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1 Holocene East African monsoonal variations recorded in wave-
2 dominated clastic paleo-shorelines of Lake Abhe, Central Afar
3 region (Ethiopia & Djibouti)

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19
20 **Abstract**

21 In tropical Africa, Late Pleistocene -Holocene climatic fluctuations heavily impacted the continental hydro-systems. The timing
22 and magnitude of 'African Humid Period' hydrological dynamics (AHP; ~14.8 – ~5.5 ka BP) are not yet fully understood due to
23 the extreme variability in African geomorphic patterns and the complex network of past air convergence boundary shifts. The
24 investigation of the functioning of specific hydro-sedimentary basins is needed to improve our knowledge on the AHP spatial
25 and chronological patterns over the continent. In this paper we present a revised Holocene lake level curve of Lake Abhe: an
26 endorheic basin situated in the Afar Depression at the northern extremity of the East African Rift System (EARS). Located at
27 the boundary of the Red Sea and the Indian and Atlantic Oceans' air moisture mass fluxes, the Lake Abhe watershed

28 represents a system sensitive to changes in Northeast African hydro-climate. Based on numerous littoral lacustrine
29 geomorphic features and 53 related radiocarbon ages on stromatolites and nearshore deposits, we have defined the main
30 Holocene lake highstands proposing a detailed paleo-shoreline record. The first rise in water level is dated to ~11.1 ka cal. BP,
31 while the Maximal Holocene Highstand Shoreline (~420 m a.s.l.) was rapidly reached around ~10 ka cal. BP. Evidence of the
32 impact of the 8.2 ka North Atlantic cooling event is recognised with slow lake level regression until ~8.3 ka cal. BP and with an
33 abrupt drop at ~8 ka cal. BP. The resurgence of humid conditions was recorded from ~7.7 ka cal. BP to ~4.6 ka cal. BP.
34 Radiocarbon ages on littoral materials allow us to propose a Holocene subsidence rate of the Abhe basin axial valley.
35 Furthermore, multiphase sill overflow incisions towards the downstream Hanle basin and transient fluvial connectivity with
36 the upstream Ziway-Shala basin, indicate strong geomorphic controls on lake highstand elevations. In summary, Lake Abhe
37 Holocene fluctuations are the result of combined hydro-climatic, tectonic and local geomorphic controls. However, the
38 confrontation with others East African basins allows us to discuss the transport of moisture masses dynamics into the
39 northern EARS during the Holocene.

40

41 Keywords: East African Rift System, Central Afar Region, Lake Abhe, Paleo-shorelines, African Humid Period,
42 Holocene, 8.2 ka event, Tectonic deformation

43

44 **1 – Introduction**

45

46 Today, the tropics concentrate the maximum amount of solar energy and redistribute it in the form of
47 air and moisture masses to latitudes between 30° N and 30° S by atmospheric and oceanic circulation
48 (<https://earthobservatory.nasa.gov/features/EnergyBalance/page3.php>). The intertropical solar-
49 irradiation-induced precipitation/evaporation balance is at the base of Earth's atmospheric convection,
50 and is thus considered one of the main driving processes of global climate evolution (Mohtadi et al.,
51 2016).

52 From the Late Pleistocene to the Mid-Holocene periods, Africa was the theatre of significant changes
53 in hydro-climatic conditions, which mainly followed precessional pacing (i.e., every 21,000 years;
54 (Rossignol-Strick et al., 1982; Mohtadi et al., 2016; Skonieczny et al., 2019). From ca. 14.8 to 5.5 ka, the
55 precession-driven increase of the northern hemisphere insolation magnitude seems to have been at

56 the origin of the northward migration of the rainfall belt associated with the Intertropical Convergence
57 Zone (ITCZ; deMenocal et al., 2000; Haug, 2001; Viste and Sorteberg, 2013). This period, referred to as
58 the African Humid Period (AHP; deMenocal et al., 2000; Tierney et al., 2011), is characterized by
59 enhanced precipitation in the tropics, the northward spread of vegetation cover (Hopcroft et al., 2017)
60 and the massive reactivation of fluvial systems (Revel et al., 2015; Schefuß et al., 2016; Mologni et al.,
61 2020) leading to the development of numerous lakes (Kutzbach and Street-Perrott, 1985; Gasse, 2000;
62 Holmes and Hoelzmann, 2017). However, the general timing (onset and termination) and timing
63 offsets of the AHP between Western and Eastern Africa, as well as between the Sahelian and
64 Equatorial latitudes, need to be fully understood (deMenocal et al., 2000; Shanahan et al., 2015;
65 Collins et al., 2017). The establishment of short-term (200-400 yr) African arid episodes, likely
66 synchronous with the North Atlantic cooling events (HE1, YD, 8.2, 4.2 ka events; Rohling and Pälike,
67 2005; Collins et al., 2013, 2013; Rasmussen et al., 2014), bring to question the causal link between
68 high and low latitudes, the role of the tropics in these cooling episodes (Clement et al., 2001; Davies et
69 al., 2014), the mechanisms of the AHP (Shanahan et al., 2015), and the local non-linear response of
70 African sedimentary basins to global climatic trends.

71 The numerous lakes of the 3000 km-long and N-S-tending East African Rift System (EARS) have proven
72 to be excellent recorders of past climate evolution, providing both long and high-resolution archives
73 (Gasse, 2000). However, the extensional tectonics (Déprez et al., 2013; Stamps et al., 2014, 2020)
74 coupled with the extreme contrast between high precipitation rates in the highland catchment zones
75 and high evaporation in the lowlands ('amplifier lake' patterns; Trauth et al., 2010) may induce a
76 modulated response of lacustrine hydro-sedimentary systems to atmospheric inputs from continental
77 climatic change. During the AHP, the enhanced atmospheric pressure gradient between the East
78 African Monsoon and the Indian Summer Monsoon (ISM) would have led to an eastward migration of
79 the Congo Air Boundary (CAB; Camberlin, 1997; Braconnot et al., 2008). These patterns induced an
80 incursion of the West African Summer Monsoon (WAM) air masses along the EARS, causing a non-
81 linear response of hydro-sedimentary systems to ITCZ north-south oscillations (Fig. 1a; Hailemichael et

82 al., 2002; Tierney et al., 2011; Costa et al., 2014; Junginger et al., 2014). The acquisition of paleo-
83 hydrological data from southern to northern EARS ranges adds relevant information to our
84 understanding of moisture mass convergence zone dynamics during the AHP, which would have
85 impacted, sometimes over the course of a few hundred years, the equilibrium and paleoenvironmental
86 stability of ancient ecosystems.

87 The Abhe Lake basin, located at the northern extremity of the EARS (~12° N), is the widest and longest
88 rift-controlled sedimentary basin of the Afar Depression. It derives its source in the Ethiopian
89 Highlands (Fig. 1b, 1c). Currently reduced to a small (~350 km²) and shallow (max. depth <30 m)
90 waterbody (Wood and Telling, 1988), Lake Abhe experienced a highstand resulting in a large
91 freshwater lake during the Early Holocene (~7000 km² and ~230 m max. depth; Gasse, 1975; Khalidi et
92 al., 2020). The fluctuations of Lake Abhe's water level seem to follow the main AHP dry-wet oscillations
93 (Gasse, 2000), producing some very marked littoral geomorphic features. Although they represent a
94 reliable indicator of the paleo-hydrological evolution history of the basin, these littoral geomorphic
95 features have never been documented or investigated in the Abhe region prior to this study.
96 Deciphering the precise control of geomorphic systems over both hydro-climatic and geodynamic
97 forcing is essential for a more robust interpretation of past climate-change data.

98 In the 1970s, Gasse and colleagues documented the paleoecological and lake level history of Lake
99 Abhe based on lacustrine nearshore and offshore deposits (Rognon and Gasse, 1973; Gasse, 1975,
100 1977; Gasse and Street, 1978; Gasse et al., 1980). In this study we update and improve knowledge of
101 Lake Abhe's paleo-hydrological fluctuations with better chronological and spatial control, and through
102 a consideration of the effects of tectonic and geomorphic processes on past hydrological variations.
103 We present a hydro-climatic record established using multiple paleo-shoreline geomorphic features
104 spanning the last ~11 ka of Holocene Lake Abhe. Our record is age-controlled by 53 new and existing
105 dates on foreshore/nearshore carbonate deposits such as stromatolites and *Melanooides tuberculata*
106 shells. With the aim of determining the precise paleo-hydrological history of Holocene Lake Abhe, our
107 results are combined with tectonic deformation kinematics data in order to take neo-tectonic control

108 and deformation of the investigated geomorphic markers into account. New archaeological and
109 perilacustrine sedimentary sequence chronologies provided in Khalidi et al. (2020) are cross-
110 referenced with our data to better constrain lake-level reconstruction. Furthermore, earth surface
111 processes which may have affected lake hydrological fluctuation dynamics (e.g. overflow sill incision,
112 temporary fluvial interconnection with adjacent basins) are documented and discussed in the text.

113 Finally, our study is compared with other regional lacustrine paleo-hydrological records with the aim
114 of estimating the influence of monsoonal variations on a N-S transect of the EARS during the Holocene
115 period.

116

117 **2- Regional settings**

118

119 **2.1 – Volcano-Tectonic control of the Abhe basin formation**

120

121 Lake Abhe is located in the southeastern part of the Tendaho-Gobaad graben, which is the widest and
122 longest basin within the Afar Depression. The extensional tectonics occurring in the Afar are associated
123 with the triple junction of the divergent Somalian, Arabian and Nubian plate boundaries (Fig 1b). Over
124 its 1.8 Myr-long development, the Tendaho-Gobaad graben has been the locus of interactions
125 between three extensional systems related to the EARS, the Red Sea and the Aden ridges (Manighetti
126 et al., 2001; Audin et al., 2004). It is considered the southwestern tip of the Red Sea propagating on
127 land (Nubia-Arabian boundary), overlapping with the Aden Ridge (Somalia-Arabian boundary), while
128 the extensional structures related to the northern tip of the EARS are observed in its southern margin
129 along the Main Ethiopian Rift (MER; Fig 1c; Acton et al., 1991).

130 East African continental rifting began between ~31 and ~28 Ma following the emplacement of the
131 flood basalts traps forming the Ethiopian and Yemeni high plateaus (current highlands, Hofmann et al.,
132 1997; Tiercelin and Lezzar, 2002; Macgregor, 2015). Soon after, starting around ~25 to ~20 Ma,

133 continental extension formed the large escarpments bounding the Afar Depression and developed the
134 filling of large Miocene basins. The Stratoid series, a 1.5-km-thick basaltic sequence which covers most
135 of the Afar region, was dated between 4.4 Ma and 0.4 Ma (Kidane et al., 2003; Varet, 2018). From 1.8
136 Ma, continental extension in the Central Afar was focused across several NNW to NW-trending active
137 rift segments, concentrating tectonic deformation and, in most cases, magmatic activity. They formed
138 a complex network of normal faults and several sub-rifts, horsts and grabens, to which the Tendaho-
139 Gobaad graben belongs (Abbate et al., 1995). Many fissural volcanic flows related to the emplacement
140 of the Stratoid series occurred between ~ 2 and ~ 0.5 Ma across the Tendaho-Gobaad graben,
141 evidencing the link between the outflows of Stratoid magma and the collapse of the Tendaho-Gobaad
142 graben (Acocella, 2010). Holocene rifting activity is evidenced by some volcanic cones and domes
143 associated with NNW-trending sub-rifts in its northern part (Fig 1c; Barberi and Varet, 1977;
144 Manighetti et al., 1998), and by the probable NE tilting of the basin plain proposed by Gasse (1975).
145 This recent surface subsidence would have induced the capture of the Awash River (Gasse, 1975; Fig 1)
146 leading to the current hydrological basin patterns. From recent geodetic studies (Dobre et al., 2017),
147 the current NE-trending extension across the Tendaho-Gobaad graben does not exceed 7 mm/yr. The
148 subsidence velocity of the inner floor cannot be easily estimated and no clear current slip on the
149 bounding normal faults of the graben is detected from long InSAR time series (Pagli et al., 2014;
150 Dobre et al., 2017). Temtime et al. (2018) identified an elongated NE-trending zone of present day
151 subsidence reaching ~ 4 cm/yr over the 2004-2010 period from InSAR data, which they attributed to
152 magmatic and hydrothermal activity below the basin.

153 These data indicate how tectonic control represents the primary driver of large-scale landform-
154 formation-processes of the Abhe Lake basin. The strong extensional deformation of the crustal layer
155 would have affected the sedimentogenetic processes and the organization of the lacustrine-induced
156 morphological elements investigated in this study. However, the strong impact of past hydrological
157 dynamics on sedimentary East African systems produced a considerable imprint on landscape
158 morphologies (Nilsson, 1940; Garcin et al., 2012; Melnick et al., 2012).

159

160 **2.2 – Hydro-climatic control of the Abhe basin and previous paleo-lacustrine studies**

161

162 The Abhe basin sedimentary system receives its water catchment from the Awash River which derives
163 its source in the region of highland Ethiopia southwest of Addis Ababa. Freshwater supplies are mainly
164 generated by African monsoon precipitation on the Ethiopian Highlands during the rainy season from
165 June to September (200 to 300 mm/month). Air moisture flows moving into northern Ethiopia
166 originate from the Gulf of Guinea, the Indian Ocean, the Red/Mediterranean Seas and the Arabian
167 Peninsula (Viste and Sorteberg, 2013).

168 Descending the highlands, the Awash River is joined by many affluent rivers until it meets the Lower
169 Awash Plain (4000 km²), which coincides with the northwestern part of the Tendaho-Gobaad graben.
170 Here, the river course continues, without any perennial freshwater supply, to Gamari and Afambo
171 Lakes (339 m a.s.l.) and to its final destination, endorheic Lake Abhe (Fig 1c; 240 m a.s.l.). Over the
172 course of one year, the Abhe basin is seasonally supplied by smaller local catchments at the
173 extremities of the graben, namely the Gobaad (South-East) and the Logiya (North-West) Rivers.
174 Presently, the Central Afar region is characterized by a hyper-arid climate with 200 mm/yr precipitation
175 and with an evaporation rate of 3500 mm/yr (Rognon and Gasse, 1973; Gasse, 1990; Varet, 2018).

176 At the tropics, monsoonal rainfall magnitude variations over the last 1 Ma have been attributed to the
177 northward migration of the rainfall belt associated with the ITCZ together with precession-driven
178 insolation changes (deMenocal et al., 2000; Gasse, 2000; Tuenter et al., 2003; Mohtadi et al., 2016;
179 Lamb et al., 2018; Skonieczny et al., 2019). The large offset between the local precipitation and
180 evaporation rates indicates that the orbital-induced monsoon activity in the Ethiopian highlands was
181 primarily responsible for the hydro-sedimentary balance which generated the Late Pleistocene-
182 Holocene synthem of the Abhe basin.

183 Except for a few Mio-Pliocene sequences located along the margins of the Afar Depression (Tiercelin,
184 1981; Gasse, 1990), the lacustrine sedimentary accumulation in the Central Afar region began during
185 the Early Pleistocene (Gasse et al., 1980; Boucourt et al., 1985), which seems synchronous with the
186 Tendaho-Gobaad graben formation. However, sedimentary deposits developed within a topography
187 similar to the current one started around 130 ka BP (Fontes and Pouchan, 1975; Gasse, 1975).

188 The first research conducted on Abhe basin sedimentary formations occurred in the 1970s and was
189 focused on paleoecological reconstruction from deep (offshore) lacustrine deposits (Faure et al., 1971;
190 Rognon and Gasse, 1973; Gasse, 1975, 1977). Three main lacustrine sedimentary sequences spanning
191 100 to 3 ka BP, were identified by Gasse (1977) based on study of exposed sections and offshore
192 sediment cores: ~100-70 ka BP (Abhe I), 60-18 ka BP (Abhe II - Abhe III), and 10-4 ka BP (Abhe IV -
193 Abhe V; Fig. 1c) corresponding to the AHP (~14.8 and ~5.5 ka; Shanahan et al., 2015).

194 Gasse (1975, 1977) reconstructed past Lake Abhe highstands on the basis of conventional radiocarbon
195 ages and elevations from nearshore to offshore carbonate deposits, with little link to paleo-shoreline
196 morpho-sedimentary structures and dynamics. A recent revision of these previous data with the
197 integration of new calibrated ages performed on littoral carbonates and archaeological assemblages
198 has led to the definition of a new water-level variation curve for Holocene Lake Abhe (Khalidi et al.
199 2020). In this study, we have further developed this approach with the aim of reconstructing
200 Holocene monsoon-induced paleo-hydrological variations in the Lake Abhe basin with higher precision.
201 During three survey campaigns (2018-2020), we documented different littoral morpho-sedimentary
202 features with specific regard for a wide range of littoral-related geomorphic and depositional
203 environments such as wave abrasion platforms/notches, overflow sill incisions, beach ridges, Gilbert-
204 deltas, pedological horizons, gravity-induced deposits and fluvial or fluvio-lacustrine deposits. This
205 approach, combined with a chronology assured by more than fifty radiocarbon ages on littoral
206 carbonates, allows us to indicate the precise paleolake outlines and elevations and their relation to
207 volcano-tectonic rifting-related deformation.

208

209 **3 – Methodology**

210

211 **3.1 – Differential GPS (DGPS) survey and Digital Elevation Model (DEM) calibration**

212

213 3.1 – DGPS survey and DEM calibration

214

215 During the 2018-2020 field campaigns, a large survey over the entire Abhe basin focused on the
216 identification and GPS measurement of its littoral morpho-sedimentary and geomorphic elements. The
217 detailed description and definition of the precise geomorphic elements considered as paleo-shoreline
218 markers are provided in the results section 4.1 and in Sup. Mat. A. Additionally, in order to provide
219 more accurate elevations from new GPS plots, and with the aim of testing and validating our previous
220 measurements and calibrating the elevation data from Gasse (1975), a systematic DGPS survey was
221 performed using the Trimble Geo 7X differential GPS, with a double frequency receptor (L1 = 1575.46
222 MHz, L1 = 1227.60 MHz). The vertical standard deviation of our measurements range between 1 and 2
223 cm. This analytical value error is similar to the Trimble error value. Due to the quasi-absence of
224 geodetic reference points in the investigated region and with the aim to validate our elevation
225 measurements, we compared our DGPS measurements with elevations from 2 DEMs: a total of 54
226 DGPS measurements were compared with SRTM® (www2.jpl.nasa.gov/srtm) and TanDEM-X
227 (<https://tandemx-science.dlr.de>; Rizzoli et al., 2017) DEMs (Fig 2). Results show generally accurate
228 elevations from the SRTM® (30 m resolution), however, they do not follow the homogeneity of DGPS
229 measurements (Fig 2). In contrast, elevations extracted from the TanDEM-X data (12m resolution;
230 Rizzoli et al., 2017) are more coherent with the DGPS data but with a systematic offset of -10 m (Fig
231 2). The elevation differences between our DGPS and the TanDEM-X data (with a +10 m offset applied)
232 never exceed ~5 m (1.7 m average), while the differences with the STRM data sometimes reach ~10 m
233 (2.3 m average). Taking into account the precision required for the elevation measurements, as well as
234 the coarse sampling resolution of the SRTM DEM, the TanDEMx (+10 m offset) was preferred for the

235 calibration of the GPS measurements obtained from three years of field campaigns and from previous
236 location data provided in Gasse (1975). For each surveyed site (see 4.2.1), a topographic profile was
237 generated with the location of geomorphic features (Supp. Mat. A), providing support for each paleo-
238 shoreline level definition (see 4.2.2).

239

240 **3.2. - Age control**

241

242 With the aim of proposing a robust age model of the main paleo-shorelines recorded across the Lake
243 Abhe basin, a calibration was performed on 18 AMS ¹⁴C conventional ages on stromatolites, 9 on
244 nearshore deposits and 26 on selected ages published in Gasse (1975).

245 The revision of previous Holocene Abhe paleolake water level data is based on the calibration and the
246 re-calculated DEM elevations of ¹⁴C ages on nearshore carbonate deposits published by Gasse (1975).

247 The 26 published ages were selected according to the following criteria: a) only ages from littoral
248 lacustrine (nearshore) material from the Abhe basin were considered; b) using stratigraphic
249 information provided by Gasse (1975), inversed ages or ages from reworked sediments were not
250 considered for the age model building; c) ages from paleo-shores of residual lakes were not considered
251 (e.g., Gamari or Gargori residual lakes, Fig 4a); d) ages with incomplete geographic location
252 (incomplete coordinates) do not allow a proper correlation with the DEM processing data and were
253 not used for the model building; e) ages on archaeological material were not considered. The selected
254 results are presented in [Tab 1](#).

255 The age calibration ([Tab 1](#)) was performed using the OxCal 4.3 program (Bronk Ramsey, 2001). For
256 conventional ages on littoral lacustrine carbonates, the calibration parameters were selected from the
257 IntCal13 atmospheric calibration curve (Reimer et al., 2013). Considering that the estimated reservoir
258 age effect for Early Holocene biogenic lacustrine carbonates is <100 yrs (Khalidi et al., 2020), which
259 remains lower than the majority of the age standard deviations, no reservoir correction was applied.

260 We are aware that a reservoir age definition based on a single parallel dating is insufficient. However,
261 comparison and cross-reference with archaeological data from littoral sequences presented in Khalidi
262 et al. (2020) helped us to constrain possible reservoir age effect anomalies (see Section 5.2). The ages
263 considered here used 95.4% ranges and are reported in median ka cal. BP (calibrated kilo annum
264 before 1950).

265

266 **4 – Results**

267

268 **4.1 - Paleo-shoreline geomorphic features of the Lake Abhe basin**

269

270 Over the ~200-km-wide Abhe basin, more than one hundred littoral lacustrine geomorphic features
271 were identified and mapped during the survey campaigns (2018-2020; [Sup. Mat. A](#)). The detection of
272 more or less pronounced wave-cut abrasion features, typical beach pebble band concentrations, and
273 continuous paleo-beach ridges, combined with other shore-related morpho-sedimentary facies
274 (Gilbert-type deltas), ascribes the Abhe paleolake to the wave-dominated clastic shoreline lake-type as
275 defined by Schuster and Nutz (2018).

276 In this section, we first describe and define the geomorphic marker types observed in the field. We
277 then present each surveyed site together with the list of shorelines. Finally, we summarize the dataset
278 of paleo-shorelines recognized across the Abhe basin.

279

280 **4.1.1 – Wave-cut and overflow abrasion platforms**

281

282 Wave energy enhanced by wind direction and magnitude can produce more or less developed shore
283 platforms, abrasion ramps or cliff notches (Schuster et al., 2003; Trenhaile, 2015). The maximal

284 estimated extension of the Holocene Lake Abhe waterbody (7000 km²; Gasse 1975) presents the
285 conditions to produce well-defined littoral abrasion features on clastic supports.

286 Throughout the Abhe basin, abrasion platforms and notches vary in relation to lateral changes in rock
287 resistance. Wave-cut abrasion platforms that developed in acidic volcanic contexts such as rhyolitic
288 domes and associated eruptive material (ignimbrite and pyroclastic rocks) are more pronounced,
289 horizontally shaped, and frequently associated with well-developed cliff notches (Fig 3.a). Most wave-
290 cut abrasion platforms that developed on basaltic stratoid series outcrops present a sub-horizontal or
291 seaward-sloping shore morphology (gently sloping abrasion platforms as defined by Sunamura (1992)
292 without any cliff notch features; Fig 3.b). The width of sub-horizontal abrasion platforms on rock
293 outcrops can vary between ~10 and ~200 m, while when wave-cut incision occurs on pre-existing
294 sedimentary deposits, the platform can achieve up to 1 km in width (Fig 3.c). Abrasion platform shape
295 and dimension are even determined by long-term exposure to wave erosion, lake waterbody
296 amplitude, which coupled with wind action influences the magnitude of waves, and by local
297 topographical morphology. For example, a short-term wave-cut abrasion on a plunging cliff could
298 produce a slight incision observable in a slope failure feature (1 to 3 m high). We refer to this kind of
299 slight pronounced abrasion feature with the term 'abrasion terrace'. Abrasion platforms on overflow
300 sill ridges represent important additional geomorphic paleo-shoreline markers which are likely
301 produced by the mixing of wave abrasion and traction current action in the overflow direction.

302 Because of the homogeneous composition of the basaltic stratoid series layers, vertical differential
303 erosion is reduced. Consequently, shore platforms as described in Bird (2008) are almost absent across
304 the Abhe basin. The great number of normal faults which characterize the Tendaho-Gobaad semi-
305 graben could highly affect the recognition and the preservation of continuous morphologic elements
306 of paleo-shorelines. For this reason, during surveys, sites with clear structural patterns were preferred
307 over sites where faulting has probably been involved in sedimentary/erosion dynamics.

308 The high extensional deformation kinematics of the Tendaho-Gobaad semi-graben do not constitute
309 the only obstacle to abrasion platform recognition. The Late Holocene to current hyper-arid

310 environmental conditions in the Central Afar region, which combine high silicate weathering, wind
311 action, thermoclastism and pronounced slope erosion due to soil/vegetation cover denudation, can
312 also deform the original abrasion features. For this reason, coupling the study of wave-cut and
313 abrasion platforms with supplementary geomorphic and morpho-sedimentary markers is needed for
314 the definition of paleo-shorelines.

315

316 **4.1.2 – Flat-pebble/cobble beaches**

317

318 Beach ridges, with their related clastic components, are commonly used as large-scale wave-related
319 geomorphic shoreline features (Hesp, 2006). Across tropical Africa, these continuous, mound-shaped,
320 and linear accumulations of clastic sediments parallel to coastlines are widely used as markers of
321 regressive paleo-shorelines at the end of the AHP (Garcin et al., 2012; Forman et al., 2014; Schuster
322 and Nutz, 2018). In the Central Afar region, beach ridge features are sporadic due to the adverse
323 climatic conditions which prevent their preservation. The total absence of vegetation and soil cover on
324 the basin margins can induce high-energy rill erosion during the short-lived summer rainfalls, causing
325 the vertical displacement of beach ridge clastic components. In addition, heavy silicate weathering and
326 enhanced thermoclastism can produce a high amount of fresh detrital material both from outcrops
327 and from inherited littoral clastic material even on gentle sub-horizontal slope contexts. This
328 pronounced formation process of *in situ* clastic material can lead to some uncertainty in the
329 recognition of continuous ancient beach ridges. Thus, only the coarser components, namely isolated
330 and continuous beach ridges, were considered as paleo-shoreline markers (Sup. Mat. A). Flat-
331 pebble/cobble beaches are often associated with wave-cut and/or abrasion platform surfaces
332 confirming the pertinence of these clastic geomorphic elements as wave-related shoreline features
333 across the Lake Abhe basin.

334 In this study, the majority of the paleo-shore/foreshore zones are defined by the detection of pebble
335 (4-64 mm diameter) and cobble (64-256 mm diameter; Bird, 2008) concentration bands (Sup. Mat. A).

336 These can be found as occasional (a few meters long), intermittent or continuous bands. The
337 roundness, flat morphology and smooth surface of clasts allow them to be distinguished from fluvial or
338 slope detrital material. Often, identification of the mineralogical cortege allows us to differentiate local
339 beach clastic products from allochthonous fluvial assemblages. Indeed, some beaches supplied with
340 the mineralogical assemblage of the Awash River were documented (see section 4.2, WK site). The
341 width of the pebble/cobble concentration bands ranges between 2 and 30 meters, and their upper
342 limit is considered a paleo-shoreline elevation marker.

343

344 **4.1.3 – Gilbert-type deltas and foreshore morphologies**

345

346 Deltas are shore-related morpho-sedimentary features. Their evolution thus directly reflects changes
347 in water level. In particular, Gilbert-type deltas (first described by Gilbert, 1885) are widely considered
348 to be sensitive recorders of short-term water-level changes (Massari and Colella, 1988; Colella and
349 Prior, 1990; Breda et al., 2007; Gobo et al., 2014, 2015). The distinctive tripartite architecture permits
350 their formation context to be recognized: the subhorizontal bottomset corresponds to the prodelta
351 formation processes, the inclined foreset indicates a subaqueous delta-slope progradation, while the
352 horizontal topset suggests a mouth-bar (fluvial) or slack-water deposition context (Smith and Jol, 1997;
353 Bridge, 2003).

354 Across the Abhe basin, the Gilbert-type delta is a recurrent landform developed at the mouth of
355 feeder canyons or at the distal end of dejection cones along the basin margins (Fig 3d). Gilbert-type
356 delta facies are generally well-defined and easily recognized (Fig 3e). These peculiar characteristics
357 allow us to easily recognize this kind of paleo-shoreline morphology preserved in the landscape at
358 distinct elevations. The majority of the tributaries of the lower Awash valley are characterized by a
359 high-energy regime due to the high altitudinal gradient between the marginal horsts and the axial
360 alluvial plain. This gradient is at the origin of the formation processes which give deltas a well-defined

361 shape during lake-level highstands: a pronounced slope failure and well-sorted clastic material
362 between the feeder mouth and the delta topset (Fig 3d).

363 In the EARS lakes, deltas are the combined product of fluvial and wave influences (Schuster and Nutz,
364 2018). Their shape evolution is driven by wave-induced erosion, and by transport and deposition of
365 river-derived sediment (Anthony, 2015). In the Abhe basin, wave action on delta landforms is
366 evidenced by the frequent association of a flat delta apex associated with abrasion platforms (Fig 3d).
367 The specific description of each Gilbert-type delta facies used as a paleo-shoreline marker in this study
368 is provided in the description of the surveyed sites (section 4.2).

369

370 **4.2 –A detailed record of the Holocene paleo-shorelines of Abhe basin**

371

372 **4.2.1 – Description of the surveyed sites**

373

374 Sixteen sites, presenting an assemblage of geomorphic feature types described in section 4.1, were
375 selected from the survey dataset and are summarized in this section. The spatial location of these sites
376 is provided in Fig 4a, while a detailed documentation of each surveyed site is provided in Sup. Mat. A
377 (specific site map, photos, and topographic profiles with the location of the inventoried geomorphic
378 markers). Our results demonstrate the existence of multiple paleo-shoreline sequences from 300 to
379 420 m a.s.l., which are summarized in Fig 4b. Here, we summarize the characteristics of the paleo-
380 shorelines recognized at the 16 surveyed sites, which we present in alphabetic order.

381 The AFBO site ('Afambo'; Sup. Mat. A .2) is a thin ridge located on the foot wall of the stratoid series
382 which separates the Gamari and Afambo lake waterbodies. Survey profiles were carried out on the S-
383 W flank of the hill and on a headland crest that drops into Afambo Lake. Two well pronounced sub-
384 horizontal abrasion platforms associated with beach pebble concentrations were recognized at 399
385 and 380 m a.s.l. A stromatolitic cover was observed at 370 m a.s.l.

386 The ALEK site ('Ara le Koma'; [Sup. Mat. A .3](#)) is located on a fault scarp on the south ridge of the
387 Gobaad valley. Two paleo-shoreline levels were recognized: the upper one (402 m a.s.l.) is
388 characterized by a sorted line of beach clastic material (cobbles), which likely forms a beach ridge; the
389 lower (389 m a.s.l.) is clearly marked by a well-developed abrasion platform on lacustrine deposits, and
390 by a beach ridge feature composed of cobbles and sub-rounded blocks. Additionally, one stromatolitic
391 level was documented at 441 m a.s.l., and a second stromatolitic cover at 385 m a.s.l.

392 The AYSO site ('Aysilo'; [Sup. Mat. A .4](#)) is located on the northern flank of the Gobaad valley as a foot
393 wall plateau mainly composed of acidic volcanism products (rhyolitic flows and ejecta) which are
394 intercalated in the stratoid series. Three main paleo-shorelines were recognized. The upper level, at
395 411 m a.s.l., takes the form of a Gilbert-delta apex deposit associated with an overlying sub-horizontal
396 and well-developed wave-cut platform. The median level is void of any clear upper foreshore
397 geomorphic markers (wave-cut features), however a beach pebble concentration band at 382 m a.s.l.
398 indicates a paleo-shoreline level overlying this elevation. A thick and wide stromatolite cover combined
399 with reworked littoral lacustrine material (379 m a.s.l.) is probably associated with the medial paleo-
400 shoreline level. The lower paleo-shoreline level is characterized by a wave-cut abrasion terrace and by
401 abundant beach pebble/cobble concentration lines at 332 m a.s.l. A thin littoral carbonate crust line
402 (microbialite) at 327 m a.s.l. is clearly associated with the lower paleo-shoreline of the AYSO site.

403 The BE site ('Borawli East'; [Sup. Mat. A.5](#)) is situated on the east flank of the Borawli rhyolitic volcano.
404 The BE surveyed profile crosses the limit between the rhyolitic flows and the stratoid series, in which
405 an elevated sedimentary flat-fan deposit formed (as a plateau). It is delimited by the rhyolitic slopes in
406 the west, north and south, and by a basaltic rampart in the east. The basin is supplied by the
407 mountain's detrital load through two feeder canyons and is drained by two dejection channels towards
408 the plain. The sub-horizontal shape of the fan against the steep mountain slopes suggests a Gilbert-
409 type delta morphology. This hypothesis is confirmed by the horizontal topset layers observed at the
410 medial-proximal part of the fan and by the horizontal shape of the fan apex. Thus, we consider the
411 proximal-apex part of the deltaic fan as a marker of an ancient foreshore-line located at 420 m a.s.l. In

412 addition, delta sediments are heavily incised by the current lateral streams which enable the sequence
413 to be easily observed and suggest a post-lacustrine torrential and erosive transport regime, indicating
414 a drop in the water level from the top to the base of the basin, thus confirming our interpretations.
415 Finally, at the BE site, a wave-cut abrasion terrace is documented on the eastern stratoid fault scarp at
416 402 m a.s.l., which is clearly associated with a stromatolitic cover-line at 400 m a.s.l.

417 The BSW site ('Borawli South-West'; [Sup. Mat. A.6](#)), like the BE site, benefits from reduced structural
418 faults, which favored the detection of morphologic features. Additionally, the rhyolitic and ignimbritic
419 outcrops of the southwest flank of Borawli Mountain provide more pronounced wave erosion features
420 (see Section 4.1.1). For this reason, we counted several well-developed paleo-shorelines at this site. At
421 415 m a.s.l. some beach cobbles were observed without any clear upper-shore morphologic marker. At
422 398 m a.s.l., well-developed wave-cut notches and abrasion platforms indicate an ancient shoreline. A
423 few meters below (393 m a.s.l.) a concentration-line of beach cobbles associated with an abrasion
424 platform could belong to the low foreshore zone of the overlying paleo-shoreline or indicate an
425 additional paleolake level. Downward, several wave-cut notches associated with abundant beach
426 pebbles/cobbles and large and well-developed abrasion platforms indicate a series of paleo-shorelines
427 at 387, 381, 372 and 367 m a.s.l. Only one stromatolite level is documented, at 354 m a.s.l.

428 The BW site ('Borawli West'; [Sup. Mat. A.7](#)) is characterized by well-preserved Gilbert-type delta
429 deposits located on the west flank of Borawli Mountain ([Fig 3d](#)). At 391 m a.s.l. a sub-horizontal delta
430 apex, associated with an abrasion terrace feature, indicates a paleo-shoreline at this level. On the delta
431 section, the offshore-transition is clearly observable from fine lacustrine deposits, through the
432 bottomset layers to the foreset progradation ([Fig 3e](#)), suggesting that these deposits and the related
433 paleo-shoreline were probably developed following a regression from the earlier higher lake level. A
434 lower level was also located at 379 m a.s.l., in the form of a well-developed wave-cut abrasion
435 platform on nearshore sediments.

436 The DDa site ('Deversoir Djibouti A'; [Sup. Mat. A.8](#)) is located along the first part of the southern
437 overflow sill channel of the Abhe basin, which drained the lake water towards the Hanle basin ([Fig 1c](#),

438 4a). Along the sill, only abrasion platform features were documented, probably due to the high energy
439 of the drainage that occurred through the channel. This is confirmed by the thick fan deposit
440 composed of coarse sediments located at its mouth. Four overflow abrasion terraces were detected on
441 the flanks, at 403, 392, 385, and 380 m a.s.l.

442 The DSa site ('Deversoir Serdo A'; [Sup. Mat. A.9](#)) corresponds to the proximal part of the northern
443 overflow sill of the Abhe basin, which drained the lake water towards the Hanle basin ([Fig 1c, 4a](#)). The
444 first and the second part of the sill is intercalated by a small alluvial plain. This sill is higher in elevation
445 than the DDa sill, thus it potentially records only the more ancient and higher drainage phases of the
446 paleolake highstands. Our results confirm this hypothesis with the recognition of two overflow
447 abrasion platforms at 419 and 397 m a.s.l.

448 The DSb site ('Deversoir Serdo B'; [Sup. Mat. 10](#)) is located at the beginning of the distal part of the
449 northern overflow sill of the Lake Abhe basin ([Fig 1c, 4a](#)). The overflow channel is not exploitable for
450 paleo-shoreline reconstruction because of the crossing of several faults. However, at the canyon
451 entrance, three paleo-shorelines were documented: the first (403 m a.s.l.) characterized by a well-
452 developed abrasion platform on the stratoid outcrops associated with abundant beach
453 cobbles/pebbles, and the lower two (390, 387 m a.s.l.) characterized by a well-developed abrasion
454 platform on previous lacustrine sediments associated with beach cobbles/pebbles ([Fig 3c](#)).

455 The GTU-AFI site ("Goma'tu-Afa'si"; [Sup. Mat. A.11](#)) is the only site surveyed on the western scarp of
456 the Tendaho-Gobaad graben that provided some paleo-shoreline geomorphic information. The limited
457 data are a result of difficult access to the Magenta massif zone. Nevertheless, two paleo-shorelines
458 were documented. The upper one, recognized at 381 m a.s.l. by some beach ridges and the lower,
459 recognized at 371 m a.s.l. by a large wave-cut abrasion platform on lacustrine sediments. One oncolite
460 level is observed at 369 m a.s.l.

461 The LAK site ('Lakora'; [Sup. Mat. A.12](#)) corresponds to the northern extension of the Borawli massif: a
462 large plateau composed of basaltic stratoid series and crossed by few northwest/southeast faults. One
463 paleo-shoreline level was detected at 385 m a.s.l. in the form of an abrasion platform underlined by a

464 beach pebble concentration (383 m a.s.l.). This site is distinguished by a large stromatolitic shelf
465 positioned on the Lakora plateau and ranging between 376 and 372 m a.s.l.

466 The MN site ('Manda North'; [Sup. Mat. A.13](#)) is located on the eastern fault scarps of the Abhe basin,
467 on well-defined northwest/southeast faults. This site is distinguished by the continuity of a well-
468 developed wave-cut abrasion platform located at 385 m a.s.l. and extending over more than 10 km.
469 This shoreline platform is clearly discernable from the faults because it is incised on previous
470 sedimentary deposits placed on the scarp margin. A stromatolitic dome level ranging between 380 and
471 377 m a.s.l. is likely associated with the shoreline level, while a thin and sporadic littoral carbonate
472 crust (microbialite) was detected at 417 m a.s.l.

473 The MYTA site ('Mokoyta'; [Sup. Mat. A.14](#)) is located on the northern side of the Gobaad valley and is
474 characterized by a stromatolitic cover level at 376 m a.s.l., while no clear paleo-shoreline was
475 detected. A continuous platform surface is located on topographic profiles around 420 m a.s.l., but no
476 other littoral markers were observed.

477 The RSO site ('Raso'; [Sup. Mat. A.15](#)) corresponds to a semi-enclosed basin delimited by
478 northwest/southeast oriented foot-wall basaltic outcrops, located at the eastern margin of the Abhe
479 basin axial valley. This site is distinguished by a gulf or semi-enclosed bay paleo-morphology, where
480 onshore winds would have maintained stable high-water levels producing well-developed littoral
481 geomorphic markers (Bird, 2008). These favorable morphological factors, coupled with systematic field
482 survey allowed us to recognize multiple ancient paleolake levels. Well-defined wave-cut abrasion
483 platforms combined with beach pebble assemblages observed both on the east and west margins of
484 the semi-enclosed bay indicate paleo-shorelines at 419, 410, 403, 393, 389 and 384 m a.s.l. The lower
485 level (384 m a.s.l.) is clearly associated with a thin stromatolitic line located at the same elevation,
486 while two other stromatolitic levels were detected westward at 382 and 376 m a.s.l.

487 The SD site ('Serdo'; [Sup. Mat. A.16](#)) is the northernmost surveyed area which provided some littoral
488 geomorphic markers. The extended plain areas in the north of the basin, which range between 380
489 and 370 m a.s.l., would have produced large foreshore and shoreline geomorphic markers, making it

490 more difficult to recognize paleo-shore limits. However, in the Serdo area, in the northeastern part of
491 the basin, a continuous abrasion platform and a beach pebble ridge permit the definition of two paleo-
492 shorelines at 400 and 389 m a.s.l.

493 The SHEK site ('Shekaito Koma'; [Sup. Mat. A.17](#)) is located on a small basaltic stratoid outcrop at the
494 center of the Abhe basin. Several paleo-shorelines were documented at the surveyed site, which
495 offered the same morphological advantages of a semi-enclosed basin as described for the RSO site.
496 The highest level at 418 m a.s.l. is marked by abundant beach pebble/cobble concentrations belonging
497 to the paleo-shore/foreshore zone. The underlying levels were detected thanks to abundant clastic
498 beach components and well-developed abrasion platforms at 411, 399, 380 and 369 m a.s.l. The
499 platform at 380 m is incised in foreshore deposits. One stromatolitic line was observed between 369
500 and 365 m a.s.l.

501 The WK site ('Wedelli Koma'; [Sup. Mat. A.18](#)) is located in the Det Bahari plain on the right bank of the
502 Awash River. It is represented by a hill composed of Tertiary deposits. A wave cut abrasion platform at
503 418 m a.s.l. is well developed on these deposits and is associated with a large amount of pebbles
504 which indicate an allochthonous origin. The allochthonous mineralogical assemblage could be part of
505 an ancient (Plio-Pleistocene) delta of the Awash River developed at the entrance of the Abhe basin,
506 not far from WK.

507

508 **4.2.2 – Synthesis of the Abhe basin paleo-shoreline sequence**

509

510 A synthesis of results from more than one hundred geomorphic shoreline markers has allowed us to
511 define the characteristics and elevations of the main Holocene lake highstands ([Fig 4b](#)). First, we
512 defined the Maximum Holocene Highstand Shoreline (MHHS), corresponding to the maximum
513 elevation of well-pronounced geomorphic features recognized in each surveyed site. The MHHS
514 elevations are similar in the central and eastern part of the basin and range between 420 m a.s.l.
515 (central-northern basin) and 410 m a.s.l. (Gobaad basin branch, AYSO site). Secondly, we defined other

516 main paleo-shorelines at 403-400, 385-380, 368 and 330 m a.s.l. (Fig 4b, 7b). These levels are
517 intercalated by minor intermediate levels probably corresponding to lacustrine regression steps.

518

519 **4.3 – The Lake Abhe stromatolites**

520

521 Stromatolite structures are the result of the bioaccumulation and precipitation of micritic/sparitic beds
522 from cyanobacteria microbial activity (Casanova, 1991). They are the most used datable markers of
523 lacustrine paleo-shorelines in Eastern Africa (Gasse, 1977; Gasse and Street, 1978; Gillespie et al.,
524 1983; Hillaire-Marcel et al., 1986; Casanova, 1987; Cohen et al., 1997; Garcin et al., 2012). Their
525 structure, mineralogy and isotopic composition provide precise information about their littoral
526 depositional environment, chemical paleo-conditions of the waters, and paleo-hydrological variations
527 (Abell et al., 1982; Casanova and Hillaire-Marcel, 1992; Ghinassi et al., 2012). In this study we present
528 for the first time the macro-/meso-/micro-structure of stromatolites from Lake Abhe with the aim of
529 defining their ecological zonation (shallow/deep-waters), which combined with their radiocarbon
530 measurements adds crucial information to the age model of Lake Abhe paleo-shorelines.

531 A preliminary investigation of over fifty stromatolite specimens allows us to define at least two main
532 facies. Facies-A is characterized by a micritic poorly-laminated microstructure, organized in vertical
533 micro-column growing patterns. Between the micro-columns, several vertical-oriented voids are often
534 intercalated, and form a complex porosity network from the base to the top of the stromatolite. These
535 bioherms form a well-defined macro-structure of large, linked domes (Fig 5a) which can extend over a
536 few kilometers (LAK site, Sup. Mat. A.12), while the diameter and the thickness of a single dome can
537 reach 1.5 m and 60 cm, respectively. The multiple dome superimposition and their vertical fissured
538 surface form complex morphologies which are attributable to the cerebroïd-like facies as described by
539 Casanova and Hillaire-Marcel (1992), Casanova and Thouin (1990), and Cohen et al. (1997). The large
540 thickness of the stromatolitic linked-dome macro-structure and their micro-/meso-structure indicates
541 a deep water zonation at ~20 m depth (Cohen et al., 1997).

542 Facies-B is characterized by a well-developed and continuous laminated micro-structure built of
543 alternating Si-Mg-rich and Ca-rich layers (Fig 5.d). Si-Mg-rich laminations are frequently ripple-shaped
544 over the entire bioherm paleo-surface (Fig 5.c). These stromatolites are a few centimeters thick (max
545 10 cm) and cover clastic blocks or cobbled surfaces, frequently forming oncolite beds (GTU-AFI site,
546 Sup. Mat. A.11). The wave-ripple facies, the oncolite morphology, as well as their irregular surface,
547 indicate a shallow water environment (Casanova, 1986; Schuster and Nutz, 2018). Taking into account
548 the silicic-ferromagnesian regional bedrock sources, Si-Mg-rich layers suggest an increase of detrital
549 delivery events in the lake basin enhancing flow currents, while Ca-rich layers probably suggest periods
550 of lacustrine-dominant biogenic productivity or the onset of evaporitic processes.

551

552 **4.4 – Paleo-shoreline age control**

553

554 During field survey, several stromatolites and near-shore lacustrine carbonate products were
555 inventoried and sampled along the entire Abhe Lake basin. The chronological evolution of lake paleo-
556 shorelines is based on the combination of radiocarbon ages measured on 11 stromatolites of both
557 Facies-A and Facies-B and on 9 new and revised published ages on nearshore deposits (Tab 1; Khalidi
558 et al., 2020). Stromatolites were observed at different altitudes ranging from 420 m to 320 m with a
559 concentration between 380 and 360 m.

560 Stromatolites located in the elevation range of ~381 to ~375 m a.s.l. (Fig 4b, 7b) are often
561 characterized by a superimposition of the well distinguished Facies A and B: Facies-A as a lower and
562 older stromatolite generation and Facies-B as an upper and younger stromatolite generation (Fig 5b,
563 c). The discontinuity between the two facies is marked by a sharp surface of subaerial exposure often
564 filled by a layer of fine sand which can reach 20 cm in thickness as evidenced at the StrIS site
565 (Stromatolis; Khalidi et al., 2020). This facies separation is even characterized by two chronological
566 groups. The older chronological group ranges in date between ~11.1 and ~8.3 ka cal. BP, and only
567 includes ages from Facies-A (Tab 1; ~20 m water depth), suggesting they are related to the paleo-

568 shorelines located around 400 m a.s.l. The more recent facies ranges in date between ~7.7 and ~4.7 ka
569 cal. BP and includes only ages from Facies-B (Tab 1; 0-2 m water depth), suggesting they are related to
570 wave-induced clastic paleo-shorelines located around 385 m a.s.l.

571 A third chronological group was located at the Str20 site (~369 m a.s.l.), which is characterized by the
572 superimposition of three stromatolitic layers similar to Facies-A, and lying on a palustrine deposit
573 (Str20-SU6). The lower layer (Str20-SU5; 11125 ± 113 cal. BP) is separated by a stone-line in the middle
574 (Str20-SU4; 10530 ± 64 cal. BP) and the upper (Str20-SU1; 10447 ± 73 cal. BP) layers, suggest the
575 existence of subaerial exposure conditions in between.

576 ~420 m (MN site) and ~330 m (AYSO site) a.s.l. paleo-shorelines (Fig 4b; Sup. Mat. A.3, A.4, A.13)
577 present characteristics similar to Facies-B, and thus likely belong to the shallow waters ecological zone.
578 A stromatolite specimen from ~440 m (ALEK site) was dated to 30534 ± 248 cal. BP, thus belonging to
579 the last Pleistocene lake highstand (Abhe-III; Gasse, 1977).

580 New and published ages on nearshore deposits (lumachelles, shell-rich deposits, diatomites) follow the
581 stromatolite chronological ranges quite well (Tab 1, Fig 7.a). The published ages also allow for the
582 definition of a lower lake level between 3 and 2.8 ka cal. BP, which corresponds to the paleo-shoreline
583 observed around 330 m a.s.l. (site AYSO; Sup. Mat. A.4). Only three nearshore deposits are located
584 between 440 and 400 m a.s.l., and are dated between 10.4 and 9.8 ka cal. BP. They are probably
585 contemporaneous with the geomorphic features and stromatolites of the MHHS located at 440 and
586 420 m a.s.l. (Fig 4.b, 7.b). Following these results and the respective dating standard deviations, we
587 suggest that the Holocene age of the MHHS was in the range of 10 ± 2.5 ka BP. The offsets of dated
588 littoral features suggest the effect of subsidence due to tectonic deformation which will be discussed
589 below.

590

591 **5 – Discussion**

592

593 **5.1 –Holocene tectonic subsidence of the Abhe basin**

594

595 Because of the large rates of horizontal extension that characterize the Central Afar region and the
596 rifting-related vertical motion at the surface (Manighetti et al., 2001; Doubre et al., 2017), we consider
597 it essential to estimate the potential tectonic impact of Holocene rifting on the Lake Abhe paleo-
598 shoreline sequence before discussing its hydro-climatic relevance. Despite the limited number of
599 available data and dates attesting to vertical slipping of the graben, we attempt to find evidence for
600 variations in the rate of subsidence both in space and time.

601 In the Gobaad branch of the basin (SE, Fig 6a) the MHHS elevation, previously estimated to have
602 occurred ~ 10 ka cal. BP (see section 4.4), differs by ~ 10 m from the northern part of the basin (Fig 4b).
603 Several faults are identified and if the relative vertical gap is due to tectonics, the Holocene fault
604 activity would have corresponded to a mean vertical slipping rate of ~ 1 mm/yr cumulated for the faults
605 observed in this area. This value is of the same order of magnitude as the vertical slipping rate of the
606 faults in the Central Afar determined from topographic profiles across fans by Manighetti et al. (2001).
607 Over the same period of time, the authors estimated a vertical slipping rate of 1 to 3 mm/yr for faults
608 located in the basins to the north, namely Hanle and Gaggade (Fig 1c). Their measurement along the
609 Gobaad fault agrees with a vertical rate of 0.6 mm/yr for the last 1.8 Ma. However, even if the rate
610 over the last 10 ka period could not be measured, it is probably higher by a factor of 2 or 4, as has
611 been demonstrated for most of the faults in the Central Afar (Manighetti et al., 2001).

612 In the Logiya-Saha branch of the basin (NW, Fig 6a), we can compare the littoral carbonate samples of
613 the same age (~ 10.4 ka cal. BP, Gasse_94 and Strom20-SU1; Tab 2; Fig 6), which show an elevation
614 offset of 49.2 m. The sample Gasse_94 is located on the left bank of the Logiya River valley, while
615 Strom20-SU1 was sampled within the Saha plain (Fig 6a, b). Between the sites, several normal faults
616 are dissecting the surface and mainly correspond to the southern end of the active Manda-Hararo
617 volcano-tectonic rift segment (Fig 6; Abbate et al., 1995; Manighetti et al., 2001; Doubre, 2017). The
618 elevation offset is therefore consistent with a mean vertical subsidence rate of 4.7 mm/yr in the
619 central part of the rift with respect to its margin. A high amplitude of subsidence rate is expected in

620 the area, since this rift is affected by regular crustal magma intrusion events within its rift zone, which
621 induce its subsidence as was the case during the last 2005-2010 dyke sequence event (Grandin et al.,
622 2009, 2011).

623 The vertical offset between the samples Gasse_50 (right bank of the Logiya River valley) and Strom14
624 (Poz-124783; Magenta ridge; Tab 2), both dated around ~ 7.7 ka cal. BP, is more complex because they
625 are located ~ 60 km apart (Fig. 6a). However, these dates suggest a mean differential vertical rate
626 reaching ~ 9.1 mm/yr over this period. Acocella (2010) mapped the down-sagging of the surface of the
627 Tendaho graben due to the withdrawal of magma from the deep elongated reservoir. This collapse of
628 the surface was particularly important over the period 2 to 0.5 Ma, but may have persevered at a
629 lower rate over the Holocene period. The recent subsidence can be related to the current extension
630 taking place across the graben estimated with GNSS measurements. The rifting across the Tendaho
631 graben occurs with both an opening rate of 7 mm/yr between the southwestern margin of the graben
632 and the Gamari plateau (Fig. 6a, Doubre et al., 2017), and a related subsidence of the central part
633 which is not clearly estimated using a geodetic approach. However, because the Tendaho graben is not
634 considered a very tectonically and magnetically active rift segment, our subsidence rate within the
635 central part of the Tendaho graben (~ 9.1 mm/yr) is probably overestimated if we compare it to the
636 subsidence rate of the inner floor of the active volcano-tectonic segment of Asal, which reached ~ 8
637 mm/yr for the last 6 ka, and 8-9 mm/yr for the last 35 ka (Stein et al., 1991).

638 All these results show that the Holocene subsidence taking place within the Tendaho graben varies in
639 time and in space. A striking result is that all the MHHS measurements in the central and the
640 northeastern parts of the Tendaho-Gobaad graben are at the same elevation (~ 420 m) which suggests
641 that most of this area was affected by quite a homogeneous subsidence during the Holocene period
642 (Fig 4). This evidence is confirmed by several stromatolites at the same elevation and age throughout
643 the central and northeastern basin (Tab 1, Fig 7a, 7c), which indicates a maximum vertical
644 displacement of ~ 5 m between them over the last 5-6 ka. However, our results show a relatively higher
645 subsidence rate (~ 4.7 mm/yr) in the northern part of the Tendaho graben (Logiya-Saha branch; Fig 6a,

646 6b) and a lower subsidence rate of ~1 mm/yr in the Gobaad valley (Fig 6a, 6b). This variation in
647 subsidence rate is also consistent with the variations in opening rate measured by GNSS across both
648 ends of the graben (Dobre et al., 2017). This is also consistent with the model proposed by
649 Tapponnier et al. (1990) and Manighetti et al. (2001, 1998) where the Tendaho graben corresponds to
650 the tip of the Red Sea ridge propagating southeastward into the Afar, and overlapping with the Aden
651 Ridge. The opening and related subsidence rates are larger in the northwest than the southeast, near
652 Lake Abhe (Fig 6a), where the lack of evidence of active deformation matches with smaller vertical
653 displacements.

654 In light of these results, the low subsidence rates cannot support the hypothesis of a tectonic-driven
655 lacustrine transgression between the Early and the Middle Holocene, since a subsidence rate of the
656 basin that is larger than the 'evaporation rate' cannot be envisaged.

657 Taking into account the paleo-shoreline elevation offsets due to the tectonic vertical slipping between
658 the SW graben margin and the central/NE basin (Fig. 6a), we can correct the elevation of samples with
659 the aim of producing a more accurate Holocene lake-level fluctuation model. We consider the
660 subsidence rate measured along the Logiya-Saha transect reliable, and have therefore calibrated
661 elevations of samples from the SW ridge of the basin (see Tab 1).

662

663 **5.2 - The last 11 ka of Abhe Lake fluctuations and their hydro-climatic relevance**

664

665 On the basis of the documentation of 107 littoral geomorphic features and their chronological control
666 obtained from 53 ¹⁴C ages on shallow-/deep-water stromatolites and nearshore deposits, we propose
667 a robust Holocene water-level reconstruction model of Lake Abhe. The lake-level variation curve takes
668 paleo-shoreline tectonic deformations into account thanks to the offset correction calculated from
669 Holocene subsidence rates, which are estimated from stromatolite and paleo-shoreline ages. In
670 addition, the integration of archaeological excavation datasets from lake-side prehistoric occupations

671 (Khalidi et al., 2020) enables us to add an altitudinal control to lake-level reconstructions that improves
672 the paleo-hydrological model and constrains some reservoir age effect anomalies (Fig 7a).

673 After the hyperarid episode coinciding with the Younger Dryas cooling event (Garzin et al., 2007; Gasse
674 et al., 2008), the first evidence for lake-level rise in the Holocene is dated on a stromatolite to ~11.1 ka
675 cal. BP (Strom20-SU5), after which the Maximal Holocene Highstand Shoreline (MHHS, ~420 m a.s.l.)
676 reached its peak sometime around 10 ka BP. Following a probable water-level stabilization at ~400 m
677 a.s.l., dated between ~9.8 and ~8.8 ka cal. BP, the lake level regressed slowly to ~375 m a.s.l. until ~8.3
678 ka cal. BP and abruptly dropped to ~250 m a.s.l. at ~8 ka cal. BP. This Early Holocene transgression
679 (between ~11.1 and 8.3 ka cal. BP) generally corresponds to the Abhe-IV lacustrine phase defined by
680 Gasse (1977). During this phase the overflow sills were certainly active, as evidenced by the DDa, DSa
681 and DSb sites (see 4.2.1 and [Sup. Mat. A.8, A.9, A.10](#)).

682 The resurgence of Mid-Holocene humid conditions as revealed by paleo-shoreline markers was
683 recorded at ~7.7 ka cal. BP. The lake reached the highest Mid-Holocene water level between ~6 and
684 ~5.6 ka cal. BP (~380-385 m a.s.l.). This lacustrine transgression corresponds to the Abhe-IV lacustrine
685 phase defined by Gasse (1977). After a slow lowering of the waterbody between ~5.6 and ~4.6 ka cal.
686 BP, the absence of littoral deposits at lower elevation suggests a drastic and rapid drop in level. Finally,
687 the lake rose gently from 2.8 ka cal. BP until it reached its maximal Late Holocene level (330 m a.s.l.) at
688 ~1.5 ka cal. BP.

689 Lake-level variations are generally attributed to changes in monsoon precipitation conditions over the
690 catchment area (i.e., the Ethiopian Highlands; Gasse, 1977). Paleoclimatic interpretation of lake-level
691 changes are rightly based on the analysis and modeling of the hydrological precipitation and
692 evaporation balance (Junginger and Trauth, 2013) coupled with paleolimnological studies. However,
693 changes in lake level and sedimentation can be affected by local geomorphic factors which can
694 mitigate the climatic signal, leading to an underestimation or overestimation of the modeled data. The
695 420, 400 and 385 m a.s.l. Lake Abhe highstands (Fig 7b) are considered to be precipitation-induced
696 lake fillings from Ethiopian Highland summer rainfalls, regulated by overflows towards the Hanle basin

697 (Fig 1c, 6, 7a). We observe several regression steps over the course of the Holocene that were the
698 result of incision of the overflow sills and subsequent water-level stabilizations, and which are
699 demonstrated by the multi-level abrasion platform sequences documented at the DDa and DSa/DSb
700 overflow sites (see 4.2.1 and Sup. Mat. A.8, A.9, A.10). Consequently, the MHHS doesn't directly imply
701 the highest precipitation or water inflow period of the Holocene, because was regulated by these
702 incision processes. However, in future studies, the estimation of the sill overflow incision rates could
703 provide an additional parameter for the calculation of hydro-balance models with the aim to give
704 precise water inflow or precipitation rates from the Awash River catchment.

705 Holocene changes in water level are not always the result of balanced source-to-sink mechanisms. A
706 non-linear response of the hydrological system could be conditioned by the interconnections between
707 basin catchments. During the Early/Middle Holocene, the Awash River catchment area was temporarily
708 supplied by freshwater inputs from adjacent basins (Fig 8d). The Ziway-Shala paleolake overflow
709 towards the Awash catchment was demonstrated for the first time by Nilsson (1940) at 1670 m a.s.l.,
710 and later revised by Grove and Goudie (1971), Gasse and Street (1978) and Gillespie et al. (1983; Fig
711 8c). The possible transient fluvial connection to the Awash River is in phase with the beginning of the
712 Early Holocene Lake Abhe highstand, coinciding with the maximal solar irradiation at 10° N around 11
713 ka cal. BP (Fig 8b, c). Similarly, the Mid-Holocene maximal Abhe Lake level, which occurred between 6
714 and 5.6 ka cal. BP, is in phase with the observed Mid-Holocene Ziway-Shala Lake overflow, which could
715 have made a significant contribution to the Lake Abhe water-level balance. The Holocene connectivity
716 between these two basins is even supported by the apparition of the *Cyprinidae sp.* fish in the Lake
717 Abhe basin around ~10 ka BP probably introduced from the Ziway-Shala basin where it was attested
718 since 13 ka BP (Coudert, 2019). Additionally, Sagri et al. (2008) demonstrated that during the Early
719 Holocene, the Awash River contributed to the water filling of the Ziway-Shala Basin, confirming the
720 mutual interconnection between these two basins.

721 After excluding the possibility of tectonic-induced Holocene lake transgressions (sec 4.2) and providing
722 some elements that demonstrate the primary local geomorphic control of lacustrine fluctuations

723 (overflow sill incision), we can consider the Holocene Abhe lake-level reconstruction as a reliable
724 record of monsoon-induced hydrologic changes over the Awash basin.

725

726 **5.3 – Implications for Holocene East African moisture mass dynamics**

727

728 Thanks to Lake Abhe's geographic position, located in the northern EARS extremity (Fig 1a), our results
729 add new elements to the discussion of the timing of the AHP transitions in Eastern Africa and more
730 particularly across the EARS basins.

731 We compared our results with other southernmost lacustrine basins along the rift axis at latitudes
732 between 12° N and the Equator, with the aim of studying Holocene hydrological feedback across a
733 coarse N-S transect (Fig 10). From north to south, we included lake-level paleo-data from Abhe (11°-
734 12° N; this study; Fig 9c), Ziway-Shala (7.5°-8.5° N; Gillespie et al., 1983; calibrated to calendar age; Fig
735 9d), Suguta (1.5°-2.5° N, Garcin et al., 2009; Junginger and Trauth, 2013; Fig 9e) and Nakuru basins
736 (0.5° N; Richardson and Dussinger, 1986; calibrated to calendar age; Fig 9f). With the aim to constrain
737 the hydro-climatic significance of these water-level changes, we added δD_{wax} data - considered a proxy
738 for precipitation/air-mass dynamics – from the northern and southern extremities of the investigated
739 area (Tana Lake, 11.5°-12.5° N, Costa et al., 2014; Challa Lake, 3°S, Tierney et al., 2011; Fig 9a, 9g).

740 Our results show a resurgence of humid conditions after the YD cooling event of the Northern
741 Hemisphere (Younger Dryas; Alley, 2000; Rasmussen et al., 2014) around ~11.1 ka cal. BP and a
742 reduction of the high Early Holocene hydrological regime from ~8.7 ka cal. BP. A shift and reduction in
743 hydrographic regimes as early as ~8.7 ka cal. BP has already been pointed out by recent studies, and
744 interpreted as a consequence of changes in monsoonal dynamics over the Ethiopian Highlands
745 (Blanchet et al., 2014; Ménot et al., 2020; Mologni et al., 2020). Despite a possible chronological
746 uncertainty of the compared records, a northward installation of Mid-to-Late Holocene aridification
747 with a ~1000 yr delay with respect to the southern EARS basins can be suggested (Fig 9).

748

749 Indeed, the observed time-latitude delays of the Mid-to-Late Holocene aridification events are in
750 conflict with the ITCZ oscillation processes: Mid-Holocene ITCZ southward migration should
751 correspond to a southward aridification, while we observe the contrary (Fig 9). However, changes in
752 precipitation are not only the result of atmospheric convection over the rain belt, but could have also
753 been produced by the migration of other convergence zones.

754 As proposed by (Hailemichael et al., 2002), and later by Tierney et al. (2011) and Junginger et al.
755 (2014), the incursion of the Congo Air Boundary (CAB) in the specific morpho-tectonic context of the
756 EARS water-sources would have contributed additional moisture to persistent humid conditions, in
757 contrast to the Mid-Holocene southern migration of the ITCZ. This process would have been the result
758 of a high pressure gradient between the CAB and the Indian Summer Monsoon (ISM) air masses,
759 which enhanced the influence of the West African Summer Monsoon (WAM) over Northeast Africa
760 (Camberlin, 1997; Marzin and Braconnot, 2009; Junginger et al., 2014). Similarly, the intensified
761 summer Indian monsoon of the Early Holocene can be hypothesized to have pulled both components
762 of the Ethiopian monsoon eastward (Overpeck et al., 1996). The same hypothesis has been advanced
763 to explain the lower and prolonged leaf-wax values at Tana Lake during the Mid-Holocene period
764 (Costa et al., 2014; Fig 10b). The intensified ISM favored an east-shifting of the CAB, fueling humid
765 conditions over the Ethiopian Highlands. In this scenario, thanks to its Eastern Ethiopian Highland
766 sources, Lake Abhe's hydrology would have been particularly subjected to the CAB moisture mass
767 influences from the beginning to the end of the AHP. In contrast, Challa Lake, which lies beyond the
768 CAB boundary (Fig 10; Tierney et al., 2011), presents a short Holocene humid regime and follows the
769 subsequent progressive southern retreat of the rain belt (Fig 9g).

770 In summary, the persistence of ISM-CAB pressure regulation combined with the geomorphic patterns
771 of the EARS, enhanced the influence of the CAB moisture in the northeast African Rift until the end of
772 the AHP producing a progressive northward development of the arid Mid-Holocene conditions across
773 the EARS (Fig 10). These patterns show an inverse air-moisture dynamic compared with Central and

774 Western intertropical Africa, where the development of Middle-Holocene arid conditions followed the
775 progressive southward retreat of the ITCZ (Shanahan et al., 2015; Fig 9h).

776

777

778 **6 – Conclusion**

779 The documentation, interpretation and dating of more than one hundred littoral geomorphic features
780 have permitted the reconstruction of a detailed paleo-shoreline sequence that elucidates the
781 Holocene lake-level rise and fall of the widest lacustrine system in the Afar region, Lake Abhe.

782 We observe that after the Younger Dryas arid event (~12.9 to ~11.7 ka) Lake Abhe began to rise around
783 11.1 ka cal. BP peaking at ~9.8 ka cal. BP, in phase with the maximum insolation values over the tropics.

784 A gradual decrease in lake level is recorded from ~9 ka cal. BP until a drastic drop in level occurred at
785 ~8 ka cal. BP. A resurgence of humid conditions is confirmed from ~7.7 ka BP with the establishment of
786 a stable lake highstand until ~4.6 ka cal. BP when the large waterbody was definitively reduced by a
787 prominent desertification process in the Afar region. Nonetheless, during the Late Holocene, a minor
788 lake transgression was recorded between ~2.4 and ~1.4 ka cal. BP, followed by a lake lowstand which
789 persists until the present day.

790 The documentation of multiphase sill overflow incisions towards the adjacent Hanle basin
791 demonstrates geomorphic control on Holocene highstand elevations. In addition, new ages on
792 carbonate littoral material allow us to propose a Holocene subsidence rate of the northern Tendaho-
793 Gobaad graben.

794 Past Lake Abhe level fluctuations are thus the result of combined monsoonal dynamics concentrated in
795 the Ethiopian Highlands (Viste and Sorteberg, 2013) and local earth surface processes. While long-
796 term hydrological changes are mainly triggered by orbitally-driven insolation pacing, our results
797 suggest that short-term events, as well as the variability of humid/arid period durations, may be

798 related to internal/regional climatic and geomorphic forcing mechanisms which are not yet fully
799 understood.

800 Comparison of our results with other lacustrine archives indicates a northward Mid-Holocene
801 aridification along the EARS, which disagrees with the classic scenario of the ITCZ southern retreat as a
802 possible cause of the southern Mid-Holocene aridification over Africa. This evidence could be
803 explained by the incursion of air moisture masses in the EARS induced by a high pressure gradient that
804 formed between the Congo Air Boundary and the Indian Summer Monsoon air masses, resulting in
805 enhanced precipitation over Northeastern Africa. These conditions would have supported the
806 persistence of large waterbodies and their related eco- and anthropo-systems in the northern EARS
807 lowlands until ~4.6 ka cal. BP.

808

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835 **Captions**

836

837 **Fig 1: a)** Location of the East African Rift System (EARS; red color) and the study area, corresponding
838 to the Abhe Lake basin (red star). Summer position (June) of present African tropical frontal systems:
839 the Intertropical Convergence Zone (ITCZ) and Congo Air Boundary (CAB; Viste and Sorteberg, 2013);
840 and the wind directions of the Western African Summer Monsoon (WAM) and the Indian Summer
841 Monsoon (ISM). **b)** Simplified morpho-tectonic patterns of the Afar Triangle depression (light gray),
842 enclosed by the Arabian, Nubian and Somalian plates (pre-rift groups, dark gray) with the location of
843 the present Lake Abhe (red star). **c)** Structural map of the Afar depression (modified from Doubre et
844 al., 2017 in which the present Abhe Lake (red star), the extension of the Abhe Lake sedimentary system
845 (light green), other Central Afar sedimentary basins (light yellow), the present natural waterbodies
846 (blue), the ancient Abhe basin overflow sills (blue arrows), recent lava flows (light red), active (red
847 lines) and ancient (dashed red lines) rift segments (modified from Barberi and Varet, 1972) are

848 indicated. The dashed dark green line indicates the Tendaho-Gobaad discontinuity, while the large and
849 small faults are indicated by bold and thin lines, respectively.

850

851 **Fig 2:** Graphic comparison between littoral geomorphic feature elevations measured with the GEO7x
852 differential GPS in the field (red triangles) and the same GPS point elevations processed with a
853 TanDEMx (12.5 m; green dots) and SRTM (30 m; black squares).

854

855 **Fig 3:** Selected photographs of the surveyed littoral geomorphic features described. **a)** Well-
856 developed wave abrasion notches combined with abrasion platform and beach pebble/cobble
857 concentration features at the BSW site (Sup. Mat. A.). **b)** Ridge made on the foot wall of the basalt
858 stratoid series which separates the Gamari and Afambo lake waterbodies with a well-developed
859 seaward-sloping shore morphology (Sunamura, 1992) and the location of the relative paleo-shoreline
860 (dashed black line) at the Afbo site (Sup. Mat. A.2). **c)** Well-developed abrasion platforms on basaltic
861 bedrock and on sediments at the DSb site (Sup. Mat. A.). **d)** Gilbert-type delta morphology at the BE
862 site with black arrows indicating: the feeder canyon, the apex delta and the connected abrasion
863 platform (Sup. Mat. A.). **e)** Gilbert-type delta facies at the BE site (Sup. Mat. A.).

864

865 **Fig 4:** Digital Elevation Model of the Central Afar Region base map (SRTM) with the **a)** location map of
866 surveyed sites over the Abhe Lake basin (black line squares) showing the Maximal Holocene Highstand
867 Shoreline (MHHS, 420 m a.s.l.; white line), location of the ancient overflow sills (black arrows) and the
868 Hanle basin. **b)** presented from the western to the eastern side of the Abhe Lake basin, a synthesis of
869 the surveyed sites showing the MHHS (bold and dotted blue lines), the paleo-shorelines (black arrows),
870 the stromatolite deposits (bold red lines) and corresponding calibrated median ¹⁴C ages (red text). The
871 dashed black lines indicate the main paleo-shorelines while the dashed red line indicates the
872 connection between stromatolites with similar calendar ages.

873

874 **Fig 5:** Selected macro-/meso-/micro-photographs of the Lake Abhe stromatolites. **a)** Aerial photo of
875 the StrIS site showing the macro-structure of connected large stromatolitic domes, the sampling zone
876 of the sample Strom31 (Poz-113925; Tab 1) and the location of the Stromatolis archaeological site
877 (Khalidi et al., 2020), where the dated Stromatolis-SU1 (Poz-113841) was sampled (Tab 1). Photo
878 courtesy of E. Soteras **b)** Photograph showing the superimposition of stromatolitic Facies-A and -B at
879 the site LAK (Sup. Mat. A.). **c)** Photograph of a polished section of a stromatolite (Strom_11) showing
880 the meso-structure superimposition of stromatolitic Facies-A and -B. **d)** SEM-EDX analysis mapping
881 showing the Si-Mg concentration in the stromatolite Facies-B rippled laminations.

882

883 **Fig 6: a)** Digital Elevation Model of the Central Afar Region (SRTM) showing the current horizontal
884 velocities with respect to the stable Nubian plate deduced from GNSS field measurements (black
885 dashed arrows; after Doubré et al. 2017), the location of samples used for the calculation of the
886 subsidence rate (black arrows, this study), the MHHS (red line), the location of the ancient overflow
887 sills (blue arrows), of new (green dots) and published (red dots) dated nearshore deposits, of the
888 stromatolitic deposits (blue triangles), the topographic profile represented in Fig. 6b (black line, A-A'),
889 and the littoral geomorphic features (yellow dots) inventoried during the survey. **b)** Topographic profile
890 A-A' with the main graben faults, and with the location of samples used for the calculation of the
891 subsidence rate (black arrows, this study) and relative offset (red text).

892

893 **Fig 7:** Holocene Abhe Lake -level reconstitution. **a)** Curve of Abhe Lake -level fluctuations (black line)
894 showing the main lake highstands (light blue squares), the dated Facies-A (blue triangles), Facies-B
895 (inverted blue triangles) and undetermined facies (empty blue triangle) stromatolites, the new (black
896 squares) and published (black dots) dated nearshore deposits (Gasse 1975 revised), and the main
897 human prehistoric occupations (empty red rhombuses). **b)** Histogram of the totality of the inventoried
898 paleo-shoreline features (red lines) and stromatolite deposits (blue lines) showing counts against the

899 elevations in ranges of 5 m. **c**) Longitude against elevation graph of the totality of stromatolites
900 inventoried (empty blue triangles) showing the main Holocene Abhe Lake highstand levels (light blue
901 bands).

902

903 **Fig 8:** Holocene interconnectivity between the Ziway-Shala and Awash River catchments. **a**)
904 Paleovariations in June insolation at 10° N (after Berger and Loutre, 1991). **b**) Abhe Lake-level
905 fluctuations (for the legend see Fig 7). **c**) Ziway-Shala Lake-level fluctuations from calibrated ages (after
906 Gillespie et al. 1983). Overflow elevations are indicated with blue lines, while the transient fluvial
907 interconnectivity between the two basin catchments are indicated with light blue bands. **d**) Map of the
908 Ziway-Shala basin and Awash catchment areas (black bold line) showing the Awash River course (bold
909 blue line), the secondary rivers (light blue lines), the present (light blue) and Early-Mid-Holocene (blue)
910 lake surfaces, and the Early-Mid-Holocene overflows (red arrows).

911

912 **Fig 9:** Paleovariations in **a**) June insolation at 10° N (Berger and Loutre, 1991); **b**) δD_{WAX} from Lake
913 Tana (Costa et al., 2014); **c**) water level of Abhe Lake (this study); **d**) human subsistence strategies in
914 the Central Afar region (Khalidi et al., 2020); and water levels of **e**) Ziway-Shala Lake (Gillespie et al.,
915 1983), **f**) Suguta Lake (Junginger and Trauth, 2013) and **g**) Nakuru Lake (Richardson and Dussinger,
916 1986). Original radiocarbon dates were calibrated and converted into calendar years. Paleovariations in
917 **h**) δD_{WAX} from Lake Challa (Tierney et al., 2011) and June insolation at the equator (Berger and Loutre,
918 1991; yellow lines).

919

920 **Fig 10:** East Africa map of monthly precipitation average (mm/month) in June, from the CRU-TS 4.03
921 dataset (Harris et al., 2014) downscaled with WorldClim 2.1 (Fick and Hijmans, 2017). Blue stars
922 indicate the location of lacustrine basins reported in Fig 9: b – Lake Tana; c – Lake Abhe; e – Ziway-
923 Shala Lake; f – Lake Suguta; g – Lake Nakuru; and h – Lake Challa. Dashed green line indicates the

924 suggested position of the CAB before 8.5 ka cal. BP, while the dashed red lines indicate the position of
925 the CAB and ITCZ after 8.5 ka cal. BP. Black arrows correspond to the air moisture mass direction of the
926 WAM and of the ISM.

927

928 **Table 1 :** New and previous ^{14}C ages calibration and elevation correction: ‡ = Ages on
929 nearshore/offshore deposits; † = ages for reservoir effect estimation; * = Selected and revised ages on
930 nearshore/offshore deposits from Abhe basin, from Gasse (1975), the ID number refers to the sample
931 number indicated in Gasse 1975 and the Lab ID refers to the Laboratory indicated in Gasse 1975 : LGD
932 = Laboratoire de Géologie Dynamique, Paris; LFR = Laboratoire des Faibles Radioactivités CEA-CNRS,
933 Gif-sur-Yvette; GU= Gakushuin University, Tokyo. Str = Stromatolite undetermined; Str-A = stromatolite
934 Facies-A; Str-B = Stromatolite Facies-B; *Mt* = *Melanoides tuberculata*; Ch = Charcoal; hcc =
935 hydrothermal carbonate crust; Sh = Shells mix; U = Unios bivalve shell.

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937 **Table2:** Subsidence rate calculation table from dated littoral carbonate samples.

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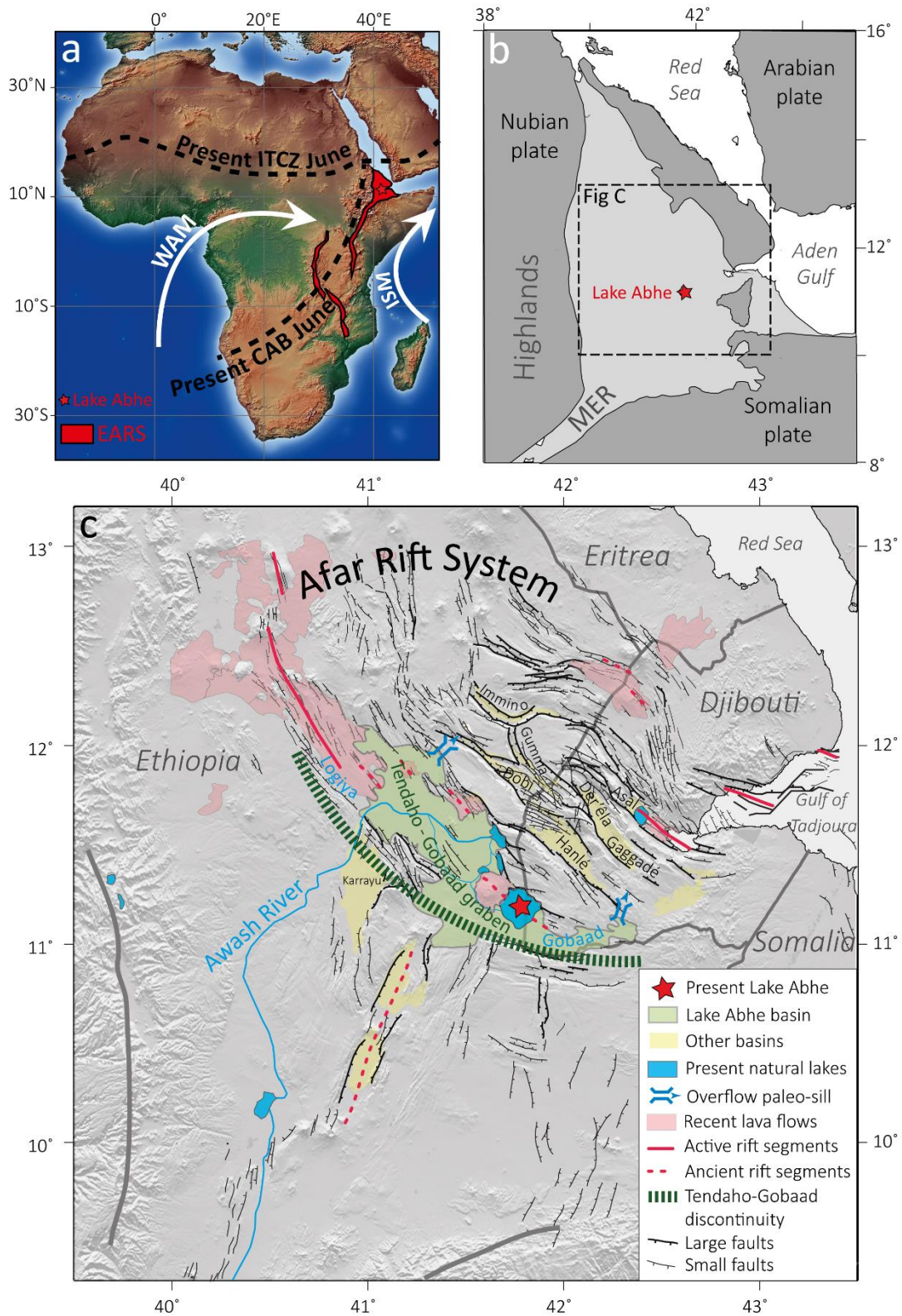
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1315 **Fig 1**

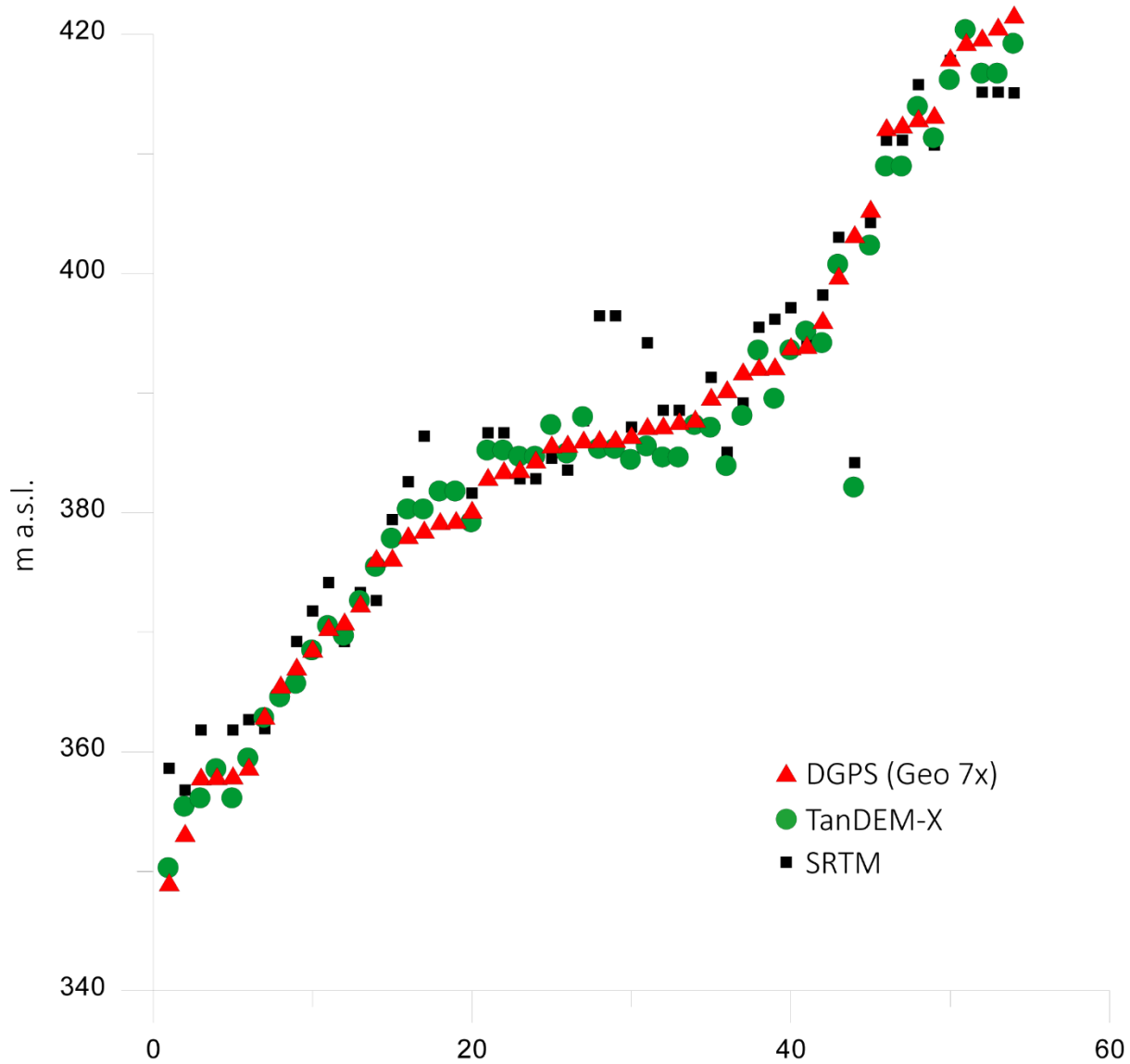


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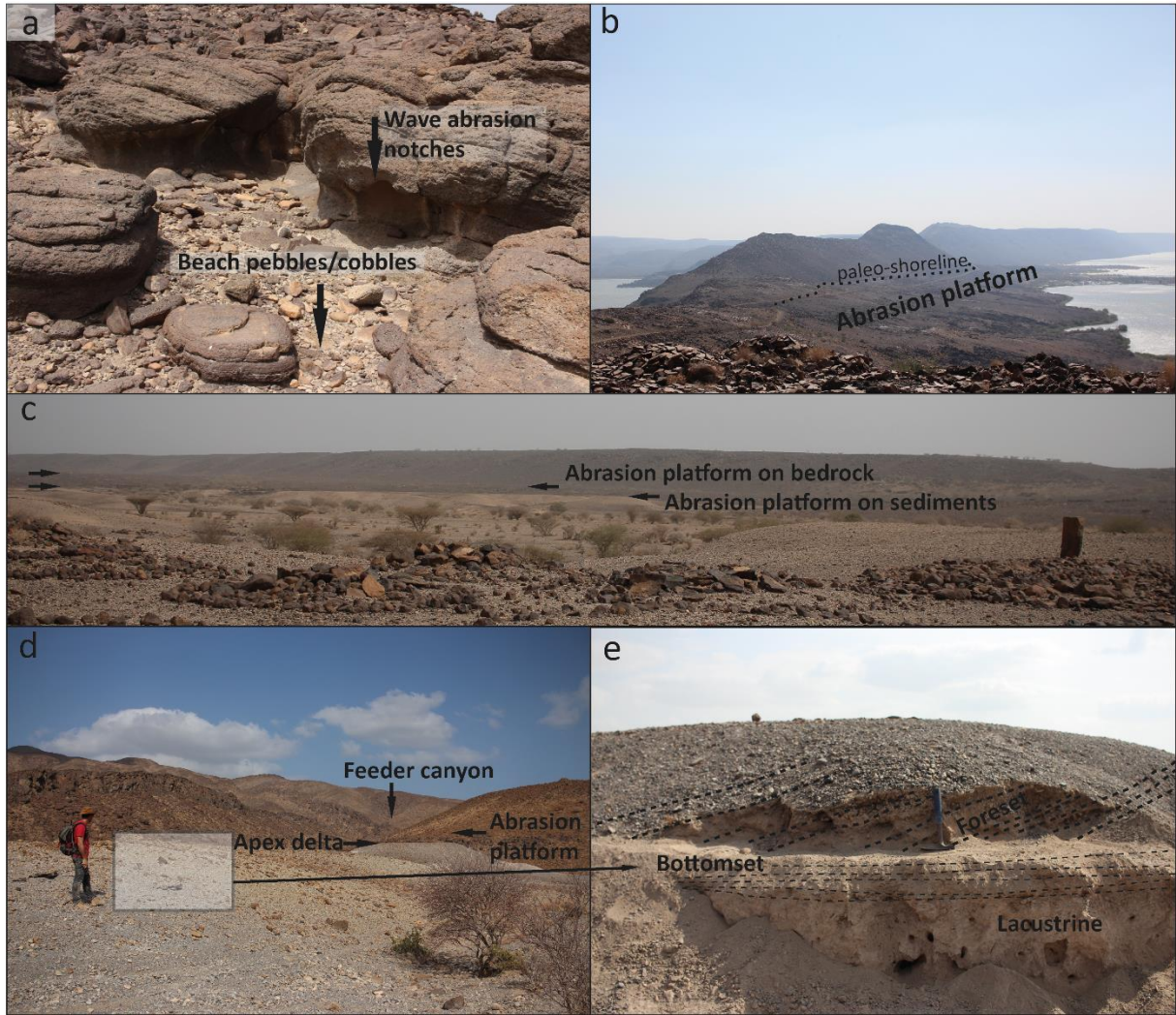
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1319 **Fig 2**



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Fig 3



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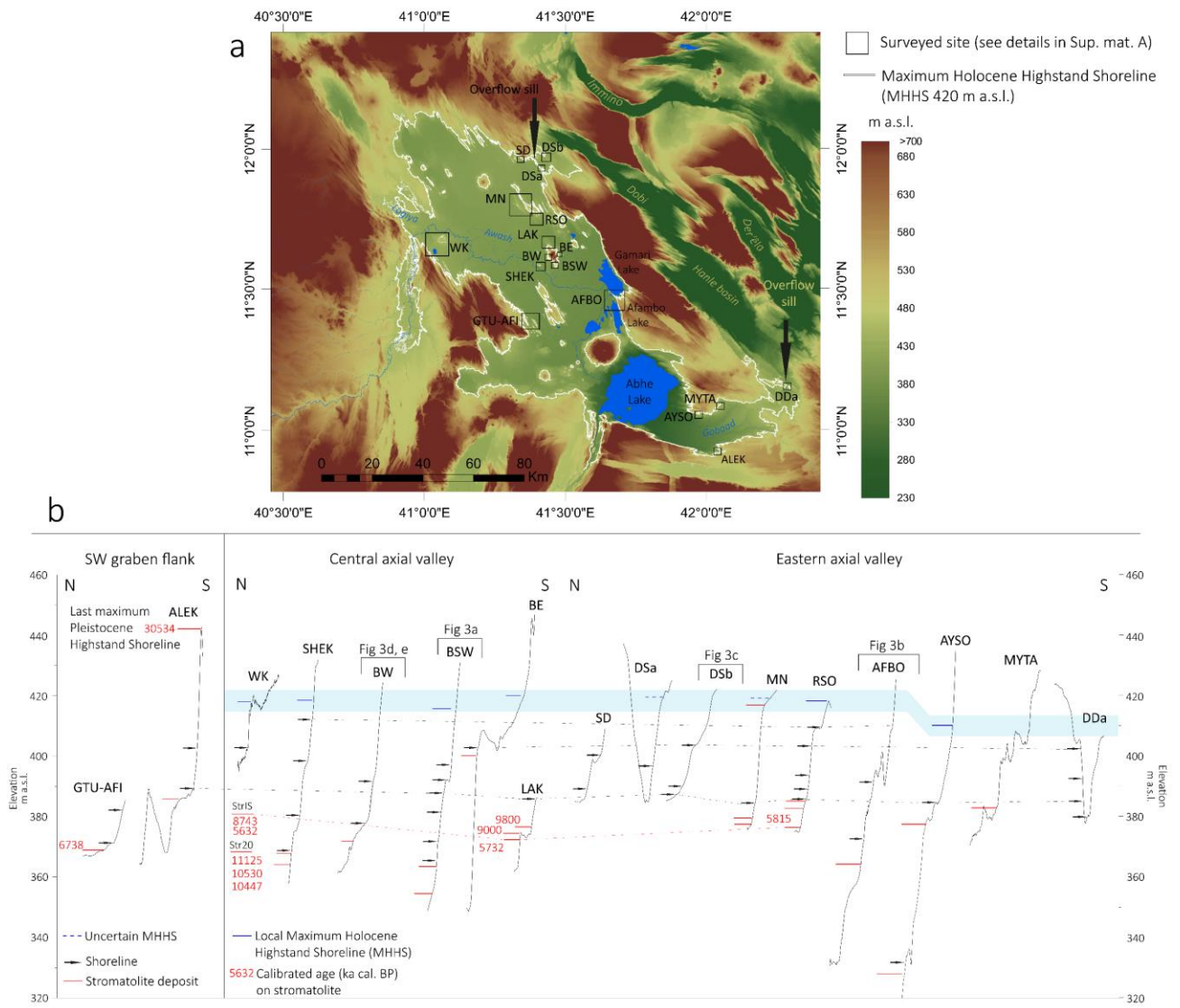
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1338 **Fig 4**



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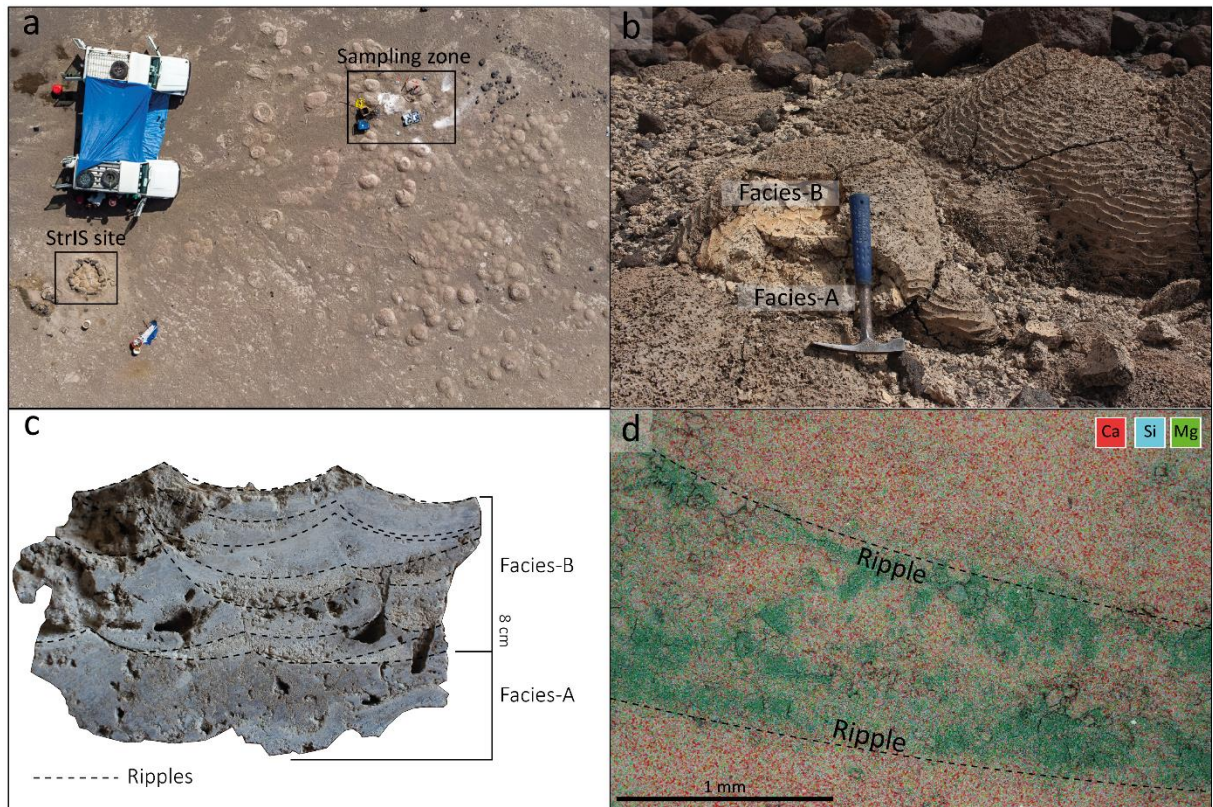
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1346 **Fig 5**



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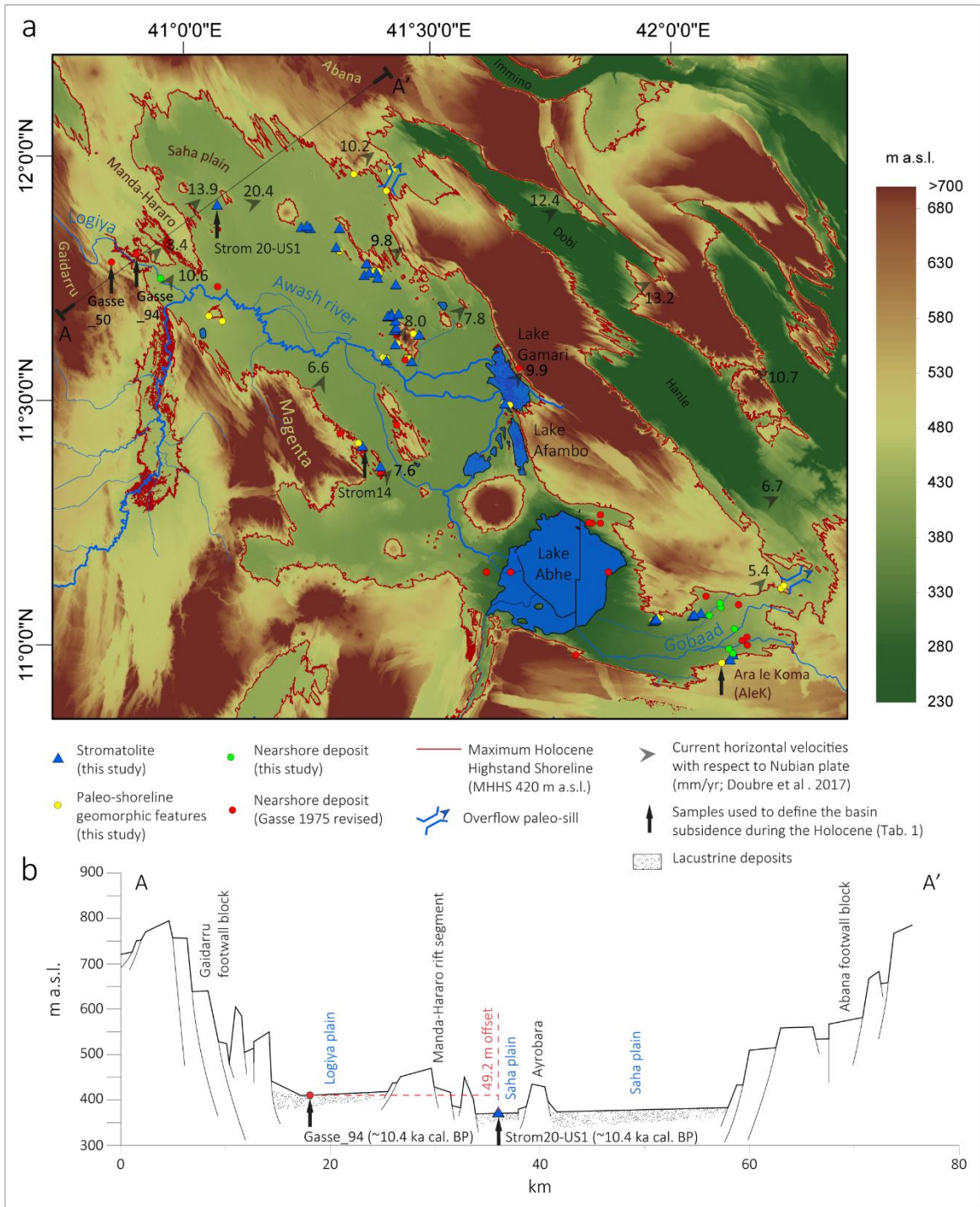
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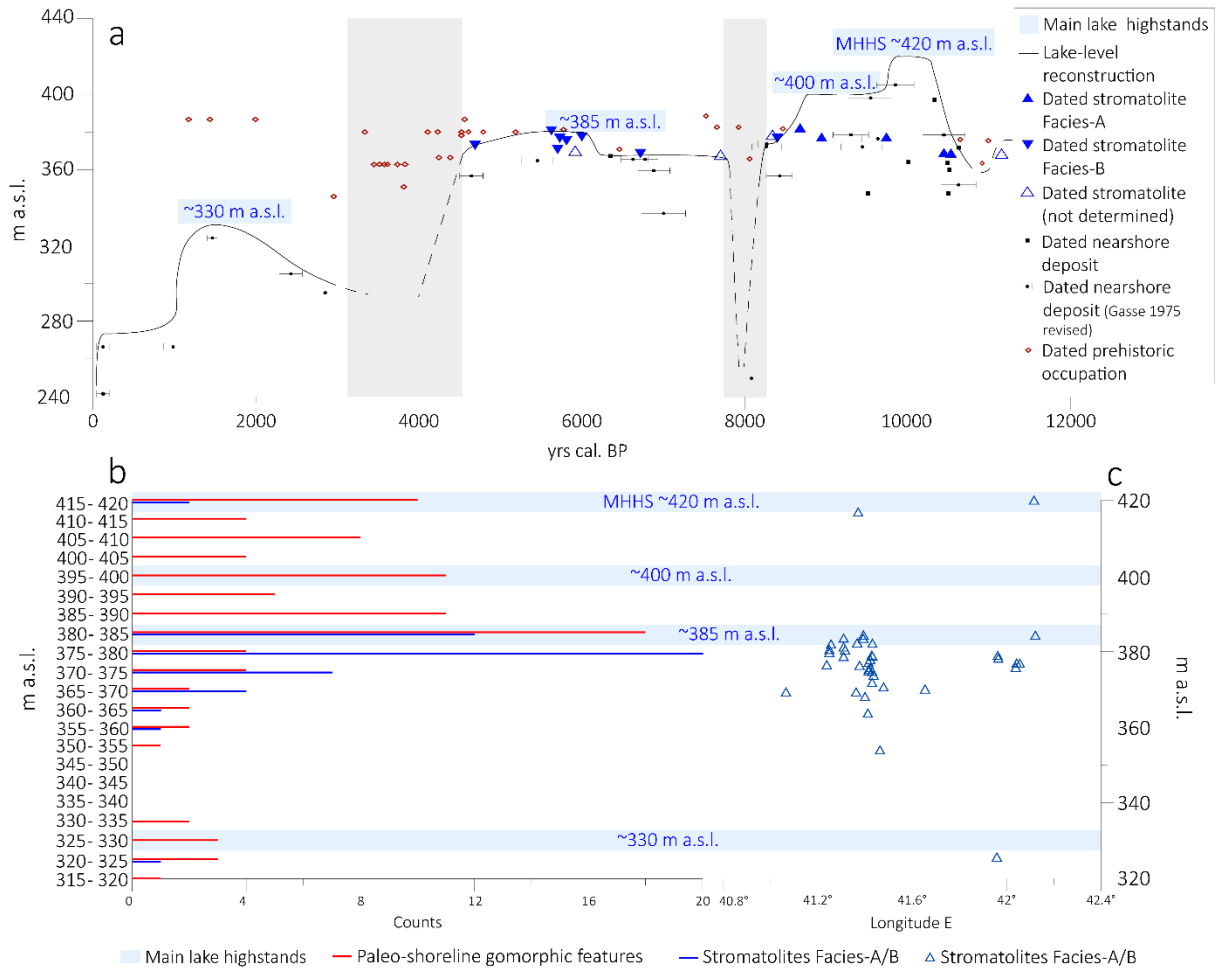
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1360 **Fig 6**





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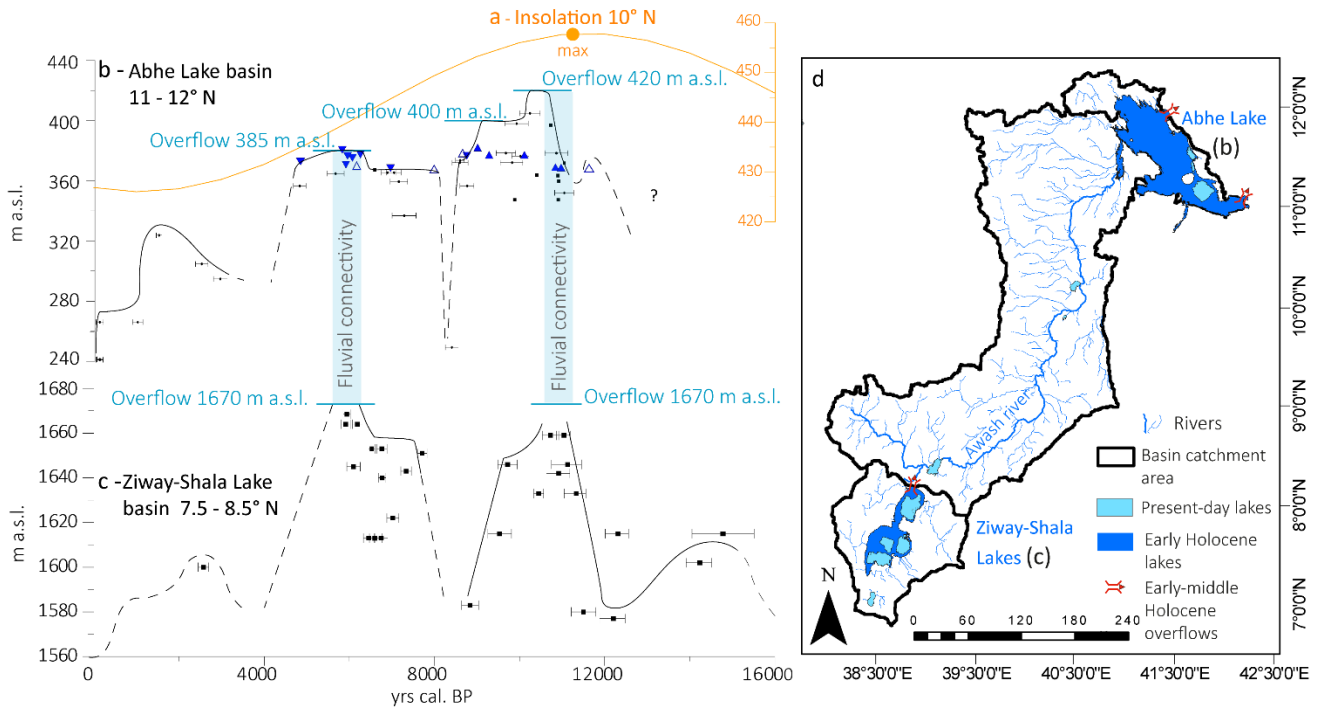
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1372 **Fig 8**



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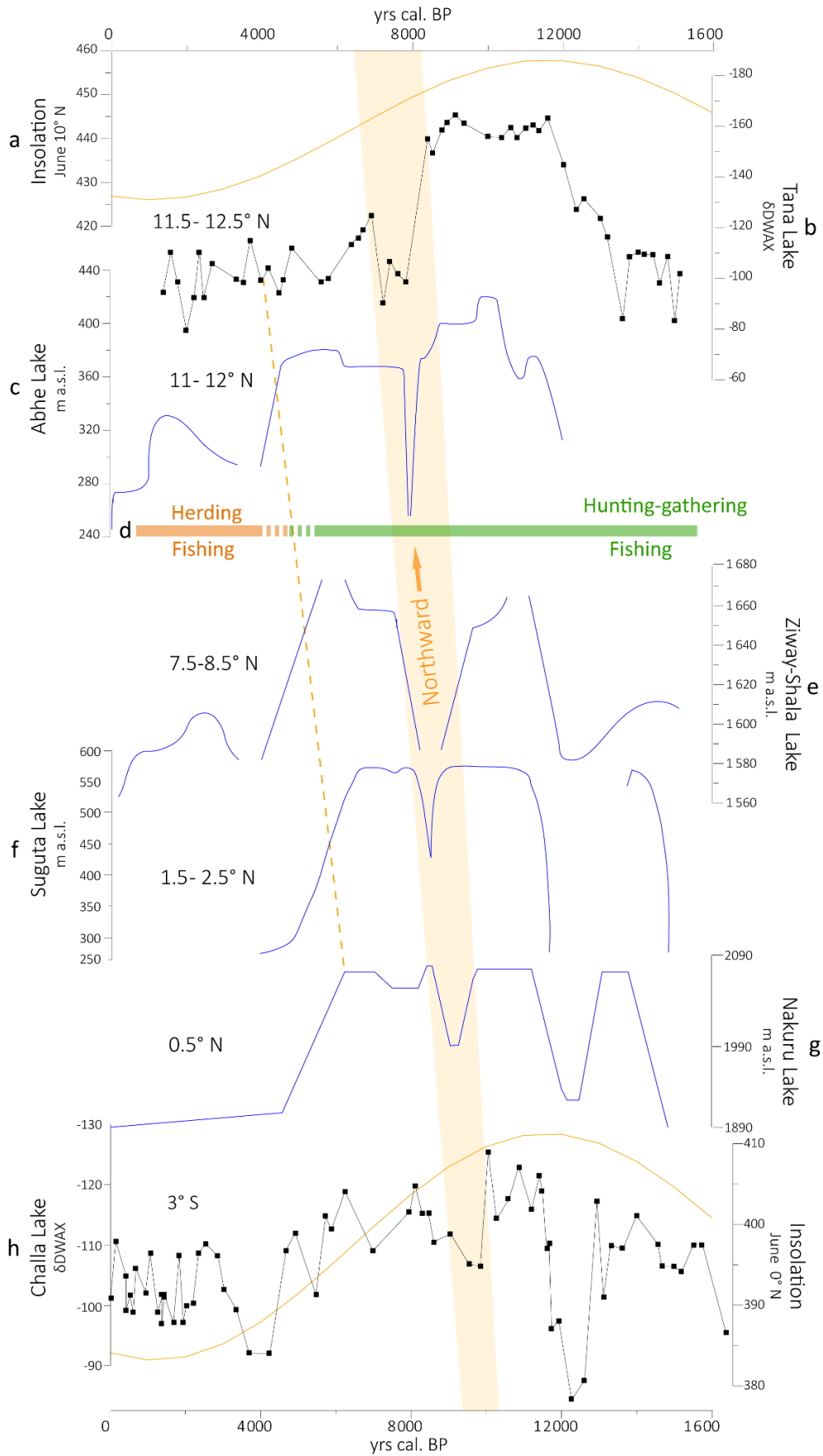
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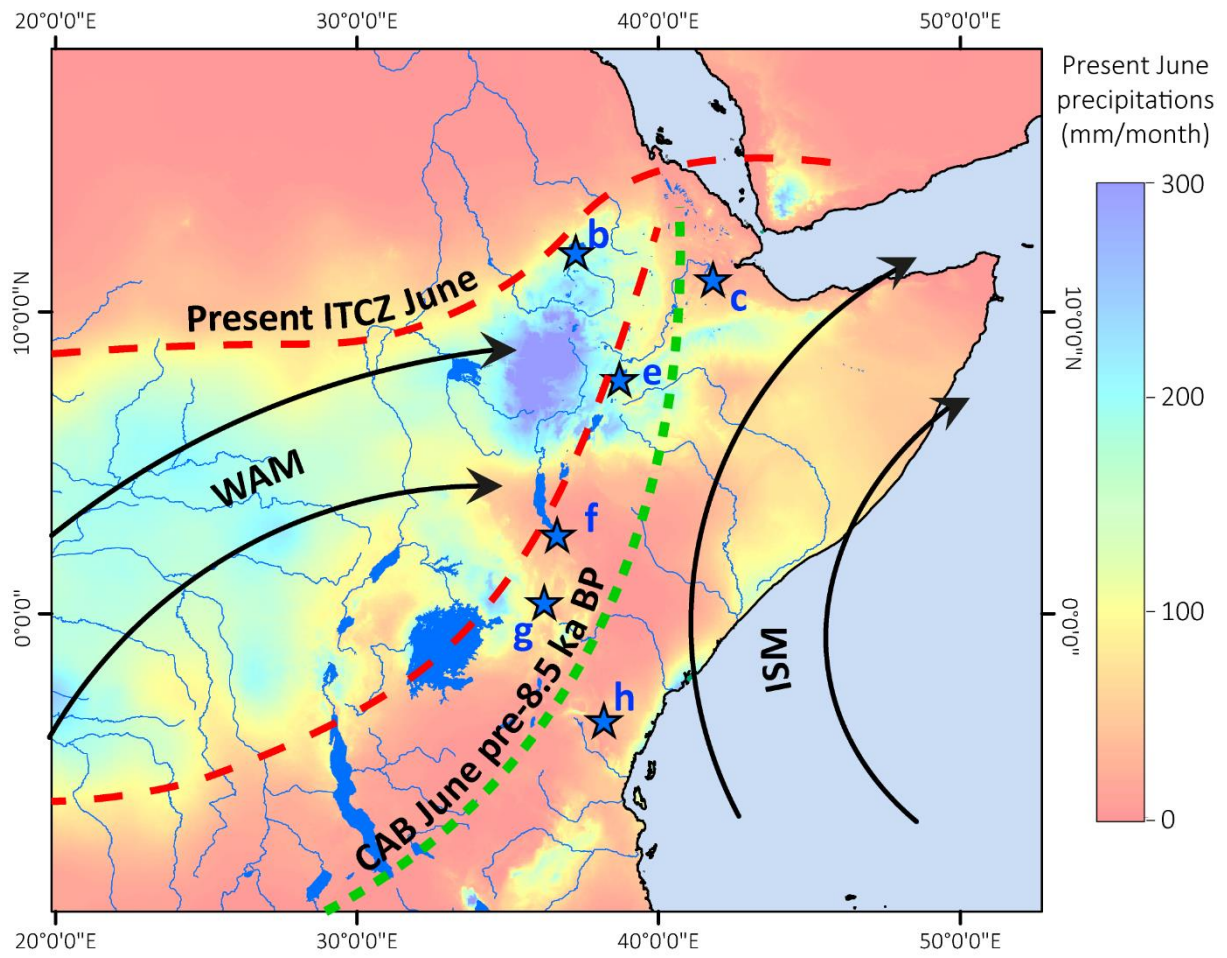
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1390 **Table 1**

ID	Lab ID	Age yr BP	Material	Cal ages yr BP (range 95.4 %)	Cal ages yr BP (median)	Elevations (TanDEMx +10m a.s.l.)	Subsidence rate (mm/yr)	Subsidence correction (m)	Corrected elevation (m a.s.l.)
Strom-ALEK-2	SacA62534	26250 ± 240	Str	30995-30082	30534 ± 248	440	/	/	/
Strom20-SU5	SacA58238	9689 ± 40	Str	11210-10805	11125 ± 113	369	/	/	/
Strom20-SU4	SacA58239	9320 ± 40	Str-A	10660-10404	10530 ± 64	369	/	/	/
Strom20-SU1	SacA58240	9265 ± 40	Str-A	10565-10295	10447 ± 73	369	/	/	/
Strom10a	SacA55608	8790 ± 30	Str-A	10112-9679	9807 ± 81	376.8	/	/	/
Strom10b	SacA55609	8075 ± 35	Str-A	9123-8785	9008 ± 70	376.8	/	/	/
Strom31	Poz-113925	7910 ± 50	Str-A	8978-8598	8743 ± 109	381.7	/	/	/
Strom49-bloc	Poz-124776	7620 ± 50	Str-B	8537-8350	8418 ± 46	377.6	/	/	/
Strom49-col	Poz-124778	7540 ± 50	Str	8419-8200	8360 ± 54	377.6	/	/	/
Strom14	Poz-124783	6870 ± 40	Str	7791-7614	7701 ± 44	367.7	/	/	/
Strom13	SacA55611	5919 ± 30	Str-B	6829-6666	6738 ± 40	369	/	/	/
Strom49-lam1a	Poz-124779	5255 ± 35	Str-B	6179-5929	6015 ± 73	377.6	/	/	/
Strom44	Poz-124781	5155 ± 35	Str	5995-5756	5918 ± 58	369.7	/	/	/
Raso2-Strom	Poz-113843	5085 ± 35	Str-B	5913-5746	5815 ± 51	376	/	/	/
Strom49-lam2a	Poz-124777	5015 ± 35	Str	5895-5608	5752 ± 71	377.6	/	/	/
Strom7	SacA55610	5005 ± 30	Str-B	5889-5653	5732 ± 69	371,5	/	/	/
Stromatolis-SU1	Poz-113841	4910 ± 30	Str-B	5712-5591	5632 ± 29	381.7	/	/	/

Strom51	Poz-124782	4160 ± 30	Str-B	4828-4580	4705 ± 69	373.3	/	/	/
2016-PSPCA-2	Poz-82850 [‡]	9310 ± 50	<i>Mt (nearshore)</i>	10662-10298	10513 ± 82	360.3	/	/	/
2016-PSPCA-3	Poz-82851 [‡]	9300 ± 50	<i>Mt (nearshore)</i>	10653-10296	10498 ± 83	348.1	/	/	/
2016-PSPCA-1	Poz-82849 [‡]	8770 ± 50	<i>Mt (nearshore)</i>	10116-9552	9773 ± 126	364.5	/	/	/
2016-PSPCA-4	Poz-83625 [‡]	5570 ± 35	<i>Mt (nearshore)</i>	6410-6294	6354 ± 33	367.6	/	/	/
2017-PSPCA-2	SacA55422 [‡]	9290 ± 35	<i>Mt (nearshore)</i>	10583-10301	10491 ± 62	364	/	/	/
2017-PSPCA-3	SacA55425 [‡]	9395 ± 35	<i>Mt (nearshore)</i>	10712-10520	10627 ± 46	372.1	/	/	/
2017-PSPCA-6	SacA55426 [‡]	8520 ± 35	<i>Mt (nearshore)</i>	9542-9479	9514 ± 18	348	/	/	/
Logiya_1A	Beta 446441 ^{††}	9180 ± 30	<i>Mt (nearshore)</i>	10477-10245	10330 ± 55	397.4	/	/	/
Logiya_1B	Beta 446442 ^{††}	9100 ± 30	<i>Ch (nearshore)</i>	10477-10245	10241 ± 27	397.4	/	/	/
Gasse_1*	LGD	98.8 ± 1.2	hcc	255 – 31	121 ± 78	241.7	2.5	- 0.3025	241.3975
Gasse_2*	LGD	100.6 ± 0.9	hcc	255 – 31	122 ± 78	241.7	/	/	/
Gasse_6*	LGD	110 ± 50	hcc	276 – 9	124 ± 81	266.4	/	/	/
Gasse_7*	LGD	1030 ± 100	hcc	1178 – 738	984 ± 115	266.4	/	/	/
Gasse_9*	LGD	1570 ± 60	hcc	1597 – 1337	1464 ± 63	324	/	/	/
Gasse_11*	LGD	2360 ± 80	hcc	2715 – 2160	2429 ± 140	305	/	/	/
Gasse_13*	LGD	2720 ± 120	hcc	3172 – 2490	2850 ± 55	295	/	/	/
Gasse_16*	LFR	4120 ± 110	<i>Sh (nearshore)</i>	4874 – 4297	4643 ± 144	356.8	/	/	/
Gasse_18*	LGD	4740 ± 150	<i>Mt (nearshore)</i>	5881 – 4983	5455 ± 93	364.9	/	/	/
Gasse_30*	LGD	5830 ± 140	<i>Sh (nearshore)</i>	6972 – 6316	6647 ± 167	365.6	/	/	/
Gasse_33*	LFR	5930 ± 130	<i>Sh (nearshore)</i>	7156 – 6468	6769 ± 167	365.6	/	/	/
Gasse_34*	LFR	6020 ± 160	<i>Sh (nearshore)</i>	7256 – 6501	6884 ± 196	359.6	/	/	/

Gasse_35*	LGD	6130 ± 250	<i>Mt (nearshore)</i>	7498 – 6448	7002 ± 271	337	/	/	/
Gasse_49*	LGD	6835 ± 160	<i>Mt (nearshore)</i>	7965 – 7432	7697 ± 142	381.3	/	/	/
Gasse_50*	LGD	6910 ± 180	<i>Mt (nearshore)</i>	8161 – 7429	7761 ± 161	438.2	9.1	- 70.6	367,5710
Gasse_57*	LGD	7250 ± 150	<i>Mt (nearshore)</i>	8375 – 7795	8082 ± 148	249.7	/	/	/
Gasse_62*	LFR	7450 ± 100	<i>Mt (nearshore)</i>	8416 – 8037	8263 ± 99	373.7	/	/	/
Gasse_63*	LFR	7460 ± 180	<i>Mt (nearshore)</i>	8630 – 7934	8268 ± 181	372.2	/	/	/
Gasse_64*	LFR	7610 ± 140	<i>Sh (nearshore)</i>	8770 – 8054	8420 ± 158	356.8	/	/	/
Gasse_76*	LFR	8340 ± 180	<i>U (nearshore)</i>	9704 – 8770	9302 ± 219	381.8	/	/	/
Gasse_80*	LFR	8450 ± 190	<i>U (nearshore)</i>	10129 – 8996	9434 ± 255	372.2	/	/	/
Gasse_82*	LFR	8530 ± 190	<i>U (nearshore)</i>	10154 – 9090	9545 ± 263	379.6	/	/	/
Gasse_85*	LGD	8600 ± 140	<i>U (nearshore)</i>	10151 – 9305	9633 ± 194	376.5	/	/	/
Gasse_90*	LFR	8770 ± 190	<i>U (nearshore)</i>	10273 – 9436	9848 ± 230	442.4	4.7	- 46.2	396,1144
Gasse_94*	GU	9230 ± 180	<i>Mt (nearshore)</i>	11084 – 9917	10444 ± 260	418,2	4.7	- 49	369,1132
Gasse_96*	LGD	9380 ± 130	<i>Mt (nearshore)</i>	11086 – 10253	10623 ± 216	352.1	/	/	/

1391 **Table 2**

SW Footwall block				NE Axial valley				Offset	Age	Subsidence
Sample ID	Cal.	error	Elevation	Sample ID	Cal.	error	Elevation	(m)		Rate
	Yr BP		(m)		Yr BP		(m)			(mm/yr)
Gasse_94	10444	± 260	418,2	SacA58240	10447	± 73	369	49,2	10447	4,70948
Gasse_90	9848	± 230	442,4	SacA55608	9807	± 81	376,8	65,6	9800	6,69387
Gasse_50	7764	± 160	438,2	Poz-124783	7701	± 44	367,8	70,4	7700	9,14286

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