^{18}O$ and $\delta^{13}C$ data from 5.4-3.0 ka BP were first resampled to constant 1-year intervals using linear interpolation. A loess (locally weighted) smoothing function was then applied to the data, using a 210-year moving window as described by Staubwasser et al. (2003). Loess smoothing uses weighted least squares, which places more importance on the data points closest to the center of the smoothing interval. The bandwidth of 210 years was considered an optimal time window for capturing the overall trends in the dataset (other time windows are shown for comparison in Supplemental Figure S2).

Statistical tests were applied to the $\delta^{18}O$ and $\delta^{13}C$ time series, including the package SiZer (Sonderegger et al., 2009) in R software (2016) that calculates whether the derivative of a time series exhibits significant changes given a range of timespans. We applied a range of smoothing windows (bandwidth of 20-500 years) to assess the significance of changes in the isotope records throughout the time series. Where step changes occur in $\delta^{18}O$ we also conducted a Student's t-test to determine if the mean population of $\delta^{18}O$ is significantly different before and after the change.

4. Results

The new $\delta^{18}O$ measurements of *G. ruber* (400-500µm) parallel the published record of *G. ruber* (315-400µm) (Staubwasser et al., 2003), but the $\delta^{18}O$ of the specimens from the larger size fraction is offset by -0.23‰ on average (Figure 3). The two records, produced in different laboratories by different investigators, display a weak correlation for the raw data (R = 0.2, slope 0.27, intercept -1.33) and a strong correlation for the 210-year smoothed records (R = 0.7, slope 0.53, intercept -0.77). When comparing the two *G. ruber* records, it is apparent that the increasing trend in $\delta^{18}O$ starts well before ~4.2 ka BP – perhaps as early as ~4.9 ka BP. This trend is also observed with the SiZer analysis, which identifies a significant increase in $\delta^{18}O$ anywhere from 4.9 to 4.2 ka BP depending on which smoothing window is selected (Figure 4). The new $\delta^{18}O$ record of *G. ruber (*400-500µm) shows additional detail after the ~4.2 ka BP event – i.e. specifically, a double-peak maximum occurring at 4.1 and 3.95 ka BP that is related to seven discrete measurements with high $\delta^{18}O$ values. These maxima are offset from the average $\delta^{18}O$ value by +0.18‰ (smoothed average), or up to +0.38‰ when considering the maximum individual measurement at 4.1 ka BP. The offsets from the average values exceed one standard deviation of the entire record from 5.4-3.0 ka BP, which is 0.13‰. A Student's t-test comparing the means of pre- and post-4.1 ka BP indicates that the +0.07‰ shift in mean $\delta^{18}O$ values of *G. ruber* (315-400µm) is statistically significant (t value = 2.9, $p < 0.01$), and the +0.04‰ shift in mean $\delta^{18}O$ values of *G. ruber* (400-500µm) is weakly significant (t value = 1.7, $p < 0.1$).

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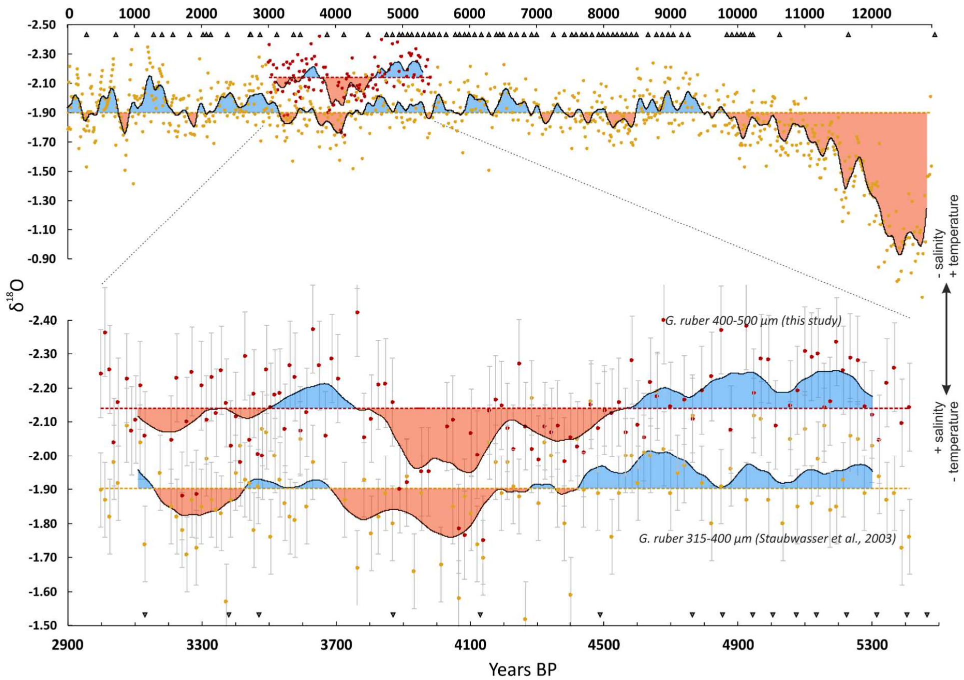

Figure 3. Core 63KA $\delta^{18}O$ *G. ruber* from two size fractions: 400-500µm (red) (this study), 315-400µm (orange) (Staubwasser et al., 2003), shown zoomed in over 5.4-3.0 ka BP. Data shown with a 210-year loess smoothing and ±1σ error bars. Individual AMS radiocarbon dates are denoted by triangles near the timeline.

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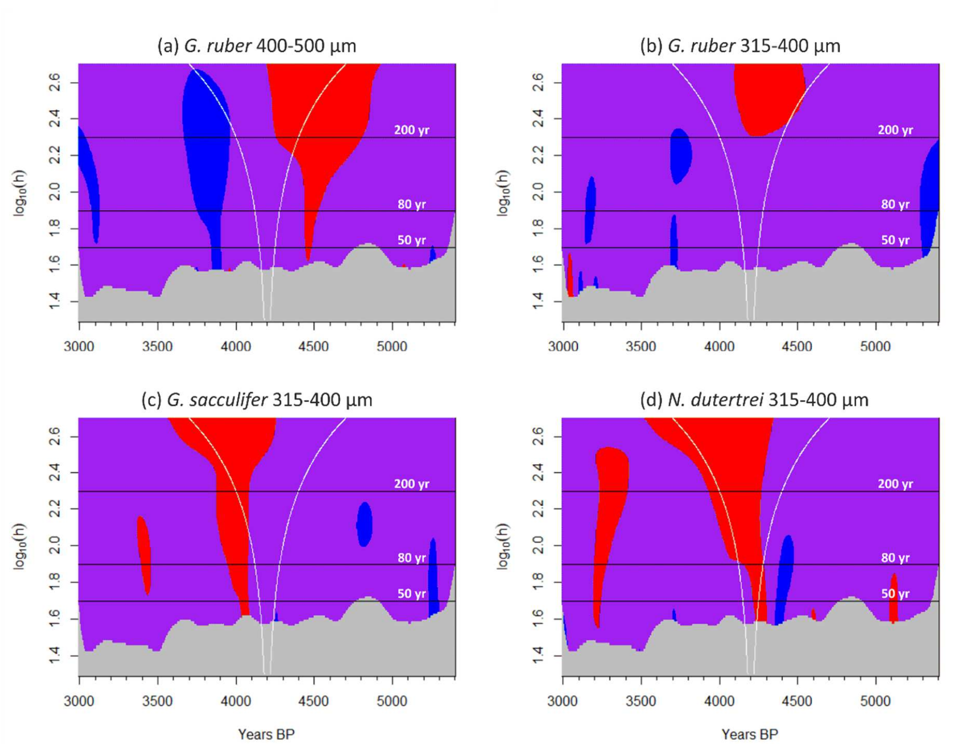

Figure 4. SiZer 1st derivative analysis (Sonderegger et al., 2009) applied to $\delta^{18}O$ of **a.** *G. ruber* 400-500µm, **b.** *G. ruber* 315-400µm, **c.** *G. sacculifer* 315-400µm, **d.** *N. dutertrei* 315-400µm. Statistically significant increases (red) and decreases (blue) are shown alongside times with no significant change (purple). The smoothing bandwidth (h) ranges from 20 to 500 years (denoted by the width between the white lines), and three bandwidths have been highlighted by black horizontal lines at h = 50, 80, and 200 years.

The relative differences in $\delta^{18}O$ of the planktonic species studied (*G. ruber*, *G. sacculifer* and *N. dutertrei*) reflect the temperature and salinity of their habitat in the water column: $\delta^{18}O$ *G. ruber* < $\delta^{18}O$ *G. sacculifer* < $\delta^{18}O$ *N. dutertrei* (Figure 5). *G. sacculifer* is offset from *G. ruber* (315-400µm) by approximately +0.57‰, whereas *N. dutertrei* is offset by +1.14‰. The larger size fraction of *G. ruber* (400-500µm) is offset from *G. ruber* (315-400µm) by -0.23‰. The offsets among species are maintained throughout the entire record (Figure 5). We also measured $\delta^{18}O$ values near the top of the core (approximately the last 200 years) for all three species in the 315-400µm size fraction, which continue to show the same offsets (Supplemental Figure S3). The $\delta^{18}O$ of *G. ruber* shows the greatest variance and *N. dutertrei* shows the least (Supplemental Figure S4, Table S1).

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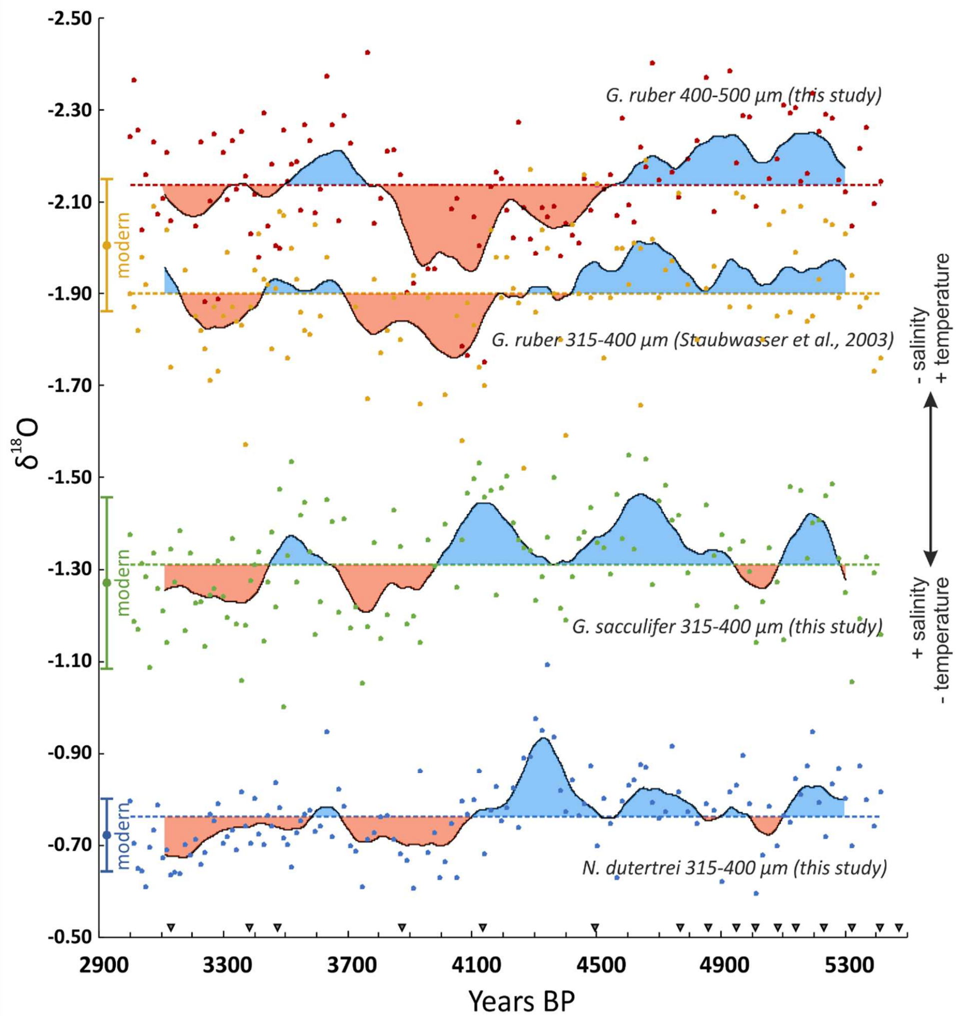

Figure 5. Core 63KA $\delta^{18}O$ *G. ruber* from two size fractions: 400-500µm (red) (this study), 315-400µm (orange) (Staubwasser et al., 2003), *G. sacculifer* 315-400µm (this study), and *N. dutertrei* 315-400µm (this study). Data shown with a 210-year loess smoothing, modern surface values ±1σ shown for comparison. Individual AMS radiocarbon dates are denoted by triangles near the timeline.

Equilibrium calcite calculations based on the salinity and temperature measurements from the September 1993 CTD profile of station 11 of the PAKOMIN Cruise (von Rad, 2013) show the expected depth habitats of the three foraminifer species (Figure 6). *G. ruber* is generally found at 0-30 m, *G. sacculifer* at 15-40 m, and *N. dutertrei* at 60-150 m (Farmer et al., 2007). Using the CTD profile from our core location, we compare these depth ranges with the measured $\delta^{18}O$ values. The calculated depths ranges agree well with those expected on the basis of other studies, placing *G. ruber* in the upper 10 m, *G. sacculifer* 10-40 m, and *N. dutertrei* 40-140 m.

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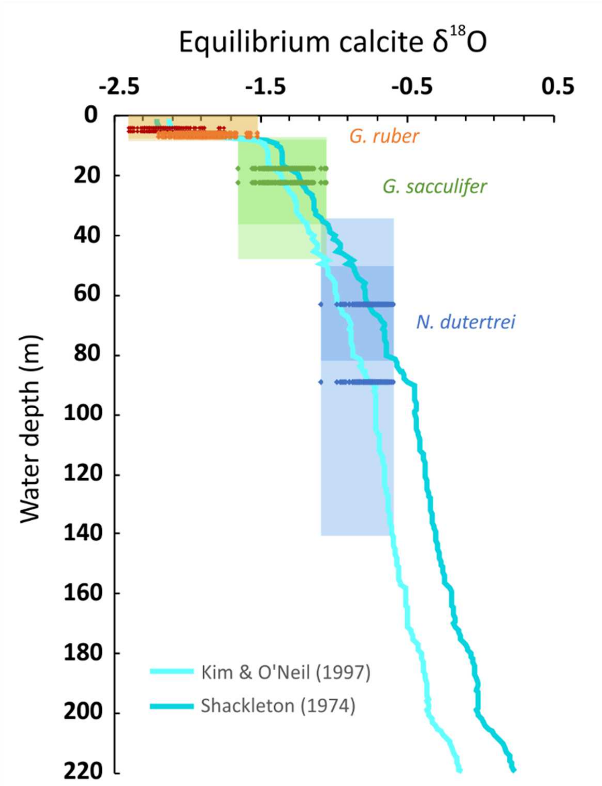

Figure 6. $\delta^{18}O$ of equilibrium calcite calculated from CTD profile at station 11 (von Rad, 2013) with projected depth ranges of *G. ruber* 400-500µm (red), *G. ruber* 315-400µm (orange), *G. sacculifer* 315-400µm (green), *N. dutertrei* 315-400µm (blue). We show estimated values using both the original paleotemperature equation of Shackleton (1974) (dark teal), and Kim & O'Neil (1997) (turquoise). Horizontal ranges show the measured $\delta^{18}O$ values of each species between 5.4-3.0 ka BP.

The most obvious trend in the *G. sacculifer* $\delta^{18}O$ is the increase around 4.1 ka BP. A Student's t-test comparing the means of pre- and post-4.1 ka BP indicates that the +0.07‰ shift in mean $\delta^{18}O$ values is statistically significant (t value = 3.9, $p < 0.01$). SiZer analysis also points to a statistically significant increase at ~4.1-3.9 ka BP, when considering all smoothing time windows between 20 and 500 years (Figure 4).

The dominant trend in the $\delta^{18}O$ of *N. dutertrei* is a mean increase at 4.1 ka BP (Figure 7). SiZer analysis also identifies a significant decrease in $\delta^{18}O$ occurring mainly between 4.45 and 4.35 ka BP, followed by a significant increase between 4.3 and 4.1 ka BP (Figure 4). A Student's t-test comparing the means of pre- and post-4.1 ka BP indicates that the +0.08‰ shift in mean $\delta^{18}O$ values is statistically significant (t value = 6.3, $p < 0.01$).

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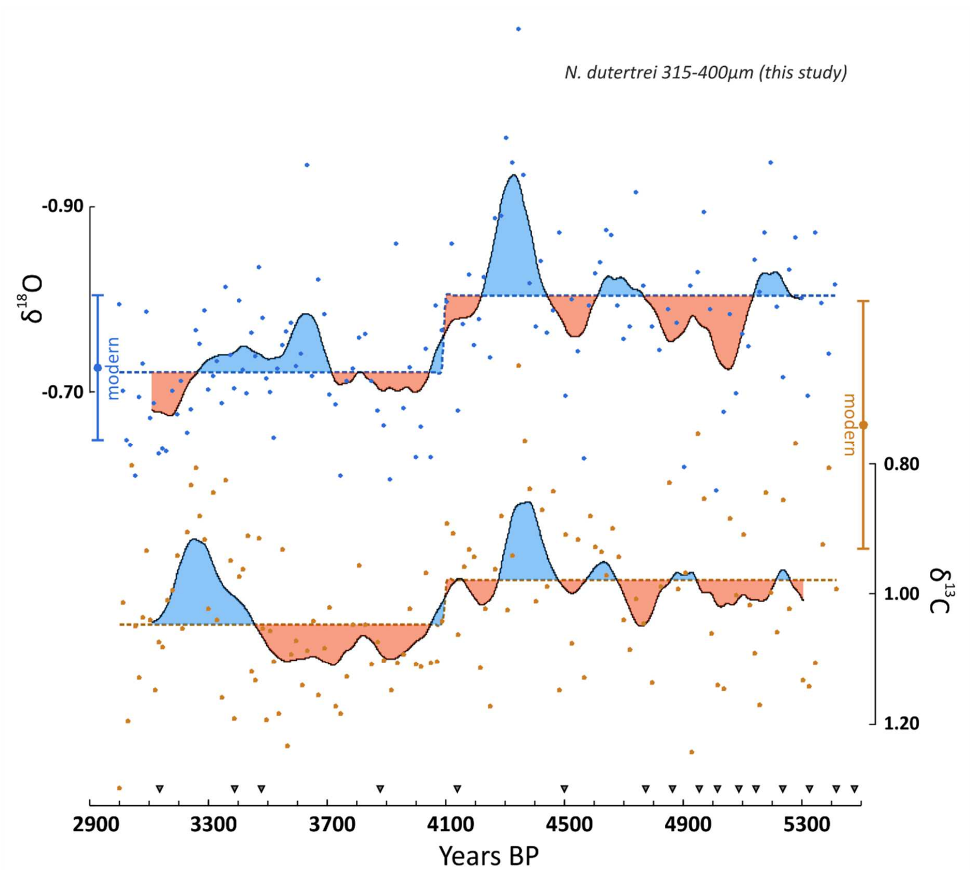

Figure 7. Core 63KA $\delta^{18}O$ **(top)** and $\delta^{13}C$ **(bottom)** of *N. dutertrei* 315-400µm. Data shown with a 210-year loess smoothing, modern surface values ±1σ shown for comparison. Individual AMS radiocarbon dates are denoted by triangles near the timeline.

Differencing $\delta^{18}O$ of foraminifera (expressed as $\Delta^{18}O$) in the same sample can sometimes improve the signal-to-noise ratio (Figure 8). The $\Delta^{18}O$ of *G. ruber* 400-500µm and *G. ruber* 315-400µm size fractions shows increasing similarity between ~4.8 and 3.9 ka BP during the period of overall higher $\delta^{18}O$. The $\Delta^{18}O$ of *N. dutertrei* and both size fractions of *G. ruber*, designated $\Delta\delta^{18}O_{d-r}$, reveals a period of more similar values between ~4.5 and 3.9 ka BP, with two minima at 4.3 and 4.1 ka BP. The $\Delta^{18}O$ of *G. sacculifer* and both size fractions of *G. ruber* ($\Delta\delta^{18}O_{s-r}$), shows a period of similar values between 4.3 and 3.9 ka BP, with a minimum difference at 4.1 ka BP. In contrast, the $\Delta^{18}O$ of *N. dutertrei* and *G. sacculifer* ($\Delta\delta^{18}O_{d-s}$), shows the most similarity between 4.5 and 4.2 ka BP with a minimum at 4.3 ka BP, followed by the maximum differences between 4.2 and 3.9 ka BP that peaks at 4.1 ka BP.

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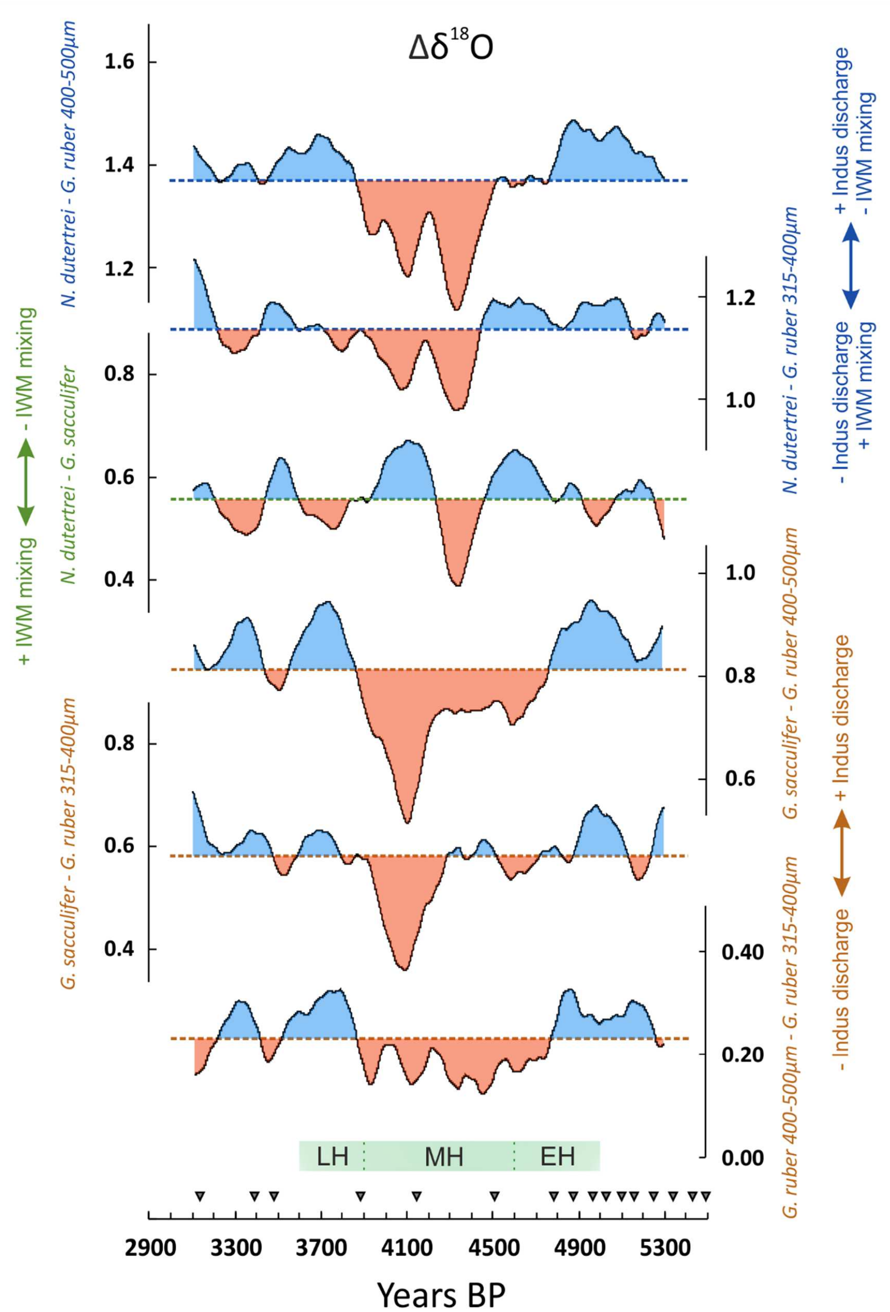

Figure 8. Core 63KA $\Delta\delta^{18}O$. Data shown with a 210-year loess smoothing. Individual AMS radiocarbon dates are denoted by triangles near the timeline. *G. ruber* 315-400µm size fraction data from Staubwasser et al. (2003). The green band near the timeline showing EH, MH, and LH refers to Early Harappan (~5.0-4.6 ka BP), Mature Harappan (~4.6-3.9 ka BP), and Late Harappan (~3.9-3.6 ka BP), respectively.

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5. Discussion

5.1 Interpretation of foraminifer $\delta^{18}O$

The original $\delta^{18}O$ record of *G. ruber* (315-400µm) by Staubwasser et al. (2003) is confirmed by our independent $\delta^{18}O$ measurements of *G. ruber* in a larger size fraction (400-500µm). The larger size fraction is offset by approximately -0.2‰, which is similar to the size-related fractionation of -0.3‰ per +100µm for *G. ruber* reported by Cayre and Bassinot (1998), and could be attributed to size-related vital effects. Alternatively, the offset might be explained by interlaboratory calibration considering the data were produced using two different methods and mass spectrometers.

The observed 4.1 ka BP maximum in $\delta^{18}O$ of *G. ruber*, living near the surface during summer months, could be attributed to either decreased SST or increased surface water salinity (Bemis et al., 1998). Staubwasser et al. (2003) acknowledged that a decrease in SST could cause the increase in $\delta^{18}O$ in the *G. ruber* record, but argued that this explanation is unlikely because a *G. ruber* $\delta^{18}O$ record from core M5-422 in the northwestern Arabian Sea shows opposing trends over the same time period (Cullen et al., 2000), and a local alkenone SST proxy record shows relatively higher temperatures in the same period (Doose-Rolinski et al., 2001). If the ~0.2‰ (relative to mean) increase in $\delta^{18}O$ of *G. ruber* at 4.1 ka BP was caused by temperature change rather than salinity, a ~1° C cooling of surface water is implied (Kim and O'Neil, 1997).

Following Staubwasser et al. (2003), we interpret the $\delta^{18}O$ variations of *G. ruber* to be predominantly a salinity signal. Salinity at the core site is dependent on changes in Indus River discharge, local run-off, and direct precipitation. Although the ISM would be the main influence on direct precipitation and run-off at the coring location, changes in the IWM could also influence Indus River discharge, because snowmelt is a significant contributor in the upper Indus catchment (Karim et al., 2002) and peaks during the summer months (Yu et al., 2013).

The thermocline-dwelling foraminifera *N. dutertrei* shows maximum abundances during winter, and are interpreted to reflect winter mixing. Weaker IWM winds are expected to result in a shorter duration and/or less intense upper ocean mixing, although how this signal is ultimately related to the amount or distribution of winter rainfall in the Indus River catchment has not been demonstrated conclusively. Dimri (2006) studied Western Disturbances for the time period 1958-1997, and noted that surplus years of winter precipitation are linked to significant heat loss over the northern Arabian Sea, which is mainly attributed to intensified westerly moisture flow and enhanced evaporation. Such conditions would promote deeper winter mixing, and provide a basis for relating thermocline depth with IWM intensity. During weak IWM conditions, colder unmixed water would result in higher $\delta^{18}O$ values of *N. dutertrei*, whereas enhanced mixing and homogenization of the water column under strong IWM conditions would decrease $\delta^{18}O$. The minimum of $\delta^{18}O$ in *N. dutertrei* occurs between 4.5 and 4.3 ka BP, pointing to a period of strengthened IWM. We interpret the stepped increase in $\delta^{18}O$ of *N. dutertrei* at 4.1 ka BP to represent a decrease in IWM wind-driven mixing. Similarly, $\delta^{13}C$ of *N. dutertrei* increases significantly after 4.1 ka BP (Figure 7), which would also suggest reduced upwelling of low $\delta^{13}C$ intermediate water under

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a weaker IWM (Lynch-Stieglitz, 2006). According to the $\delta^{18}O$ signal of *N. dutertrei*, the temperature pattern in the thermocline implies surface cooling between 4.5 and 4.3 ka BP and surface warming after 4.1 ka BP interrupted only by a period of cooling between 3.7 and 3.3 ka BP, which is in broad agreement with records of alkenone sea-surface temperature estimates from cores in the northeastern Arabian Sea ("E" in Figure 1) (Doose-Rolinski et al., 2001; Staubwasser, 2012).

5.2 Interpretation of foraminifer $\Delta^{18}O$

By using $\Delta\delta^{18}O$ between foraminifer species, we can distinguish additional processes affecting the surface waters and thermocline (Ravelo and Shackleton, 1995). This technique has been used previously to infer changes in the strength of the East Asian Winter Monsoon (EAWM) in the South China Sea (Tian et al., 2005), as well as mixed layer and thermocline depth in other studies (Billups et al., 1999; Cannariato and Ravelo, 1997; Norris, 1998). Here we use the difference in the $\delta^{18}O$ of *G. ruber* and *N. dutertrei* ($\Delta\delta^{18}O_{d-r}$) to track changes in the surface-to-deep gradient. This gradient can be driven by either $\delta^{18}O$ changes in the surface-dwelling (*G. ruber*) and/or the thermocline-dwelling species (*N. dutertrei*). During times of a strengthened winter monsoon, $\Delta\delta^{18}O_{d-r}$ will decrease as surface waters are homogenized and the thermocline deepens. Similarly, $\Delta\delta^{18}O_{d-r}$ will also decrease during times of a weakened summer monsoon, as decreased Indus River discharge will increase surface water salinity and $\delta^{18}O$ of *G. ruber* will become more similar to *N. dutertrei*.

G. sacculifer is also a surface dweller, but has a slightly deeper depth habitat than *G. ruber*. We thus expect *G. ruber* to be more influenced by surface salinity variations than *G. sacculifer*, and suggest the $\delta^{18}O$ difference between the two species ($\Delta\delta^{18}O_{s-r}$), reflects the influence of Indus River discharge on near surface salinity. The greatest difference in $\Delta\delta^{18}O_{s-r}$ occurs at 4.1 ka BP, which is interpreted as an increase in surface water salinity (Figure 8).

The difference in $\delta^{18}O$ between *G. sacculifer* and *N. dutertrei* ($\Delta\delta^{18}O_{d-s}$) also reflects surface mixing and thermocline depth, but *G. sacculifer* is less affected by surface salinity changes than *G. ruber*. Thus, the responses of $\Delta\delta^{18}O_{d-s}$ and $\Delta\delta^{18}O_{d-r}$ can be used to differentiate between surface water salinity changes and wind-driven mixing. Accordingly, simultaneously low $\Delta\delta^{18}O_{d-s}$ and $\Delta\delta^{18}O_{d-r}$ indicate a period of increased surface water mixing and increased IWM (such as the period between 4.5 and 4.3 ka BP), but times of relatively low $\Delta\delta^{18}O_{d-s}$ but high $\Delta\delta^{18}O_{d-r}$ and $\Delta\delta^{18}O_{s-r}$ (around 5.0 ka BP) indicate periods of increased Indus discharge and strength of the ISM and IWM.

However, the following period of low $\Delta\delta^{18}O_{d-r}$ but high $\Delta\delta^{18}O_{d-s}$ from 4.1-3.9 ka BP is likely driven by increased salinity of surface water. This distinction becomes clearer when examining the $\Delta\delta^{18}O_{s-r}$, where increased similarity from 4.8-3.9 ka BP (with a sharp increase at 4.1 ka BP) reflects the effect of increased sea surface salinity that reduces the $\delta^{18}O$ difference between *G. ruber* and *G. sacculifer*. At the same time, weakened winter mixing increases $\Delta\delta^{18}O_{d-s}$, which occurs from 4.2-3.9 ka BP. Importantly, the proxies also indicate that increased IWM mixing is generally positively correlated with increased Indus discharge, and vice versa. The single time period when this does not hold true is 4.5-4.25 ka BP, when increased IWM mixing is coupled with decreased Indus discharge.

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In summary, our multi-species approach using $\delta^{18}O$ of *G. ruber*, *G. sacculifer*, and *N. dutertrei* allows us to differentiate between strength of the IWM and freshwater discharge of the Indus River. We suggest that ISM strength decreased gradually from at least 4.8 ka BP, while the IWM strength peaked around 4.5-4.3 ka BP and then weakened afterwards. It is unlikely that the abrupt increase in *G. ruber* $\delta^{18}O$ at 4.1 ka BP and low $\Delta\delta^{18}O_{s\text{-}r}$ could be caused solely by the decrease in IWM strength, even though IWM contributes to Indus River discharge. Weakening of the ISM must have played a substantial role in the 4.1 ka BP shift as well, indicated by the period 4.5-4.25 ka BP, when Indus discharge reflected a weak ISM ($\Delta\delta^{18}O_{s\text{-}r}$) despite a phase of strengthened IWM.

5.3 Comparison to marine records

The interpretation of core 63KA relies on proxies that directly link surface water salinity to ISM precipitation and Indus River discharge, and thermocline shifts to related IWM-driven mixing. Additionally, there is an established mechanism relating mixing with IWM strength, as anomalously cool and evaporative conditions over the northern Arabian Sea (promoting deeper winter mixing) correlates with increased winter precipitation in the western Himalayas (Dimri, 2006). The strength of the 63KA core lies in its highly resolved age model, high sedimentation rates and its position in particularly saline surface waters (ASHSW) close to highly contrasting freshwater sources. Additionally, both the ISM and IWM are co-registered in proxies in the same laminated core with no bioturbation, thereby permitting an explicit evaluation of the relative timing of the two monsoons.

Other marine records from the Arabian Sea also suggest a gradual decrease in ISM strength since ~5 ka BP (Gupta et al., 2003; Overpeck et al., 1996). Cullen et al. (2000) observed an abrupt peak in aeolian dolomite and calcite in marine sediments in the Gulf of Oman from 4.0-3.6 ka BP, and Ponton et al. (2012) also showed a shift to weaker ISM after 4.0 ka BP in the Bay of Bengal, based on $\delta^{13}C$ of leaf waxes. Marine IWM reconstructions are not particularly coherent: although Doose-Rolinski et al. (2001) find a decrease in evaporation and weakening of the ISM between 4.6 and 3.7 ka BP, they argue this was accompanied by a relative increase in IWM strength. Giosan et al. (2018, in review) inferred enhanced winter monsoon conditions from 4.5-3.0 ka BP based on a planktic paleo-DNA and % *Globigerina falconensis* record close to our coring site ("C" in Figure 1), which contradicts our finding of decreased upper ocean mixing after 4.3 ka BP. We suggest the chronological resolution of core 63KA paired with a multi-species foraminifer $\delta^{18}O$ record can provide more detail about the timing of changes in IWM and ISM strength.

5.4 Comparison to regional terrestrial records

The 63KA $\delta^{18}O$ record obtained from three foraminifer species highlights several important ocean-atmosphere changes over the 5.4-3.0 ka BP time period. First, a sharp decrease occurred in both summer and winter precipitation at 4.1 ka BP, which is within a broader 300-year period of increased aridity spanning both rainfall seasons between 4.2 and 3.9 ka BP. In detail, we infer a relative decrease in Indus River discharge and weakened ISM between 4.8 and 3.9 ka BP, peaking at 4.1 ka BP, while a 200-year phase of strong IWM interrupted this period from 4.5-4.3 ka BP. Furthermore, the stepped change in $\delta^{18}O$ of *N. dutertrei* suggests an enduring change in ocean-atmosphere conditions after 4.1 ka BP.

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A relatively abrupt ~4.2 ka BP climate event has been observed in several terrestrial records on the Indian subcontinent, most notably Mawmluh Cave (~4.1-3.9 ka BP) in northeastern India (Berkelhammer et al., 2012) and Kotla Dahar (~4.1 ka BP) in northwestern India (Dixit et al., 2014) (Figure 9). A less abrupt yet still arid period is documented in a peat profile (~4.0-3.5 ka BP) from northcentral India (Phadtare, 2000), at Lonar Lake (~4.6-3.9 ka BP) in central India (Menzel et al., 2014), and at Rara Lake (~4.2-3.7 ka BP) in western Nepal (Nakamura et al., 2016). Finally, a recent study of oxygen and hydrogen isotopes in gypsum hydration water from Karsandi on the northern margin of the Thar Desert showed wet conditions between 5.1 and 4.4 ka BP, after which the playa lake dried out sometime between 4.4 and 3.2 ka BP (Dixit et al., 2018). Considering terrestrial records can record more local climatic conditions than marine records, it is remarkable that the records collectively agree on a regional phase of aridity between 4.2 and 3.9 ka BP within the uncertainties of the age models that vary considerably among records.

However, not all records support this finding, such as a reconstruction from Sahiya Cave in northwestern India that shows an abrupt decrease in $\delta^{18}O$ interpreted to reflect an increase in monsoon strength from ~4.3-4.15 ka BP, followed by an arid trend after 4.15 ka BP (Kathayat et al., 2017). In addition, several other Thar Desert records do not identify a "4.2 ka BP event" *sensu stricto*, but instead suggest that lakes dried out several centuries earlier (Deotare et al., 2004; Enzel et al., 1999; Singh et al., 1990) or later (Sinha et al., 2006) than 4.2 ka BP. This discrepancy may relate to non-linear climate responses of lakes, which would not record a drought at 4.2 ka BP if they had already dried out earlier from the ongoing decrease in summer rainfall. In addition, there are also significant concerns about chronological uncertainties when using radiocarbon of bulk sediment for dating in some of these records. It is also possible that variations in the timing of climate change inferred from the terrestrial records may be real, reflecting different sensitivity to ISM and IWM rain. As a marine record, core 63KA integrates large-scale ocean-atmosphere changes, and therefore can help inform the interpretation of the more locally sensitive terrestrial records.

More distantly, several terrestrial records in the Middle East also show a decrease in winter precipitation proxies around 4.2 ka BP: Jeita Cave in Lebanon records a relatively dry period between 4.4 and 3.9 ka BP (Cheng et al., 2015) and Soreq Cave in Israel shows a period of increased aridity starting at ~4.3 ka BP (Bar-Matthews et al., 2003; Bar-Matthews and Ayalon, 2011) (Figure 10). Lake Van in eastern Turkey also records reduced spring rainfall and enhanced aridity after ~4.0 ka BP (Wick et al., 2003; Lemcke and Sturm, 1997). All records suggest a relatively arid phase in winter precipitation after ~4.3 ka BP, as inferred from core 63 KA. Qunf Cave in Oman (Fleitmann et al., 2003), which is outside the range of IWM influence, instead shows a steady mid-Holocene weakening of the ISM that closely follows trends in summer solar insolation. Core 63 KA also infers a protracted decrease in ISM since ~4.8 ka BP.

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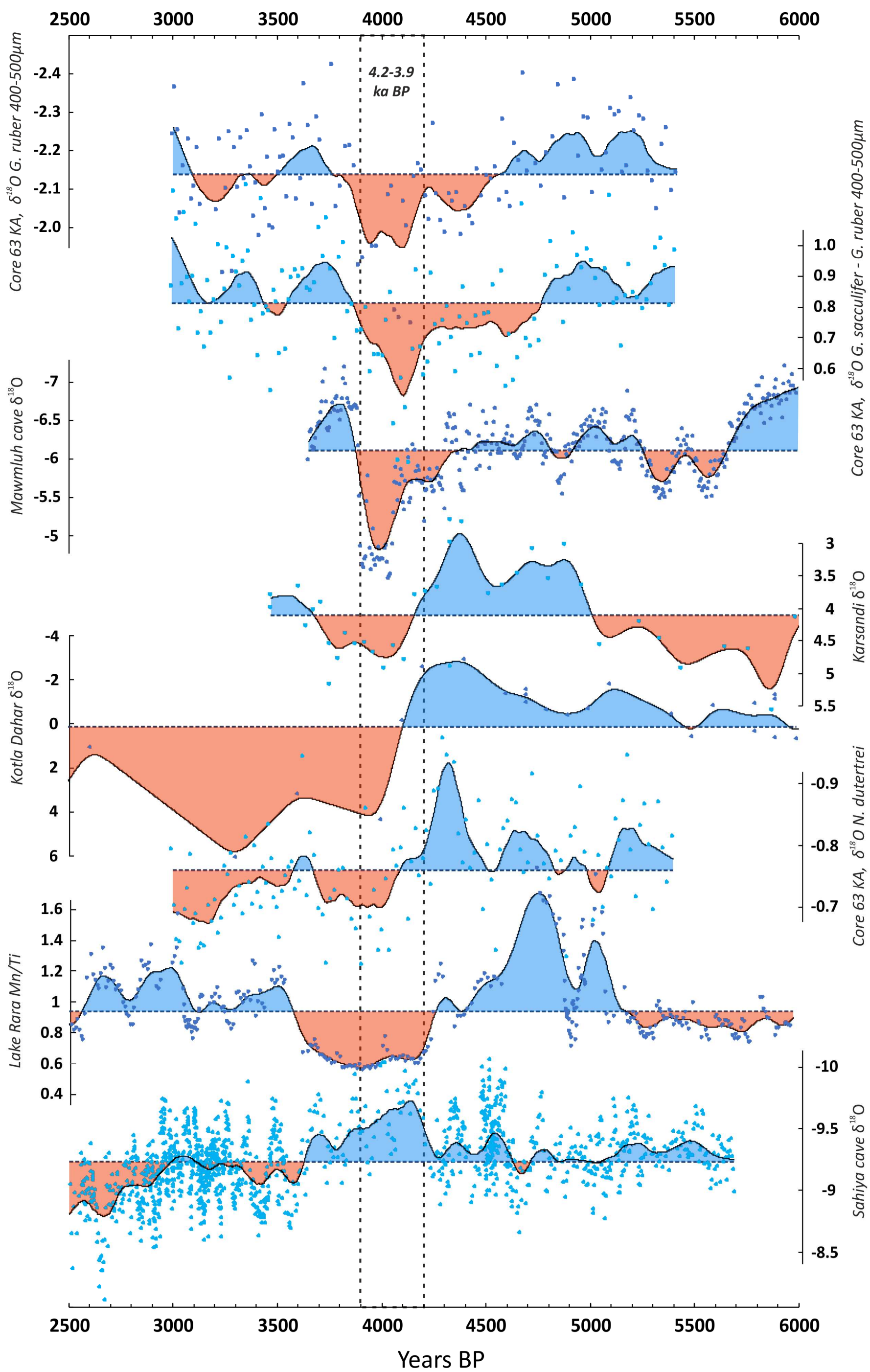

634
635 **Figure 9.** Comparison of the $\delta^{18}O$ record of core 63KA with terrestrial records from the Indian
636 Subcontinent, from top to bottom: this study (first two), Berkelhammer et al., 2012; Dixit et al., 2018;
637 Dixit et al., 2014; this study; Nakamura et al., 2016; Kathayat et al., 2017.

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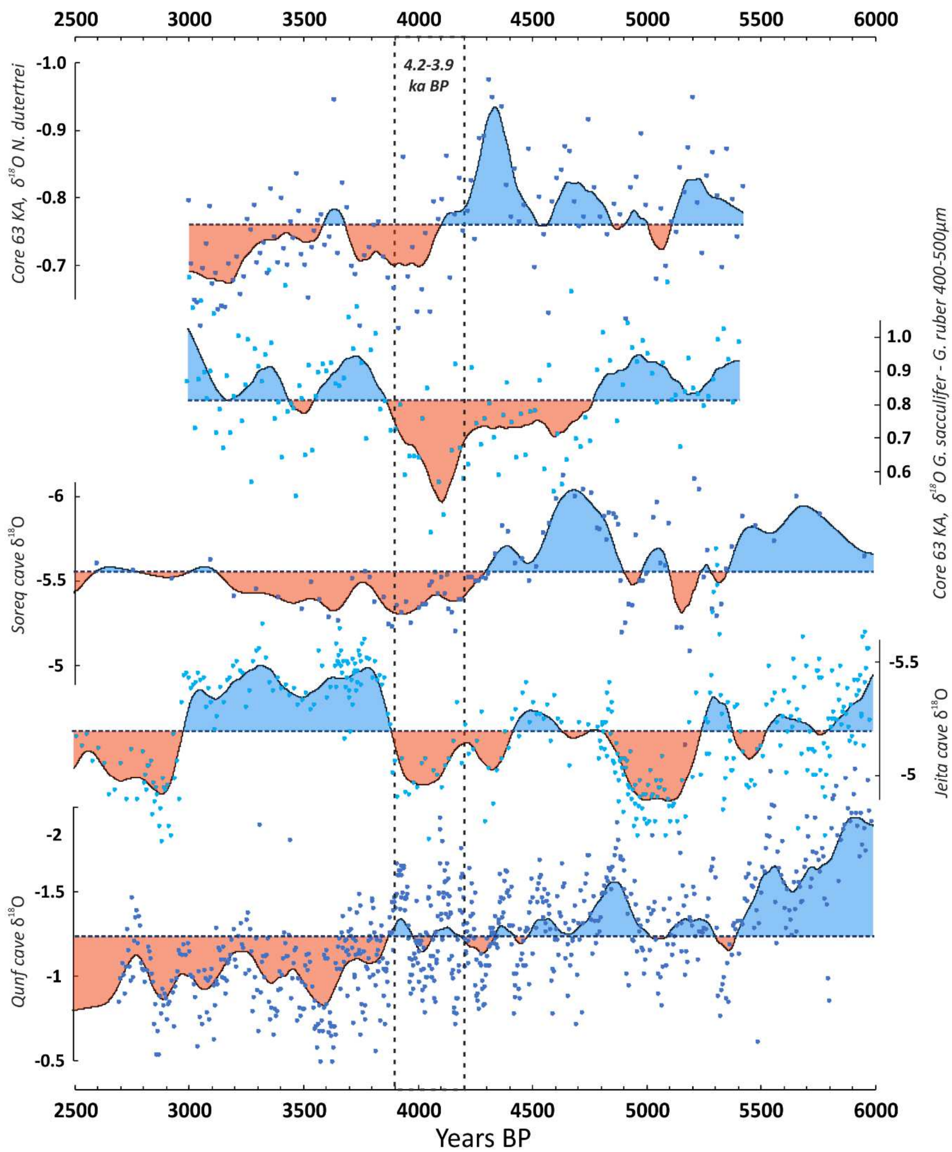

Figure 10. Comparison of the $\delta^{18}O$ record of core 63KA (topmost records) with more distant records, from top to bottom: Bar-Matthews et al., 2003; Cheng et al., 2015; and Fleitmann et al., 2003. Mean value for each record is taken for all available data between 6.0-2.5 ka BP.

5.5 Cultural impacts

Based on our reconstruction of reduced IWM mixing after 4.3 ka BP, accompanied by decreased freshwater discharge of the Indus River, it is worth considering what impacts could be expected from a reduction in IWM and ISM precipitation. A weakened IWM overlying a

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reduced or more variable ISM would likely result in a distinct climate signal over the Indus River catchment, with broad implications for seasonal river flow and water availability throughout the year. The presence of the two rainfall systems creates a complex and diverse range of ecologies across northwest South Asia (Petrie et al., 2017). In a situation when rainfall in both seasons is reduced over extended periods, step-shifts in the natural environment may occur that are difficult to reverse (desertification, lake desiccation, regional vegetation changes, decline in overbank flooding and shift in river avulsion patterns).

Societies reliant on IWM, ISM, or a combination of the two would have been vulnerable to years with monsoon failure, and a shift affecting both seasons will have challenged resilience and tested sustainability (Green and Bates et al. in prep.; Petrie et al., 2017). Archaeological research into the transition from the urban Mature Harappan phase (~4.6-3.9 ka BP) to the post-urban Late Harappan phase (~3.9-3.6 ka BP) notes progressive deurbanization through the abandonment of large Indus cities and a depopulation of the most western Indus regions, concurrent with a general trend towards an increase of concentrations of rural settlements in some areas of the eastern Indus extent (Green and Petrie, 2018; Petrie et al., 2017; Possehl, 1997) (Figure S6). The relatively limited range of well-resolved available archaeobotanical data suggests that there was a degree of diversity in crop choice and farming strategies in different parts of the Indus Civilization across this time span (Petrie et al., 2016; Petrie and Bates, 2017; Weber, 1999; Weber et al., 2010). Farmers in southerly regions appear to have focused on summer or winter crops, while the more northern regions of Pakistan Punjab and Indian Punjab and Haryana were capable of supporting combinations of winter and summer crops (Petrie and Bates, 2017). Although there is evidence for diverse cropping practices involving both summer and winter crops in the northern areas during the urban period, agricultural strategies appear to favor more drought-resistant summer crops in the Late Harappan period (Madella and Fuller, 2006; Petrie and Bates, 2017; Pokharia et al., 2017; Weber, 2003; Wright, 2010). It has previously been suggested that weakened ISM was a major factor in these shifts (e.g. Giosan et al., 2012; Madella and Fuller, 2006). On the basis of our reconstruction of decreased IWM in northwest South Asia after 4.3 ka BP with a step-shift at 4.1 ka BP, we suggest that both IWM and ISM climatic factors played a role in the redistribution of population to smaller settlements in eastern regions with more direct summer rain, as well as the observed shift to more summer crop dominated cropping strategies.

6. Conclusion

This study expanded on the $\delta^{18}O$ record of planktonic foraminifer in core 63KA of the northeastern Arabian Sea, originally published by Staubwasser et al. (2003). Using $\delta^{18}O$ of the surface-dwelling foraminifera *G. ruber*, the original study inferred an abrupt reduction in Indus River discharge at ~4.2 ka BP. Our further $\delta^{18}O$ analysis of a larger size fraction of this species confirmed maximum salinity at 4.1 and 3.95 ka BP. In addition, the $\delta^{18}O$ difference between the surface-dwelling *G. ruber* and slightly deeper-dwelling *G. sacculifer* ($\Delta\delta^{18}O_{s-r}$) reveals that surface waters were more saline than average for the period from 4.8-3.9 ka BP. By also measuring a thermocline-dwelling planktonic foraminiferal species, *N. dutertrei*, we infer an increase in the strength of the IWM between 4.5 and 4.3 ka BP, followed by reduction in IWM-driven mixing that peaks at 4.1 ka BP.

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Assuming that weaker IWM mixing implies a reduction in IWM rainfall amount or duration over northwest South Asia under past climatic conditions, the 63KA core is used to infer important changes in seasonal hydrology of the Indus River catchment. We propose a combined weakening of the IWM and ISM at 4.1 ka BP led to what has been termed the "4.2 ka BP" drought over northwest South Asia. The intersection of both a gradually weakening ISM since 4.8 ka BP and a maximum decrease in IWM strength at 4.1 ka BP resulted in a spatially layered and heterogeneous drought over a seasonal to annual timescale. Regions in the western part of the Indus River basin accustomed to relying mainly on winter rainfall (also via river run-off) would have been most severely affected by such changes. Regions in the northeastern and eastern extents benefitted more from summer rainfall, and would have been less severely affected, particularly as the ISM appears to recover strength by 3.9 ka BP.

Relatively strengthened IWM surface water mixing between 4.5 and 4.3 ka BP correlates with a period of higher precipitation recorded at Karsandi on the northern margin of the Thar Desert (Dixit et al., 2018), an area within the summer rainfall zone that is sensitive to small changes in winter precipitation. This time span also represents the beginnings of the Mature Harappan phase (Possehl, 2002; Wright, 2010), which implies that increasingly urbanized settlements may have flourished under a strengthened IWM. With a weakening of the IWM at ~4.1 ka, eastern regions with more access to ISM rainfall may have been more favorable locations for agriculture. This may also help explain the broad shift in population towards more rural settlements in the northeastern extent of the Indus Civilization that occurred by ~3.9 ka BP (Possehl, 1997; Petrie et al., 2017), and a shift to more drought-tolerant kharif (summer) season crops in Gujarat (Pokharia et al., 2017) and at Harappa (Madella and Fuller, 2006; Weber, 2003).

Given the importance of the relationships between humans and the environment during the time of the Indus Civilization, understanding the impact of the IWM on precipitation variability in northwest South Asia remains a critical area of research. We especially need a better understanding of the wind patterns and moisture pathways that controlled the IWM in the past. Disentangling both the length and intensity of seasonal precipitation is a crucial aspect of understanding the impact of climate change on past societies, particularly in a diverse region relying on mixed water sources (e.g., fluvial, ground aquifer, direct rainfall).

Data availability

Data presented in the paper can be accessed by contacting the corresponding author at ag927@cam.ac.uk. After final acceptance of the manuscript, the data will also be uploaded to an online database.

Author contributions

M.S. supplied core 63KA material, A.G. prepared the material for isotopic measurements, and A.G. and D.A.H. interpreted the results. A.G., D.A.H., and C.A.P. wrote the manuscript.

Competing interests

The authors declare that they have no conflict of interest.

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