

1 **Accepted version**

2 Last updated 25-Apr-2018

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4 **A late Holocene palaeoenvironmental ‘snapshot’ of the Angamma Delta, Lake**  
5 **Megachad at the end of the African Humid Period.**

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22 Keywords: Holocene; Palaeoecology; Sedimentology; Stable isotopes; Central  
23 Africa; Lake Megachad

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25

26 **Abstract**

27 During the African Humid Period (AHP) there was a dramatic increase in the area of  
28 lakes and wetlands. Lake Megachad, one of several huge lakes, underwent dramatic  
29 fluctuations during the AHP prior to regression in the mid Holocene. However, the  
30 timing and nature of AHP termination has been disputed. We present evidence from  
31 sediments of the Angamma Delta, from the northern end of the palaeolake, for Lake  
32 Megachad lake-level fluctuations at the end of the AHP. Delta slope deposits were  
33 deposited over 7000 cal BP at the height of the AHP. Overlying bioclastic sediments,  
34 from 4300 – 4800 cal BP and an elevation of 285 – 290 m, lie below the palaeolake  
35 highstand (339 m) but close to the elevation of the Bahr el Ghazal sill, which divided  
36 the lake's two sub-basins. Ostracod  $\delta^{18}\text{O}$  values indicate that the waters of the  
37 northern sub-basin were evaporated to levels similar to modern Lake Chad.  
38 Palaeoecological evidence suggests that the lake was perennial and evaporative  
39 enrichment is attributed to restricted circulation of lake waters as the sill emerged.  
40 The age and elevation of the bioclastic sediment, coupled with published lake level  
41 reconstructions, suggests a complex lake-level history with a major regression at the  
42 end of the AHP, followed by a short lived, lake level rise after the followed by a  
43 transgression. This new evidence for changes in lake level provide support for other  
44 geological records and some modelling experiments that suggest rapid fluctuations  
45 in hydroclimate at the end of the AHP.

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52 **1. Introduction**

53 From the late glacial until the mid Holocene, northern Africa was characterised by  
54 increases in effective moisture (precipitation minus evaporation, or P - E) as a result  
55 of orbitally-forced strengthening of the African summer monsoon (Kutzbach and Liu,  
56 1997). During this interval, known as the African Humid Period (AHP – de Menocal  
57 et al., 2000), there was an increase in the extent of lakes and wetlands over large  
58 parts of northern Africa (Holmes and Hoelzmann, 2016), and shrubland and  
59 grassland replaced desert vegetation (Hoelzmann et al. 2004; Le 2017). The AHP  
60 was interrupted by millennial-scale arid intervals during the late glacial stadial and on  
61 several occasions during the Holocene; these intervals may have been accompanied  
62 by southward shifts in the intertropical convergence zone (ITCZ), in situ weakening of  
63 the summer monsoon and accompanying reduction in rainfall, and/or reductions in  
64 the latitudinal extent of the African rain-belt that were symmetrical over both  
65 hemispheres (Shanahan et al., 2015). The AHP came to an end in the mid Holocene  
66 sometime between around 6000 and 5000 BP, with palaeoenvironmental evidence  
67 and modelling experiments variously suggesting an abrupt, regionally-synchronous  
68 termination (e.g. de Menocal et al., 2000), a gradual ending (Kroepelin et al., 2008,  
69 Francus et al., 2013) or a pattern of progressive drying on which was superimposed  
70 increased short-term climatic variability (Gasse 2006, Renssen et al. 2006).

71 Understanding the nature and timing of AHP termination is important because it  
72 sheds light on the non-linear response of the African monsoon to orbital forcing and  
73 the role of vegetation and land-surface feedbacks (Claussen, 2009) and well as  
74 having implications for the human occupation of northern Africa during the Holocene  
75 (Manning and Timpson, 2014).

76

77 During the AHP, a huge palaeolake, known as Lake Megachad, occupied a large  
78 endorheic basin in the central part of North Africa. The basin extends from 6° to 25°  
79 north, spanning present-day subtropical to arid climatic zones. Water levels in the  
80 basin have fluctuated in response to the strength of the West African summer  
81 monsoon (Armitage et al., 2015). At its early to mid-Holocene peak, Lake Megachad  
82 was around 1000 km long (N-S) and up to 600 km wide (E-W); lake levels reached  
83 330 m (above present-day sea level), above which the lake spilled south into the  
84 Benue River (Drake and Bristow 2006). Lake Chad is now greatly reduced in extent,  
85 is currently around 200 km long and up to 150 km wide and covers roughly 5% of its  
86 former area. The lake surface lies at an elevation of 280 m, is confined to the  
87 southern half of the basin and supplied with around 90% of its waters through the  
88 Chari River. The northern half of the basin, the Bodélé Depression, is deeper with a  
89 low-point of 170m. This northern sub-basin is dry despite its greater depth, and is  
90 separated from Lake Chad by a 285-m-elevation sill that currently prevents water  
91 from flowing north from Lake Chad into the Bodélé Depression, although in the past,  
92 water has flowed in this direction through a river system known as the Bahr el  
93 Ghazal when the lake level exceeded 285 m. The changes in the extent and  
94 elevation of Lake Chad that occurred during the Holocene are amongst the most  
95 dramatic climatically driven changes on Earth.

96

97 During the AHP, the Sahara-Sahel boundary shifted in central and eastern Africa as  
98 far as 23°N (Hoelzmann et al., 2004). Tropical trees and shrubs occurred about 400-  
99 500 km north of their present distribution, mainly as part of the gallery-forest  
100 communities along the abundant rivers and lakes (Watrín et al, 2009). Most pollen

101 records from the Sahelian and Sudanian zone indicate a shift towards drier  
102 vegetation accompanied by a distinct decrease in lake levels between 6000 - 3000  
103 cal BP. The two pollen records available for the Lake Chad Basin provide a rather  
104 inconsistent picture of the termination of the AHP. Whereas a record from the  
105 southern pool (Amaral et al. 2013) suggest a gradual retreat of trees and shrubs that  
106 indicated a humid climate from ca. 6050 cal BP onwards, the Tjeri sequence (Maley,  
107 1981), further to the north-east, records a comparable change in vegetation  
108 composition approximately 2000 years later. The influence of anthropogenic  
109 activities on the West African landscape at the end of the AHP is a matter of debate.  
110 Archaeological evidence suggests that with the introduction of pastoralism and  
111 agriculture, ca. 4500 years ago, West Africa experienced a significant cultural and  
112 environmental transformation along with an increase in human population (McIntosh  
113 and McIntosh, 1983). However, to date, a large-scale human impact on the Sahelian  
114 and Sudanian savanna has not been detected in the geological records (e.g.  
115 Salzmänn and Waller, 1998; Salzmänn et al., 2002; Waller et al. 2007).

116

117 We investigated sediments from the Angamma Delta, in the northern part of the  
118 palaeolake, and from the Bodélé Depression, in order to provide constraints on lake-  
119 level changes at the end of the AHP, and to characterise the environment of the  
120 Bodélé Depression as it desiccated, using a combination of sedimentology,  
121 geochronology, micropalaeontology and isotope analysis. The sediments of the  
122 Falaise d'Angamma were interpreted to be a Holocene delta by Servant et al. (1969),  
123 who described volcanic breccias and tuffs at the base of the section overlain by  
124 layers of deltaic sediments that dip gently towards the South and Southwest. They  
125 describe a 20-30m series of rhythmic alternations of silts, clays and sands with a

126 range of sedimentary structures including cross-strata, slumps and channels. They  
127 also identified fossil wood, and bones of animals including an ancient form of  
128 Elephant *Loxodonta africana* and a small form of Hippopotamus as well as a cranio-  
129 facial fragment of a hominin, and conducted some radiocarbon dating of shells and  
130 carbonate concretions that provided an early Holocene age (Servant et al. 1969).

131

## 132 **2. Materials and methods**

133 The Angamma delta is located at the northern end of palaeolake Megachad (Fig. 1).  
134 The beach ridge along the top of the delta front stands at an elevation over 330m  
135 (Drake and Bristow 2006), and the delta slopes down to the basinal sediments that  
136 are composed of mudstones and diatomite at elevations below around 240m. The  
137 morphology of the delta is very well preserved and its deposits are locally very well  
138 exposed in a series of canyons incised into the western margin of the delta front (17°  
139 36' 54" N, 17° 36' 11" E). A 25m sedimentary log was measured through the outcrop  
140 at a scale of 1:50 (Fig. 2). A sample of sand (CH36) was collected at 5m on the log,  
141 a sample of charcoal (CH37) was collected at 9m on the log, and bioclastic silty  
142 sands CH38 and CH39 were collected at 20 and 23m respectively. In addition, we  
143 have analysed a sample of the lakebed sediments from localities CH59 and CH60  
144 (16 48' 19.0" N 17 48' 35.3" E and 16° 47' 17.6" N, 17° 50' 13.2" E, respectively),  
145 which lie at an elevation of 175m and close to the base of the Bodélé Depression.

146

147 Bulk sediment samples for faunal analysis were taken from CH38, CH39 and CH60.  
148 Dried bulk sediment was dispersed in tap water, sieved through a 250µm mesh and  
149 the coarse fraction dried in an oven at 105°C: ostracod and mollusc shells were  
150 extracted from this fraction under low-power stereo microscope and stored in

151 micropalaeontological slides (ostracods) or glass vials (mollusc shells). Quantitative  
152 counts of ostracods were undertaken whereas only the presence of individual  
153 mollusc taxa was noted. Selected ostracod specimens were measured (length and  
154 height) using a calibrated reticule under a low power (18.75x magnification) stereo  
155 microscope. Selected, well-preserved ostracod shells were brush-cleaned with  
156 methanol for oxygen and carbon isotope analysis: either single or multiple-shell  
157 samples were analysed depending on the species. Stable-isotope analyses on  
158 samples in the range 15 - 200ug were undertaken using an Isorime Multiprep and  
159 dual inlet mass spectrometer system at Royal Holloway, University of London  
160 (RHUL) and NIGL Keyworth, and the results reported in standard delta units relative  
161 to V-PDB. The external analytical reproducibility was better than  $\pm 0.07$  ‰ for both  
162  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$ .

163

164 Four sediment samples (CH38, CH39, CH59, CH60) were processed for pollen  
165 analysis using standard laboratory techniques (Faegri and Iversen , 1989), including  
166 HF treatment and acetolysis. *Lycopodium clavatum* spore tablets were added to  
167 each sample to allow calculation of pollen concentration (Stockmarr, 1971). Pollen  
168 and spores have been identified using the pollen reference collection held at  
169 Northumbria University.

170

171 Radiocarbon dating was undertaken on specimens of ostracods and molluscs, and  
172 on charcoal at the NERC Radiocarbon Laboratory and Beta Analytic. Radiocarbon  
173 dates were calibrated using IntCal13 (Reimer et al., 2013).

174

175 **3. Results**

176 3.1 Angamma Delta Geomorphology

177 The northern shoreline of palaeolake Mega-Chad is dominated by the Angamma  
178 delta which is around 50 km wide. Satellite images show remarkable preservation of  
179 the delta's geomorphology, including distributary channels on the delta top, beach  
180 ridges on the western side of the delta, a clearly defined beach ridge along the delta  
181 front, and cusate forelands to the east (Fig. 2a). The delta was fed by a braided  
182 fluvial distributary that flowed into the lake from the Tibesti Mountains in the north.  
183 These channels can be picked out on the satellite image cutting through some older  
184 beach ridges preserved on the delta top (Fig. 2b). However, the channels do not cut  
185 through the beach ridge which defines the delta front, which is known as the cordon  
186 littoral (Servant et al., 1969). The fact that the beach ridge is not cut by the fluvial  
187 distributary channels indicates that the fluvial drainage from the northern catchments  
188 ceased before the lake-level fell (Armitage et al. 2016). Had the rivers continued to  
189 flow after the lake-level had fallen, then the rivers would have incised through the  
190 beach ridge to create a falling stage delta. The planform of the delta front, which is  
191 defined by the cordon littoral, shows that the Angamma delta had a cusate  
192 morphology. Cusate deltas are characteristic of wave dominated deltas (Galloway  
193 1975) indicating that sediment delivered to the delta front by rivers was reworked  
194 and redistributed by waves in the lake. The waves that impacted the northern shores  
195 of the palaeolake would have been driven by southwesterly monsoon winds (Drake  
196 and Bristow 2006) which had a maximum fetch of over 800 km from the southern  
197 shores of the lake. A topographic profile from the Angamma Delta to the Bodélé  
198 Depression preserves the lake bathymetry (Figure 2c). The cordon littoral stands at  
199 an elevation of around 339 m, which is 170 m above the lowest point in the Bodélé  
200 Depression, indicating that the lake was up to 170 m deep. Topographic profiles



201 across the delta front reveal a sigmoidal profile. The overall slope of the delta front is  
202 around 2° with the steepest part dipping at 10-12°. While most of the delta top  
203 morphology is well preserved there has been some erosion of the delta front and this  
204 is most obvious on the western side of the delta, which has been incised by steep  
205 sided gullies. These gullies, which do not cut the preserved shoreline, have steep  
206 headwalls with very small catchments and are interpreted to have formed by  
207 groundwater sapping after the lake level fell. Spring systems that might once have  
208 fed the gullies have long since dried up and the gullies and interfluves have been  
209 eroded by the north-easterly Harmattan wind forming giant yardangs (Fig. 3a). The  
210 erosion provides excellent exposure of the Angamma Delta sediments, which are  
211 described and interpreted below.

212

### 213 3.2 Angamma Delta Sediments

214 The sediments of the Angamma Delta are very well exposed in a series of NE-SW  
215 trending canyons and cliff sections that are perpendicular to the delta front. The base  
216 of the Angamma delta sediments is underlain by volcanic breccias and tuffs (Servant  
217 et al. 1969). These are overlain by diatomite deposits with shells of *Pisidium* sp. and  
218 *Valvata* sp., which have been radiocarbon dated at  $9260 \pm 140$  <sup>14</sup>C yr BP and  $10160$   
219  $\pm 160$  <sup>14</sup>C yr BP (Servant et al. 1969). A fine-grained carbonate concretion within the  
220 overlying sediments has a radiocarbon age of  $6050 \pm 150$  <sup>14</sup>C yr BP (Servant et al.  
221 1969). The section described in this paper is on the western side of the delta where  
222 25 m of Holocene sediments are exposed. This corresponds with the 20-30 m series  
223 of rhythmic alternations of silts, clays and sands described by Servant et al. (1969).  
224 The grainsize of the sediments, bed thickness and sedimentary structures are  
225 recorded on a graphic sedimentary log (Fig. 4). The sediments are composed of silts

226 and very-fine to fine-grained sands with a few thin layers of intraformational  
227 conglomerates (Fig. 4). Beds are generally thinner at the base of the section and  
228 thicker towards the top. Bed contacts are mostly sharp, many have erosional bases  
229 and a few fine up with gradational tops. Sedimentary structures include: current  
230 ripple lamination, wave ripple lamination, hummocky and swalley cross-stratification,  
231 bioturbation and soft sediment deformation (Fig. 4). Current ripple lamination is very  
232 common in the lower half of the section (Fig. 4) with a palaeocurrent direction of 250°  
233 which is attributed to currents flowing from the delta down-slope towards the lake  
234 bed. It is possible that some of these currents could be density driven turbidity  
235 currents. Some of the fine-grained sandstone beds (4-5 m on log, Fig. 4) have sharp  
236 erosive bases with intraformational mudstone clasts and fine upwards (Fig. 3b). They  
237 show many of the features of turbidite deposits including a sharp erosive base, fining  
238 upwards, and planar lamination and current ripple lamination (Bouma 1962).  
239 Turbidite deposits are common in lake sediments (e.g. Dyni and Hawkins 1981,  
240 Sturm and Matter 1978) and it is possible that dense, sediment-laden, flows from  
241 flood events on the rivers that supplied sediment to the delta continued to flow  
242 across the lake bed as turbidity currents, because the sediment-laden river water  
243 was denser than the freshwater within the lake. One palaeocurrent direction trending  
244 towards 250° is consistent with gravity driven flows down the southwest-facing delta  
245 slope. Wave driven current flows within palaeolake Mega-Chad have been modelled  
246 by Bouchette et al. (2010); their model suggests westward flowing surface currents  
247 and weak bottom currents driven by the north easterly Harmattan wind, and  
248 northeast flowing surface currents and weak bottom currents driven by the south-  
249 westerly monsoon wind around the Angamma delta.

250

251 An isolated set of wave ripple lamination is recorded close to the base of the section  
252 (1.3 m on the log, Fig. 4), a bed of wave ripple lamination is also recorded at 17 to 18  
253 m on the log (Fig. 4). Wave ripple lamination indicates that the lake-bed is within  
254 wave-base and hence exposed to the oscillatory currents set up by surface wind-  
255 driven waves.

256

257 Hummocky cross-stratification is recorded at 5.5, 6, and 11 - 12, m on the log (Fig.  
258 4), while swaley cross-stratification is recorded at 16.5 and 18m (Fig. 3c).

259 Hummocky Cross-stratification (HCS) is an indication of storm conditions most often  
260 associated with shallow marine environments (e.g. Duke 1985, Cheel and Leckie  
261 1993), but has also be described in lacustrine sediments (Eyles and Clark 1986).

262 Wave tank experiments by Dumas and Arnott (2006) demonstrate that HCS can be  
263 developed under combined oscillatory and unidirectional currents, which are  
264 believed to occur in nature during storms when waves interact with unidirectional  
265 (offshore) currents. The preservation of HCS is aided by deposition of fine sand  
266 eroded from the upper shoreface during storm conditions. Dumas and Arnot (2006)  
267 suggest that swaley cross stratification can be formed under similar flow condition to  
268 HCS, but with lower rates of aggradation which preserve the swales rather than the  
269 hummocks. In their model swaley cross-stratification is found in slightly shallower  
270 water, closer to the shore than HCS, which is consistent with the observations that  
271 swaley cross-stratification occurs above the HCS in the Angamma delta log (Fig. 4).  
272 The possible wave ripple lamination on the top of one of the sharp-based fining-  
273 upwards beds at 4.5 m on the log (Fig. 4) suggests that these could be tempestites  
274 rather than turbidites.

275

276 Soft sediment deformation is very common and includes dewatering structures (3 m  
277 on log, Fig. 4), load structures (9.8 m on log), folded cross-strata (7.8 and 14 – 14.5  
278 m on log) as well as extensional slides and injection structures. Servant et al (1969)  
279 also noted the presence of contorted beds from slumping in nearby sections.  
280 Overlying the interbedded sandstones and siltstones at the top of the measured  
281 section are brown, silty sands with abundant ostracods and gastropods. They are  
282 poorly indurated, and thus eroded more, and less well exposed than underlying  
283 beds.

284

### 285 3.3 Sandbody geometry

286 The canyons incised into the delta reveal a dip-section perpendicular to the delta  
287 front. The bedding geometry is lens-like, with a series of low-angle erosion surfaces  
288 cutting down to the west (Fig. 5). Some of the lenses are formed by channels  
289 because both channel banks can be observed in the field (Fig. 5). However, other  
290 erosion surfaces cut down from west to east and the opposite 'channel' bank is  
291 missing. The succession off-laps towards the west, into the lake. However, many of  
292 the beds are truncated by asymmetric and lens-like scour surfaces that cut down to  
293 the west (Fig. 5). Although the succession is broadly progradational there is an  
294 absence of obvious progrades and the origin the erosion surfaces is not certain, they  
295 might have been driven by changes in lake level, storm events or slumping. Similar  
296 looking, but slightly smaller scours within heterolithic distal lower shoreface  
297 sediments have been interpreted as formed by storm-generated currents coincident  
298 with riverine sediment influx 'storm floods' (Onyenanu et al. 2018). A similar scenario  
299 for coincident riverine flooding and storms during an enhanced monsoon might  
300 explain the scours on the Angamma delta front.

301

### 302 3.4 Chronology

303 Radiocarbon dates (Table1) indicate that the middle part of the Angamma delta  
304 sequence dates to around 7300 cal BP whereas the fossiliferous upper unit dates to  
305 around 4300 – 4800 cal BP. Previous radiocarbon dates from Servant et al. (1969)  
306 suggest that the base of the Angamma Delta sequence dates to the earliest part of  
307 the Holocene. The sediments from the residual pool in the Bodélé Depression,  
308 represented by CH59 and CH60, date to around 1000 cal BP Table 1: Armitage et  
309 al., 2015).

310

### 311 3.5 Palaeontology and geochemistry

312 Ostracods were present in three of the four samples investigated and most abundant  
313 in CH38, even allowing for the larger size of that sample (Table 2). The assemblage  
314 in CH38 is dominated by *H. giesbrechtii* and, with the exception of *S. aculeata*, which  
315 is represented by a single specimen, adults and juveniles are present (Fig. 6, for *S.*  
316 *bicornis*). The other samples are characterised by lower abundances and diversity  
317 (Table 2), although many *S. bicornis* specimens were removed from CH39 for dating  
318 prior to enumeration. No ostracods were found in CH59. Molluscs were also present  
319 in three out of the four samples investigated, although the lack of material in CH39  
320 probably reflects the sample processing methods rather than genuine absence of  
321 molluscs. Sample CH38 is the most diverse; CHG59 contains two gastropod species  
322 and one bivalve (*Coelatura aegyptiaca*); CH60 is dominated by *Coelatura aegyptiaca*  
323 (Table 3).

324

325 Stable isotope values in the ostracod shells show large variability both within and  
326 between levels for oxygen and within levels for carbon (Fig. 7, Table 4). Maximum and  
327 minimum  $\delta^{18}\text{O}$  values are seen in CH38 (+0.03 ‰) and CH60 (+9.88 ‰), respectively;  
328 corresponding values for  $\delta^{13}\text{C}$  are -3.18 ‰ and +2.73 ‰ (both in CH38). Values in  
329 CH38 reveal some inter-species differences for isotope signatures (Fig. 7). For oxygen  
330 the 1.3 to 1.5 ‰  $^{18}\text{O}$ -enrichment in *Candona* compared with the other three species  
331 analysed is the most marked difference. For carbon, *Sclerocypris bicornis* appears  
332  $^{13}\text{C}$ -deplete compared with *H. giesbrechtii*, *Cytheridella tepida* and *Candona* cf.  
333 *neglecta*, have  $\delta^{13}\text{C}$  values that broadly fall between these two species. There is no  
334 covariance amongst  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values (Fig. 7), either for the individual species or  
335 for the dataset as a whole. The small number of trace-element determinations on  
336 shells of *L. inopinata* gave values of between 2.73 mmol/mol (CH60) and 6.36  
337 mmol/mol (CH38) (Table 4).

338

339 All processed sample residues show a high organic matter content. Identifiable  
340 pollen and spores, however, are only preserved in sample CH60. As the other  
341 samples have been taken from outcrops, it is very likely that palynomorphs have  
342 been destroyed by post-sedimentary processes. Sample CH60 has a concentration  
343 of 21,778 pollen/g dry weight. Diversity is very low with *Typha* (53.3%) and  
344 *Cyperaceae* (34.9) being most abundant, followed by *Poaceae* (5.3%) and  
345 *Chenopodiaceae* (6.6). Green algae, such as *Pediastrum*, occur in high numbers.

346

#### 347 **4. Discussion**

348 We combine the stratigraphical and sedimentological information from the Angamma  
349 Delta section and the Bodélé Depression with palaeoecological and geochemical

350 data, in order to develop a palaeoenvironmental synthesis for the middle and late  
351 Holocene intervals that these sediments represent.

352

#### 353 4.1 Stratigraphy and sedimentology

354

355 Deposition of the fine-grained, heterolithic, sediments of the delta front during the  
356 early to middle Holocene AHP is consistent with stratigraphic models for lacustrine  
357 sedimentation where sediment supply and the input of water from rivers is intimately  
358 linked so that high lake levels are coincident within increased fluvial sediment input  
359 e.g. (Bohacs et al. 2000). Sharp-based fine-grained sandstone beds that fine  
360 upwards are interpreted as the deposits of turbidity currents formed when dense  
361 sediment-laden flood waters entered the lake from the rivers that flowed across the  
362 top of the Angamma Delta. Lacustrine turbidites have been linked to storm events  
363 within lake catchments (e.g. Osleger et al., 2009), and it is possible to speculate that  
364 these might be driven by annual monsoon rains, but equally they could be due to  
365 heavy rainfall and floods from convective thunderstorms. The scour surfaces  
366 observed within the heterolithic deltaic sediments might also be associated with  
367 storm-generated currents coincident with riverine sediment influx (Onyenanu et al.  
368 2018). Additional work on lateral continuity of beds, and the geometry of erosive  
369 scour surfaces, as well as correlation along strike around the delta front would be  
370 needed to test the potential for seasonality. Such reconstructions are likely to be  
371 complicated by erosion on the delta front and switching of distributary channel  
372 across the delta top that will have created breaks in deposition and diachronous  
373 changes in facies. Another possible explanation is that the turbidites were  
374 seismically triggered (Moernaut et al. 2014, 2017), which would be consistent with

375 the possible seismic triggering of widespread soft-sediment deformation. One  
376 explanation for the slumping is gravitational instability on the gently inclined delta  
377 slope. However, soft sediment deformation can also be triggered by seismic shock  
378 and similar deposits in Pleistocene deltaic and lake sediments have been interpreted  
379 as seismogenic (e.g. Gilbert et al. 2005, Moretti and Ronchi 2011). Moretti and  
380 Ronchi (2011) rejected an internal, autokinetic, trigger because the deformed  
381 sediments are similar to other beds in the succession that lack evidence for  
382 liquefaction or fluidisation. On this basis, it is possible to argue for an allokinetic  
383 (external) trigger such as an earthquake but alternative allokinetic triggers related to  
384 lake level changes or storm events cannot be ruled out. The interbedded sandstones  
385 and mudstones that include possible turbidite and tempestite deposits along with  
386 slumps and channels are interpreted as a delta-slope facies. The overlying  
387 bioclastic-rich silty-sands are less well exposed and the depositional environment is  
388 not as easy to reconstruct from the sediments alone. In order to reconstruct their  
389 depositional environment, we consider evidence from the faunal assemblage,  
390 oxygen and carbonate isotopes below.

391

## 392 4.2 Ostracods

393 Published information on the ecology of ostracod taxa is used for the  
394 palaeoecological interpretation of the ostracod assemblages from the Angamma  
395 Delta and Bodélé Depression sediments.

396

### 397 4.2.1 Taxonomic and ecological notes on the ostracod taxa

398 *Limnocythere inopinata* (Baird, 1843) (Fig. 8)



399 This species is a widespread benthic taxon found in the littoral of large lakes and in  
400 small lakes and ponds (Geiger, 1990; Meisch, 2000; Rossi et al., 2010; Van der  
401 Meeren et al., 2010). In the Holarctic, the species is almost always parthenogenetic,  
402 although pockets of sexual populations are found geographically and  
403 stratigraphically (Griffiths and Horne, 1999): previously-reported African occurrences  
404 are parthenogenetic (Martens, 1990). Sexual populations are common in North  
405 America (Delorme, 1971) and China (Yin et al., 1999; Zhang et al., 2015). North  
406 American sexual populations are commonly referred to *L. sappaensis*, although  
407 many authors regard this as a junior synonym of *L. inopinata*, a view subscribed to  
408 here. *Limnocythere inopinata* is strongly euryhaline, although in saline lakes is  
409 restricted to waters with an alkalinity/Ca ratio >1 (Forester, 1983). The species is  
410 intolerant of low dissolved oxygen (Geiger, 1990) but can tolerate seasonal  
411 desiccation (Rossi et al., 2010; Van der Meeren et al., 2010). In the Angamma Delta  
412 samples, the species is moderately common, with both males and females present; it  
413 also occurs in CH60 from the Bodélé Depression, in which it is the most common  
414 species.

415

416 *Cytheridella* cf. *tepida* Victor, 1987 (Fig. 8)

417 The genus *Cytheridella* is most commonly found in North and South America (Park  
418 et al., 2002) but has also been reported from Africa (Klie, 1944; Rome and De  
419 Deckker, 1977; Victor, 1987; Karanovic, 2009). The Angamma Delta specimens  
420 show some resemblance to *Cytheridella tepida* Victor 1987, which is known from  
421 Nigeria, where it is associated with vegetation-rich, gently flowing streams and  
422 springs. It is moderately abundant in the Angamma Delta samples

423

424 *Darwinula stevensoni* Brady and Robertson, 1870) (Fig. 8)

425 This is a common, cosmopolitan species (Mesich, 2000) also reported from East and  
426 North Africa (Martens, 1984a) in a wide range of habitats. It is characteristic of  
427 freshwater, although can tolerate elevated salinity up to 30 gL<sup>-1</sup> (Gandolofi et al.,  
428 2001; Van Doninck et al., 2003) and is found in bicarbonate waters as well as those  
429 dominated by chloride and sulphate (Mezquita et al., 1999). It shows brood care and  
430 so cannot tolerate desiccation (Griffiths and Butlin, 1994). It is moderately abundant  
431 in the Angamma Delta samples and a single specimen was recovered from sample  
432 CH60 from the Bodélé Depression.

433

434 *Candona cf. neglecta* Sars, 1887 (Fig. 8)

435 Members of the genus *Candona* are not widely reported from Africa (Martens,  
436 1984b) although *C. neglecta* has been recorded from North Africa (Martens, 1984b).  
437 Given the morphological variability within *C. neglecta* and the similarity of its shell to  
438 that of several other species (Meisch, 2000) it is possible that the specimens from  
439 the Angamma Delta belong to another species, although they are referred to  
440 *Candona cf. neglecta* here. *Candona neglecta sensu stricto* is a commonly  
441 freshwater ostracod that prefers colder water but can tolerate elevated temperature  
442 and brackish coastal and continental water (Meisch, 2000). It is moderately common  
443 in the Angamma Delta samples.

444

445 *Heterocypris giesbrechtii* (G. W. Müller, 1898) (Fig. 8)

446 This species has been found in Central and East Africa, in waters that are temporary  
447 or that fluctuate in volume and salinity (Martens, 1984b), and in permanent saline  
448 ( $\leq 9.4$  ‰) waterbodies on Aldabra (McKenzie, 1971). It was also abundant as part of

449 a low diversity ostracod assemblage in late Holocene lake sediments from NE  
450 Nigeria (Holmes et al., 1998) and the Faiyoum, in Egypt (Keatings et al., 2010). In  
451 the Angamma Delta samples, it is the most abundant ostracod taxon; it also occurs  
452 in CH60 from the Bodélé Depression.

453

454 *Sclerocypris cf. bicornis* (G. W. Müller, 1900) (Fig. 9)

455 The specimens show some similarity to both *S. bicornis* (G. W. Müller, 1900) and *S.*  
456 *excerta* Sars 1924. Compared to *S. excerta*, the specimens from the Angamma  
457 Delta are more quadrate and show a less prominent posterior point; moreover, this  
458 species has not been found from the Sahara or Sahel region (K. Martens, pers.  
459 comm. 2017). Compared to *S. bicornis*, the specimens from the Angamma Delta are  
460 also more subquadrate; moreover, the juveniles lack the lateral tubercles seen in this  
461 species, although both tuberculate and non-tuberculate forms of the species have  
462 been recorded (K. Martens, pers. comm. 2017). Furthermore, *S. bicornis* has been  
463 reported from West Africa (Gauthier, 1929, 1951) and Egypt (Keatings et al., 2010).  
464 On these bases, the Angamma Delta specimens are referred to *Sclerocypris cf.*  
465 *bicornis*.

466

467 Along with other members of the genus (Martens, 1986, 1988), *Sclerocypris bicornis*  
468 is most commonly found in small pools, which may be ephemeral, although it has  
469 also been found as a minor component of the deepwater fauna of Lake Turkana and  
470 in the Late Holocene sediments of lake Qarun in the Faiyum, Middle Egypt, probably  
471 in association with shallow ( $\geq 8$  m), saline, permanent water (Flower et al., 2006;  
472 Keatings et al., 2010). The species is quite common in the Angamma Delta samples.

473

474 *Sarscypridopsis aculeata* (Costa, 1847) (Fig. 8)

475 This species is typical of smaller waterbodies and can tolerate seasonal desiccation;  
476 it is common in slightly saline waters, with an optimum salinity range of 5 – 10 ‰ and  
477 preference for Na-Cl-type waters (Ganning, 1971; Meisch and Broodbakker, 1993).  
478 In the Angamma Delta samples the species is represented by a single specimen.

479

#### 480 4.2.2 Interpretation of the ostracod assemblages

481 Previous work on ostracods from past and present Lake Chad is sparse. Gauthier  
482 (1939) described living ostracods from several sites on Lake Chad. Zamar and  
483 Tukur (2015) described a small collection of ostracods from sediments of the Bama  
484 Ridge, a beach ridge that lies between 320 and 338 m.a.s.l. to the west of the  
485 present-day lake and marks a mid Holocene highstand; however, some or all of the  
486 material appears to have been misidentified. Within the Lake Chad Basin, but  
487 beyond the Holocene extent of the megalake, Holmes (1997) described a small  
488 collection of wetland ostracod species and Holmes et al. (1998) examined mid to late  
489 Holocene ostracod assemblages from inter-dunal lake sediments. Despite the dearth  
490 of previous studies on ostracods in the Lake Chad region, a reasonable amount of  
491 ecological information is available for the species encountered, as noted above.

492

493 The presence of adult and juvenile ostracod shells suggests that the assemblages  
494 are in situ and have not been subjected to significant post mortem reworking. The  
495 taxa present are all essentially freshwater species although those for which  
496 information is available are also able to tolerate elevated salinity. The occurrence of  
497 *L. inopinata* suggests that if the water were saline, it must have had an alkalinity/Ca  
498 ratio >1 (Forester, 1983). Although the species present are found in a range of

499 habitats, the association of several of the taxa, especially *S. bicornis*, *H. giesbrechtii*  
500 and *C. tepida*, with small, shallow and fluctuating waterbodies, is notable.

501

#### 502 4.3 Molluscs

503 Brown (1994) has summarised studies on the modern molluscs from Lake Chad,  
504 based on Lévêque (1967), Mandahl-barth (1968) and Brown (1974) and this can be  
505 used to interpret the fossil assemblages reported here. Van damme (1984) reports  
506 mollusc assemblages from exposures of sediments from the Falaise d'Angamma  
507 that are attributed to early or Middle Pleistocene sediments, but which are probably  
508 Holocene. Böttcher et al (1972) also report mollusc assemblages the Falaise  
509 d'Angamma, but of early Holocene age, around 10,000 – 9200 cal BP. In both  
510 instances, many of the species are similar to those encountered in our  
511 investigations.

512

513 The molluscs present in the Angamma samples reported here inhabit a wide range  
514 of aquatic habitats (Table 3). Half of the taxa are absent from, or not normally found  
515 in, water that desiccate seasonally and at least two of the taxa are able to tolerate  
516 elevated salinity. Overall, the mollusc assemblages suggest that the palaeo-  
517 waterbody was permanent.

518

#### 519 4.4 Pollen

520 The absence of pollen from all but sample CH60 means that inferences about past  
521 vegetation are restricted to the Bodélé Depression during the late Holocene interval.  
522 The pollen assemblage and frequent *Pediastrum* in sample CH60 indicates that the  
523 sediment was deposited in a shallow water body with fringing bulrush (*Typha*) and

524 sedges (Cyperaceae). Abundant Chenopodiaceae pollen point to the presence of an  
525 arid environment with halophytic vegetation. The low pollen taxa diversity might have  
526 partly been caused by the small basin size and catchment. However, the absence of  
527 any trees or shrubs in the pollen assemblage suggests that the lake was located in a  
528 desert environment.

529

## 530 4.5 Ostracod shell chemistry

### 531 4.5.1 Oxygen isotopes

532 We use the ostracod-shell oxygen-isotope data to estimate the oxygen-isotope  
533 values for the palaeo-lake waters. The oxygen-isotope ratio of aquatic carbonate is  
534 determined by the temperature and water isotope composition of the water from  
535 which the carbonate precipitated and the factors can be described quantitatively  
536 using empirical equations such as that of Kim and O'Neil, (1997). Such equations  
537 assume equilibrium precipitation, yet it is well known that ostracods form shells that  
538 show offsets from equilibrium that are positive, but which vary between taxa (von  
539 Grafenstein et al., 1999; Decrouy, 2012). Of the species analysed here, members of  
540 the subfamily Candoninae have the best constrained vital offset, which is  $+2.2 \pm 0.15$   
541 ‰ (von Grafenstein et al., 1999). Although higher offsets for Candoninae have been  
542 reported (up to +3; Decrouy, 2012), a value of +2.2 is probably appropriate for this  
543 particular location based on the likely chemical composition of the palaeo-lake water.  
544 Offsets for the other taxa are less certain. A value of  $\sim +1$  ‰ has been suggested for  
545 *Heterocypris* (Perez et al., 2013), which accords with data in Burn et al., 2016,  
546 although higher (+1.7 ‰: Lawrence et al., 2008) and lower (+0.54 ‰: Schwalb et al.,  
547 2002) values have also been suggested, albeit with a small sample size in the former  
548 case and large uncertainties over calcification conditions in the latter. For the

549 purpose of the present study, we use a value of +1 ‰. For *Cytheridella* a value of +1  
550 ‰ has been suggested by Meyer et al. (2017) and between 0.1 and 1‰ by Perez et  
551 al. (2013), although poorly constrained in both cases. There are no published  
552 estimates of vital offsets for *Sclerocypris*; comparison of  $\delta^{18}\text{O}$  values for this species  
553 and co-occurring *Candona cf neglecta* in CH38 suggests a value of 0.88 ‰: a similar  
554 approach for *Heterocypris* and *Cytheridella* yields values of +0.4 and +0.6,  
555 respectively, both of which are lower than the values quoted above. However, the  
556 use of co-occurring ostracods of different species to calculate a vital offset is  
557 problematical because it assumes that the individuals calcified under the exact same  
558 conditions of water temperature and water isotope composition, which may not be  
559 the case. For this reason, we prefer to use the values quoted from the literature and  
560 cited above for *Heterocypris* and *Cytheridella*: the value of +0.88 is used for  
561 *Sclerocypris* in the absence of better data, but with the caveat that it is highly  
562 uncertain.

563

564 The offset-corrected  $\delta^{18}\text{O}$  values are used to calculate  $\delta^{18}\text{O}_{\text{water}}$  values using Kim  
565 and O'Neil (1997). In the absence of temperature data for central Africa during the  
566 late Holocene, we assume present-day values taken from Faya-Largeau, Chad  
567 (minimum = January, 20°C, maximum = 34°C, June: IAEA/WMO, 2018). The results  
568 of this exercise (Fig. 10) show large variability, which is unsurprising given  
569 uncertainties in the calcification temperature and the large variation in  $\delta^{18}\text{O}$  values in  
570 ostracod specimens at each level. However, the reconstructions do suggest that the  
571 ambient lake water was moderately to strongly evaporated compared to estimated  
572 rainfall and rainfall-derived runoff for the present-day. Modern water isotope data for  
573 the greater Lake Chad region are summarized in Bouchez et al. (2016). The

574 weighted mean annual  $\delta^{18}\text{O}_{\text{ppt}}$  value for the nearest IAEA GNIP station at  
575 N'Djamena, Chad, is  $-3.8 \pm 1.7$  ‰ (IAEA/WMO, 2018).  $\delta^{18}\text{O}$  values for fluvial inputs  
576 vary between -6 and +3 ‰ for the Chari-Logone (weighted mean annual value = - 3  
577 ‰) an -4 to +8 ‰ (no weighted mean annual value quoted) for the Komadugu Yobe  
578 (Bouchez et al., 2016). The most negative values are associated with the summer  
579 monsoon and the most positive values with dry-season evaporative enrichment.  $\delta^{18}\text{O}$   
580 values in the modern lake vary spatially, with the lowest values (-1 to +4 ‰) in the  
581 Southern Pool, intermediate values (+4 to +7‰) in the Archipelagos and highest  
582 values (+6 to +8 ‰) in the Northern Pool.

583

#### 584 4.5.2 Carbon isotopes

585 The carbonate-isotope ratio of aquatic carbonate is determined by the  $\delta^{13}\text{C}$  value of  
586 total dissolved inorganic carbon (TDIC), which in turn is controlled by equilibration  
587 between TDIC and atmospheric  $\text{CO}_2$ , the balance between aquatic productivity and  
588 decay and inputs of soil-derived carbon from the catchment. The  $\delta^{13}\text{C}$  values  
589 therefore provide information about sources and cycling of carbon within the lake.

590

591 Complete equilibration between TDIC and the atmosphere would yield a  $\delta^{13}\text{C}_{\text{TDIC}}$   
592 value that is between 8.5 ‰ (at 20°C) and 7 ‰ (at 34°C) higher than atmospheric  
593  $\text{CO}_2$ , (Mook et al, 1974), which had a  $\delta^{13}\text{C}$  value of about -6.5 ‰ in the late  
594 Holocene: Leuenberger et al., 1992). Complete equilibration tends to occur in waters  
595 with long residence times. Aquatic plants utilizing TDIC for photosynthesis will  
596 preferentially fix  $^{12}\text{C}$ , leaving the residual TDIC  $^{13}\text{C}$ -enriched, whereas decay or  
597 organic matter releases  $^{12}\text{C}$ -enriched carbon to the TDIC pool (Kelts and Talbot,  
598 1990). Soil-derived carbon reflects catchment vegetation: in an area of



599 predominantly C4 plants, it will lie in the range -9 to -16 ‰ (Smith and Epstein,  
600 1971).

601

602 Variation in the carbon-isotope signatures probably reflects the ecological  
603 preferences of the different ostracod species, because  $\delta^{13}\text{C}_{\text{TDIC}}$  values vary over  
604 small distances in lakes. *Heterocypris giesbrechtii*, which (based on better-studied  
605 congeners, such as *H. incongruens*: Rossi and Menozzi, 1990) is most likely to be a  
606 swimming species, will have a carbon-isotope signature that reflects open-water  
607 TDIC. Members of the genera *Candona*, *Cytheridella* and *Sclerocypris* are  
608 epibenthic, and thus their lower  $\delta^{13}\text{C}$  values (especially those for *Sclerocypris* and  
609 *Cytheridella*), most likely reflect TDIC  $\delta^{13}\text{C}$  values that are influenced by the  
610 mineralization of  $^{13}\text{C}$ -deplete organic matter, pointing to an organic-rich substrate.

611

612 There is lack of covariance amongst the ostracod  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values. Although  
613 this is often taken to indicate the existence of a hydrologically open system (Talbot,  
614 1990), there is often a lack covariance amongst  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values specifically for  
615 biogenic carbonates, such as ostracods, even in closed systems (e.g. Holmes et al.,  
616 1997), probably owing to local habitat controls on the isotopic composition of the  
617 dissolved inorganic carbonate from which the biogenic carbonate precipitated  
618 (Talbot, 1990),

619

#### 620 4.5.3 Trace elements

621 In Kajemarum Oasis, NE Nigeria, Sr/Ca ratios in shells of *Limnocythere inopinata*  
622 have been used as a proxy for past salinity. Values for the Angamma Delta samples  
623 and for the Bodélé Depression sample are towards the low end of the range seen in

624 Kajemarum Oasis (<1 to >10 mmol/mol). However, although it is tempting use this  
625 information to suggest that the Angamma waters were of low salinity, it is not certain  
626 that the same Sr/Ca – salinity relationship observed for Kajamarum also prevailed for  
627 Angamma.

628

#### 629 4.6 Palaeoenvironmental synthesis

630 The interbedded sandstones and mudstones that characterise the lower part of the  
631 Angamma delta deposits appear to be part of an early to middle Holocene  
632 progradational delta-slope succession deposited during the AHP. The age of 7253-  
633 7416 cal BP ( $1\sigma$ ) is consistent with the radiocarbon ages for the base of the delta  
634 published by Servant et al. (1969). It is reasonable to suggest that rivers flowing from  
635 the north were active at that time and this is supported by the presence of lakes in  
636 the Tibesti, such as the Trou au Natron, during the early to middle Holocene  
637 (Kroepelin et al. 2016). However, river discharge must have ceased before 5600  $\pm$   
638 300 BP because the river channels on the delta top are truncated by the beach ridge  
639 along the edge of the delta top (Armitage et al. 2015).

640

641 The bioclastic silty sands that overlie the interbedded delta-slope facies are around  
642 3000 years younger (Table 1) and appear to postdate the end of the (AHP) between  
643 about 6000 and 5000 cal BP (de Menocal et al. 2000; McGee et al., 2013). The  
644 palaeoenvironmental interpretation of this facies has important implications for the  
645 lake-level history of Lake Megachad at the end of the AHP. The occurrence of a  
646 number of ostracod species that are desiccation tolerant coupled with an inferred  
647 water isotope composition that is consistent with evaporative enrichment (Fig. 10),  
648 could indicate that the sediments had been deposited in small, seasonally-

649 desiccated waterbodies that had been isolated following the regression of the mega  
650 lake. This interpretation, if correct, would suggest that the lake had undergone  
651 regression by this time to an altitude lower than 285 – 290 m, the height of samples  
652 CH38 and CH39. However, some of the molluscs found in these samples cannot  
653 tolerate desiccation (Table 3) and while some of the ostracods are desiccation  
654 resistant (Table 2), none is restricted to such environments and some are unable to  
655 tolerate desiccation. The oxygen-isotope values point to evaporative enrichment of  
656 the waterbody, although not to values that are any greater than in the modern lake.  
657 The elevation of the bioclastic sediments coincides with the elevation of the Bahr el  
658 Ghazal sill. We suggest that the Bahr el Ghazal sill would have restricted the  
659 circulation of waters between the southern and northern sub-basins. Given the  
660 geomorphological evidence that influent streams from the north had dried up prior to  
661 5700 BP, and almost all of the water flowing into the palaeolake were derived from  
662 the south through the Chari delta, with restricted flow from south to north through the  
663 Bahr el Ghazal, it is proposed that the northern sub-basin became slightly  
664 evaporated with positive  $\delta^{18}\text{O}$  values similar to those of the northern sub-basin of  
665 Lake Chad today. On balance then, the sedimentological, stratigraphical,  
666 palaeoecological and isotopic evidence, when considered collectively, is best  
667 interpreted as representing deposition in the littoral zone of the megalake. Given the  
668 elevation CH 38 and CH 39 well below the 339m Angamma highstand shoreline, it is  
669 possible that they represent a regressive deposit, formed as the lake level fell.

670

671 According to the reconstruction of Armitage et al. (2015) the level of palaeolake  
672 Megachad was even lower between 4.7 and 3.2 ka. The evidence for the low lake-  
673 level comes from OSL ages of dune sands in the Erg du Jourab (Mauz and Felix

674 Henningsen 2005) of  $4700 \pm 200$  OSL BP,  $4700 \pm 300$  OSL BP,  $3900 \pm 400$  OSL  
675 BP,  $3400 \pm 200$  OSL BP,  $3100 \pm 200$  OSL BP and  $3100 \pm 200$ ; the two older ages  
676 are from elevations of 242 m and 266m respectively. Their ages overlap with the  
677 calibrated radiocarbon ages for the shells from the Angamma delta, while their  
678 elevations are around 20 to 45 m lower than the bioclastic silty sands on the  
679 Angamma delta, and both locations included layers of diatomite above the dune  
680 sands demonstrating that the dunes had been flooded by a lake transgression. The  
681 younger dune sand samples are from elevations between 278 and 289 m, which are  
682 within 5 to 10m of the elevation of the bioclastic silty sands and the scenario that  
683 best fits the observations is that the lake level fell abruptly after 5500 cal BP to an  
684 elevation beneath that of the dunes (less than 242m) causing them to be reactivated.  
685 Using the published OSL dates of  $5400 \pm 500$  BP at 333 m a.s.l. and  $4700 \pm 200$  BP  
686 at 224 m a.s.l. (Armitage et al., 2015) suggests a lake regression of  $\sim 16$  cm yr<sup>-1</sup>,  
687 which is within the interannual range of lake-level variations for the recent past. The  
688 dune age of 4700 BP overlaps the calibrated radiocarbon age for the bioclastic silty  
689 sands and this can be reconciled if the dune ages are interpreted as dune  
690 stabilisation ages, caused by a lake level rise and transgression at 4700 cal BP. This  
691 scenario is common in endorheic basins in the Sahara (Bristow and Armitage, 2016)  
692 but contrasts with groundwater-fed lakes, such as the Ounianga, that persists to this  
693 day (Kroepelin et al., 2008).

694

695 During the later stages of the lake within the Bodélé depression, the  
696 palaeoecological remains point to the existence of a small and shallow waterbody,  
697 which the oxygen-isotope data confirm was evaporated. Pollen assemblages  
698 indicate that the lake was fringed with emergent macrophytes and set within an arid

699 catchment. Despite evaporative enrichment, there is no palaeoecological evidence  
700 that this waterbody was saline, however, and it is possible that the very final stages  
701 of the lake are not represented by the samples that were investigated.

702

## 703 **5. Conclusions**

704 Stratigraphic, palaeoecological and isotopic evidence from the Angamma Delta  
705 confirms that there were complex lake-level and palaeohydrological changes in  
706 Lake Megachad at the end of the African Humid Period. The cusate geomorphology  
707 of the Angamma delta indicates that it was wave dominated. However, within the  
708 delta front sediments, wave ripples are relatively rare and a combination of storms  
709 and fluvial floods are interpreted to control sedimentation. Storm waves shaped the  
710 delta top forming a beach and interacted with unidirectional (offshore) currents and a  
711 supply of fine grained sand to generate HCS within on the delta front. The  
712 succession appears to shallow up with HCS being replaced by swalley cross-strata  
713 higher in the section, possibly associated with increased sediment supply and  
714 deposition during fluvial floods. Storm events and the presence of contorted and  
715 slumped beds could be an indication of neo-tectonic movements within the Chad  
716 Basin. However, there are many possible trigger mechanisms for liquefaction in the  
717 Angamma delta including sediment loading, unloading due to erosion or changes in  
718 lake level, or storm events, and it is not possible to determine the trigger mechanism  
719 without further work. Integrating the chronology of the deltaic sediments with other  
720 published ages for palaeolake Mega-Chad suggests that lake level was at its  
721 maximum elevation around 5700 cal BP. The level then fell sharply, to an elevation  
722 well below that of the sill between Lake Chad and the Bodélé depression allowing  
723 reactivation of dunes at elevations of 242 and 266m. The lake level then rose again

724 flooding across the sill between Lake Chad and the Bodélé depression,  
725 transgressing across the dunes around 4700 cal BP and depositing the bioclastic  
726 silty sands on the Angamma delta. The radiocarbon ages of shells suggest that the  
727 lake transgression lasted for around 500 years before the water level fell beneath the  
728 level of the sill. During this time the ostracods and molluscs suggest a littoral lake  
729 environment with moderate evaporation of shallow waters. The final stages of the  
730 lake in the Bodélé depression , around 1000 cal BP were marked by a shallow,  
731 evaporated waterbody fringed by emergent macrophytes. Our results suggest that  
732 Lake Megachad did not undergo a single regression during the mid Holocene, but  
733 rather experienced a series of abrupt fluctuations. These findings may be consistent  
734 with evidence for fluctuating climate and environment at the end of the AHP (Liu et  
735 al., 2007; Kroepelin et al., 2008; Amaral et al., 2013).

736

### 737 **Acknowledgements**

738 We thank Richard Preece, Tom White and Jon Ablett for help and advice with  
739 mollusc identifications; Koen Martens, Dave Horne, Patrick De Deckker and Andy  
740 Cohen for discussions about the ostracod taxa and their ecology; Melanie Leng for  
741 facilitating some of the stable isotope analyses; Miles Irving for drafting most of the  
742 figures; two anonymous reviewers for their constructive comments. Radiocarbon  
743 dates were supported by the UK Natural Environment Research Council  
744 Radiocarbon Facility Allocation 1216.0407.

745

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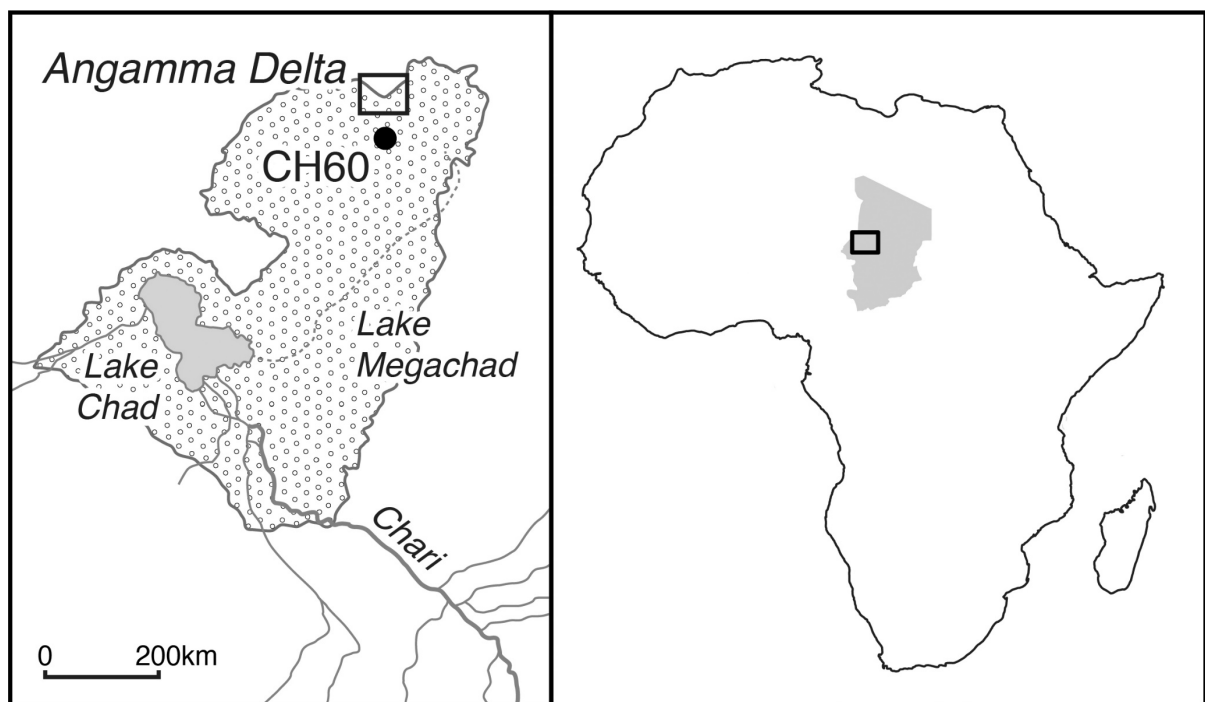
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1094 **Figures**



1095

1096 Fig. 1. Map of palaeolake Mega-Chad showing the location of the Angamma Delta at  
1097 the northern end of the lake with a map of Africa inset.

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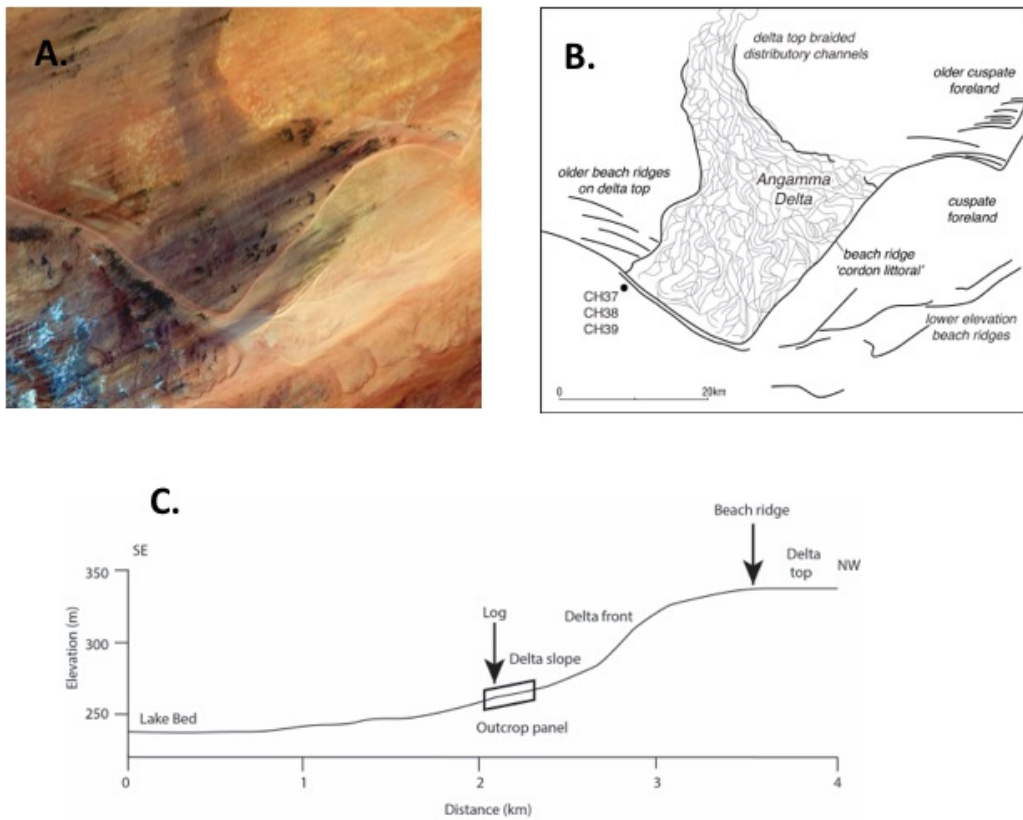
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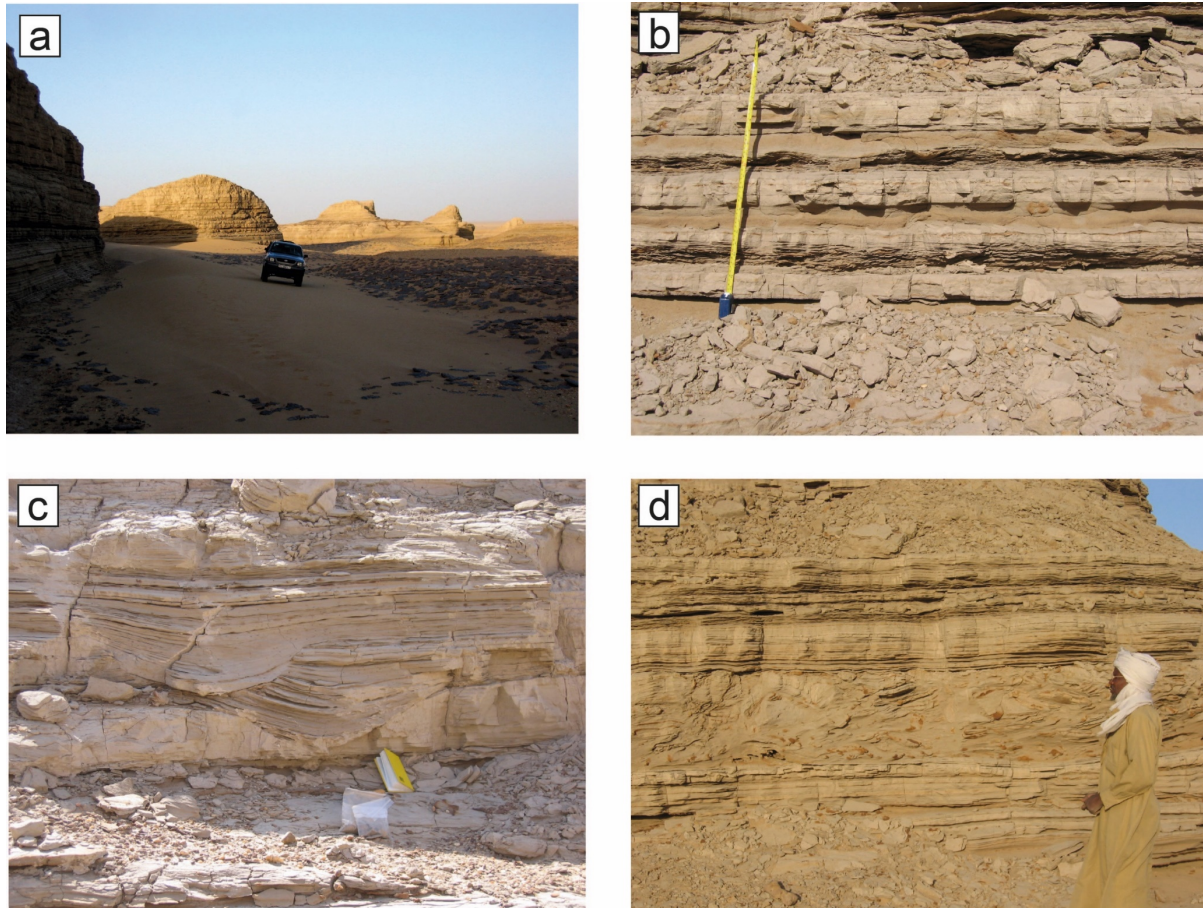
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1114 Fig. 2. A) Satellite image of the Angamma Delta. B) Geomorphological interpretation

1115 of the satellite image above showing beach ridges and the braided fluvial channels

1116 that are preserved on the delta top, modified from Drake and Bristow (2006) and

1117 Schuster et al. (2005). C) Topographic profile across the Angamma delta and the  
1118 northern part of the Bodélé depression that represents the palaeobathymetry of the  
1119 northern margin of the lake. Inset box shows the location of the sedimentary log and  
1120 the outcrop photograph.  
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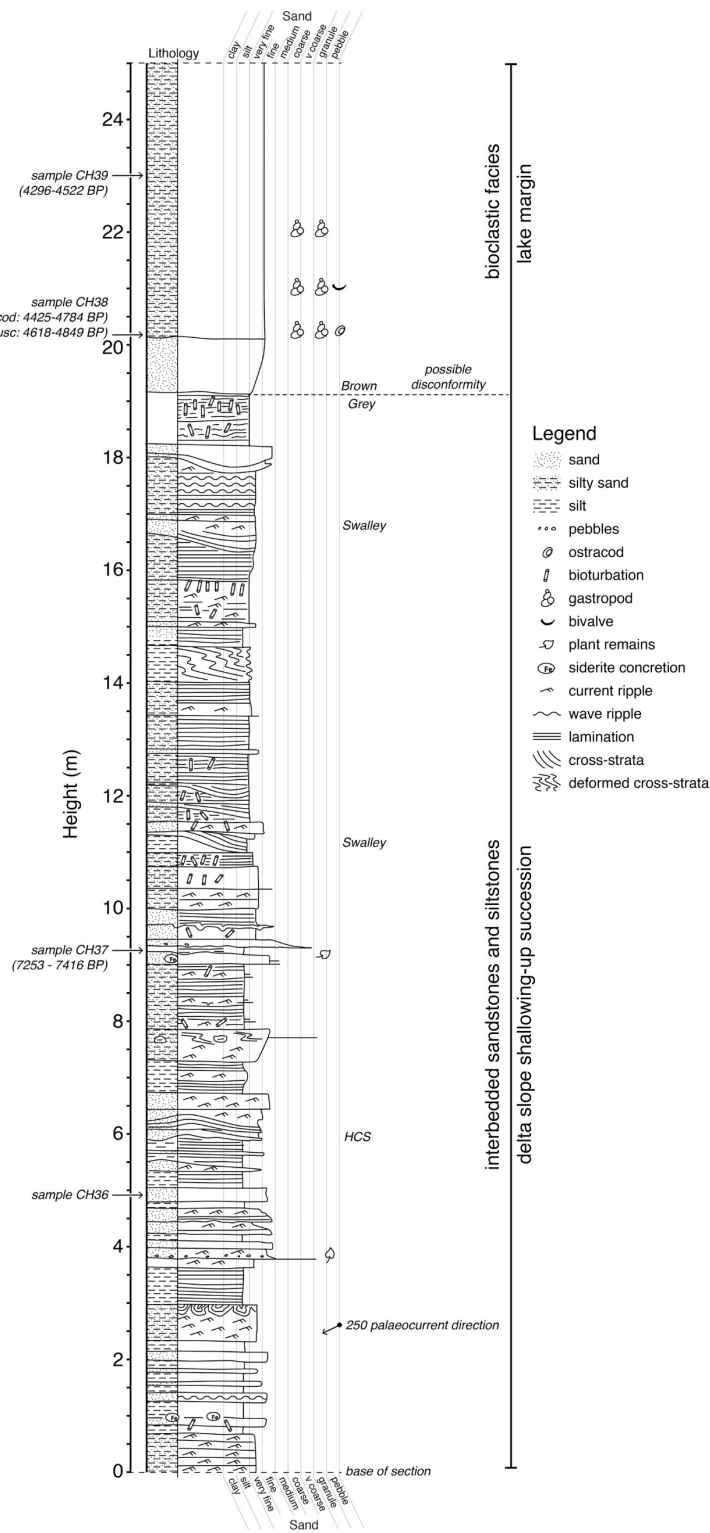
1123 Fig. 3. Geomorphology and sedimentary structures and bedding within the  
1124 Angamma Delta.

1125 (a) Field photograph of giant yardangs eroded from the western margin of the  
1126 Angamma Delta, pick-up truck for scale.

1127 (b) Horizontal laminated siltstones interbedded with sharp based current ripple  
1128 laminated sandstones that are possible turbidite deposits, tape measure with  
1129 0.6 m rule for scale.

- 1130 (c) Swaley cross-strata interpreted to be formed during storm events, 19 cm field  
1131 note book for scale.
- 1132 (d) Partially fluidised bed with slump folds, disrupted and contorted bedding  
1133 underlain by horizontal sandstone beds and overlain by horizontal sandstone  
1134 beds, person for scale.

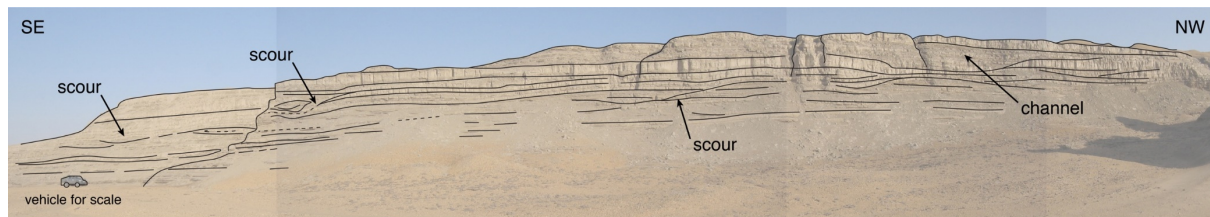




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1136 Fig. 4. Graphic sedimentary log through 25 m of sediments on the western side of  
 1137 the Angamma Delta showing the grainsize of the sediments, bed contacts, bed  
 1138 thickness and sedimentary structures.

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1141 Fig. 5. Annotated photograph of canyon wall trending southeast - northwest and  
1142 incised into the delta slope showing the geometry of the deltaic deposits that are cut  
1143 by erosion surfaces defining lense-like sandbody geometries from channels (one of  
1144 which is marked). Vehicle for scale lower left.

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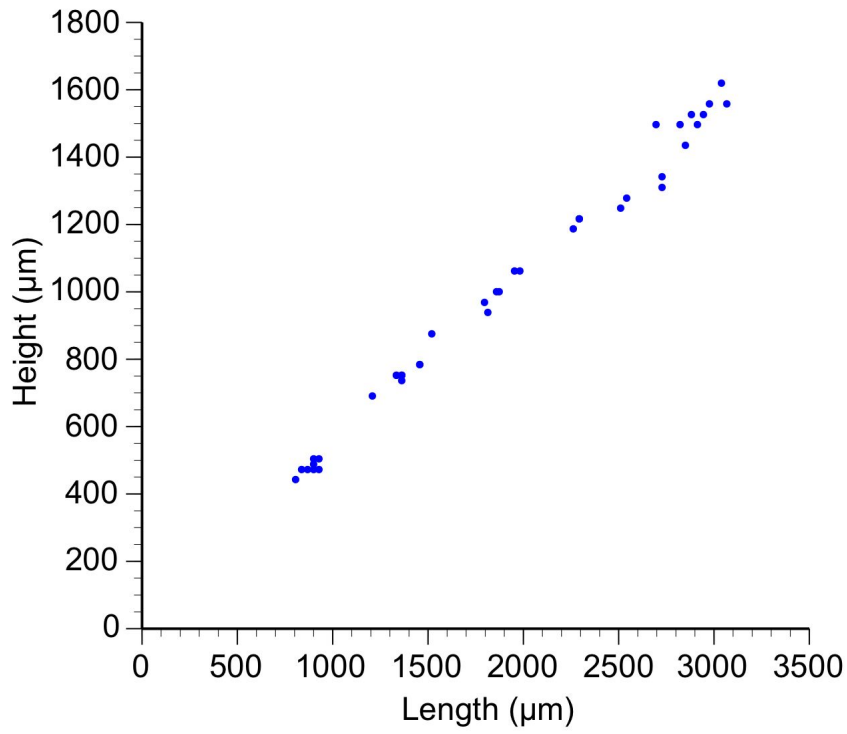
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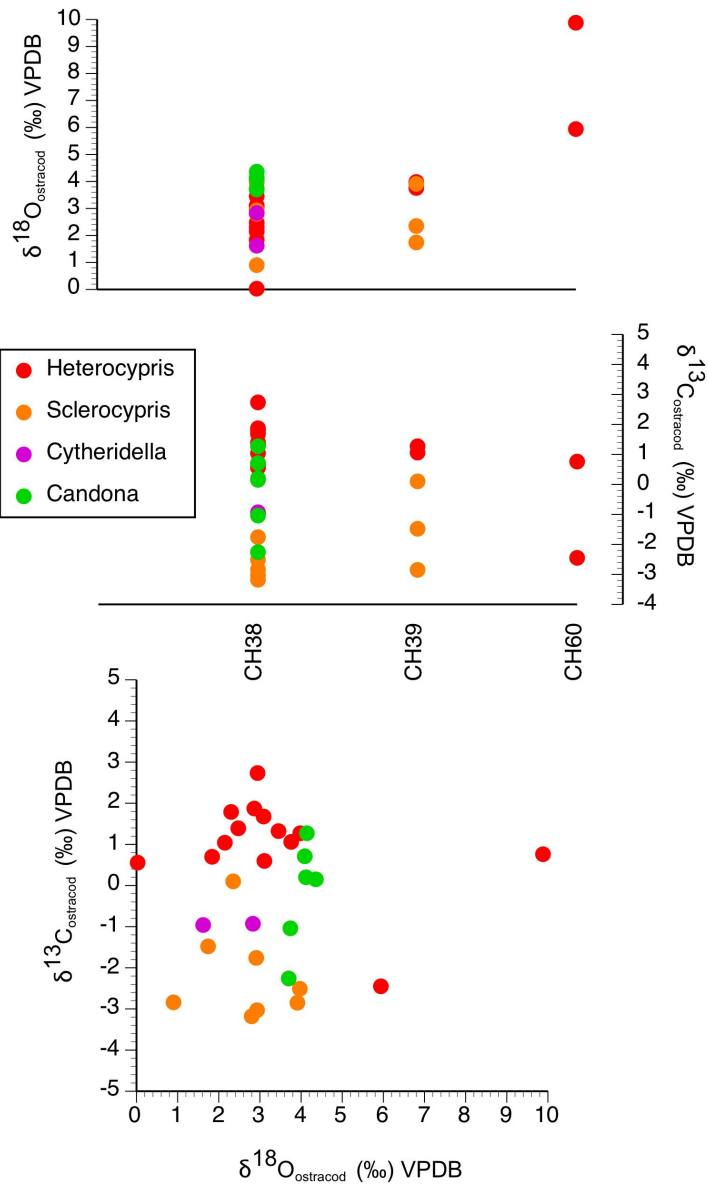
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1157 Fig. 6. Length-height plot for specimens of *Sclerocypris* cf. *bicornis* from CH38

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1163 Fig. 7. Oxygen and carbon isotopes in ostracods and oxygen v carbon isotope cross

1164 plot

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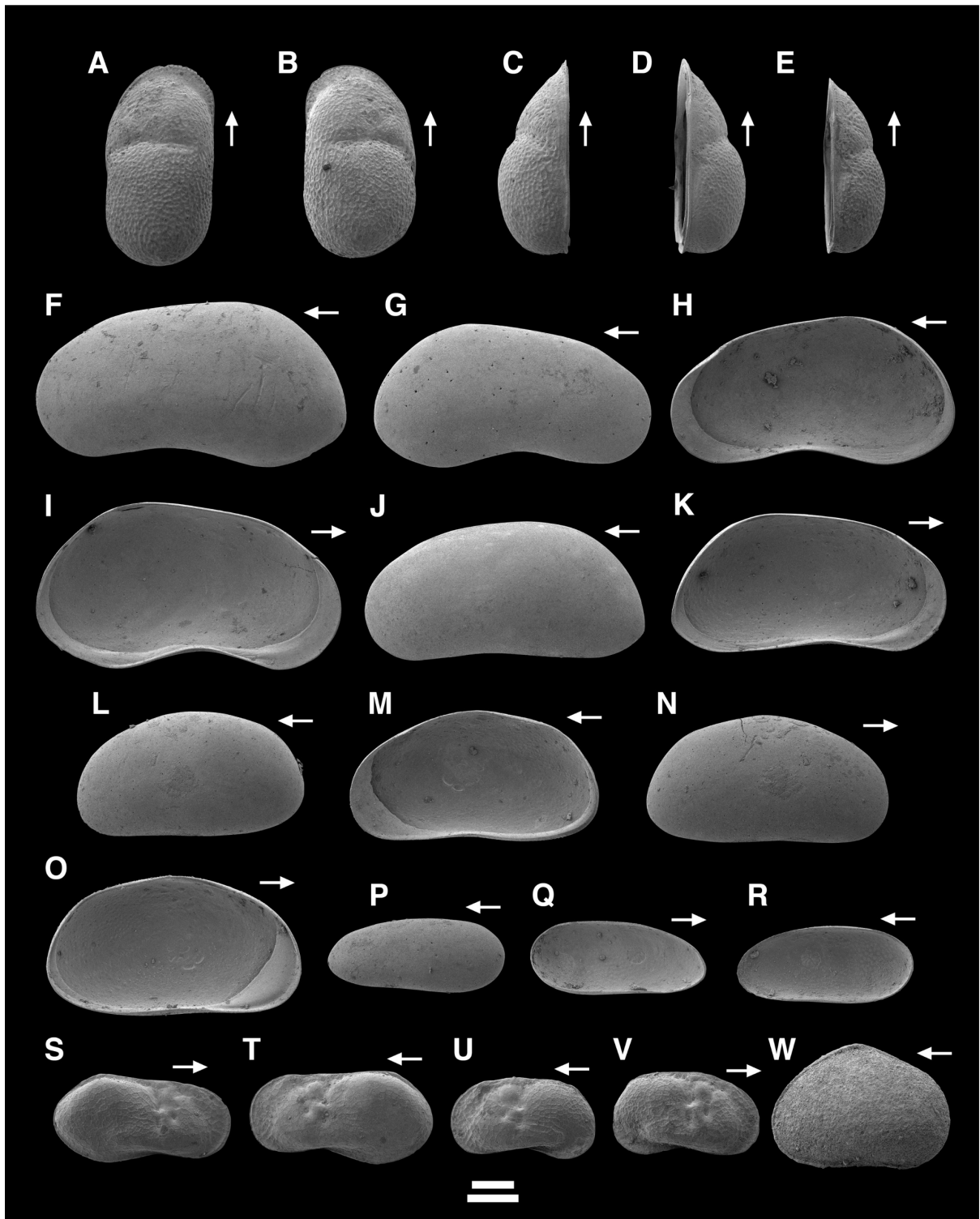
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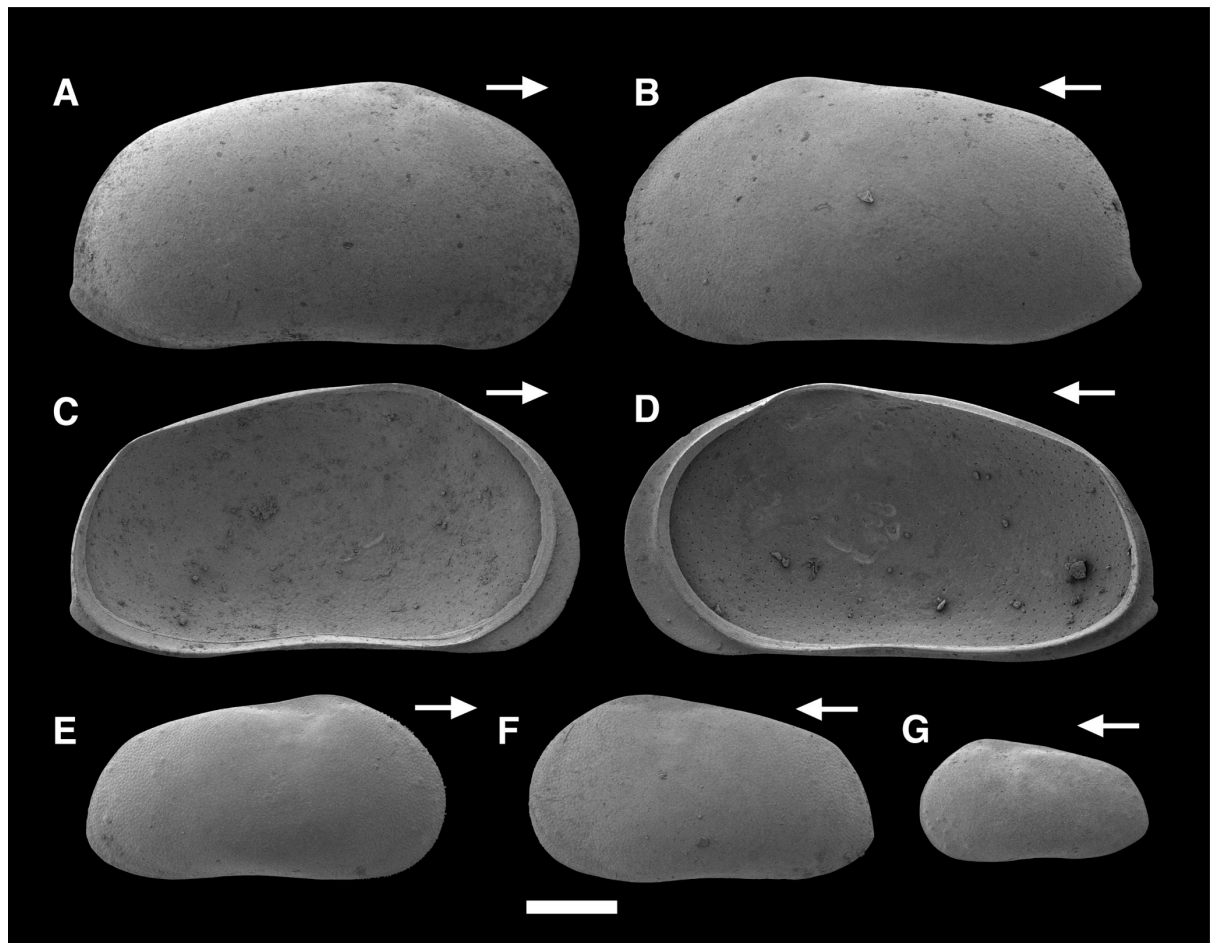
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- 1175 Fig. 8. SEM images of ostracods. All specimen from sample CH38. Scale bars:  
1176 200µm. Upper, A-O; lower P-W. Arrows point in anterior direction.
- 1177 (A) *Cytheridella tepida*. External lateral view of female right valve.  
1178 (B) *Cytheridella tepida*. External lateral view of female left valve.  
1179 (C) *Cytheridella tepida*. Dorsal view of female left valve.  
1180 (D) *Cytheridella tepida*. Dorsal view of female right valve.  
1181 (E) *Cytheridella tepida*. Dorsal view of A-1, right valve.  
1182 (F) *Candona cf neglecta*. External lateral view of left valve.  
1183 (G) *Candona cf neglecta*. External lateral view of right valve.  
1184 (H) *Candona cf neglecta*. Internal lateral view of right valve.  
1185 (I) *Candona cf neglecta*. Internal lateral view of left valve.  
1186 (J) *Candona cf neglecta*. External lateral view of left valve.  
1187 (K) *Candona cf neglecta*. Internal lateral view of left valve.  
1188 (L) *Heterocypris giesbrechtii*. External lateral view of left valve.  
1189 (M) *Heterocypris giesbrechtii*. Internal lateral view of right valve.  
1190 (N) *Heterocypris giesbrechtii*. External lateral view of right valve.  
1191 (O) *Heterocypris giesbrechtii*. External lateral view of left valve.  
1192 (P) *Darwinula stevensoni*. External lateral view of left valve.  
1193 (Q) *Darwinula stevensoni*. Internal lateral view of left valve.  
1194 (R) *Darwinula stevensoni*. External lateral view of left valve.  
1195 (S) *Limnocythete inopinata*. External lateral view of male right valve.  
1196 (T) *Limnocythete inopinata*. External lateral view of male left valve.  
1197 (U) *Limnocythete inopinata*. External lateral view of female left valve.  
1198 (V) *Limnocythete inopinata*. External lateral view of female right valve  
1199 (W) *Sarscypridopsis aculeata*. External lateral view of carapace from left side.



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1201 Fig. 9. SEM images of ostracods (continued) All specimen from sample CH38 Scale

1202 bar: 500 $\mu$ m. Arrows point in anterior direction.

1203 (A) *Sclerocypris cf bicornis*. External lateral view of right valve

1204 (B) *Sclerocypris cf bicornis*. External lateral view of left valve

1205 (C) *Sclerocypris cf bicornis*. Internal lateral view of right valve

1206 (D) *Sclerocypris cf bicornis*. Internal lateral view of left valve

1207 (E) *Sclerocypris cf bicornis*. Internal lateral view of A-1, right valve

1208 (F) *Sclerocypris cf bicornis*. Internal lateral view of A-1, left valve

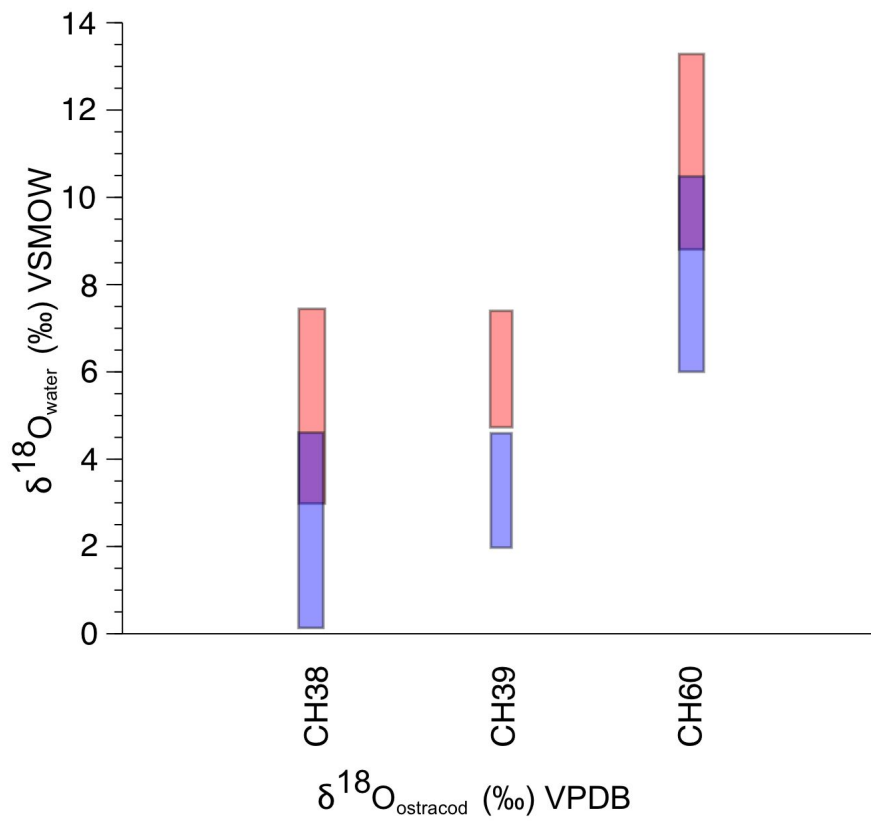
1209 (G) *Sclerocypris cf bicornis*. Internal lateral view of A-2, left valve

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1216 Fig. 10.  $\delta^{18}\text{O}_{\text{lake water}}$  values inferred from  $\delta^{18}\text{O}_{\text{ostracod}}$  at 21.2 ° (blue shading) and 31.4

1217 °C (red shading) water temperature.

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1227 **Tables**

1228 Table 1. New radiocarbon dates from the Angamma Delta section (CH37-CH39) and  
 1229 published date (Armitage et al., 2015) from the Bodélé Depression (CH60)

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Sample code	Laboratory reference	Material	Radiocarbon age ( $^{14}\text{C}$ BP)	Calendar age range ( $2\sigma$ ) (BP)	$\delta^{13}\text{C}$ ‰ VPDB
CH37	Beta-480211	Charcoal	6370 ± 30	7253 - 7416	-28.3
CH38-ostracod	Beta-480212	Ostracod shells - <i>Sclerocypris</i> cf. <i>bicornis</i>	4050 ± 30	4425 - 4784	-2.0
CH38-gastropod	SUERC-17169	Gastropod shells	4204 ± 37	4618 - 4849	-4.5
CH39	SUERC-20101	Ostracod shells - <i>Sclerocypris</i> cf. <i>bicornis</i>	3962 ± 37	4296 - 4522	-2.6
CH60	SUERC-18366	Bivalve shells - <i>Coelatura aegyptiaca</i>	1061 ± 37	926 - 1055	+0.3

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1233 Table 2. Ostracod occurrences in Angamma Delta samples (numbers of valves  
 1234 counted)

	CH38	CH39	CH60
Dry weight of sediment (g)	30.8	9.4	6.9
<i>Limnocythere inopinata</i>	15	0	11
<i>Cytheridella tepida</i>	42	10	0
<i>Darwinula stevensoni</i>	13	0	1
<i>Candona</i> cf. <i>neglecta</i>	20	0	0
<i>Heterocypris giesbrechtii</i>	231	6	3
<i>Sclerocypris</i> cf. <i>bicornis</i>	54	6	0
<i>Sarscypridopsis aculeata</i>	1	0	0
Total	376	22	15

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1237 Table 3. Mollusc occurrences in Angamma Delta samples. For occurrences, √  
 1238 denotes presence, no symbol denotes absence. For ecological preferences, √  
 1239 indicates that the species is associated with that environment or condition, no  
 1240 symbol denotes no information and x indicates that the species is absent from that  
 1241 environment or is not known to tolerate the condition. Ecological data from Brown  
 1242 (1994), Van Damme (1984) and Ibrahim et al. (1999).

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	CH38	CH59	CH60	River	Lake	Pond	Elevated salinity	Dessication
Gastropoda								
<i>Biomphalaria pfeifferi</i>	√	√		√	√	√		x
<i>Bulinus cf. jousseaumei</i>	√			√				x
<i>Corbicula consobrina</i>	√			√	√	√		
<i>Valvata nilotica</i>	√			√	√	√		
<i>Gabiella tchadensis</i>	√			√	√			
<i>Cleopatra bulimoides</i>	√			√	√	√		
? <i>Lymnaea natalensis</i>	√			√	√	√		(x)
<i>Melanoides tuberculata</i>	√	√		√	√	√	√	x
<i>Bellanya unicolor</i>	√			√	√	√		x
Bivalvia								
<i>Sphaerium hartmanni courteti</i>	√				√			
<i>Pisidium pirothi</i>	√			√	√	√	(√)	x
<i>Coelatura aegyptica</i>		√	√	√	√	√		

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1264 Table 4. Ostracod shell stable-isotope and trace-element data  
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	<i>H. giesbrechtii</i>		<i>S. bicornis</i>		<i>C. tepida</i>		<i>Candona cf. neglecta</i>		<i>L. inopinata</i>
	$\delta^{13}\text{C PDB}$	$\delta^{18}\text{O PDB}$	$\delta^{13}\text{C PDB}$	$\delta^{18}\text{O PDB}$	$\delta^{13}\text{C PDB}$	$\delta^{18}\text{O PDB}$	$\delta^{13}\text{C PDB}$	$\delta^{18}\text{O PDB}$	Sr/Ca
	‰ VPDB	‰ VPDB	‰ VPDB	‰ VPDB	‰ VPDB	‰ VPDB	‰ VPDB	‰ VPDB	mmol/mol
CH38	0.70	1.84	-2.51	3.97	-0.93	2.83	-2.26	3.70	4.30
CH38	1.87	2.86	-3.03	2.93	-0.96	1.62	0.15	4.36	6.36
CH38	1.79	2.30	-3.18	2.80			1.27	4.14	4.95
CH38	0.56	0.03	-2.84	0.90			0.71	4.09	4.96
CH38	1.04	2.15	-1.76	2.91			-1.04	3.74	6.07
CH38	1.39	2.47					0.20	4.12	
CH38	1.68	3.09							
CH38	1.32	3.45							
CH38	2.73	2.94							
CH38	0.60	3.11							
CH39	1.27	3.98	-2.85	3.91					
CH39	1.06	3.76	0.10	2.35					
CH39			-1.48	1.74					
CH60	1.27	3.98							2.97
CH60	1.06	3.76							2.77
CH60									2.73

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