

**Interactive comment on “Re-examining the 4.2 ka BP event in foraminifer isotope records from the Indus River delta in the Arabian Sea” by Alena Giesche et al.**

**Anonymous Referee #2**

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Giesche, et al. “Re-examining the 4.2 ka BP event in foraminifer isotope records from the Indus River delta in the Arabian Sea

*Summary:*

*A precise characterization of the timing, duration and climatic expression of the 4.2 ka event in the Indian subcontinent is crucial in order to assess its potential linkage with the transformation of the Indus valley civilization. Giesche et al. provide new foraminiferal isotope data from the northeast Arabian Sea that spans a critical interval of time (~ 3.3 to 5.3 ka), an interval of considerable interest from both climatic and societal points of view. The new oxygen isotope data of surface-and-thermocline-dwelling foraminiferal species is (ultimately) interpreted by the authors in the terms of temporal variations in the strength of Indian summer and winter monsoons, respectively. The results (and proxy interpretations) while complements the results from Staubwasser et al (2003) study, I think this manuscript provides an improved (and less-speculative) discussion of plausible climatic and societal linkages. The manuscript is fairly well-written (but on the verbose side), logically-structured, but the graphical presentation of figures requires some attention. I recommend this manuscript be accepted for publication in *Climate of the Past* after some revisions.*

We thank reviewer #2 for taking the time to provide a review with helpful comments on our manuscript.

*Specific Comments: I think the title “Re-examining the 4.2 ka. . .Arabian Sea” would seem to suggest that the focus of this manuscript lies in the “re-examining” part but in reality, it constitutes a rather small component of this manuscript. While generally complementary, I think there are also important differences in the results between the two studies (see figure 3b), which may have important bearing in assessing the timing and abruptness of the 4.2 event. I encourage authors to find a title that is more representative of the main conclusions/results.*

Both reviewers 2&3 have addressed this point in their comments. We agree with this assessment, and have proposed a new title for the revised manuscript: **“Indian winter and summer monsoon strength over the 4.2 ka BP event in foraminifer isotope records from the Indus River delta in the Arabian Sea”**

*Text and other information on Figure 1 are hard to follow. A different color scheme would also help.*

We have edited Figure 1 to increase font size where possible, as well as increased the contrast between colors of the text and background. These changes can be viewed in the revised version of the manuscript, at the end of this file.

*Figures with scatter plots (with the exception of Figure # 3b) are also not particularly helpful. Why not show these figures like figure S2?*

We have added lines between data points in Figures 3/5/7, which have been consolidated into a single figure (now called Figure 3) following the comments of Reviewer #1. We have also added lines between data points in Figures 7 & 8 (previously called 9 & 10).

*There is also a fair bit of redundancy between sections 2, 5.1, 5.2, and 5.3. I think the manuscript will benefit from a more concise presentation.*

We have now reduced redundancies by only mentioning elements of site description in section 2. For example, we removed a paragraph at the start of section 5.3, and removed or edited several sentences in section 5.1.

*Line 134-35: Consider rephrasing “northeastern regions close to New Delhi”*

Done.

*Line 141: “while many paleoclimate studies...” Provide some reference here.*

Done.

*Line 180: Fleitmann et al 2007 is a tertiary reference. Remove it.*

Done.

*Line 280-281: Provide a statement that explicitly compares the age models between Staubwasser 2003 and this study. I think there is no difference in age models between the two studies but it needs to be spelled out here. Also, indicate the temporal resolution of the new dataset.*

A new age model was made using updated (Bayesian) modeling software and updated radiocarbon calibrations (IntCal13), but there is no major difference between the age models except during the period 13-11 ka BP, which is out of the range of interest for this paper (a visual comparison of age models can be seen in supplemental figure S5b). The temporal resolution of the interval of interest (18 years/cm) has now been noted in two places - Section 2.2 and 3.1.

*Line 333: “The bandwidth of 210 years was considered optimal time window for capturing the overall trends in the dataset”. What constitutes optimal? Why 210 years? It seems like a large window to smooth data to discern the onset and termination of the 4.2 event.*

We trialed several smoothing windows for the data, which can be viewed in Supplemental figure S2. We agree that “optimal” is subjective and rephrased it to “reasonable”. 210 years were picked for several reasons: 1) It was the original smoothing window applied to the dataset in Staubwasser et al., 2003. 2) This represents the solar Gleissberg cycle of ~210 years, and helps smooth out any influence of the century-scale solar cycles if present. 3) Considering the relatively low signal to noise ratio of the record, we wanted to ensure that we were not exaggerating the significance of only a few data points – 210 years incorporates ~12 cm of the record, and the resulting smoothing shows the most obvious trends in the data considering the variability. However, as we have now added lines between the raw data points in figures 3, 7 and 8, the reader is better able to assess the data independently of its smoothing.

*Line 351-352: Please quantify “weak” and “strong” correlation and provide statistical significance.*

We no longer report the correlation for smoothed *G. ruber* data (possible autocorrelation), but have included the p value for the Pearson's correlation for the raw data: weak positive correlation ( $R = 0.25$ ,  $p < 0.01$ ,  $n = 109$ , slope 0.26, intercept -1.36).

*Line 415 and 421 : consider rephrasing "trend" with "change" or something similar.*

Done.

*Line 432-433: Support this by a suitable reference and briefly explain why "differencing" will improve the signal-to-noise-ratio.*

We understand how this phrasing might be misleading, because differencing will not reduce the errors associated with each individual measurement. However, differencing can help emphasize signals of interest. If we wish to isolate the freshwater plume salinity signal, differences may remove the varying influence of other factors such as water temperature (without requiring an actual reconstruction of the temperature). While temperature changes would affect the surface dwellers *G. sacculifer* and *G. ruber* equally, the surface freshwater plume would primarily affect *G. ruber*. The advantages of differencing are discussed in more detail by Tian et al. (2005) and Ravelo and Shackleton (1995). We have rephrased this sentence in the manuscript to better reflect the intended meaning.

*Lines 556-565: Some of this should be in the "Introduction" and not here.*

We have removed this paragraph entirely from the section as the same information is provided in Section 2.2.

References mentioned in this response:

Ravelo, A.C., and Shackleton, N.J.: Evidence for surface-water circulation changes at Site 851 in the eastern Tropical Pacific Ocean, in: Proceedings of the Ocean Drilling Program, Scientific Results, College Station, TX (Ocean Drilling Program), Pisias, N. G.; Mayer, L. A.; Janecek, T. R.; Palmer-Julson, A.; van Andel, T. H. (Eds.), 138, 503-514, doi: 10.2973/odp.proc.sr.138.126, 1995.

Tian, J., Wang, P., Chen, R., and Cheng, X.: Quaternary upper ocean thermal gradient variations in the South China Sea: Implications for east Asian monsoon climate, *Paleoceanography*, 20, 2005.

Staubwasser, M., Sirocko, F., Grootes, P. M., and Segl, M.: Climate change at the 4.2 ka BP termination of the Indus valley civilization and Holocene south Asian monsoon variability, *Geophysical Research Letters*, 30, 2003.

Revised manuscript below.

1 **Re-examining Indian winter and summer monsoon strength over the 4.2 ka BP event in**  
2 **foraminifer isotope records from the Indus River delta in the Arabian Sea**

3  
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52 **Abstract**

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54 The plains of northwest South Asia receive rainfall during both the Indian Summer (June-September)  
55 and Winter (December-March) Monsoon. Researchers have long attempted to deconstruct the  
56 influence of ~~both these~~ precipitation regimes in paleoclimate records, in order to better understand  
57 regional climatic drivers and their potential impact on human populations. The Mid-Late Holocene  
58 transition between 5.3-3.3 ka BP is of particular interest in this region because it spans the period of  
59 the Indus Civilization from its early development, ~~through its urbanization and on-~~to eventual  
60 transformation ~~into a rural society~~. ~~The An~~ oxygen isotope record of the surface-dwelling planktonic  
61 foraminifer *Globigerinoides ruber* from the northeast Arabian Sea provided evidence for an abrupt  
62 decrease in rainfall and reduction in Indus River discharge at 4.2 ka BP, which the authors linked to  
63 the decline of the urban phase of the Indus Civilization (Staubwasser et al., 2003). Given the  
64 importance of this study, we used the same core (63KA) to ~~replicate the oxygen isotope profiles of a~~  
65 ~~larger size fraction of G. ruber than measured previously and, in addition, we measured~~ measure the  
66 oxygen isotope profiles of two other foraminifer species at decadal resolution over the interval from  
67 5.4 to 3.0 ka BP, ~~and replicate a larger size fraction of G. ruber than measured previously~~. By selecting  
68 both thermocline-dwelling (*Neogloboquadrina dutertrei*) and shallow-dwelling (*Globigerinoides*  
69 *sacculifer*) species, we provide enhanced detail of the climatic changes that occurred over this crucial  
70 time interval. We found evidence for a period of increased surface water mixing, which we suggest  
71 was related to a strengthened winter monsoon with a peak intensity over 200 years from 4.5 to 4.3  
72 ka BP. The time of greatest change occurred at 4.1 ka BP when both the summer and winter monsoon  
73 weakened, resulting in a reduction in rainfall in the Indus region. The earliest phase of the urban  
74 Mature Harappan period coincided with the period of inferred stronger winter monsoon between 4.5-  
75 4.3 ka BP, whereas the end of the urbanized phase ~~followed~~ occurred some time after the decrease  
76 in both the summer and winter monsoon strength by 4.1 ka BP. Our findings provide evidence that  
77 the initial growth of large Indus urban centers ~~was coincident~~ coincided with increased winter rainfall,  
78 whereas the contraction of urbanism and change in subsistence strategies followed a reduction in  
79 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 44<sup>th</sup> 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 both winter and summer hydroclimate over the Indian Subcontinent.

The  $\delta^{18}\text{O}$  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 urban phase 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). This is because the winter rain zone partially overlaps with the summer rain zone (Figure 1), and provides a critical supply of rain and snowfall for the Indus River basin. However, we currently understand much less about the behavior of the IWM than the ISM.

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-locations in the central Thar desert receive as little as 100 mm yr<sup>-1</sup>, which is only about 10% of the amount of direct annual rainfall ~~seen in the northeastern region close~~ compared 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 (references A-C, I, K-M, S, and U in Figure 1) 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

148 contributions and timing of seasonal rainfall from both the winter and summer monsoons.  
149 Previously, it has not been possible to reliably differentiate winter and summer rain in  
150 reconstructions from the Indus region.

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152 In this study, we re-examined the same marine core (63KA) used in the original [research of](#)  
153 Staubwasser et al. (2003) ~~paper~~. We first assessed the reproducibility of the *Globigerinoides*  
154 *ruber*  $\delta^{18}\text{O}$  record using a larger size fraction of the same species for the time period 5.4-3.0  
155 ka BP. We also measured the  $\delta^{18}\text{O}$  of two additional foraminifer species, *G. sacculifer*  
156 (*Globigerinoides sacculifer*) and *N. dutertrei* (*Neogloboquadrina dutertrei*), which live deeper  
157 than *G. ruber* in the water column. The different ecologies of these ~~two~~ three species provide  
158 additional information with which to evaluate the multiple  $\delta^{18}\text{O}$  records and assess seasonal  
159 changes in the paleoceanography of the northeastern Arabian Sea near the mouth of the  
160 Indus River.

161  
162 The  $\delta^{18}\text{O}$  of foraminifera has been widely applied as an indicator of temperature and salinity  
163 changes (Duplessy et al., 1992; Maslin et al., 1995; Wang et al., 1995; Rohling, 2000; [among](#)  
164 [others](#)). Measuring the  $\delta^{18}\text{O}$  of species calcifying at different depths can provide further  
165 information about upper ocean seasonal hydrography such as surface water mixing, depth of  
166 the thermocline, and upwelling (Ravelo and Shackleton, 1995). ~~Such~~ Similar methods have  
167 been applied by several other studies (Billups et al., 1999; Cannariato and Ravelo, 1997;  
168 Norris, 1998; Steinke et al., 2010; Steph et al., 2009; among others), including a reconstruction  
169 of East Asian Winter Monsoon strength in the South China Sea (Tian et al., 2005). ~~We~~ Here  
170 we ~~applied~~ a ~~similar-comparable~~ method to samples from core 63KA in the northeastern  
171 Arabian Sea because surface waters at this location are influenced by freshwater discharge  
172 from the Indus River and direct precipitation during the summer monsoon months, whereas  
173 enhanced upper ocean mixing occurs during the winter monsoon. We hypothesized that our  
174 new measurements of  $\delta^{18}\text{O}$  of *G. sacculifer* and *N. dutertrei* would allow us to track changes  
175 in upper ocean mixing. Weaker IWM winds are expected to result in a shorter duration and/or  
176 less intense upper ocean mixing, although how this signal is ultimately related to the amount  
177 or distribution of winter rainfall in the Indus River catchment has not been demonstrated  
178 conclusively. Dimri (2006) studied Western Disturbances for the time period 1958-1997, and  
179 noted that surplus years of surplus winter precipitation are linked to significant heat loss over  
180 the northern Arabian Sea, which is mainly attributed to intensified westerly moisture flow  
181 and enhanced evaporation. Such conditions would promote deeper winter mixing, and  
182 provide a basis for relating thermocline depth with IWM intensity. By comparing the  $\delta^{18}\text{O}$  of  
183 multiple species of foraminifera we seek to infer variations in the relative strengths of the  
184 summer and winter monsoons, and by comparing the 63KA record to other nearby marine  
185 and terrestrial records we evaluate the potential role that climate played in cultural  
186 transformation of the Indus Civilization.

## 187 188 **2. Site Description**

### 189 190 *2.1 Monsoon – land-based processes*

191  
192 Today, most of the annual precipitation over northwest South Asia stems from the ISM, and  
193 occurs mainly between June and September. The pressure gradient between the low-  
194 pressure Tibetan Plateau and high-pressure Indian Ocean is accompanied by the ITCZ

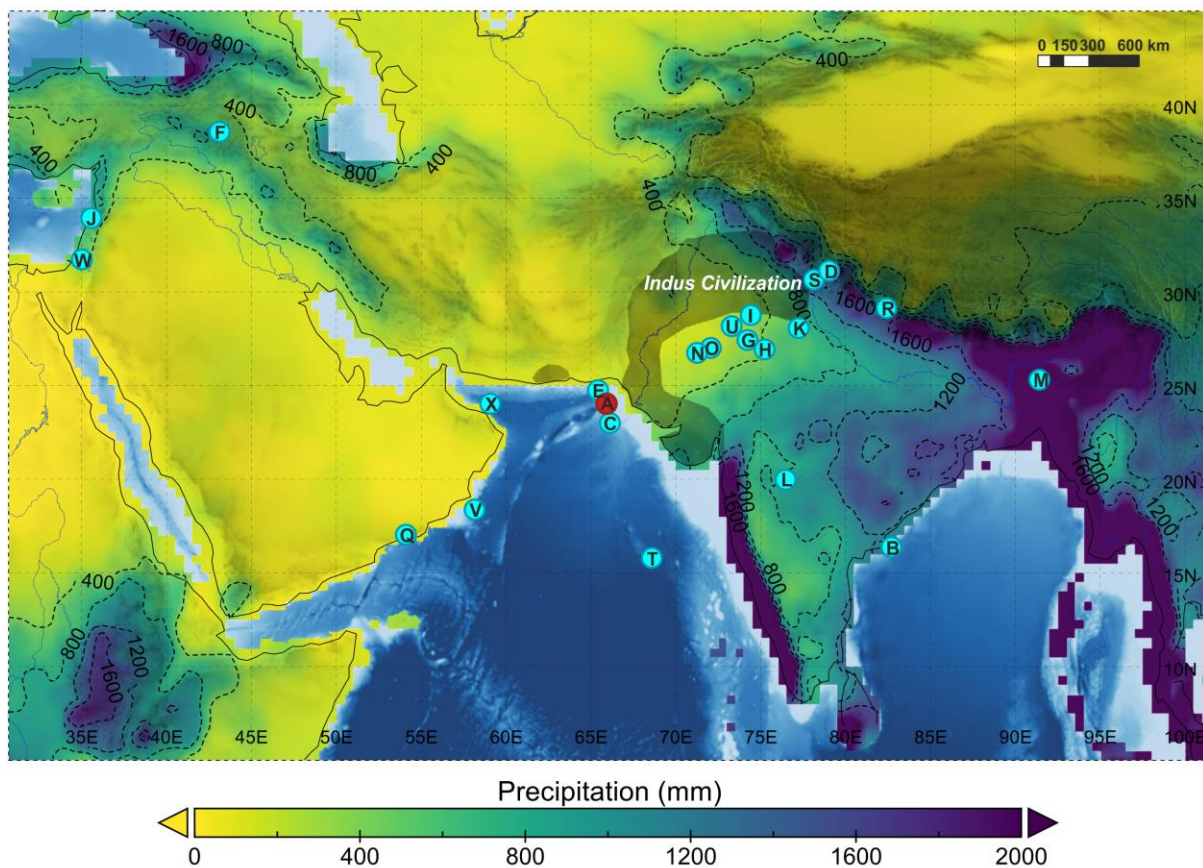
195 (Intertropical Convergence Zone) reaching its northward maximum in summer, which draws  
196 in moisture over the subcontinent via southwesterly winds from the Indian Ocean ([Fleitmann  
197 et al., 2007](#); Gadgil, 2003). The summer rainfall gradient increases from the central Thar  
198 Desert (as little as 100 mm direct summer rainfall per year) to the Himalaya mountains in the  
199 north (>1000 mm) and the Aravalli range to the west (>500 mm) (Figure 1b).

200  
201 The IWM rain falls between December through March, and is mainly the result of atmospheric  
202 Western Disturbances (Dimri and Dash, 2012; Yadav et al., 2012) originating over the  
203 Mediterranean and Black Sea (Hatwar et al., 2005) that allow for moisture incursion from the  
204 Arabian Sea (Rangachary and Bandyopadhyay, 1987). During the IWM, the pressure gradient  
205 is reversed from the summer condition, allowing the passage of Western Disturbances when  
206 the ITCZ moves southward. As winter transitions to spring, predominantly northeasterly  
207 winds shift to westerly winds (Sirocko, 1991) that result in peak winter rainfall over the plains  
208 of northwest India in February and March. [Anomalously cool, evaporative conditions over the  
209 northern Arabian Sea \(promoting deeper winter mixing\) also correlates with increased winter  
210 precipitation in the western Himalayas \(Dimri, 2006\)](#). The winter rainfall gradient increases  
211 from the southern Thar Desert (<10 mm per year) up to the Himalayas in the northwest (>400  
212 mm) (Figure 1c). Overall, the IWM contributes between roughly 10 to 50% of the total annual  
213 rainfall of northwest South Asia today.

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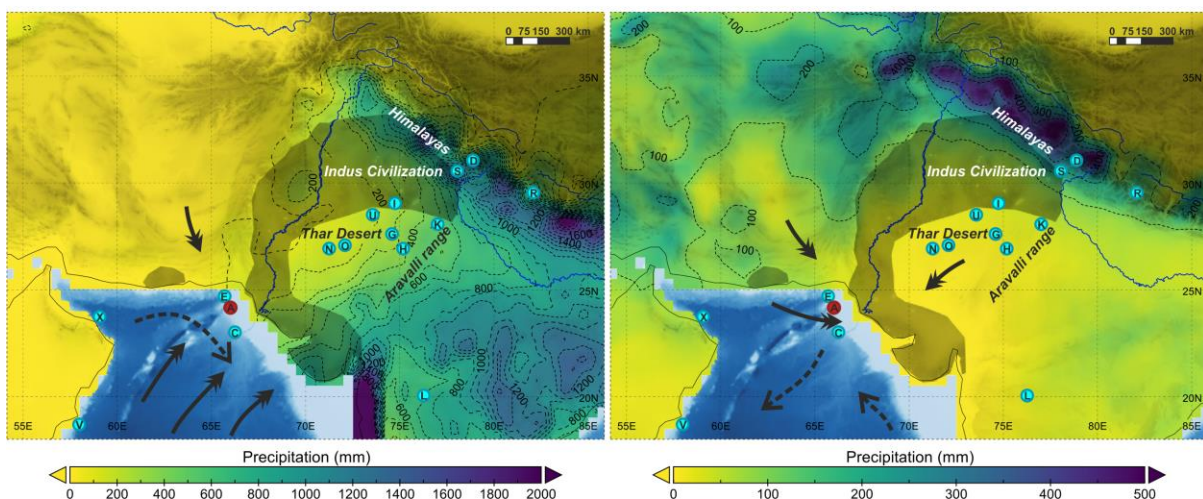


(a) Longterm mean (1981-2010) annual precipitation



(b) Longterm mean (1981-2010) summer precipitation

(c) Longterm mean (1981-2010) winter precipitation



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**Figure 1. a.** Annual **b.** ISM (JJAS) **c.** IWM (DJFM) mean precipitation (1981-2010) isohyets taken from the GPCC V7 global gridded dataset ( $0.5^\circ \times 0.5^\circ$  resolution) (Schneider et al., 2015); 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 (Weatherall et al., 2015), 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:

- 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 – (Dose-Rolinski 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)

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224 The Indus and the other rivers that make up Punjab are partly fed by winter snow and ice melt  
 225 from their upper mountain catchment areas. Melting peaks during the summer months  
 226 around July-August (Yu et al., 2013), which coincides with the peak of ISM rainfall, and Indus  
 227 River discharge reaches its maximum during August (Karim ~~et al.,~~ and Veizer, 2002). The  
 228 proportion of winter to summer precipitation contributing to the Indus River is not entirely  
 229 clear, although one study has estimated a 64-72% contribution of winter precipitation from  
 230 the deuterium excess of Indus River water (Karim ~~et al.,~~ and Veizer, 2002), whereas a previous  
 231 study estimated a lower 15-44% contribution of snowmelt to Indus tributaries (Ramasastrri,  
 232 1999). Since the 1960s, the Indus River has seen ~~a~~ more than a 50% reduction in discharge  
 233 because of the construction of barrages as well as the diversion of water for agricultural uses  
 234 (Ahmad et al., 2001).

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## 236 2.2 Hydrography – core site and ocean-based processes

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238 Core 63KA was obtained by the PAKOMIN cruise in 1993 (von Rad et al., 1995). The laminated  
 239 core from the northeastern Arabian Sea (24° 37' N, 65° 59' E) was taken at 316 m water depth  
 240 on the continental shelf, ~100 km west of the Indus River delta. The core has high  
 241 sedimentation rates (equivalent to a temporal resolution of around 18 years/cm in the period  
 242 of interest, 5.4-3.0 ka BP), and all foraminifer proxies were produced from the same laminated  
 243 core with no bioturbation. An important aspect of core 63KA is that different components of  
 244 the monsoon system are co-registered in the same sediment core, thereby permitting an  
 245 explicit evaluation of the relative timing of different parts of the climate system (e.g., ISM and  
 246 IWM).

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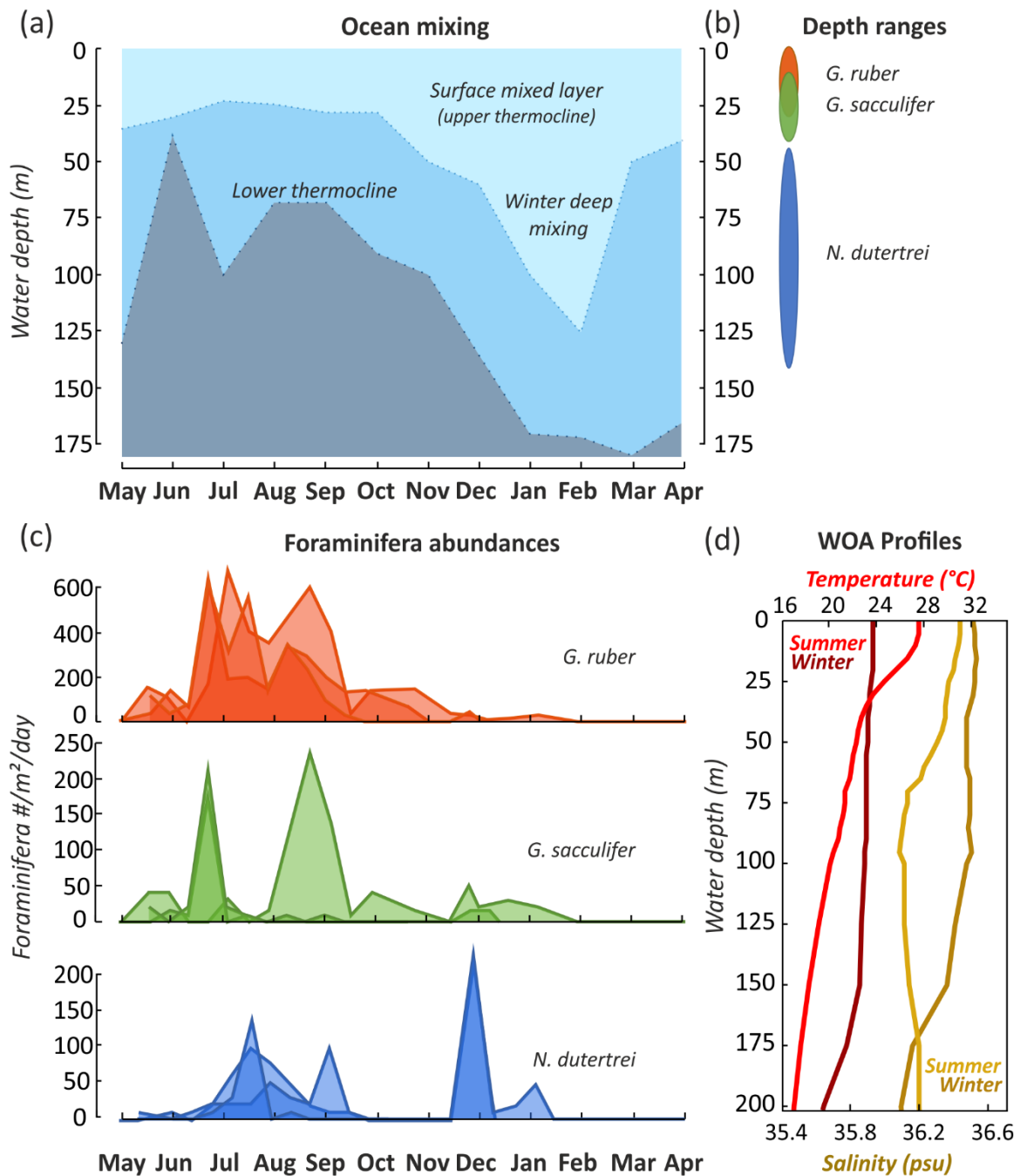
248 Modern hydrographic conditions in the northeastern Arabian Sea are highly influenced by the  
 249 seasonal monsoon. During summertime, highest sea surface temperatures (SSTs) are  
 250 observed along with a shallow mixed layer depth <25 m (Schulz et al., 2002) (Figure 2a). A low  
 251 salinity plume surrounds the Indus River delta and shoreline extending as far as the coring  
 252 location (Supplemental Figure S1). The reverse occurs in winter when the lowest SSTs are  
 253 accompanied by surface water mixing to >125 m, resulting in warming of the deeper waters  
 254 (Schulz et al., 2002). Northeasterly winds promote convection in the northeastern Arabian  
 255 Sea by cooling and evaporation of surface water (Banse, 1984; Madhupratap et al., 1996),  
 256 and during the transition from winter to spring, wind directions shift from northeasterly to  
 257 westerly (Sirocko, 1991), promoting a period of upwelling in the northeastern Arabian Sea  
 258 (Staubwasser et al., 2002; Rao, 1981).

259

260 The northern Arabian Sea is dominated by highly saline (up to 37 psu) surface waters ~~of~~  
 261 the known as Arabian Sea High Salinity Water ~~Mass~~ (ASHSW), which extends s from the surface

262 ~~down~~ to 100 m depth (Joseph and Freeland, 2005). ~~The~~is high salinity ~~can be~~is explained  
263 by the high evaporative rates over this region. ASHSW forms in the winter, but is prevented  
264 from reaching our coring site on the shelf by northerly subsurface currents until the summer  
265 (Kumar and Prasad, 1999). Along coastal areas, the ASHSW is starkly contrasted by the fresh  
266 water discharge of the Indus River, combined with direct precipitation. In contrast, surface  
267 waters in the Bay of Bengal on the eastern side of India have much lower surface water  
268 salinity, because of overall higher precipitation and stronger stratification from weaker winds  
269 (Shenoi et al., 2002). The heightened evaporative conditions and highly saline surface waters  
270 of the northeastern Arabian Sea make it a sensitive study location to observe changes in  
271 discharge of the entire Indus River catchment area – ultimately tracking changes in monsoon  
272 strength. Unlike individual terrestrial records, which may be affected by local climatic  
273 processes, the marine record from core 63KA is more likely to integrate regional changes of  
274 the large-scale ocean-atmosphere system.

275  
276 Planktonic foraminifera complete their life cycle within a few weeks (Bé and Hutson, 1977).  
277 Peak abundances indicate the time of year when each species tends to calcify, thereby  
278 recording the  $\delta^{18}\text{O}$  and temperature of the seawater in their  $\text{CaCO}_3$  shells primarily during  
279 certain seasons. Foraminifer abundances in the eastern Arabian Sea have been studied by  
280 Curry et al. (1992) using sediment traps deployed at shallow (~1400 m) and deep (~2800 m)  
281 water depths (“T” in Figure 1a). ~~Peak abundances for~~ *G. ruber* and *G. sacculifer* ~~occur~~ have  
282 peak abundances during the summer months (June-September), whereas *N. dutertrei* ~~peak~~  
283 lives mainly during the winter ~~as well as~~ and has a secondary peak in ~~with a secondary peak~~  
284 in summer months (Figure 2c). Preferred depth ranges for each species reflect their ecological  
285 niches, including requirements for nutrients and tolerance for ranges of temperature and  
286 salinity (Bé and Hutson, 1977; Hemleben et al., 2012). *G. ruber* lives in the upper surface  
287 waters (0-10 m), *G. sacculifer* is found in slightly deeper surface waters (10-40 m), and *N.*  
288 *dutertrei* inhabits the base of the mixed layer near the thermocline (40-140 m) (estimates  
289 based on ranges from Farmer et al. (2007) and the local CTD profiles) (Figure 2d).  
290



291  
 292 **Figure 2.** a. Seasonal surface water mixing depth based on station EPT-2 located nearby the coring site  
 293 of 63KA (adapted from Schulz et al., 2002 who also used data from Hastenrath and Lamb, 1979) b.  
 294 Foraminifer depth ranges based on CTD profile c. Foraminifer abundances from EAST traps  
 295 (overlapping peaks indicate data from multiple traps): *G. ruber* (orange), *G. sacculifer* (green), and *N.*  
 296 *dutertrei* (blue) (adapted from Curry et al., 1992 using Zaric, 2005) d. CTD-World Ocean Atlas (WOA)  
 297 mean (1955-2012) temperature (red) and salinity (yellow) profiles at 24.875°N, 65.875°E from  
 298 station 11 at coring location, shown for summer (JAS) and winter (JFM) seasons (Locarnini et al., 2013;  
 299 Zweng et al., 2013) taken September 1993 (von Rad, 2013).

300

### 301 3. Materials and Methods

302

#### 303 3.1 Age model

304

305 The radiocarbon dates from Staubwasser et al. (2002, 2003) were obtained from 80 samples  
306 of mainly the foraminifer *G. sacculifer* and three samples of *O. universa*. In the interval of  
307 interest (5.4-3.0 ka BP), there are 15 radiocarbon dates with a 95% confidence range of 30-  
308 130 years. The average sample resolution is 18 years/cm. Bayesian age modelling software,  
309 BACON v2.3.3 (Blaauw and Christen, 2011), was used as an R-package to update the age  
310 model of core 63KA. No major difference exists between the old and new age models, except  
311 for the period 13-11 ka BP (Supplemental Figure S5, Table S2). IntCal13 was used for  
312 radiocarbon calibration (Reimer et al., 2013) with marine reservoir ages provided by  
313 Staubwasser et al. (2002, 2003).

314

### 315 3.2 Stable isotope analysis

316

317 Oxygen and carbon isotopes were measured on three species of foraminifera selected from  
318 washed samples at 1-cm intervals throughout 132 cm of the core covering 5.4-3.0 ka BP: *G.*  
319 *ruber* (white, *sensu stricto*), *G. sacculifer*, and *N. dutertrei*. For *G. ruber*, 12 ± 8 foraminifera  
320 were picked from the 400-500µm size fraction with an average weight of 21.4 ± 2.5µg. The  
321 400-500µm size fraction was picked because too few specimens remained in the size fraction  
322 315-400µm used by Staubwasser et al. (2003). For *G. sacculifer*, 34 ± 7 foraminifera were  
323 picked from the 315-400µm size fraction with an average weight of 21.9 ± 2.6µg. For *N.*  
324 *dutertrei*, 34 ± 4 foraminifera were picked from the 315-400µm size fraction with an average  
325 weight of 25.9 ± 2.2µg. At some depth levels in the core there were insufficient foraminifera  
326 for measurement, along with outlier measurements in two cases, leaving 11-14 gaps in the *G.*  
327 *ruber* 400-500µm record, 3-4 gaps in the *G. sacculifer* record, and no gaps for *N. dutertrei*. The  
328 published *G. ruber* is from the 315-400µm size fraction and contains 17 gaps in the depth  
329 range examined (Staubwasser et al., 2003).

330

331 All foraminifera were weighed, crushed, and dried at 50° C. Samples were cleaned for 30  
332 minutes with 3% H<sub>2</sub>O<sub>2</sub>, followed by a few drops of acetone, ultrasonication, and drying  
333 overnight. –Where sample weights exceeded 80µg, oxygen and carbon isotopes were  
334 measured using a Micromass Multicarb Sample Preparation System attached to a VG SIRA  
335 Mass Spectrometer. In cases of smaller sample sizes, the Thermo Scientific Kiel device  
336 attached to a Thermo Scientific MAT253 Mass Spectrometer was used in dual inlet mode. This  
337 method adds 100% H<sub>3</sub>PO<sub>4</sub> to the CaCO<sub>3</sub>, water is removed cryogenically, and the dry CO<sub>2</sub> is  
338 analyzed isotopically by comparison with a laboratory reference gas. For both measurement  
339 methods, 10 reference carbonates and 2 control samples were included with every 30  
340 samples. Results are reported relative to VPDB, and internal precision/long-term  
341 reproducibility of laboratory standards (e.g., Carrara marble) is better than ±0.08‰ for δ<sup>18</sup>O  
342 and ±0.06‰ for δ<sup>13</sup>C. External precisions/Reproducibility of foraminiferal measurements were  
343 was estimated by five triplicate (three separately picked) measurements of *G. ruber* (400-  
344 500µm) that yielded one standard deviation of ±0.12‰ (δ<sup>18</sup>O) and ±0.10‰ (δ<sup>13</sup>C). For *G.*  
345 *sacculifer* (315-400µm) the standard deviation of eight triplicate measurements were/was  
346 ±0.07‰ (δ<sup>18</sup>O) and ±0.07‰ (δ<sup>13</sup>C), and for *N. dutertrei* (315-400µm) the standard deviation  
347 of nine triplicate measurements was ±0.06‰ (δ<sup>18</sup>O) and ±0.07‰ (δ<sup>13</sup>C).

348

349 To calculate equilibrium values of δ<sup>18</sup>O<sub>calcite(PDB)</sub>, we used the CTD profile from station 11  
350 (24.62° N, 66.07° E) (Figure 2d) taken in September 1993 during PAKOMIN *Sonne* cruise no.  
351 90 (von Rad, 2013), which is nearly identical to the location of core 63KA (24.62° N 65.98° E).

352 The  $\delta^{18}\text{O}_{\text{water(SMOW)}}$  was calculated from salinity following Dahl and Oppo (2006), and  
353  $\delta^{18}\text{O}_{\text{calcite(SMOW)}}$  was further calculated using the calcite-water equation of Kim and O’Neil  
354 (1997). We also used the equation of Shackleton (1974) as a comparative method for  
355 calculating  $\delta^{18}\text{O}_{\text{calcite(PDB)}}$ .

356

### 357 3.3 Statistical treatment

358

359

360 Statistical tests were applied to the raw data from the  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  time series, including the  
361 package SiZer (Chaudhuri and Marron, 1999; Sonderegger et al., 2009) in R software (2016)  
362 that calculates whether the derivative of a time series exhibits significant changes given a  
363 range of timespans. A Pearson’s correlation test (confidence level 95%) was done on paired  
364 samples from both size fractions of *G. ruber*. We also conducted a Welch’s t-test to determine  
365 if the mean population of  $\delta^{18}\text{O}$  is significantly different before and after 4.1 ka BP.~~We applied~~  
366 ~~a range of smoothing windows (bandwidth of 20–500 years) to assess the significance of~~  
367 ~~changes in the isotope records throughout the time series.~~

368

369 As in the original data of Staubwasser et al. (2003), the oxygen isotope results show great  
370 variability and distinguishing long-term trends in these data requires benefits from statistical  
371 smoothing for visualization purposes. ~~To reduce the variance in the data and identify~~  
372 ~~trends.~~~~After completing all statistical tests and performing the differences on the raw data~~  
373 ~~(132 depths), the  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  data from 5.4–3.0 ka BP were first resampled to constant 1-~~  
374 ~~year intervals using linear interpolation. A~~ loess (locally weighted) smoothing function was  
375 ~~then~~ applied to the  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  data from 5.4–3.0 ka BP data, using a 210-year moving  
376 window as described by Staubwasser et al. (2003). Loess smoothing uses weighted least  
377 squares, which places more importance on the data points closest to the center of the  
378 smoothing interval. The bandwidth of 210 years was considered an optimal-reasonable time  
379 window for capturing the overall trends in the dataset (other time windows are shown for  
380 comparison in Supplemental Figure S2).

381

382 ~~Where step changes occur in  $\delta^{18}\text{O}$  we also conducted a Student’s t test to determine if the~~  
383 ~~mean population of  $\delta^{18}\text{O}$  is significantly different before and after the change.~~

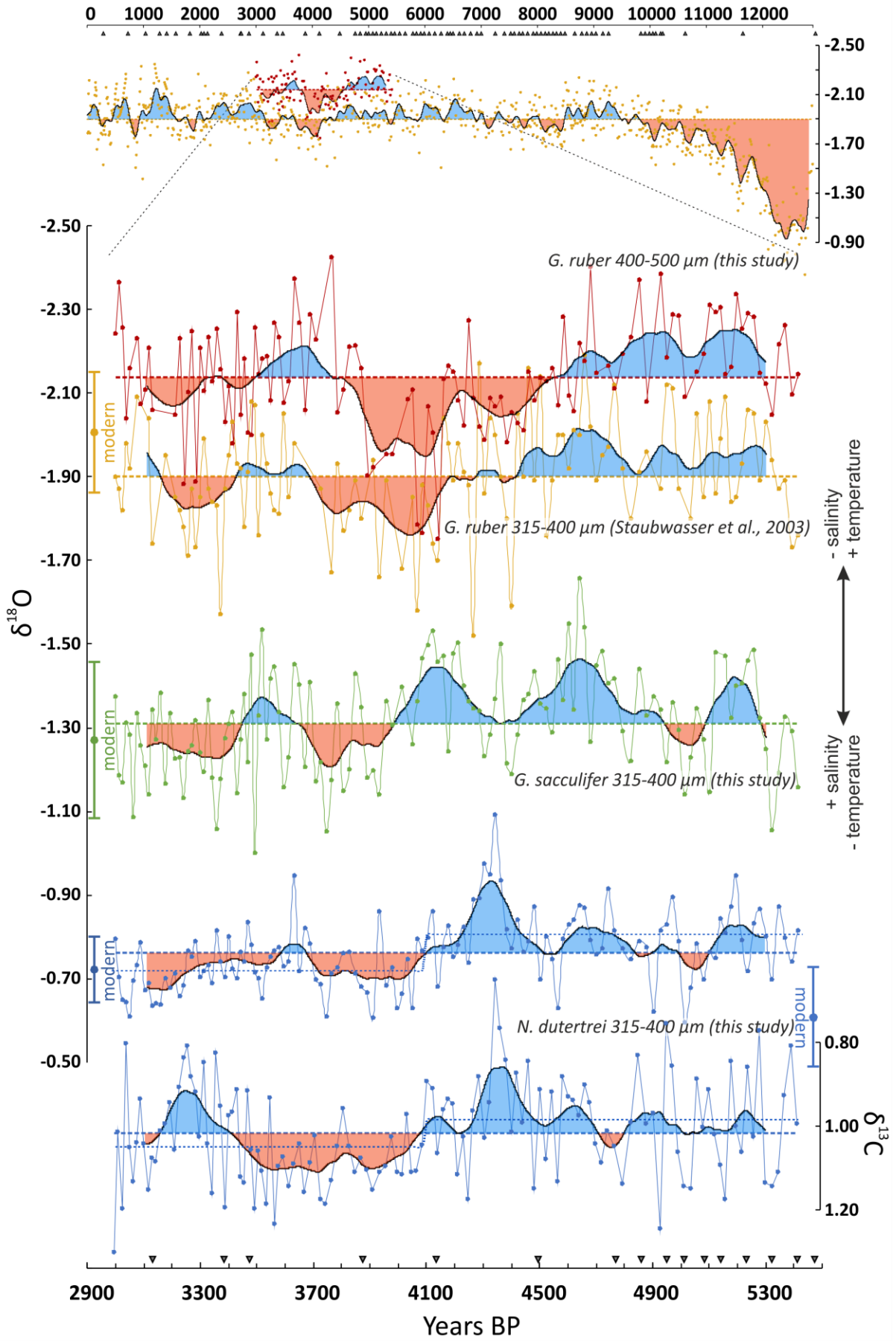
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## 385 4. Results

386

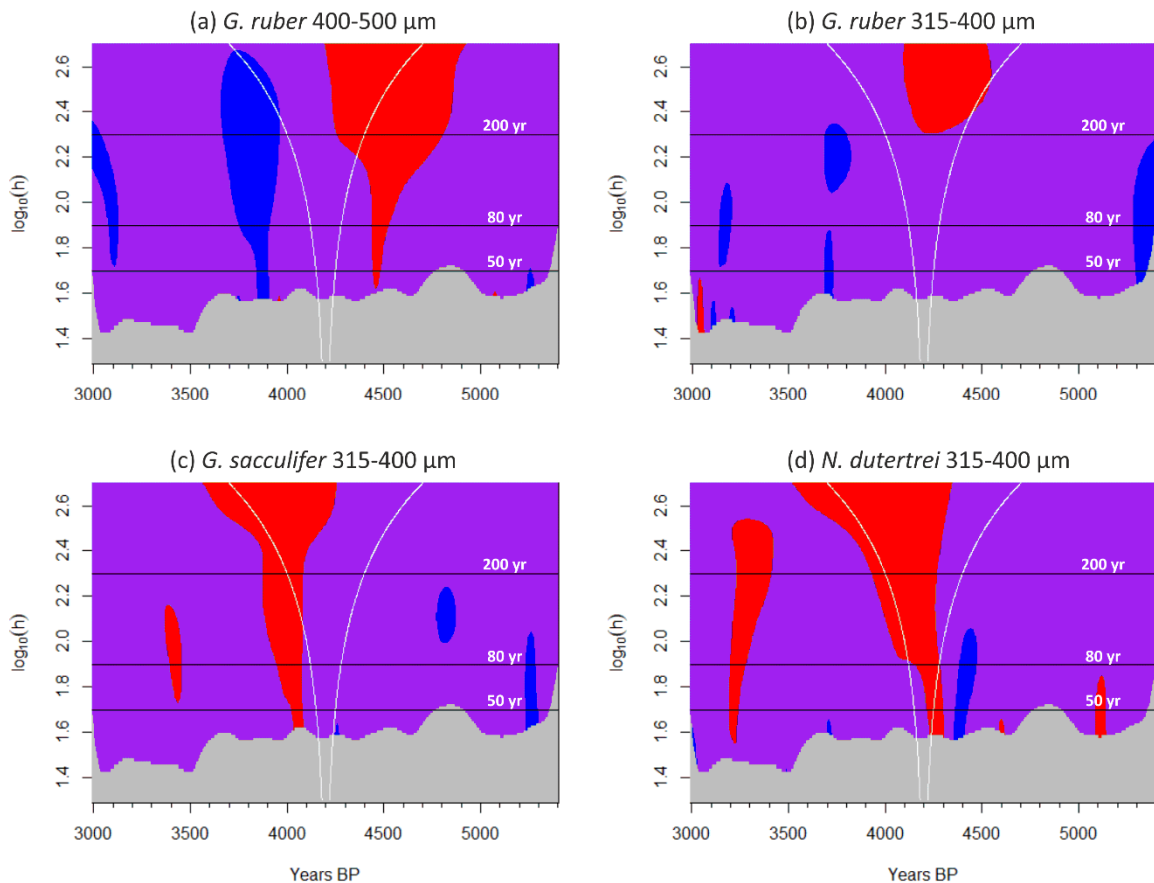
387 The new  $\delta^{18}\text{O}$  measurements of *G. ruber* (400–500 $\mu\text{m}$ ) parallel the published record of *G.*  
388 *ruber* (315–400 $\mu\text{m}$ ) (Staubwasser et al., 2003), but the  $\delta^{18}\text{O}$  of the specimens from the larger  
389 size fraction is offset by -0.23‰ on average (Figure 3). The ~~two~~ records from two size  
390 fractions, produced in different laboratories by different investigators, display a weak positive  
391 correlation for the raw data (R = 0.25, p < 0.01, n = 109, slope 0.2726, intercept -1.3336), and  
392 the 210-year smoothed records reveal good agreement in the overall trends of the data and  
393 ~~a strong correlation for the 210-year smoothed records (R = 0.7, slope 0.53, intercept -0.77).~~  
394 When comparing the two *G. ruber* records, it is apparent that the increasing trend in  $\delta^{18}\text{O}$   
395 starts well before ~4.2 ka BP – perhaps as early as ~4.9 ka BP. This trend is also observed with  
396 the SiZer analysis, which identifies a significant increase in  $\delta^{18}\text{O}$  anywhere from 4.9 to 4.2 ka  
397 BP depending on which smoothing window is selected (Figure 4). The new  $\delta^{18}\text{O}$  record of *G.*  
398 *ruber* (400–500 $\mu\text{m}$ ) shows additional detail after the ~4.2 ka BP event – i.e. specifically, a

399 double-peak maximum occurring at 4.1 and 3.95 ka BP that is related to seven discrete  
400 measurements with high  $\delta^{18}\text{O}$  values. These maxima are offset from the average  $\delta^{18}\text{O}$  value  
401 by +0.18‰ (smoothed average), or up to +0.38‰ when considering the maximum individual  
402 measurement at 4.1 ka BP. The offsets from the average values exceed one standard  
403 deviation of the entire record from 5.4-3.0 ka BP, which is 0.13‰. Although *G. ruber* shows  
404 an event at 4.1 ka BP, it does not show a permanent step change: A Student's-Welch's t-test  
405 comparing the means of pre- and post-4.1 ka BP indicates that the +0.07‰ shift in mean  $\delta^{18}\text{O}$   
406 values of *G. ruber* (315-400 $\mu\text{m}$ ) is statistically significant (t value = 2.9,  $p < 0.01$ ,  $n = 115$ ), ~~and~~  
407 but the +0.0403‰ shift in mean  $\delta^{18}\text{O}$  values of *G. ruber* (400-500 $\mu\text{m}$ ) is weakly-not significant  
408 (t value = 1.75,  $p < 0.12$ ,  $n = 118$ ).  
409





411 **Figure 3.** Core 63KA  $\delta^{18}\text{O}$  *G. ruber* from two size fractions: 400-500 $\mu\text{m}$  (red) (this study), 315-400 $\mu\text{m}$   
 412 (orange) (Staubwasser et al., 2003), shown in the context of the original record and also zoomed in  
 413 over 5.4-3.0 ka BP.  $\delta^{18}\text{O}$  of *G. sacculifer* 315-400 $\mu\text{m}$  (green), and  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  of *N. dutertrei* 315-  
 414 400 $\mu\text{m}$  (blue) are shown over the interval 5.4-3.0 ka BP. Data are shown with a 210-year loess  
 415 smoothing, and modern surface values  $\pm 1\sigma$  are plotted for comparison and  $\pm 1\sigma$  error bars. Mean  
 416 values for all species are denoted by the dotted line, and the pre- and post-4.1 ka BP mean values are  
 417 indicated by an additional dotted line for *N. dutertrei*. Individual AMS radiocarbon dates are denoted  
 418 by triangles near the timeline.  
 419



420 **Figure 4.** SiZer 1<sup>st</sup> derivative analysis (Chaudhuri and Marron, 1999; Sonderegger et al., 2009) applied  
 421 to  $\delta^{18}\text{O}$  of a. *G. ruber* 400-500 $\mu\text{m}$ , b. *G. ruber* 315-400 $\mu\text{m}$ , c. *G. sacculifer* 315-400 $\mu\text{m}$ , d. *N. dutertrei*  
 422 315-400 $\mu\text{m}$ . The red areas indicate statistically significant increases in  $\delta^{18}\text{O}$ , the blue represent  
 423 decreases, and the purple no significant change. Black horizontal lines are the smoothing bandwidths  
 424 ( $h = 50, 80,$  and  $200$  years). The distance between the white lines denotes the change in smoothing  
 425 bandwidth scaled to the x-axis.  
 426

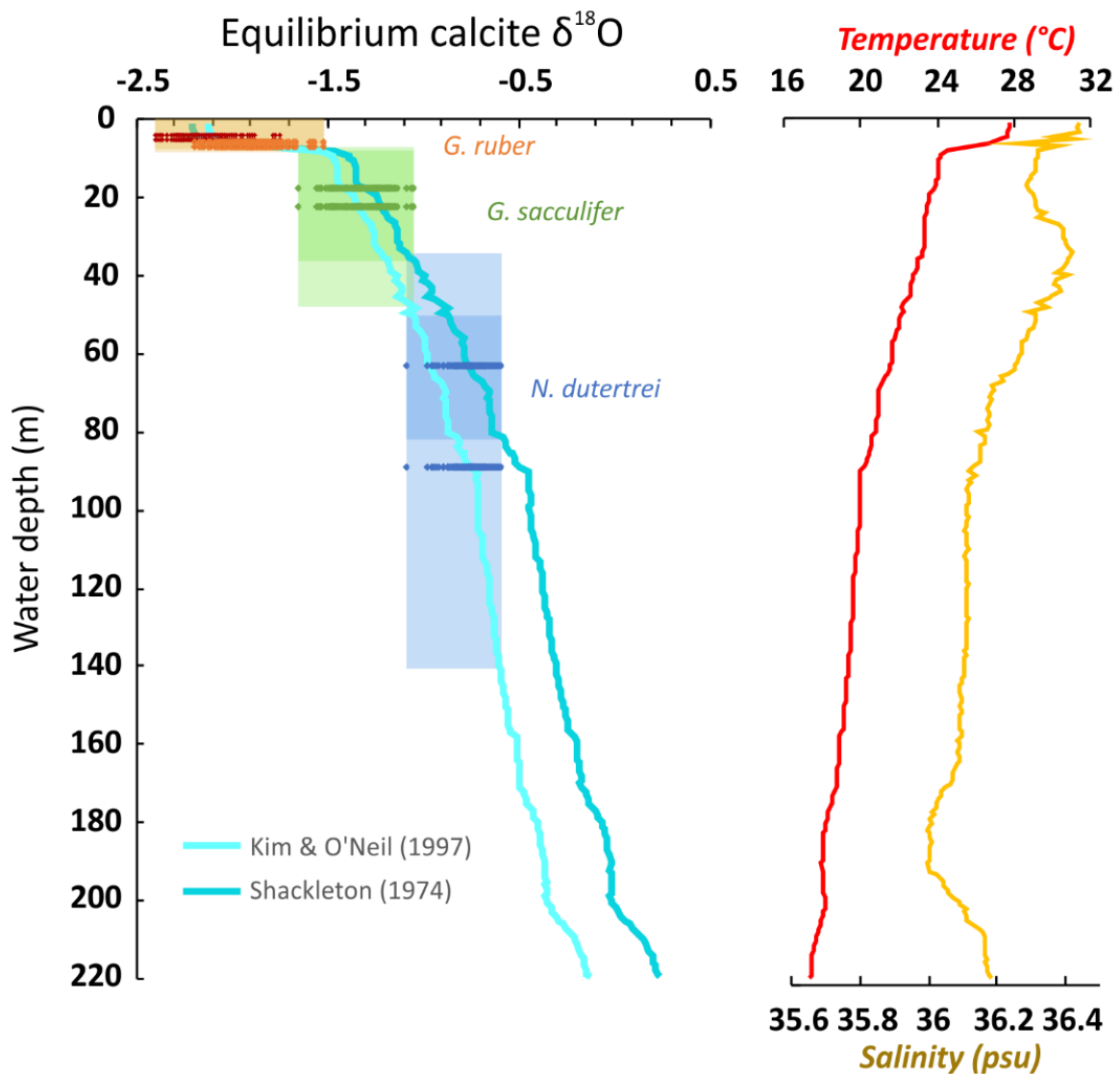
427  
 428 The relative differences in  $\delta^{18}\text{O}$  of the planktonic species studied (*G. ruber*, *G. sacculifer* and  
 429 *N. dutertrei*) reflect the temperature and salinity of their habitat in the water column:  $\delta^{18}\text{O}$   
 430 *G. ruber* <  $\delta^{18}\text{O}$  *G. sacculifer* <  $\delta^{18}\text{O}$  *N. dutertrei* (Figure 35). *G. sacculifer* is offset from *G. ruber*  
 431 (315-400 $\mu\text{m}$ ) by approximately +0.57‰, whereas *N. dutertrei* is offset by +1.14‰. The larger  
 432 size fraction of *G. ruber* (400-500 $\mu\text{m}$ ) is offset from *G. ruber* (315-400 $\mu\text{m}$ ) by -0.23‰. The  
 433 offsets among species are maintained throughout the entire record (Figure 35). We also  
 434 measured  $\delta^{18}\text{O}$  values near the top of the core (approximately the last 200 years) for all three  
 435 species in the 315-400 $\mu\text{m}$  size fraction, which continue to show the same offsets

436 (Supplemental Figure S3). The  $\delta^{18}\text{O}$  of *G. ruber* shows the greatest variance and *N. dutertrei*  
 437 shows the least (Supplemental Figure S4, Table S1).

438

439 Equilibrium calcite calculations based on the salinity and temperature measurements from  
 440 the September 1993 CTD profile of station 11 of the PAKOMIN Cruise (von Rad, 2013) show  
 441 the expected depth habitats of the three foraminifer species (Figure 65). *G. ruber* is generally  
 442 found at 0-30 m, *G. sacculifer* at 15-40 m, and *N. dutertrei* at 60-150 m (Farmer et al., 2007).  
 443 Using the CTD profile from our core location, we compare these depth ranges with the  
 444 measured  $\delta^{18}\text{O}$  values. The calculated depths ranges agree well with those expected on the  
 445 basis of other studies, placing *G. ruber* in the upper 10 m, *G. sacculifer* 10-40 m, and *N.*  
 446 *dutertrei* 40-140 m.

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448

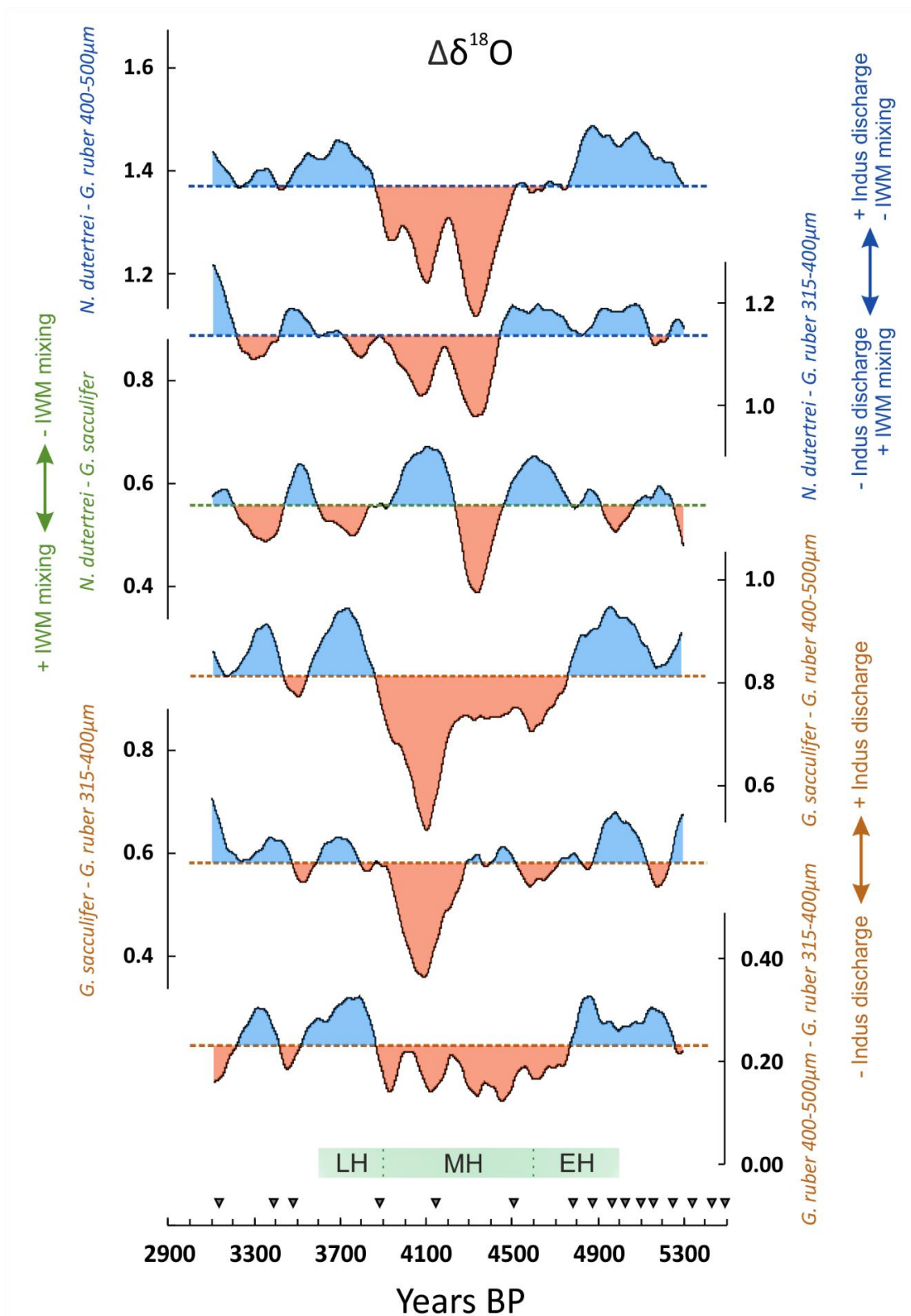
449 **Figure 65.**  $\delta^{18}\text{O}$  of equilibrium calcite (left) calculated from the CTD temperature and salinity profile  
 450 at station 11 (von Rad, 2013) (right) with projected depth ranges of *G. ruber* 400-500 $\mu\text{m}$  (red), *G. ruber*  
 451 315-400 $\mu\text{m}$  (orange), *G. sacculifer* 315-400 $\mu\text{m}$  (green), *N. dutertrei* 315-400 $\mu\text{m}$  (blue). We show  
 452 estimated values using both the original paleotemperature equation of Shackleton (1974) (dark teal),  
 453 and Kim & O'Neil (1997) (turquoise). Horizontal ranges show the measured  $\delta^{18}\text{O}$  values of each species  
 454 between 5.4-3.0 ka BP.

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~~The most obvious trend in the *G. sacculifer*  $\delta^{18}\text{O}$  is the increase~~ around 4.1 ka BP. ~~A, and a Student's-Welch's~~ t-test comparing the means of pre- and post-4.1 ka BP indicates that the +0.0708‰ shift in mean  $\delta^{18}\text{O}$  values is statistically significant (t value = 3.98,  $p < 0.01$ ,  $n = 128$ ). 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 44).

~~Likewise, the dominant trend-change~~ in the  $\delta^{18}\text{O}$  of *N. dutertrei* is a mean increase at 4.1 ka BP (Figure 37). SiZer analysis also identifies a significant decrease in  $\delta^{18}\text{O}$  occurring mainly between 4.45 and 4.35 ka BP, followed by a significant increase between 4.3 and 4.1 ka BP (Figure 44). A ~~Student's-Welch's~~ t-test comparing the means of pre- and post-4.1 ka BP indicates that the +0.08‰ shift in mean  $\delta^{18}\text{O}$  values is statistically significant (t value = 6.32,  $p < 0.01$ ,  $n = 132$ ), along with the +0.07‰ shift in mean  $\delta^{13}\text{C}$  (t value = 3.3,  $p < 0.01$ ,  $n = 132$ ).

Differencing  $\delta^{18}\text{O}$  of foraminifera (expressed as  $\Delta\delta^{18}\text{O}$ ) in the same sample can ~~better sometimes improve the signal-to-noise ratio~~ emphasize signals of interest (Figure 68). The  $\Delta\delta^{18}\text{O}$  of *G. ruber* 400-500 $\mu\text{m}$  and *G. ruber* 315-400 $\mu\text{m}$  size fractions shows increasing similarity between ~4.8 and 3.9 ka BP during the period of overall higher  $\delta^{18}\text{O}$ . The  $\Delta\delta^{18}\text{O}$  of *N. dutertrei* and both size fractions of *G. ruber*, designated  $\Delta\delta^{18}\text{O}_{\text{d-r}}$ , 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  $\Delta\delta^{18}\text{O}$  of *G. sacculifer* and both size fractions of *G. ruber* ( $\Delta\delta^{18}\text{O}_{\text{s-r}}$ ) 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  $\Delta\delta^{18}\text{O}$  of *N. dutertrei* and *G. sacculifer* ( $\Delta\delta^{18}\text{O}_{\text{d-s}}$ ) 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 86.** Core 63KA  $\Delta\delta^{18}\text{O}$  ~~Data~~ shown with a 210-year loess smoothing. Individual AMS radiocarbon dates are denoted by triangles near the timeline. *G. ruber* 315-400 $\mu\text{m}$  size fraction data come 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) periods, respectively.

## 489 5. Discussion

490

### 491 5.1 Interpretation of foraminifer $\delta^{18}\text{O}$

492

493 The trends in the original  $\delta^{18}\text{O}$  record of *G. ruber* (315-400 $\mu\text{m}$ ) by Staubwasser et al. (2003) is  
494 confirmed-reflected by our independent  $\delta^{18}\text{O}$  measurements of *G. ruber* in a larger size  
495 fraction (400-500 $\mu\text{m}$ ), although an important difference exists suggesting a decrease in  
496 freshwater discharge as early as 4.8 ka BP. The larger size fraction is offset by approximately  
497 -0.2‰, which is similar to the size-related fractionation of -0.3‰ per +100 $\mu\text{m}$  for *G. ruber*  
498 reported by Cayre and Bassinot (1998), and could be attributed to size-related vital effects.  
499 Alternatively, part of the offset might be explained by interlaboratory calibration considering  
500 the data were produced using two different methods and mass spectrometers.

501

502 The observed 4.1 ka BP maximum in  $\delta^{18}\text{O}$  of *G. ruber*, living near the surface during summer  
503 months, could be attributed to either decreased SST or increased surface water salinity  
504 (Bemis et al., 1998). Staubwasser et al. (2003) acknowledged that a decrease in SST could  
505 cause the increase in  $\delta^{18}\text{O}$  in the *G. ruber* record, but argued that this explanation is unlikely  
506 because a *G. ruber*  $\delta^{18}\text{O}$  record from core M5-422 in the northwestern Arabian Sea shows  
507 opposing trends over the same time period (Cullen et al., 2000), and a local alkenone SST  
508 proxy record shows relatively higher temperatures in the same period (Doose-Rolinski et al.,  
509 2001). If the ~0.2‰ (relative to mean) increase in  $\delta^{18}\text{O}$  of *G. ruber* at 4.1 ka BP was caused by  
510 temperature change rather than salinity, a ~1° C cooling of surface water is implied would be  
511 required (Kim and O'Neil, 1997).

512

513 Following Staubwasser et al. (2003), we interpret the  $\delta^{18}\text{O}$  variations of *G. ruber* to be  
514 predominantly a salinity signal. Salinity at the core site is dependent on changes in Indus River  
515 discharge, local run-off, and direct precipitation. Although the ISM would be the main  
516 influence on direct precipitation and run-off at the coring location, changes in the IWM could  
517 also influence Indus River discharge, ~~because snowmelt is a significant contributor in the~~  
518 ~~upper Indus catchment (Karim et al., 2002) and peaks during the summer months (Yu et al.,~~  
519 ~~2013).~~

520

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

(Figure 37), which ~~would could also suggest~~ indicate reduced upwelling of low  $\delta^{13}\text{C}$  intermediate water ~~under a weaker IWM~~ (Lynch-Stieglitz, 2006; Ravelo and Hillaire-Marcel, 2007; Sautter and Thunell, 1991); however, the interpretation of  $\delta^{13}\text{C}$  remains uncertain because of a poor understanding of the controls on the  $\delta^{13}\text{C}$  of planktonic foraminifera in this region. According to the  $\delta^{18}\text{O}$  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) (Dooze-Rolinski et al., 2001; Staubwasser, 2012).

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

By using  $\Delta\delta^{18}\text{O}$  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  $\delta^{18}\text{O}$  of *G. ruber* and *N. dutertrei* ( $\Delta\delta^{18}\text{O}_{d-r}$ ) to track changes in the surface-to-deep gradient. This gradient can be driven by either  $\delta^{18}\text{O}$  changes in the surface-dwelling (*G. ruber*) and/or the thermocline-dwelling species (*N. dutertrei*). During times of a strengthened winter monsoon,  $\Delta\delta^{18}\text{O}_{d-r}$  will decrease as surface waters are homogenized and the thermocline deepens. Similarly,  $\Delta\delta^{18}\text{O}_{d-r}$  will also decrease during times of a weakened summer monsoon, as decreased Indus River discharge will increase surface water salinity and  $\delta^{18}\text{O}$  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  $\delta^{18}\text{O}$  difference between the two species ( $\Delta\delta^{18}\text{O}_{s-r}$ ) reflects the influence of Indus River discharge on near surface salinity. The ~~greatest-smallest~~ difference in  $\Delta\delta^{18}\text{O}_{s-r}$  occurs at 4.1 ka BP, which is interpreted as an increase in surface water salinity (Figure 68).

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

~~However, t~~The following period of low  $\Delta\delta^{18}\text{O}_{d-r}$  but high  $\Delta\delta^{18}\text{O}_{d-s}$  from 4.1-3.9 ka BP is likely driven by increased salinity of surface water. This distinction becomes clearer when examining the  $\Delta\delta^{18}\text{O}_{s-r}$ , 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  $\delta^{18}\text{O}$  difference between *G. ruber* and *G. sacculifer*. At the same time, weakened winter mixing increases  $\Delta\delta^{18}\text{O}_{d-s}$ , 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

583 vice versa. The single time period when this does not hold true is 4.5-4.25 ka BP, when  
584 increased IWM mixing is coupled with decreased Indus discharge.

585

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

595

### 596 5.3 Comparison to marine records

597

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

608

609 Other marine records from the Arabian Sea also suggest a gradual decrease in ISM strength  
610 since from ~5 ka BP (Gupta et al., 2003; Overpeck et al., 1996). Cullen et al. (2000) observed  
611 an abrupt peak in aeolian dolomite and calcite in marine sediments in the Gulf of Oman from  
612 4.0-3.6 ka BP, and Ponton et al. (2012) also showed a shift to weaker ISM after 4.0 ka BP in  
613 the Bay of Bengal, based on  $\delta^{13}\text{C}$  of leaf waxes. Marine IWM reconstructions are not  
614 particularly coherent: although Dose-Rolinski et al. (2001) find a decrease in evaporation  
615 and weakening of the ISM between 4.6 and 3.7 ka BP, they argue this was accompanied by a  
616 relative increase in IWM strength. Giosan et al. (2018, in review) inferred enhanced winter  
617 monsoon conditions from 4.5-3.0 ka BP based on a planktic paleo-DNA and % *Globigerina*  
618 *falconensis* record close to our coring site ("C" in Figure 1), which ~~contradicts~~ disagrees with  
619 our finding of decreased upper ocean mixing after 4.3 ka BP. We suggest ~~the~~ that the high  
620 stratigraphic (i.e., laminated) and chronological (i.e., 15 radiocarbon dates between 5.4-3.0  
621 ka BP) resolution of core 63KA paired with a multi-species foraminifer  $\delta^{18}\text{O}$  record ~~can~~  
622 provides a more detail about robust history of the timing of changes in IWM and ISM strength,  
623 but additional studies are needed to resolve some of the discrepancies among the records.

624

### 625 5.4 Comparison to regional terrestrial records

626

627 The 63KA  $\delta^{18}\text{O}$  record obtained from three foraminifer species highlights several important  
628 ocean-atmosphere changes over the 5.4-3.0 ka BP time period. First, a sharp decrease  
629 occurred in both summer and winter precipitation at 4.1 ka BP, which is within a broader 300-

630 year period of increased aridity spanning both rainfall seasons between 4.2 and 3.9 ka BP. In  
631 detail, we infer a relative decrease in Indus River discharge and weakened ISM between 4.8  
632 and 3.9 ka BP, peaking at 4.1 ka BP, while a 200-year-long ~~phase interval~~ of strong IWM  
633 interrupted this period from 4.5-4.3 ka BP. Furthermore, the stepped change in  $\delta^{18}\text{O}$  of *N.*  
634 *dutertrei* suggests an enduring change in ocean-atmosphere conditions after 4.1 ka BP.

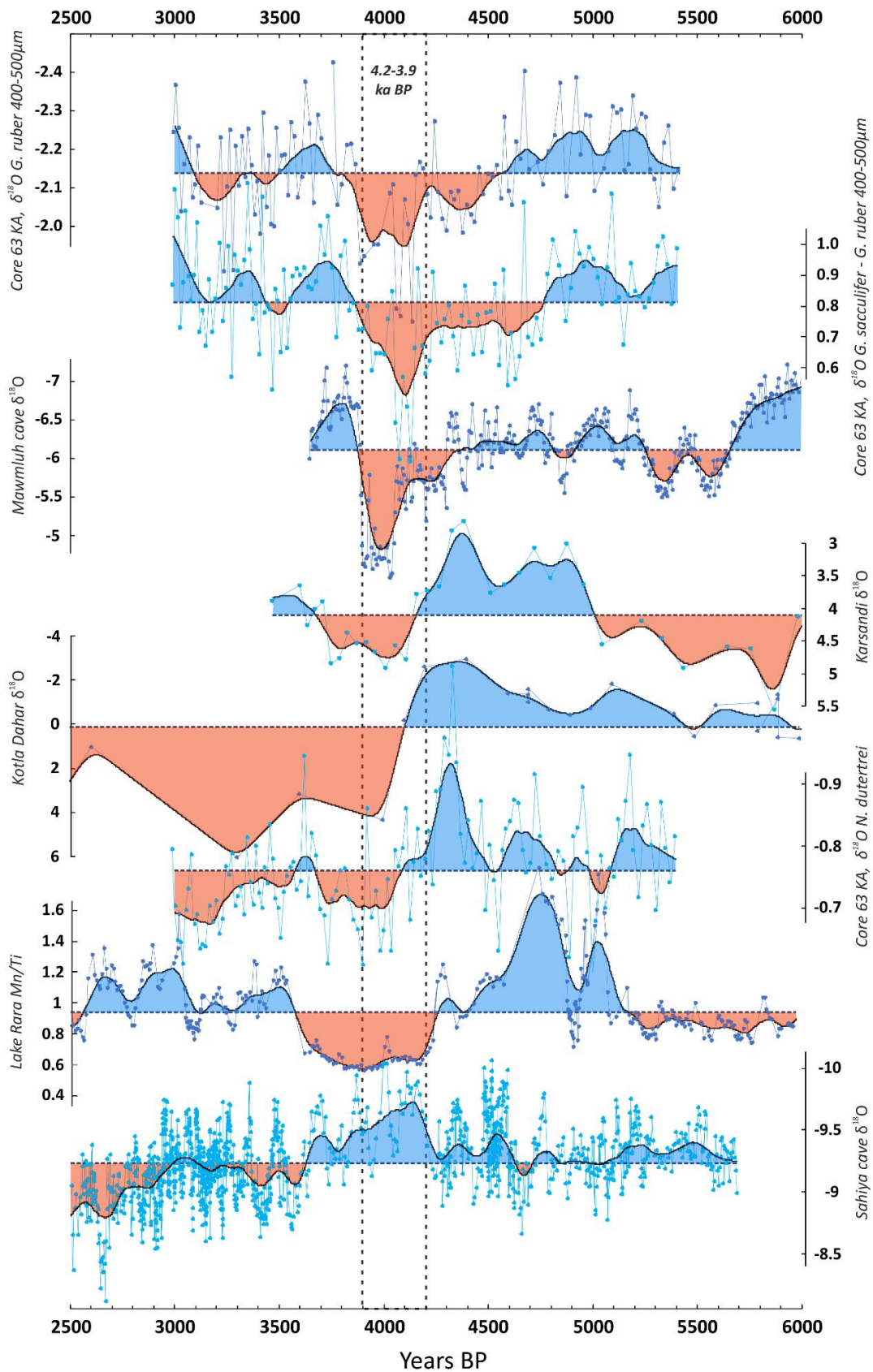
635  
636 A relatively abrupt  $\sim 4.2$  ka BP climate event has been observed in several terrestrial records  
637 on the Indian subcontinent, most notably Mawmluh Cave ( $\sim 4.1$ - $3.9$  ka BP) in northeastern  
638 India (Berkelhammer et al., 2012) and Kotla Dahar ( $\sim 4.1$  ka BP) in northwestern India (Dixit et  
639 al., 2014) (Figure 79). A less abrupt yet still arid period is documented in a peat profile ( $\sim 4.0$ -  
640  $3.5$  ka BP) from northcentral India (Phadtare, 2000), at Lonar Lake ( $\sim 4.6$ - $3.9$  ka BP) in central  
641 India (Menzel et al., 2014), and at Rara Lake ( $\sim 4.2$ - $3.7$  ka BP) in western Nepal (Nakamura et  
642 al., 2016). Finally, a recent study of oxygen and hydrogen isotopes in gypsum hydration water  
643 from Karsandi on the northern margin of the Thar Desert showed wet conditions between 5.1  
644 and 4.4 ka BP, after which the playa lake dried out sometime between 4.4 and 3.2 ka BP (Dixit  
645 et al., 2018). Considering terrestrial records can record more local climatic conditions than  
646 marine records, it is remarkable that the records collectively agree on a ~~regional phase period~~  
647 of ~~regional~~ aridity between 4.2 and 3.9 ka BP within the uncertainties of the age models that  
648 vary considerably among records.

649  
650 However, not all records support this finding, ~~such as~~ for example, a reconstruction from  
651 Sahiya Cave in northwestern India ~~that~~ shows an abrupt decrease in  $\delta^{18}\text{O}$  interpreted to  
652 reflect an increase in monsoon strength from  $\sim 4.3$ - $4.15$  ka BP, followed by an arid trend after  
653  $4.15$  ka BP (Kathayat et al., 2017). In addition, several other Thar Desert records do not  
654 identify a “4.2 ka BP event” *sensu stricto*, but instead suggest that lakes dried out several  
655 centuries earlier (Deotare et al., 2004; Enzel et al., 1999; Singh et al., 1990) or later (Sinha et  
656 al., 2006) than 4.2 ka BP. This discrepancy may relate to non-linear climate responses of lakes,  
657 which would not record a drought at 4.2 ka BP if they had already dried out earlier from the  
658 ongoing decrease in summer rainfall. In addition, there are also significant concerns about  
659 chronological uncertainties ~~when using~~ from the use of radiocarbon of bulk sediment for  
660 dating in some of these records. It is also possible that variations in the timing of climate  
661 change inferred from the terrestrial records may be real, reflecting different sensitivity to ISM  
662 and IWM rain. As a marine record, core 63KA integrates large-scale ocean-atmosphere  
663 changes, and therefore can help inform the interpretation of the more locally sensitive  
664 terrestrial records.

665  
666 More distantly, several terrestrial records in the Middle East also show a decrease in winter  
667 precipitation proxies around 4.2 ka BP: Jeita Cave in Lebanon records a relatively dry period  
668 between 4.4 and 3.9 ka BP (Cheng et al., 2015) and Soreq Cave in Israel shows a period of  
669 increased aridity starting at  $\sim 4.3$  ka BP (Bar-Matthews et al., 2003; Bar-Matthews and Ayalon,  
670 2011) (Figure 84). Lake Van in eastern Turkey also records reduced spring rainfall and  
671 enhanced aridity after  $\sim 4.0$  ka BP (Wick et al., 2003; Lemcke and Sturm, 1997). All of these  
672 records suggest a relatively arid ~~phase period in with reduced~~ winter precipitation after  $\sim 4.3$   
673 ka BP, as inferred from core 63-KA. Qunf Cave in Oman (Fleitmann et al., 2003), which is  
674 outside the range of IWM influence, instead shows a steady mid-Holocene weakening of the  
675 ISM that closely follows trends in summer solar insolation. ~~Core 63 KA also infers a protracted~~  
676 ~~decrease in ISM since  $\sim 4.8$  ka BP.~~



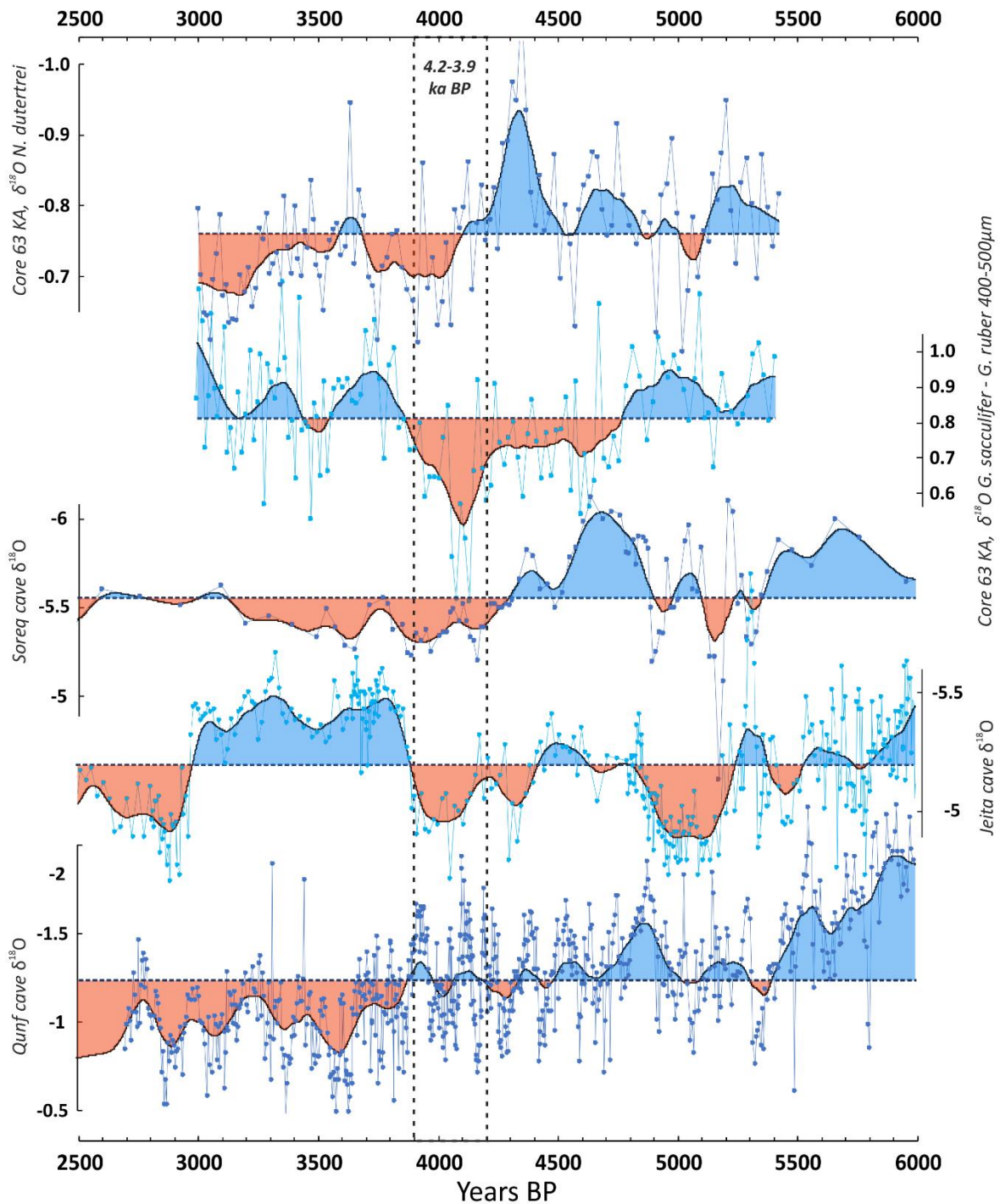
677  
678



679  
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681

**Figure 79.** Comparison of the  $\delta^{18}\text{O}$  record of core 63KA with terrestrial records from the Indian Subcontinent, from top to bottom: this study (first two), Berkelhammer et al., 2012; Dixit et al., 2018;

682 Dixit et al., 2014; this study; Nakamura et al., 2016; Kathayat et al., 2017. The mean value for each  
 683 record indicated by the horizontal dashed lines is taken for all available data between 6.0-2.5 ka BP.  
 684



685  
 686 **Figure 810.** Comparison of the  $\delta^{18}\text{O}$  record of core 63KA (topmost records),  
 687 from top to bottom: Bar-Matthews et al., 2003; Cheng et al., 2015; and Fleitmann et al., 2003. Mean  
 688 The mean value for each record indicated by the horizontal dashed lines is taken for all available data  
 689 between 6.0-2.5 ka BP.

690  
 691 5.5 Cultural impacts

692

693 ~~Based on~~ On the basis of our reconstruction of reduced IWM mixing after 4.3 ka BP,  
694 accompanied by decreased freshwater discharge of the Indus River, it is worth considering  
695 what impacts could be expected from a reduction in IWM and ISM precipitation. A weakened  
696 IWM overlying a reduced or more variable ISM would likely result in a distinct climate signal  
697 over the Indus River catchment, with broad implications for seasonal river flow and water  
698 availability throughout the year. The presence of the two rainfall systems creates a complex  
699 and diverse range of environments and ecologies across northwest South Asia (Petrie et al.,  
700 2017). In a situation when rainfall in both seasons is reduced over extended periods, step-  
701 shifts in the natural environment may occur that are difficult to reverse (e.g., desertification,  
702 lake desiccation, regional vegetation changes, decline in overbank flooding and shift in river  
703 avulsion patterns).

704  
705 Societies reliant on IWM, ISM, or a combination of the two would have been vulnerable to  
706 years with monsoon failure, and a shift affecting both seasons will have challenged resilience  
707 and tested sustainability (Green and Bates et al. in ~~prepress~~; Petrie et al., 2017).  
708 Archaeological research into the transition from the urban Mature Harappan phase (~4.6-3.9  
709 ka BP) to the post-urban Late Harappan phase (~3.9-3.6 ka BP) notes progressive  
710 deurbanization through the abandonment of large Indus cities and a depopulation of the  
711 most western Indus regions, concurrent with a general trend towards an increase of  
712 concentrations of rural settlements in some areas of the eastern Indus extent (Green and  
713 Petrie, 2018; Petrie et al., 2017; Possehl, 1997) (Figure S6). The relatively limited range of  
714 well-resolved available archaeobotanical data suggests that there was a degree of diversity in  
715 crop choice and farming strategies in different parts of the Indus Civilization across this time  
716 span (Petrie et al., 2016; Petrie and Bates, 2017; Weber, 1999; Weber et al., 2010). Farmers  
717 in southerly regions appear to have focused on summer or winter crops, while the more  
718 northern regions of Pakistan Punjab and Indian Punjab and Haryana were capable of  
719 supporting combinations of winter and summer crops (Petrie and Bates, 2017). Although  
720 there is evidence for diverse cropping practices involving both summer and winter crops in  
721 the northern areas during the urban period, agricultural strategies appeared to favor more  
722 intensive use of drought-resistant summer crops in the Late Harappan period (Madella and  
723 Fuller, 2006; Petrie and Bates, 2017; Pokharia et al., 2017; Weber, 2003; Wright, 2010). It has  
724 previously been suggested that weakened ISM was a major factor in these shifts (e.g. Giosan  
725 et al., 2012; Madella and Fuller, 2006). ~~On the basis of~~ Based on our reconstruction of  
726 decreased IWM in northwest South Asia after 4.3 ka BP with a step-shift at 4.1 ka BP, we  
727 suggest that both IWM and ISM climatic factors played a role in shaping the human landscape.  
728 This includes the redistribution of population to smaller settlements in eastern regions with  
729 more direct summer rain, as well as the ~~observed~~ shift to more-increased summer crop  
730 dominated cropping strategies.

## 731 732 **6. Conclusion**

733  
734 This study expanded on the  $\delta^{18}\text{O}$  record of planktonic foraminifer in core 63KA of the  
735 northeastern Arabian Sea, originally published by Staubwasser et al. (2003). Using  $\delta^{18}\text{O}$  of the  
736 surface-dwelling foraminifera *G. ruber*, the original study inferred an abrupt reduction in  
737 Indus River discharge at ~4.2 ka BP. Our further  $\delta^{18}\text{O}$  analysis of a larger size fraction of this  
738 species confirmed-corroborates maximum salinity at 4.1 and 3.95 ka BP. In addition, the  $\delta^{18}\text{O}$   
739 difference between the surface-dwelling *G. ruber* and slightly deeper-dwelling *G. sacculifer*

740 ( $\Delta\delta^{18}\text{O}_{\text{s-r}}$ ) reveals that surface waters were more saline than average for the period from 4.8-  
741 3.9 ka BP. By also measuring a thermocline-dwelling planktonic foraminiferal species, *N.*  
742 *dutertrei*, we infer an increase in the strength of the IWM between 4.5 and 4.3 ka BP, followed  
743 by reduction in IWM-driven mixing that ~~peaks~~ reaches a minimum at 4.1 ka BP.

744

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

756

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

769

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

777

#### 778 **Data availability**

779

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

783

#### 784 **Author contributions**

785

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

788

### 789 **Competing interests**

790

791 The authors declare that they have no conflict of interest.

792

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794

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802

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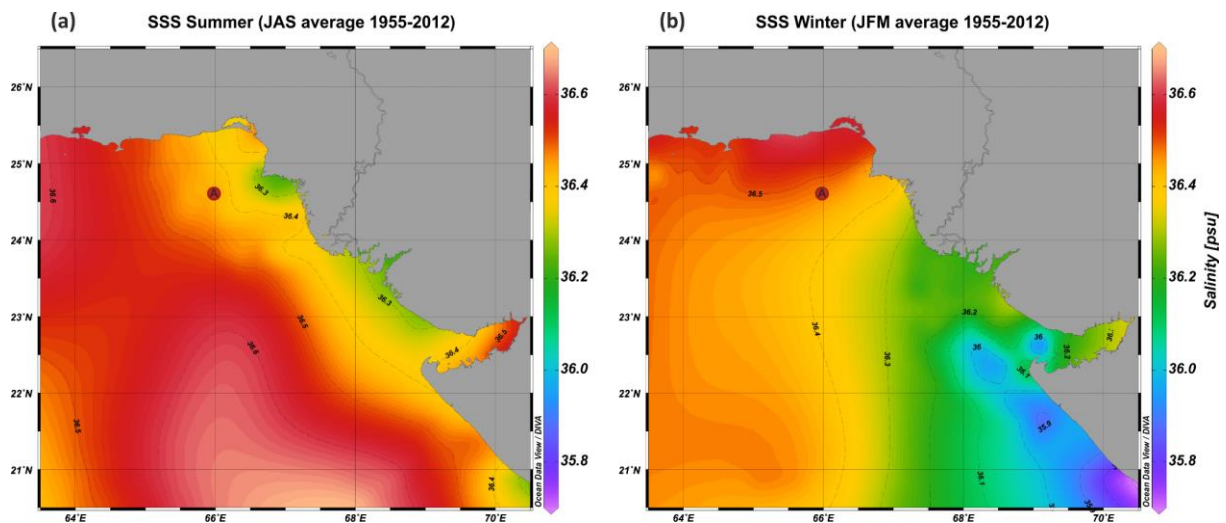
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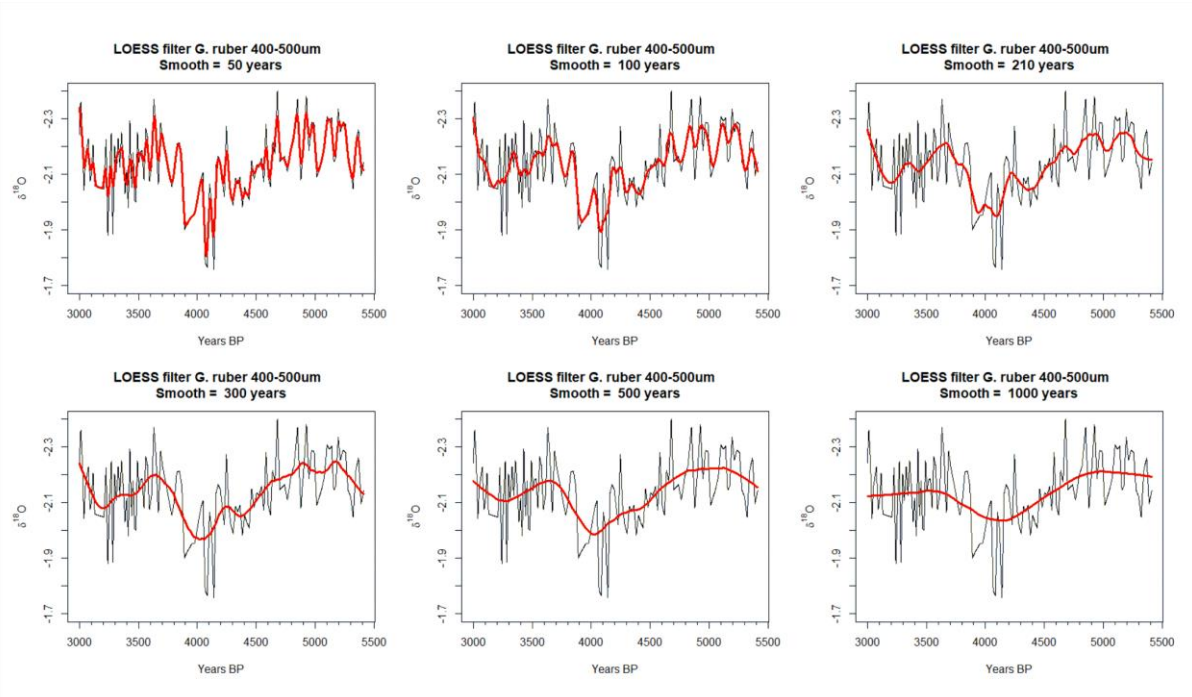
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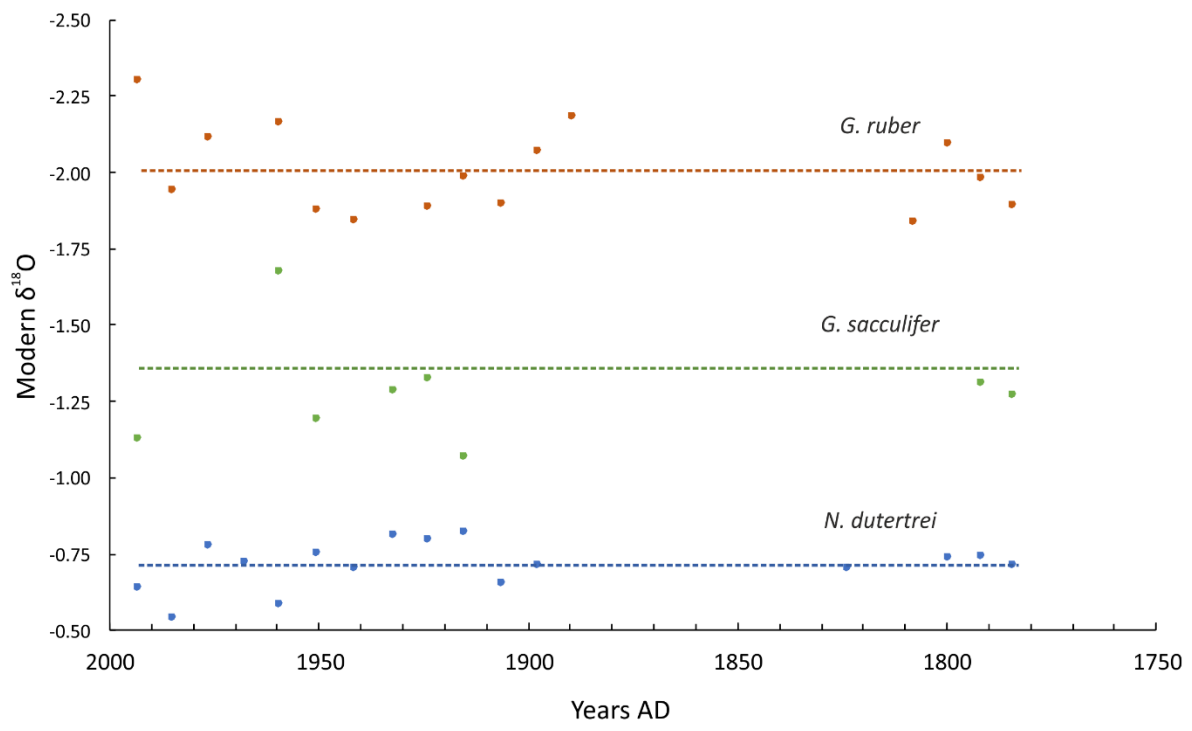


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1109 **Figure S1.** Mean surface salinity for 1955-2012, with data from the 2013 World Ocean Atlas (WOA) at  
1110 0.25° resolution (Zweng et al., 2013). Salinity contours are shown for **a.** summer (JAS) and **b.** winter  
1111 (JFM). The Indus River is outlined. Note that over the time window of this dataset, modern Indus River  
1112 discharge has been reduced by >50% due to barrages and irrigation (Ahmad et al., 2001). Plots created  
1113 with Ocean Data Viewer (ODV).  
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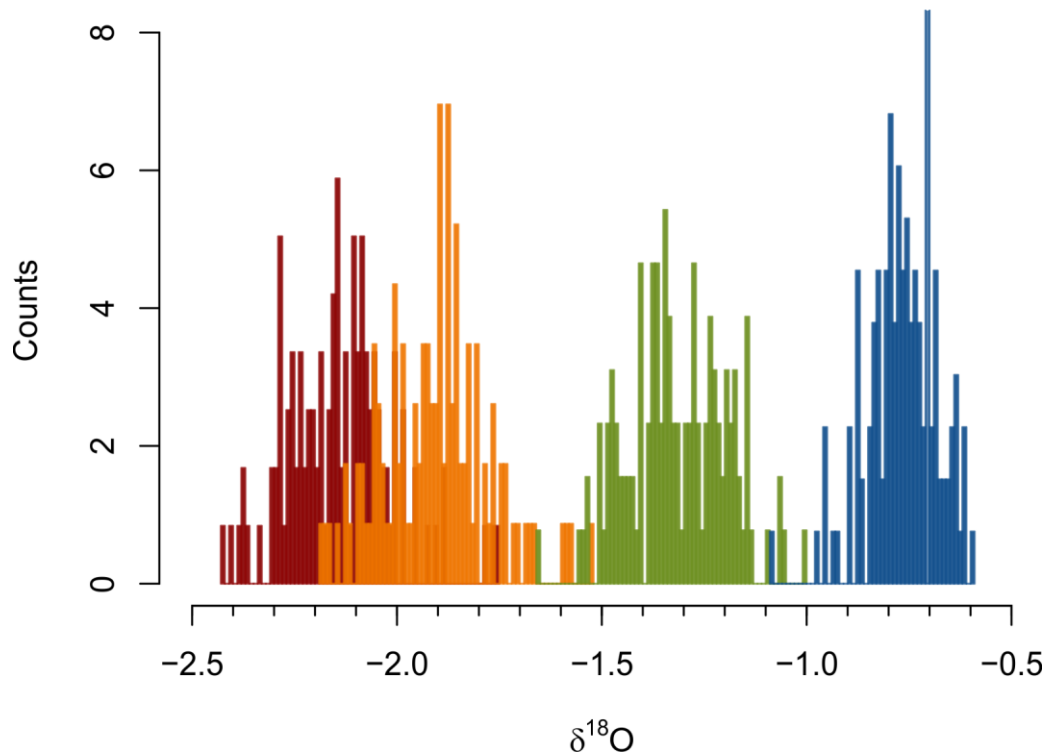
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**Figure S2.** Comparison of loess smoothing windows of 50, 100, 210, 300, 500, and 1000 years for *G. ruber* in the 400-500 $\mu$ m fraction.

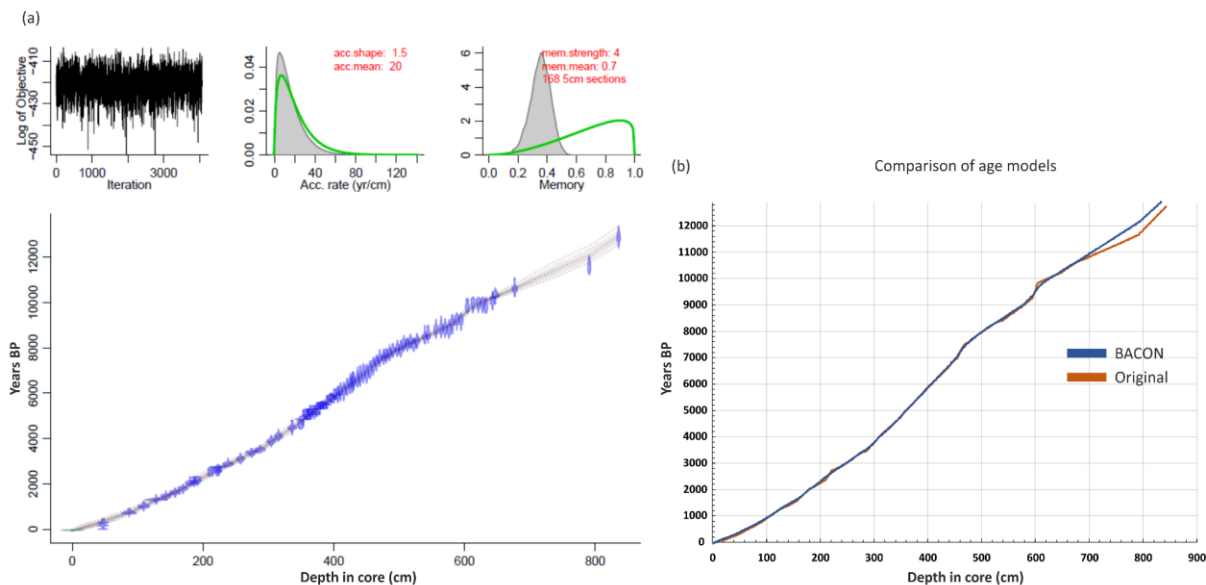


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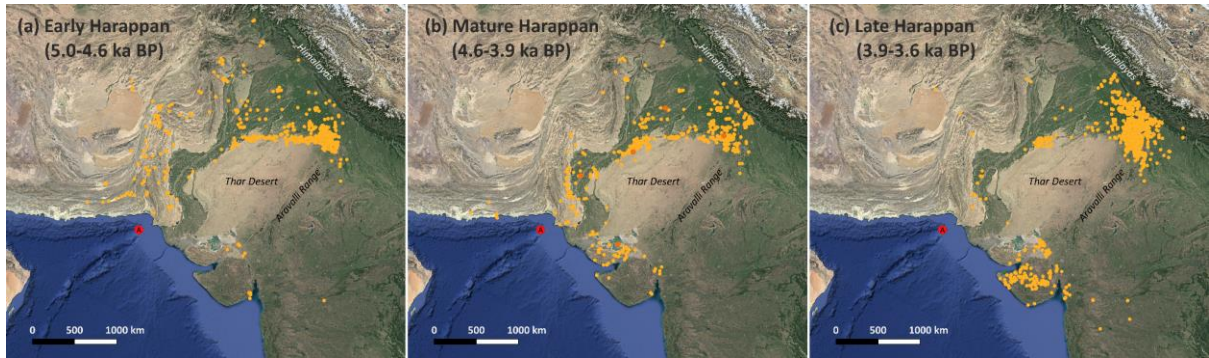
**Figure S3.** Modern  $\delta^{18}\text{O}$  values of calcite, spanning approximately the last 200 years, measured from surface sediment samples for all three species at the size fractions 315-400 $\mu$ m. Averages values for the last 200 years ( $\sim$ 1780-1993 AD) are compared to the period 5.4-3.0 ka BP: -2.01‰ (modern) and -1.90‰ (old) for *G. ruber* (orange), -1.28‰ (modern) and -1.31‰ (old) for *G. sacculifer* (green), and -0.72‰ (modern) and -0.76‰ (old) for *N. dutertrei* (blue).



1127  $\delta^{18}\text{O}$   
 1128 **Figure S4.** Frequency distributions of  $\delta^{18}\text{O}$  data during 5.4-3.0 ka BP for *G. ruber* 400-500 $\mu\text{m}$  (red), *G.*  
 1129 *ruber* 315-400 $\mu\text{m}$  (orange), *G. sacculifer* 315-400 $\mu\text{m}$  (green), *N. dutertrei* 315-400 $\mu\text{m}$  (blue).  
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1131 **Figure S5. a.** BACON age-depth model with calibrated dates shown in blue **b.** Age-depth model  
 1132 comparison with the original published age model from Staubwasser et al. (2003) (orange) and the  
 1133 new age model based on BACON software (blue).  
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**Figure S6.** Indus site distributions (yellow points) during the **a.** Early Harappan (~5.0-4.6 ka BP), **b.** Mature Harappan (~4.6-3.9 ka BP), and **c.** Late Harappan (~3.9-3.6 ka BP). Orange sites show larger Harappan cities during the Mature Harappan period (Dholavira, Mohenjo Daro, Ganweriwala, Harappa, and Rakhigarhi from bottom to top), core 63KA shown by red circle, background terrain from Google Earth.

**Table S1.** Main statistical parameters of the  $\delta^{18}\text{O}$  data.

	<i>G. ruber</i> 400-500 $\mu\text{m}$	<i>G. ruber</i> 315-400 $\mu\text{m}$	<i>G. sacculifer</i> 315-400 $\mu\text{m}$	<i>N. dutertrei</i> 315-400 $\mu\text{m}$
<i>n</i>	119	115	129	132
<i>Minimum</i>	-2.423	-2.190	-1.660	-1.090
<i>Maximum</i>	-1.752	-1.520	-1.000	-0.590
<i>1<sup>st</sup> Quartile</i>	-2.232	-1.995	-1.400	-0.810
<i>3<sup>rd</sup> Quartile</i>	-2.068	-1.830	-1.220	-0.700
<i>Mean</i>	-2.139	-1.901	-1.312	-0.761
<i>Median</i>	-2.144	-1.890	-1.320	-0.760
<i>Sum</i>	-254.58	-218.66	-169.26	-100.46
<i>SE Mean</i>	0.012	0.012	0.011	0.007
<i>LCL Mean</i>	-2.163	-1.926	-1.333	-0.776
<i>UCL Mean</i>	-2.116	-1.877	-1.291	-0.746
<i>Variance</i>	0.016	0.017	0.015	0.007
<i>Stdev</i>	0.128	0.131	0.122	0.085
<i>Skewness</i>	0.408	0.288	-0.011	-0.592
<i>Kurtosis</i>	0.511	0.174	-0.364	0.850

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1147

**Table S2.** Age-Model calibration with BACON software.

<i>Depth (cm)</i>	<i><sup>14</sup>C date</i>	<i>Error (<math>\pm 1\sigma</math>)</i>	<i>Reservoir (years)</i>	<i>IntCal13 min age BP</i>	<i>IntCal13 max age BP</i>	<i>IntCal13 mean age BP</i>
<i>surface</i>	-	-	-	-	-	-43
47	790	30	565	267	309	288
87	1370	35	565	678	780	729
109.5	1665	30	565	952	1062	1007
128.5	1955	25	565	1283	1339	1311
143.5	2115	35	565	1369	1529	1449
157.5	2270	25	565	1552	1634	1593
169.5	2430	25	565	1728	1869	1799
180.5	2640	25	565	1988	2122	2055
186.5	2675	35	565	1993	2154	2074
191.5	2720	30	565	2044	2184	2114
211.5	3000	35	565	2356	2541	2449

221.5	3110	40	565	2491	2602	2547
224.5	3145	25	565	2708	2758	2733
238.5	3340	25	565	2836	2929	2883
257.5	3510	30	565	2999	3181	3090
274.5	3730	30	565	3343	3451	3397
287.5	3850	30	565	3450	3576	3513
304.5	4145	30	565	3828	3975	3902
315.5	4310	30	565	4062	4159	4111
336.5	4570	40	565	4408	4578	4493
349.5	4655	40	565	4512	4711	4612
353.5	4870	30	565	4832	4892	4862
357.5	5005	35	565	4952	5079	5016
360.5	4980	30	565	4868	5057	4963
363.5	5080	30	565	5050	5194	5122
366.5	5105	35	565	5053	5189	5121
370.5	5070	35	565	5046	5300	5173
374.5	5160	40	565	5372	5463	5418
378.5	5210	40	565	5303	5469	5386
381.5	5315	30	565	5460	5585	5523
385.5	5315	35	565	5453	5586	5520
389.5	5420	35	565	5580	5654	5617
395.5	5635	35	565	5741	5907	5824
398.5	5610	35	565	5713	5904	5809
402	5750	40	565	5891	6008	5950
406.5	5830	35	638	5899	6002	5951
410.5	5965	40	638	5994	6210	6102
415.5	5980	45	638	5997	6216	6107
420.5	6120	45	638	6201	6351	6276
425.5	6265	45	638	6311	6490	6401
428.5	6335	55	638	6395	6639	6517
430.5	6345	60	638	6396	6657	6527
436.5	6440	40	638	6495	6678	6587
440.5	6540	55	638	6627	6883	6755
445.5	6665	45	638	6773	6984	6879
450.5	6650	40	638	6749	6948	6849
455.5	6960	45	824	6912	7162	7037
460.5	7155	45	824	7166	7331	7249
465.5	7310	45	824	7308	7480	7394
470.5	7480	55	824	7438	7606	7522
476.5	7550	50	824	7551	7670	7611
480.5	7815	55	1011	7571	7743	7657
485.5	7920	70	1011	7617	7867	7742
490.5	8070	50	1011	7788	7976	7882
497.5	8130	55	1011	7837	8027	7932
502.5	8115	55	1011	7828	8020	7924
507.5	8400	60	1011	8148	8345	8247



512.5	8350	50	1011	8020	8218	8119
517.5	8490	50	1011	8194	8381	8288
522.5	8355	60	1011	8023	8312	8168
527.5	8510	60	1011	8194	8400	8297
539.5	8790	60	1118	8384	8563	8474
544.5	8880	55	1118	8425	8631	8528
556.5	9060	50	1118	8637	8986	8812
564.5	9120	70	1118	8636	9026	8831
570.5	9110	50	1118	8698	9007	8853
576.5	9060	50	1118	8637	8986	8812
581.5	9260	50	1118	8999	9153	9076
588.5	9390	50	1118	9119	9430	9275
595	9370	60	1118	9076	9419	9248
604.5	9570	50	781	9602	9952	9777
613	9660	70	781	9736	10194	9965
621.5	9670	50	781	9884	10189	10037
628	9650	70	781	9732	10188	9960
633	9570	80	781	9581	9963	9772
643	9770	70	781	9906	10251	10079
647.5	9920	60	781	10206	10436	10321
677	10160	60	781	10480	10752	10616
791	11145	50	1095	11325	11806	11566
836	12285	55	1300	12726	12995	12861