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## Climate impact on the development of Pre-Classic Maya civilization

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1 **Climate impact on the development of Pre-Classic Maya civilization**

2  
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14  
15 **Keywords**

16 Pre-Classic Maya period, Central Maya Lowlands, climate record, beach ridges, palaeo-precipitation,  
17 500-yr periodicity, 2.8 ka event.

18  
19 **Abstract**

20 The impact of climate change on the development and disintegration of Maya civilization has long  
21 been debated. The lack of agreement among existing palaeoclimatic records from the region has  
22 prevented a detailed understanding of regional-scale climatic variability, its climatic forcing  
23 mechanisms, and its impact on the ancient Maya. We present two new palaeo-precipitation records for  
24 the Central Maya Lowlands, spanning the Pre-Classic period (1800 BCE – 250 CE), a key epoch in the  
25 development of Maya civilization. Lake Tuspan's diatom record is indicative of precipitation changes  
26 at a local scale, while a beach ridge elevation record from world's largest late Holocene beach ridge  
27 plain provides a regional picture. We identify centennial-scale variability in palaeo-precipitation that  
28 significantly correlates with the North Atlantic  $\delta^{14}\text{C}$  atmospheric record, with a comparable periodicity  
29 of approximately 500 years, indicating an important role of North Atlantic atmospheric-oceanic forcing  
30 on precipitation in the Central Maya Lowlands. The Early Pre-Classic period was characterized by  
31 relatively dry conditions, shifting to wetter conditions during the Middle Pre-Classic period, around the  
32 well-known 850 BCE (2.8 ka) event. We propose that this wet period may have been unfavorable for  
33 agricultural intensification in the Central Maya Lowlands, explaining the relatively delayed  
34 development of Maya civilization in this area. A return to relatively drier conditions during the Late  
35 Pre-Classic period coincides with rapid agricultural intensification in the region and the establishment  
36 of major cities.

37  
38 **1. Introduction**

39 During the last decades, a wealth of new data has been gathered to understand human-environmental  
40 interaction and the role of climate change in the development and disintegration of societies in the  
41 Maya Lowlands (e.g., Akers et al., 2016; Douglas et al., 2015, 2016; Dunning et al., 2012, 2015; Lentz  
42 et al., 2014; Turner and Sabloff, 2012). Previous studies have emphasized the impact of prolonged  
43 droughts and their possible link with social downturn, such as the Pre-Classic Abandonment and the  
44 Classic Maya Collapse (Ebert et al., 2017; Hoggarth et al., 2016; Lentz et al., 2014; Kennett et al.,  
45 2012; Medina-Elizalde et al., 2010, 2016; Hodell et al., 1995, 2001, 2005; Haug et al., 2003). Less  
46 attention has been given to episodes of excessive rain and floods that may also have severely impacted  
47 ancient Maya societies (e.g. Iannone et al., 2014). This may be testified by the fact that floods, as well  
48 as droughts, are an important theme depicted in the remaining ancient Maya codices (Fig. 1)  
49 (Thompson, 1972), and Mayan mythological stories (Valásquez Garcíá, 2006).

50  
51 One of the main challenges in palaeoclimatic reconstructions is to unravel climate from human induced  
52 changes. Maya societies played a key role in the formation of the landscape, but the degree of human  
53 induced impact remains highly debated (Hansen, 2017; Beach et al., 2015; Ford and Nigh, 2015). For  
54 example, it is proposed that the increase in sedimentation rate after 1000 BCE at Lake Salpeten  
55 (Anselmetti et al., 2007) and Peten-Itza (Mueller et al., 2009) is related to human induced soil erosion.  
56 However, other high resolution lake records from the area do not show a significant increase in  
57 sedimentation rate during the Pre-Classic or Classic period (e.g. Wahl et al., 2014), and past volcanic  
58 activity could have been responsible for the deposition of 'Maya Clay' (Nooren et al., 2017a).  
59 Palynological records from the Central Maya Lowlands (CML, Fig. 2) show no evidence of widespread  
60 land clearance and agriculture before ~400 BCE (Wahl et al., 2007; Islebe et al., 1996; Leyden et al.,



61 1987), and there is growing consensus that the decline in the percentage of lowland tropical forest  
62 pollen during the Pre-Classic period (Galop et al., 2004; ; Islebe et al., 1996; Leyden et al., 1987) was  
63 caused by climatic drying instead of deforestation (Torrescano and Islebe, 2015; Wahl et al., 2014;  
64 Mueller et al., 2009).

65  
66 In this paper, we present two new palaeo-precipitation records reflecting precipitation changes in the  
67 CML. The records span the Pre-Classic period (1800 BCE – 250 CE), when Maya societies in the CML  
68 transformed from predominantly mobile hunter-gatherers in the Early Pre-Classic Period (e.g. Inomata  
69 et al., 2015; Coe, 2011; Lohse, 2010), to complex sedentary societies that founded impressive cities  
70 like El Mirador by the later part of the Pre-Classic period (Hansen, 2017; Inomata and Henderson,  
71 2016). The period of rapid growth in these centralized societies likely occurred much later than  
72 previously thought, likely sometime after the start of the Late Pre-Classic period around 400 BCE  
73 (Inomata and Henderson, 2016). This raises the question for the reason behind the delayed  
74 development of societies in this area, which was to become the core area of Maya civilization during  
75 the following Classic period (250 – 900 CE). We hypothesize that climate during the Middle Pre-  
76 Classic Period (1000 – 400 BCE) may have been less stable than recently reported (Ebert et al., 2017),  
77 and could have been unfavorable for intensification of maize-based agriculture, which formed the  
78 underlying subsistence economy responsible for the development of many neighbouring Mesoamerican  
79 societies during this period.

80  
81 The CML have been intensively studied, and several well-dated speleothem, palynological, and  
82 limnological records have been obtained for this area (Díaz et al., 2017; Akers et al., 2016; Douglas et  
83 al., 2015; Wahl et al., 2014; Kennett et al., 2012; Mueller et al., 2009; Metcalfe et al., 2009;  
84 Domínguez-Vázquez and Islebe, 2008; Galop et al., 2004; Rosenmeier et al., 2002; Islebe et al., 1996)  
85 (Fig. 2 and A1). However, palaeo-precipitation signals from these records and those from adjacent  
86 areas in the Yucatan and Central Mexico exhibit large differences among records (Fig. A2), making the  
87 reconstruction and interpretation of larger-scale precipitation for the region a challenge (Lachniet et al.,  
88 2013, 2017; Douglas et al., 2016; Metcalfe et al., 2015). Existing climate reconstructions mostly  
89 represent local changes and are predominantly based on oxygen isotope variability, although some new  
90 proxies have been introduced recently (e.g. Díaz et al., 2017; Douglas et al., 2015).

91  
92 We present a regional-scale palaeo-precipitation record for the CML, extracted from world's largest  
93 late Holocene beach ridge sequence at the Gulf of Mexico coast (Fig. 2B). The beach ridge record  
94 captures changes in river discharge resulting from precipitation patterns over the entire catchment of  
95 the Usumacinta River and thus represents regional changes in precipitation over the CML (Nooren et  
96 al., 2017b). Currently the annual discharge of the Usumacinta river is approximately 2000 m<sup>3</sup>/s,  
97 corresponding to ~40 % of the excess or effective rain falling in the 70,700 km<sup>2</sup> large catchment  
98 (Nooren et al., 2017b). Mean annual precipitation within the catchment is ~2150 mm, with 80 % falling  
99 during the boreal summer, related to the North American or Mesoamerican Monsoon system (Lachniet  
100 et al., 2013, 2017; Metcalfe et al., 2015). The interpretation of the beach ridge record is supported by a  
101 new multi-proxy record from Lake Tuspan, an oligosaline lake situated within the CML, receiving  
102 most of its water from a relatively small catchment of 770 km<sup>2</sup> (Fig. 2).

#### 103 104 *Regional palaeo-precipitation signal*

105 The coastal beach ridges consist of sandy material originating from the Grijalva and Usumacinta rivers,  
106 topped by wind-blown beach sand (Nooren et al., 2017b). Although multiple factors determine the final  
107 elevation of the beach ridges, it has been shown that during the period 1775 ± 95 BCE to 30 ± 95 CE  
108 (at 1σ), roughly coinciding with the Pre-Classic period, beach ridge elevation has primarily been  
109 determined by the discharge of the Usumacinta river, in a counter-intuitive manner: low elevation  
110 anomalies of the beach ridges occur in periods with increased river sediment discharge, which in turn is  
111 the product of high precipitation within the river catchment. Under these conditions, beach ridges  
112 develop relatively rapidly, and are exposed to wind for a shorter period. In contrast, during periods of  
113 drought, sediment supply to the coast is reduced, resulting in a decreased seaward progradation rate of  
114 the beach ridge plain. This leaves a longer period for aeolian accretion on the beach ridges near the  
115 former shoreline, resulting in higher beach ridges (Nooren et al., 2017b). Hence, variations in beach  
116 ridge elevation reflect changes in rainfall over the Usumacinta catchment, and thereby represent  
117 catchment-aggregated precipitation, instead of a local signal. The very high progradation rates and the  
118 very robust age-distance model (Fig. A3), with uncertainties of the calibrated ages not exceeding 60–70  
119 years (at 1σ), effectively allow the reconstruction of palaeo-precipitation at centennial time scale.

120



### 121 *Local palaeo-precipitation signal: Lake Tuzsan record*

122 Diatom communities within oligo- to hypersaline lakes are strongly influenced by lake water salinity  
123 (Reed, 1998; Gasse et al., 1995), and we therefore determined diatom assemblage changes within the  
124 Lake Tuzsan sediment record (Fig. 3) to reconstruct palaeo-salinities of the lake water, reflecting  
125 palaeo-precipitation in the lake's catchment. During dry periods, a reduced riverine input of fresh water  
126 and a lowering of the lake level enhance the effect of evaporation and increase the salinity of the lake  
127 water. The first principal component (PC-1) of the variability in the diatom assemblages is interpreted  
128 as an indicator of lake water salinity (Fig. 3). This interpretation is supported by the fact that high PC-1  
129 values are accompanied by relatively high percentages of *Plagiotropis arizonica* (Fig. A4), a diatom  
130 species characteristic of high-conductivity water bodies (Czarnecki and Blinn, 1978).

## 131 **2. Methods**

### 132 *Lake Tuzsan*

133 Two parallel cores, Tuzsán core B and C, were taken with a Russian corer (type GYK) in shallow  
134 water near the inflow of the Rio Dulce, not far from core A which has been studied for pollen (Galop et  
135 al., 2004). Semi-quantitative analyses of Si, S, K, Ca, Ti, Mn and Fe were conducted on both cores  
136 with an X-ray fluorescence core scanner (type AVAATECH) at 0.5 cm intervals. Deposits of large  
137 floods were identified on the basis of elevated concentrations of Si, Fe, Ti and Al, with peak  
138 concentrations exceeding at least the one standard deviation threshold above the mean.

139 Core C was investigated for amorphous silica, charred plant fragments, and diatoms (Fig. 3, and A5).  
140 The core was subsampled at 4-12 cm contiguous intervals, each interval representing 25-80 years. In  
141 addition, 37 1-cm samples (representing ~6.5 yr) were processed using the method outlined by  
142 Battarbee (1973) to determine diatom concentrations and to determine short time variability (decadal  
143 scale). Subsamples were treated with HCl (10 %) to remove calcium carbonate. Large organic particles  
144 were removed by wet sieving (250 µm mesh), and charred plant fragments > 250 µm were counted  
145 under a dissection microscope. Remaining organic material was removed by heavy liquid separation  
146 using a sodiumpolywolframate solution with a density of 2.0 g/cm<sup>3</sup>. A silicious residue, denoted  
147 'amorphous silica' was subsequently removed by heavy liquid separation using a  
148 sodiumpolywolframate solution with a density of 2.3 g/cm<sup>3</sup>, and dry weight was determined after  
149 drying of the samples at 105°C.

150 Slides were prepared from the remaining material. Diatoms were identified, counted and reported as  
151 percentages of the total diatom sum, excluding the small and often dominant *Denticula elegans* and  
152 *Nitzschia amphibia* species. These species show a large variability on short time scales (Fig. A6), and  
153 are not indicative for changes at centennial time scale. We relate changes in diatom assemblages  
154 mainly to lake water salinity changes. The first principal component on the entire assemblage (PC-1) is  
155 interpreted as a palaeosalinity indicator. Diatom taxonomy is mainly after Patrick and Reimer (1966;  
156 1975) and Novelo, Tavera, and Ibarra (2007). We identified *Plagiotropis arizonica* following  
157 Czarnecki and Blinn (1978), and *Mastogloia calcarea* following Lee et al. (2014).

158 The age-depth model for core C is based on seven AMS radiocarbon dated terrestrial samples and  
159 stratigraphical correlation with core A (Fleury et al., 2014). We used a linear regression between the  
160 available radiocarbon dated samples (Fig. A7) which is comparable with the age-depth model by  
161 Fleury et al. (2014) for the time window between ~2500 BCE and 1000 CE.

### 162 *Beach ridge sequence*

163 Beach ridges elevations were extracted from a Digital Elevation Model (DEM) of the coastal plain  
164 along the transects indicated in Fig. 2 (Nooren et al., 2017b). The DEM is based on LiDAR data  
165 originally acquired in April-May 2008 and processed by Mexico's National Institute of Statistics and  
166 Geography (INEGI), Mexico. The relative beach ridge elevation is defined as the difference between  
167 the beach ridge elevation and the long-term (~500 yr) running mean (Fig. A3).

### 168 *Wavelet transfer functions*

169 The relation between our beach ridge and diatom record and other palaeo-precipitation records from  
170 the Maya Lowlands and nearby regions (figure A1 and A2) were investigated by wavelet coherence  
171 (CWT) analyses using the software developed by Grinsted et al. (2004). The record of drift ice from the  
172 North Atlantic (Bond et al., 2001) is bimodally distributed, oscillating between periods of low and high  
173 concentrations of hematite stained grains. The timeseries was therefore transformed into a record of



180 percentiles based on its cumulative distribution function to avoid leakage of the square wave into  
181 frequency bands outside the fundamental period (Grinsted et al., 2004).

182

### 183 3. Climate change in the CML during the Pre-Classic period

184

#### 185 *Early Pre-Classic Period (1800 – 1000 BCE)*

186 The Lake Tuspan diatom record (Fig. 3) indicates relatively dry conditions, comparable to those during  
187 the preceding Late Archaic Period (~5000 – 1800 BCE). Despite the predominantly dry conditions,  
188 large floods still occurred, as demonstrated by the repetitive input of fluvial material into the lake.  
189 These flood events are identifiable as distinctive dark layers of detrital sediment within the calcareous  
190 lake deposits, and are characterized by elevated concentrations of amorphous silica and charred plant  
191 fragments (Fig. 3 and A4). The average recurrence time of large floods was approximately 50 years,  
192 and periods with highest fluvial sediment input in Lake Tuspan coincided with periods of increased  
193 input of charcoal into Lake Peten-Itza (Schüpbach et al., 2015) (Fig. A2). Because the CML were still  
194 sparsely populated during the Early Pre-Classic period (Inomata et al., 2015) we relate the presence of  
195 charcoal to the occurrence of wildfires.

196

197 The beach ridge record indicates a drying trend that culminated in a prolonged dry period at the end of  
198 the Early Pre-Classic period. Although this exceptionally dry phase is less apparent from Lake  
199 Tuspan's diatom record (Fig. 3), it has been recorded at many other sites within the CML. At Lake  
200 Puerto Arturo, high  $\delta^{18}\text{O}$  values on the gastropod *Pyrgophorus* sp. indicate that this was the driest  
201 period since 6300 BCE (Wahl et al., 2014), and the recently extended and improved speleothem  $\delta^{18}\text{O}$   
202 record from Macal Chasm indicates that this dry period was probably at least as severe as any  
203 prolonged droughts during the Classic and Post-Classic Period (Akers et al., 2016). Dry conditions are  
204 reflected in high  $\text{Ca}/\Sigma(\text{Ti,Fe,Al})$  values at Lake Peten-Itza (Mueller et al., 2009), indicating elevated  
205 authigenic carbonate ( $\text{CaCO}_3$ ) precipitation relative to the input of fluvial detrital elements (Ti, Fe and  
206 Al) during this period, and water level at this large lake must have dropped by at least 7 m (Mueller et  
207 al., 2009).

208

#### 209 *Middle Pre-Classic Period (1000 – 400 BCE)*

210 Both the beach ridge and the Lake Tuspan diatom records indicate a change to wetter conditions  
211 around 1000-850 BCE, causing major changes in hydrological conditions in the CML (Fig. 3). The  
212 diatom assemblages in the Lake Tuspan record show a major change in composition. Species indicative  
213 of meso- to polysaline water almost completely disappear, and are replaced by species indicating fresh  
214 water conditions (Fig. 3 (PC1) and A4). In the lake sediments, this transition is also marked by a  
215 lithological shift from laminated to more homogeneous sediments that lack repetitive flood layers,  
216 while charred plant fragments are almost absent until ~400 BCE. Similar abrupt lithological transitions  
217 were reported from Lake Chichancanab (Hodell et al., 1995) and Lake Peten-Itza (Mueller et al., 2009),  
218 and Wahl et al. (2014) describe a regime shift at Puerto Arturo. The sudden reduction in charred plant  
219 fragments around ~1000 BCE at Lake Tuspan coincides with reduced concentrations of charcoal at  
220 Lake Peten-Itza (Fig. A2) (Schupbach et al., 2015) and Laguna Tortuguero, Puerto Rico (Burney and  
221 Pigott Burney, 1994) indicating rapid climatic changes over a large spatial scale.

222

#### 223 *Late Pre-Classic period (400 BCE – 250 CE)*

224 The diatom record at Lake Tuspan (Fig. 3) shows a general increase in lake water salinity, indicating a  
225 gradual shift to drier conditions in the Late Pre-Classic Period. The beach ridge record (Fig. 3)  
226 indicates that a relatively dry period occurred by the onset of the Late Pre-Classic period, which has not  
227 been identified in other proxy records from the region (Fig. A2), although high *Pinus* pollen  
228 percentages in the pollen record from Petapilla pond near Copan (McNeil, 2010) during this period  
229 may indicate dry conditions, as high *Pinus* pollen percentage at highland sites could be indicative for  
230 drier conditions (Domínguez-Vázquez and Islebe, 2008).

231

#### 232 *Precipitation variability over long time scales*

233 The observed general drying trend over the last thousands of years may be related to the southward  
234 shift of the ITCZ during the late Holocene. The shift occurred in response to orbitally-forced changes  
235 in insolation (Haug et al., 2001), causing a gradual Northern Hemisphere cooling versus Southern  
236 Hemisphere warming (Fig. 3), thereby shifting the ITCZ towards the warming southern hemisphere  
237 (Schneider et al., 2014). A more northerly position of the ITCZ during the Pre-Classic period may be  
238 related to stronger easterly tradewinds and the less frequent occurrence of cold fronts during the Pre-  
239 Classic period, as beach ridge morphological changes suggest (Nooren et al., 2017b).



240

#### Centennial scale precipitation variability

241 Wavelet coherence (WTC) analysis (Grinsted et al., 2004) indicates in-phase coherence between the  
242 beach ridge record and the recently extended and revised calcite  $\delta^{18}\text{O}$  speleothem record from Macal-  
243 Chasm cave (Akers et al., 2016) (Fig. A8). The in-phase relationship between the two records is  
244 significant above a 5% confidence level at centennial timescales during the Pre-Classic Period. We did  
245 not find significant relationships between the beach ridge record and other palaeo-precipitation records  
246 from the CML, nor with records from the Yucatan and Central Mexico (Fig. A2), except for a  
247 significant in-phase coherence at centennial time scale with the *Pyrgophorus* sp.  $\delta^{18}\text{O}$  record from Lake  
248 Chichancanab (Hodell et al., 1995).  
249

250

251 The coherence between the beach ridge record and the well-dated Macal-Chasm speleothem record  
252 give us confidence that these records reflect regionally coherent variability at centennial timescales  
253 during the Pre-Classic period. Interestingly, the beach ridge record is significantly in anti-phase with the  
254 North Atlantic ice drift record (Bond et al., 2001) and the Northern Hemispheric atmospheric  $\delta^{14}\text{C}$   
255 record during the Pre-Classic Period (Reimer et al., 2013) (Fig. 4), suggesting an important role of  
256 North Atlantic atmospheric-oceanic forcing on precipitation in the CML. The Northern Hemispheric  
257 atmospheric  $\delta^{14}\text{C}$  record shows a 512-yr periodicity (Stuiver and Braziunas, 1993), which is similar to  
258 the observed ~500 year periodicity of the beach ridge record during the Pre-Classic period. Such a  
259 centennial scale periodicity is not apparent in Lake Tuspan's diatom record (Fig. 3), nor in any of the  
260 other palaeo-precipitation records from the Maya Lowlands (Fig. A2), but has been identified in the Ti  
261 record from Lake Juanacatlán in the highlands of Central Mexico (Jones et al., 2015). This periodicity  
262 has been related to the intensity of the North Atlantic thermohaline circulation and variations in solar  
263 activity (Stuiver and Braziunas, 1993).  
264

265

265 The coherence with fluctuations in solar irradiance is most evident during the 2.8 ka event, related to  
266 the Homeric Grand Solar Minimum. At this time, a strong decrease in the total solar irradiance resulted  
267 in higher atmospheric  $^{14}\text{C}$  production and a change to cooler and wetter condition in the Northern  
268 Hemisphere (e.g. Van Geel et al., 1996), and apparently also a shift to wetter conditions in the CML,  
269 evident from our two new palaeo-precipitation records (Fig. 3). This correlation should not be used as  
270 an analogue for modern precipitation variability, when periods of lower solar activity are associated  
271 with lower Usumacinta River discharge and hence less precipitation in the CML (Fig. A9).  
272

273

273 A similar precipitation response to the late Holocene southward shift of the ITCZ for both Northern  
274 South America and the Maya Lowlands has previously been suggested (Haug et al., 2003), implying  
275 that the beach ridge record should be in-phase with the Cariaco Ti record (Haug et al., 2001). Although  
276 the Cariaco record indicates large centennial scale variability in precipitation over Northern South  
277 America (Fig. 3), this variability is not significantly correlated with the beach-ridge record. The  
278 correlation slightly improved using an updated age-depth model for the Cariaco record (Fig. A10), but  
279 remains insignificant, probably due to uncertainties in the chronological control of both records or due  
280 to a more prominent influence of the Northern Atlantic climatic forcing mechanisms in the Maya  
281 Lowlands.  
282

283

#### 4. Precipitation versus human development in the CML

284 Our records indicate that the Early Pre-Classic period in the CML was relatively dry. During this  
285 period, the CML were still sparsely populated by moving hunter-gatherers. It is highly likely that  
286 before maize became sufficiently productive to sustain sedentism, the karstic lowlands were less  
287 attractive for humans than the coastal wetlands along the Gulf of Mexico and Pacific coast, where  
288 natural resources were abundantly present to successfully sustain a hunting/gathering subsistence  
289 system (Inomata et al., 2015). Reliance on cultivated crops, most notably maize, rapidly increased after  
290 the onset of the Middle Pre-Classic period around 1000 BCE (Rosenswig et al., 2015). Between 1000 –  
291 850 BCE, under still dry conditions, there is evidence for increased maize agriculture in the Pacific  
292 flood basin (Rosenswig et al., 2015), and within the Olmec area at the Gulf of Mexico coast (Arnold  
293 III, 2009), and maize grains (AMS  $^{14}\text{C}$  dated to  $875 \pm 29$  BCE) have been found as far as Ceibal within  
294 the CML (Inomata et al., 2015). We speculate that wetter conditions after 850 BCE might have been  
295 unfavorable for a further development of intensive agriculture in the CML. This is supported by  
296 palynological evidence, indicating that widespread land clearance and agriculture activity did not occur  
297 before ~400 BCE (Wahl et al., 2007; Galop et al., 2004; Islebe et al., 1996; Leyden et al., 1987),  
298 despite some early local agricultural activity (Wahl et al., 2014; Rushton et al., 2013; McNeil et al.,  
299 2010; Galop et al., 2004). A return to drier conditions during the Late Pre-Classic period coincided



300 with an expansion of maize-based agriculture in the CML, and communities within the Maya Lowlands  
301 show a strong and steady development with relatively uniform ceramic and architectural styles  
302 (Hansen, 2017; Inomata and Henderson, 2016). Hence, major development of Maya civilization in the  
303 Central Maya Lowlands occurred only after the onset of the Late Pre-Classic period, when climate  
304 became progressively drier, in line with earlier findings that drier conditions were favorable for  
305 agricultural development in the CML (Wahl et al., 2014).

306

307

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314

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613

#### 614 **Figure captions**

616 Figure 1: The image on page 74 of the Codex Dresden depicts a torrential downpour probably  
617 associated with a destructive flood (Thompson, 1972).  
618

619 Figure 2: A large part of the Central Maya Lowlands (outlined with a red dashed line) is drained by the  
620 Usumacinta (Us.) River (A). During the Pre-Classic period this river was the main supplier of sand  
621 contributing to the formation of the extensive beach ridge plain at the Gulf of Mexico coast (B).  
622 Periods of low rainfall result in low river discharges and are associated with relatively elevated beach  
623 ridges. The extend of the watersheds of the Usumacinta and Dulce River is calculated from SRTM 1-  
624 arc data (USGS, 2009). Indicated are archaeological sites (squares) and proxy records discussed in the  
625 text; Tu= Lake Tuspan, Ch = Lake Chichancanab, PI = Lake Peten-Itza, MC = Macal Chasm Cave, and  
626 PA = Lago Puerto Arturo.  
627

628 Figure 3: Comparison of the Lake Tuspan and beach ridge record (A) with local and proximal records  
629 from Macal-Chasm cave (Akers et al., 2016) and the Cariaco basin (Haug et al., 2001)(B). The Cariaco  
630 record is conform updated age-depth model (Fig. A10). Climate records related to North Atlantic  
631 atmospheric-oceanic forcing are indicated in panel C, including the drift ice reconstruction from the  
632 North Atlantic (Bond et al., 2001), the Northern Hemispheric residual atmospheric  $\delta^{14}\text{C}$  content  
633 (Reimer et al., 2013), the Northern-to Southern hemispheric temperature anomaly (Schneider et al.,  
634 2014) and reconstructed Total Solar Irradiance (TSI) (Steinhilber et al., 2012).  
635

636 Figure 4: Wavelet Transform Coherence (WTC) analysis between the beach ridge record and the  
637 Northern Hemispheric atmospheric  $\delta^{14}\text{C}$  record (Reimer et al., 2013)(A) and the North Atlantic ice drift  
638 record (Bond et al., 2001)(B). The beach ridge record is significantly in anti-phase with both records at  
639 approximately 500 yr time scale, indicating an important role of North Atlantic atmospheric-oceanic  
640 forcing on precipitation in the Maya Lowlands during the Pre-Classic period. The 5% significance level  
641 against red noise is shown as a thick contour. Arrows indicate phase difference, with in-phase  
642 relationship between records if arrows point to the right.  
643

#### 644 **Appendix: Additional figures**

646 Figure A1: Location of proxy records indicated in figure A2 and/or mentioned in the main text. A:  
647 Northern Maya Lowlands (Tz=Tzabnah, PL=Punta Laguna, RS=Rio Secreto, Ch=Chichancanab and  
648 Si=Silvituc), the Central and Southern Maya Lowlands (PA=Puerto Arturo, NRL=New River Lagoon,  
649 Tu=Tuspan, PI/Sa=Petén-Itza and Salpeten, MC/CH=Macal Chasm and Chen Ha, and YB=Yok  
650 Balum), the Maya Highlands (Oc/Na= Ocotitalito and Naja, Am=Amatitlan, and Pet=Petapilla). B:  
651 Central Mexico (Jua=Juanacatlan, CdD=Cueva de Diablo, Jx=Juxtlahuacan, and Alj=Aljojuca) and the  
652 marine record from the Cariaco (C) basin. Annual precipitation (1950-2000) calculated with  
653 WorldClim version 1.4 (release3); Hijmans et al. (2005). Long term (1958-1998) mean ITCZ position  
654 and wind at 925 hPa ( $\text{m}\cdot\text{s}^{-1}$ ) for July after Amador et al. (2006), based on NCED/NCAR Reanalysis  
655 data (Kalnay et al., 1996).



656

657 Figure A2a: Palaeoprecipitation records from the Central Maya Lowlands and Yucatan; Beach ridge  
658 elevation and Tuspan diatom record (this study), compiled record of Central Peten and Yucatan  
659 (Douglas et al., 2016), Salpeten and Chichancanab dD wax-corr. (Douglas et al., 2015), Salpeten  $\delta^{18}\text{O}$   
660 (Rosenmeier et al., 2012), Peten-Itza  $\delta^{18}\text{O}$  (Curtis et al., 1998), Puerto Arturo  $\delta^{18}\text{O}$  (Wahl et al., 2014),  
661 Macal Chasm  $\delta^{18}\text{O}$  (Akers et al., 2016), Chen Ha  $\delta^{18}\text{O}$  (Pollock et al., 2016), Yok Balum  $\delta^{18}\text{O}$  (Kennett  
662 et al., 2012), Rio Secreto  $\delta^{18}\text{O}$  (Medina-Elizalde et al., 2016), Silvituc DV-pollen (Torrescano-Valle  
663 and Islebe, 2015), Chichancanab S and  $\delta^{18}\text{O}$  (Hodell et al., 1995), Punta Laguna  $\delta^{18}\text{O}$  (Hodell et al.,  
664 2007), and Tzabnah  $\delta^{18}\text{O}$  (Medina-Elizalde et al., 2010).

665

666 Figure A2b: Proxy records from the Central Maya Lowlands, the Maya Highlands and Central Mexico.  
667 Peten-Itza charcoal (Schüpbach et al., 2015), Peten-Itza pollen (Islebe et al., 1996), Amatitlan  
668 Aulacoseira and Pinus (Velez et al., 2011), Petapilla Pinus (McNeil et al., 2010), Naja Pinus  
669 (Domínguez-Vázquez and Islebe, 2008), Ocotalito Sr (Díaz et al., 2017), Aljojuca  $\delta^{18}\text{O}$  (Bhattacharya  
670 et al., 2015), Cueva del Diablo  $\delta^{18}\text{O}$  (Bernal et al., 2011), Juxtahuaca  $\delta^{18}\text{O}$  (Lachniet et al., 2015,  
671 2017), and Juanacatlan Ti -15 point running mean (Jones et al., 2015).

672

673 Figure A3: Age-distance model for beach ridge transect B (after Nooren et al., 2017b).

674

675 Figure A4: Summarized proxy record of Lake Tuspan sediment core C. The 1-4 cm thick dark  
676 palaeoflood-layers contrast with the predominantly light coloured calcareous deposits, and are  
677 characterized by elevated detrital input, resulting in elevated concentrations of Si (cps = counts per  
678 second), amorphous silica (% of dry weight), and charred plant fragments (number of particles/g dw).  
679 Only the relative abundance of 'key' diatom species are shown here and the small and often dominant  
680 *Denticula elegans* and *Nitzschia amphibia* species were excluded from the diatom sum. The first  
681 Principal Component axis (PC-1) is interpreted as a lake water salinity indicator, with low values  
682 corresponding to high salinity waters, reflecting relatively dry conditions. Notice abrupt change around  
683 1000 BCE.

684

685 Figure A5: Diatom record for lake Tuspan core C. Diatom concentration (\*1000 valves/g dw) were  
686 determined on 37 selected 1-cm samples and diatom percentages (only the 'key species' are shown  
687 here) were determined on the 123 subsamples at 4-12 cm contiguous intervals. The small and often  
688 dominant *Denticula elegans* and *Nitzschia amphibia* species were excluded from the diatom sum.

689

690 Figure A6: Detailed diatom record around one of the larger flood event ~1200 BCE

691

692 Figure A7: Age-depth model for Tuspan core C. The age-depth model is based on a linear  
693 interpolation between calibrated ages of radiocarbon dated terrestrial macroremains from core A  
694 (Galop et al., 2004) and core C (Fleury et al., 2014). The model is most reliable for ages between  
695 ~2500 BCE and 1000 CE.

696

697 Figure A8: Wavelet Transform Coherence (WTC) analysis between the beach ridge record and the  
698 Macal Chasm  $\delta^{18}\text{O}$  record (Akers et al., 2016). The 5% significance level against red noise is shown as  
699 a thick contour. Arrows indicate phase difference, with in-phase relationship between records if arrows  
700 point to the right.

701

702 Figure A9: Mean annual discharge of the Usumacinta river at Boca del Cerro (Banco Nacional de  
703 Datos de Aguas Superficiales, consulted in January 2017) compared with the total solar irradiance  
704 (TSI). The TSI is comprised of the reconstruction from 1700-2004 (Krivovo et al., 2007), concatenated  
705 with observations from the Total Irradiance Monitor (TIM) on NASA's Solar Radiation and Climate  
706 Experiment (SORCE) from 2005-2011 (Kopp and Lean, 2011). 4.56 watts are added to the TIM  
707 measurements as previous reconstructions were calibrated against less accurate measuring equipment,  
708 compared with the TIM instrument, which led to an overestimation of TSI.

709

710 Figure A10: Updated age-depth model for Cariaco core 1002D. Original model (Haug et al., 2001) has  
711 been based on a linear interpolation of calibrated ages. We applied a 4th order polynomial fit through  
712 modelled ages calculated with a P\_sequence model (Oxcal 4.2) (Bronk Ramsey, 2009, 2016):  
713  $k = 10$ , Marine13 calibration curve,  $\Delta R = 15 \pm 50$ , one outlier: NSRL-13050.

714

715



Fig. 1

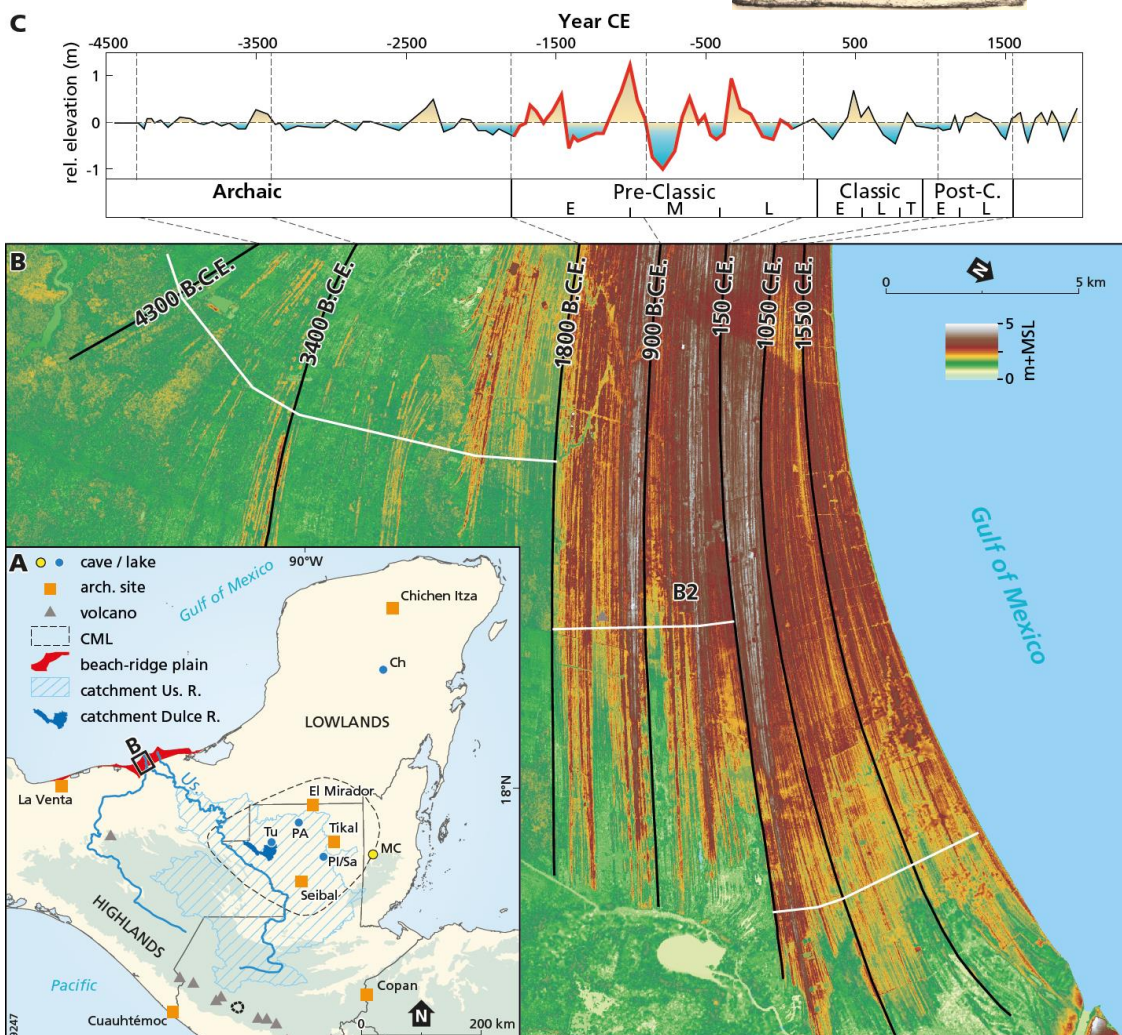


Fig. 2

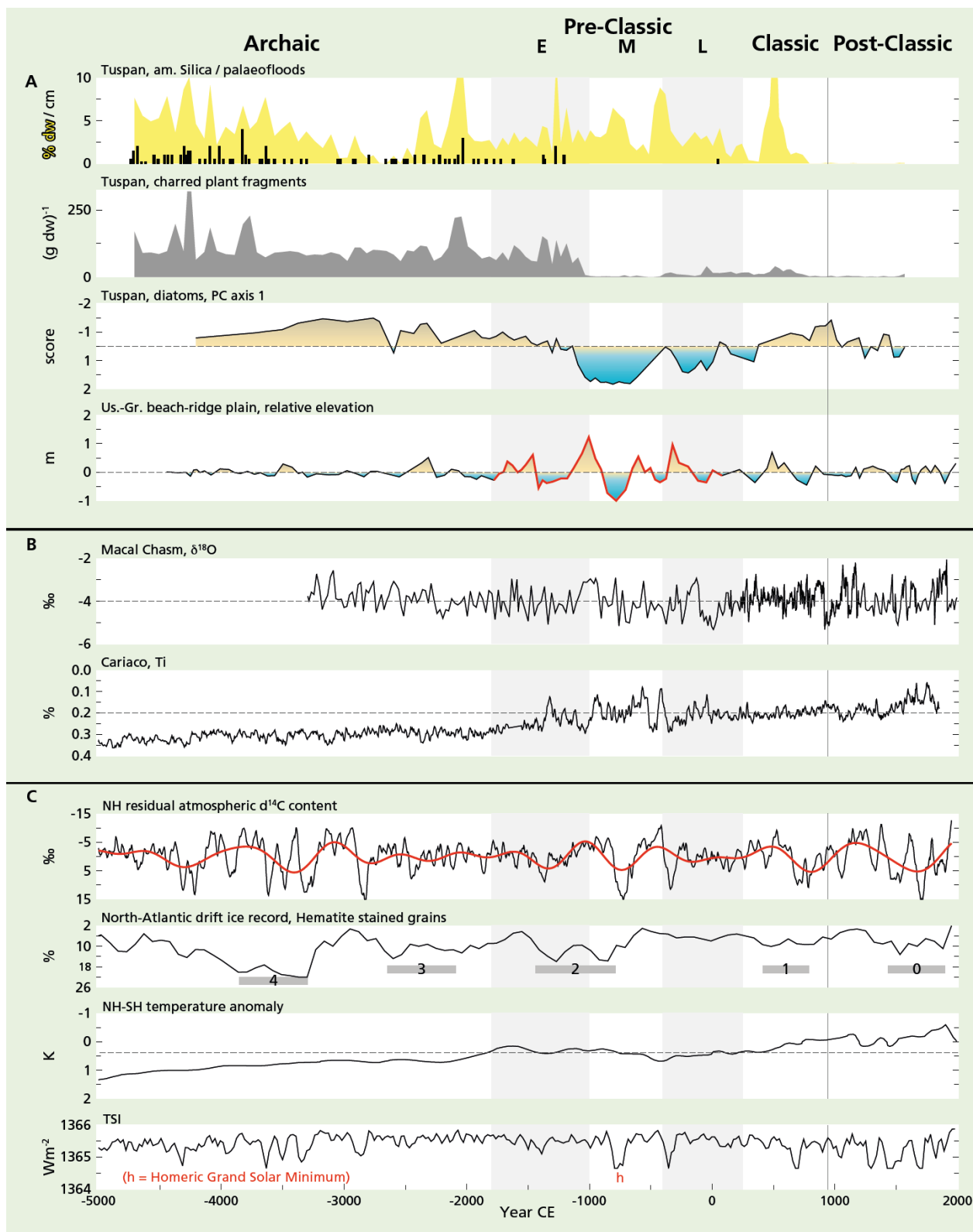


Fig. 3

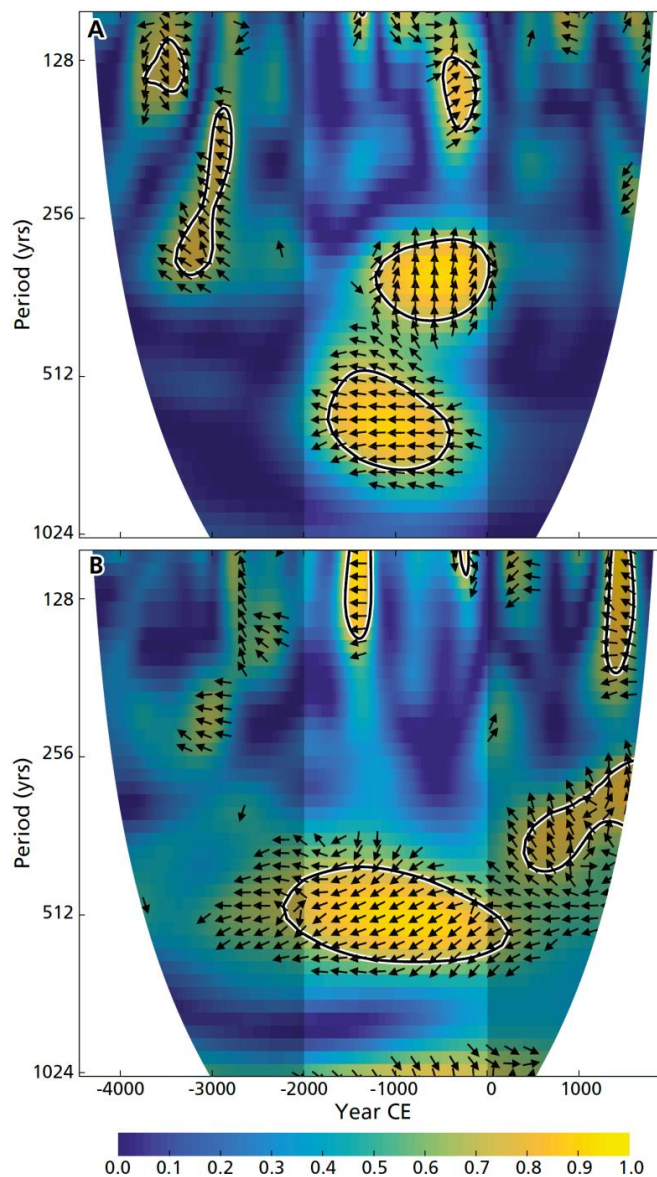


Fig. 4



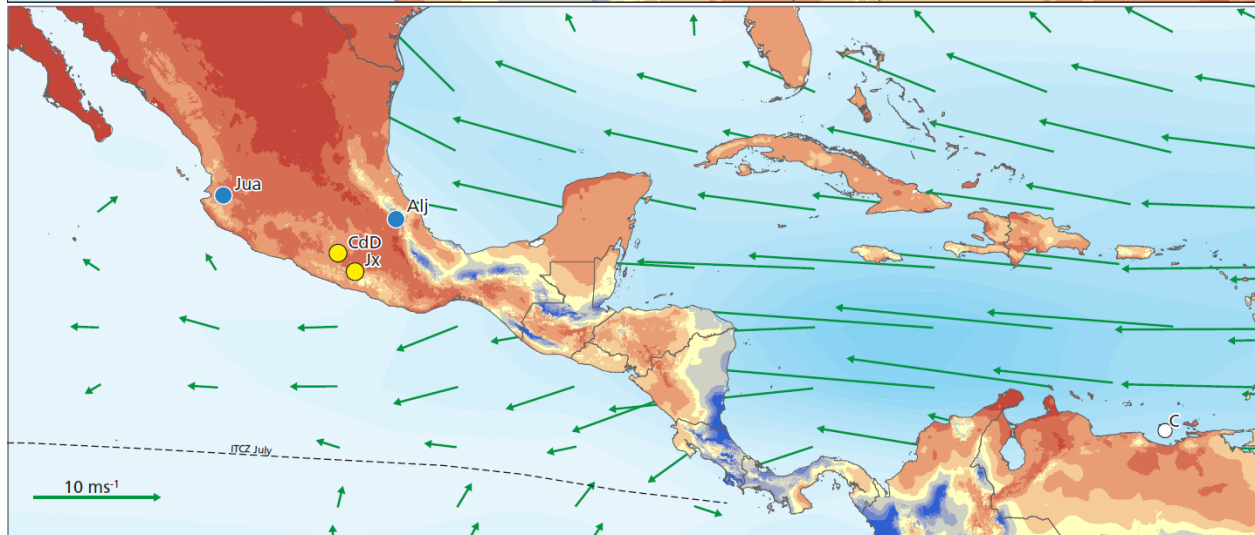
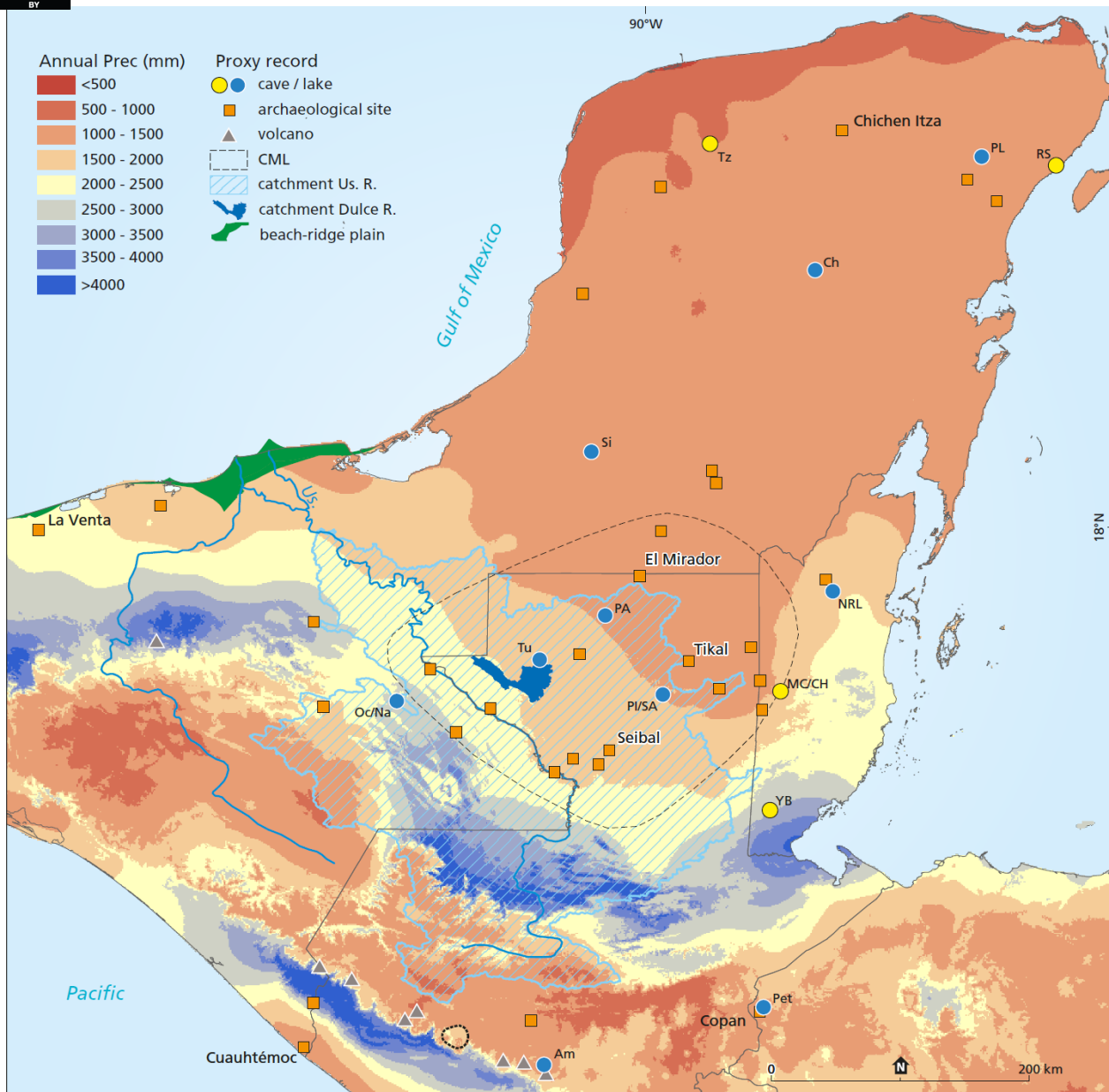


Fig. A1

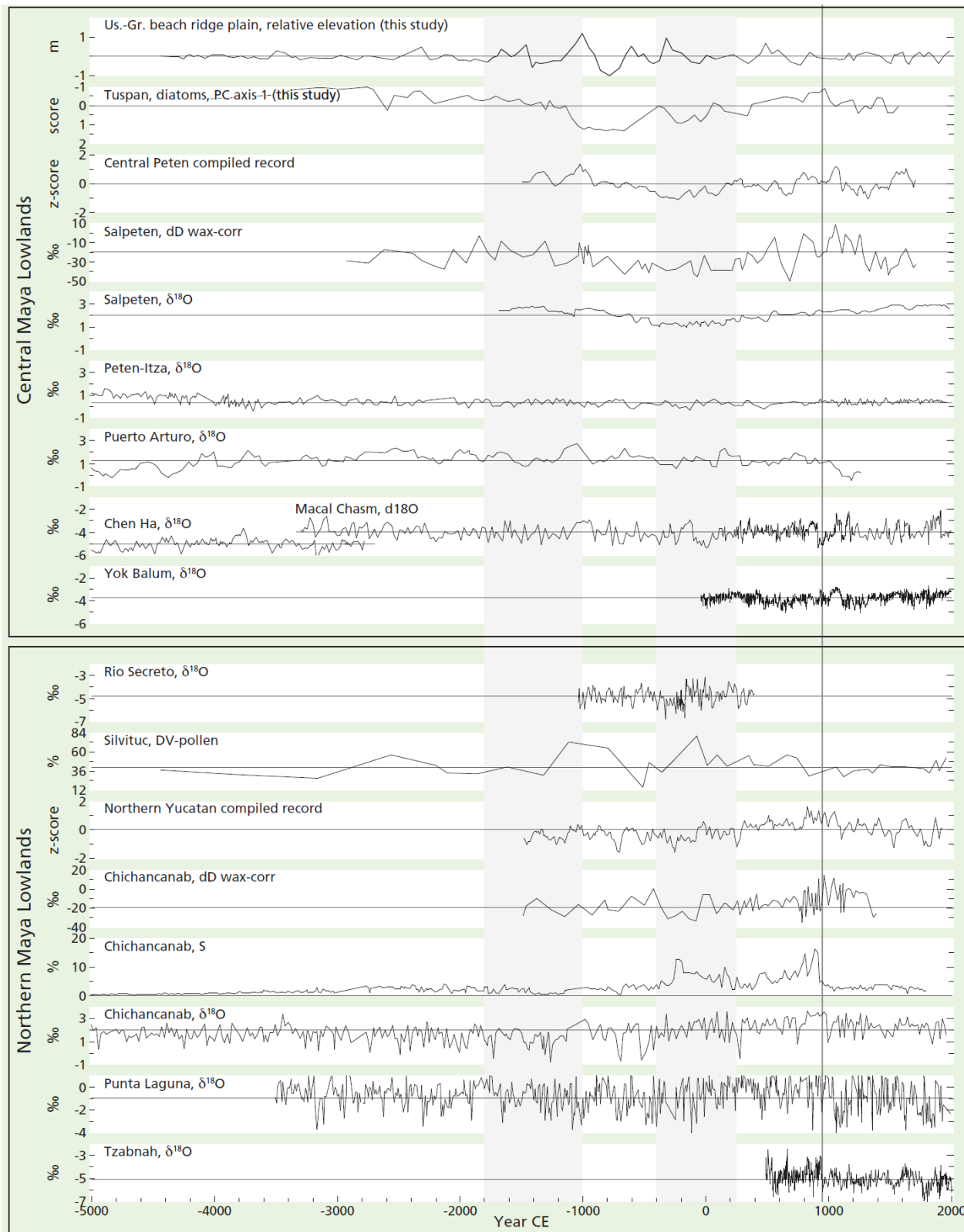


Fig. A2a

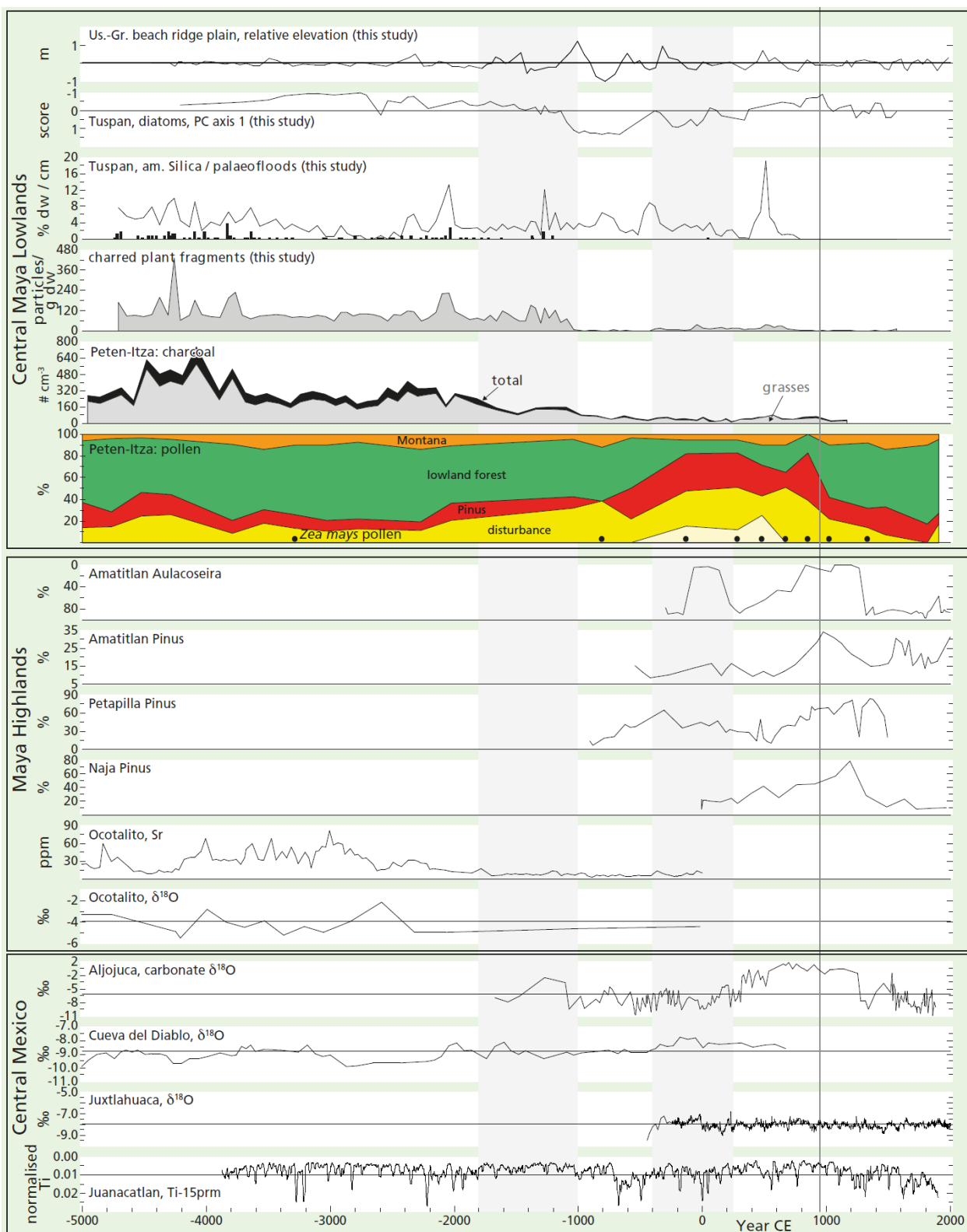


Fig. A2b

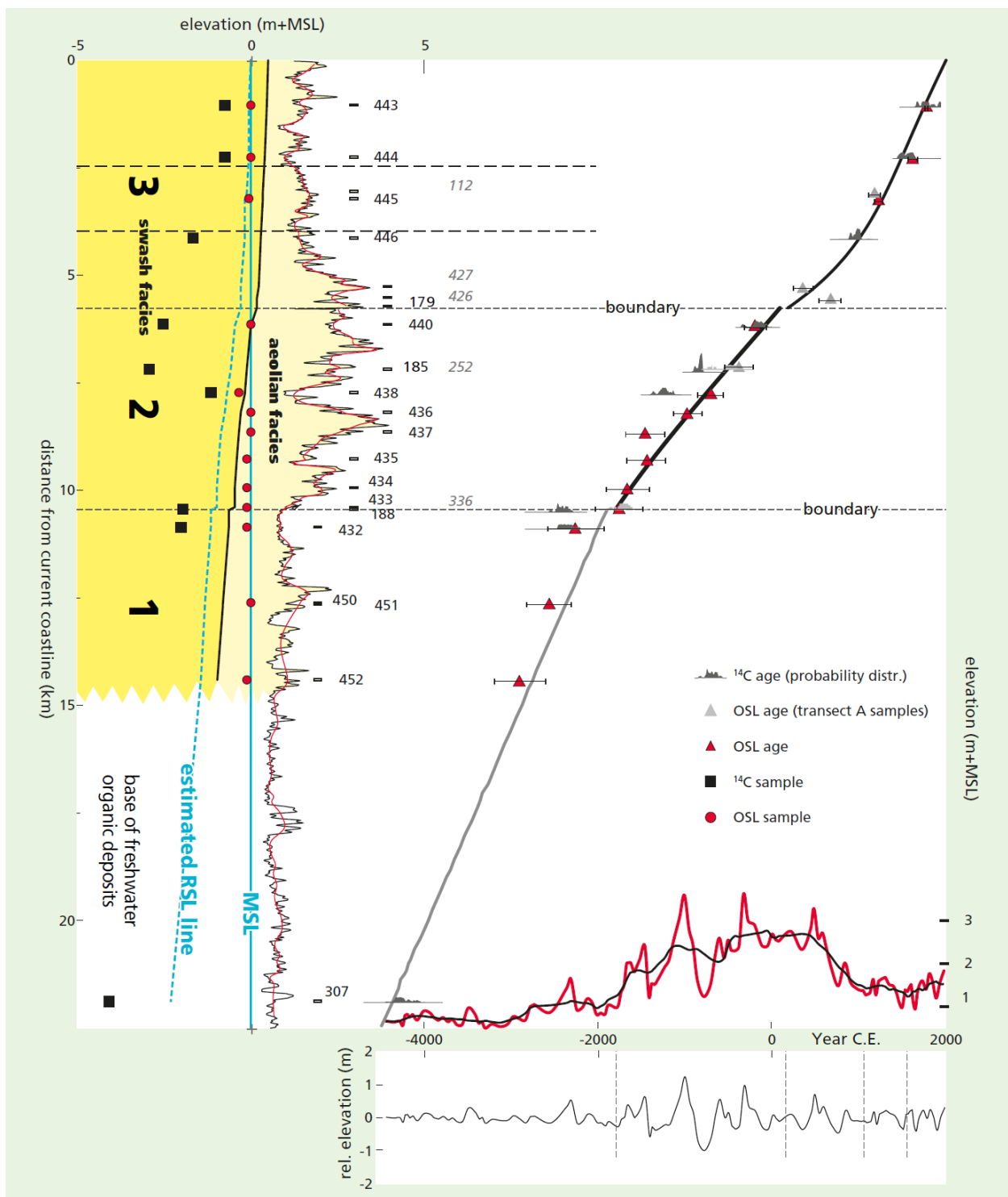


Fig. A3

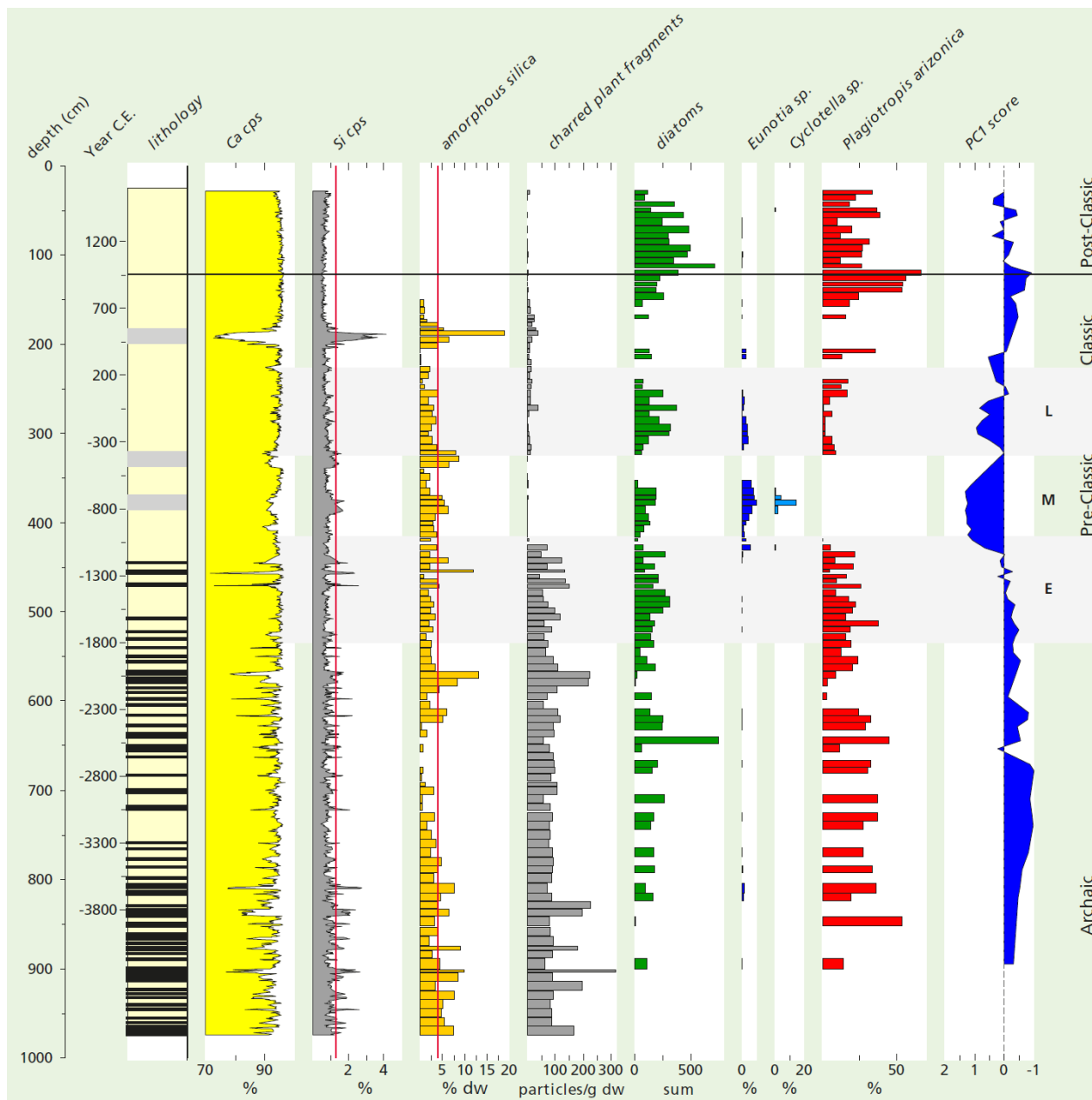


Fig. A4

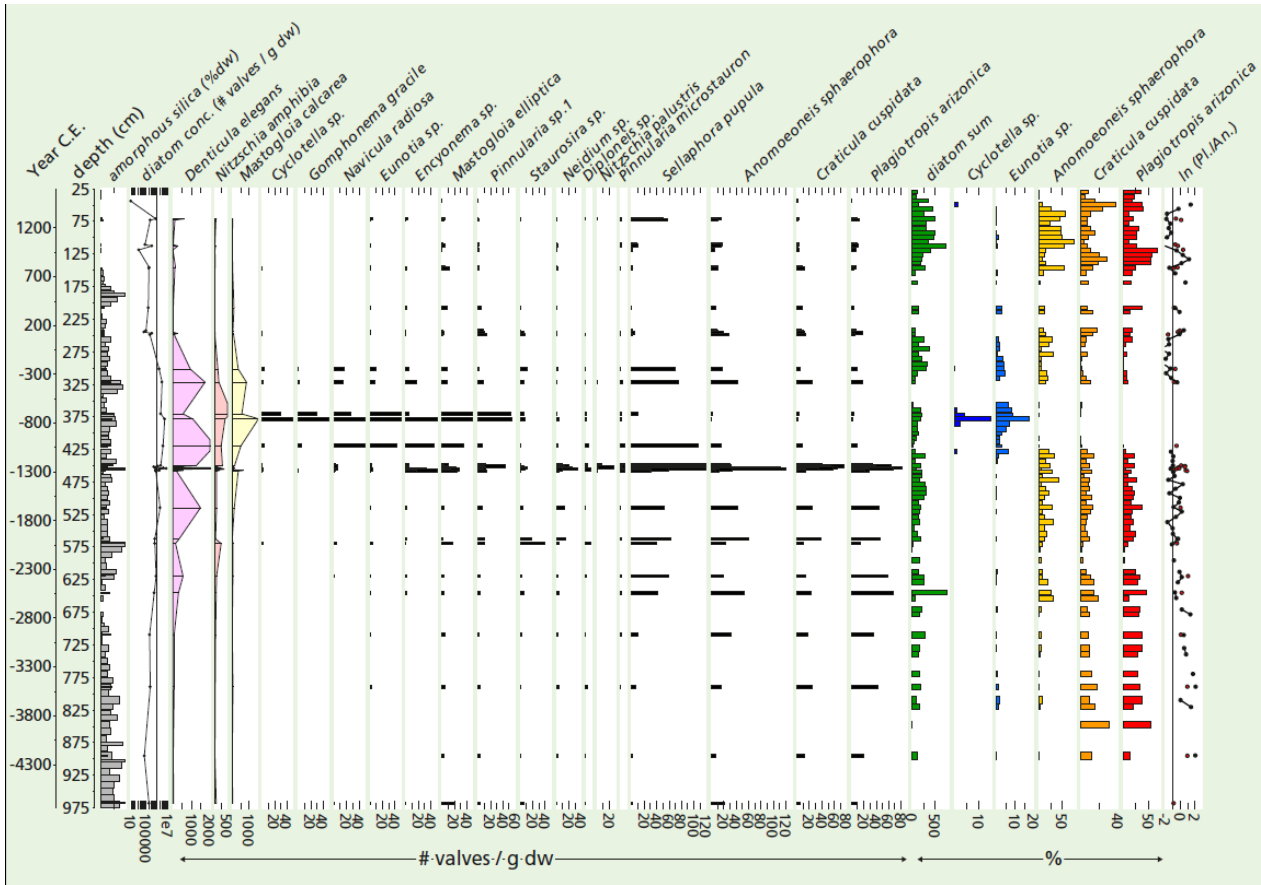


Fig. A5

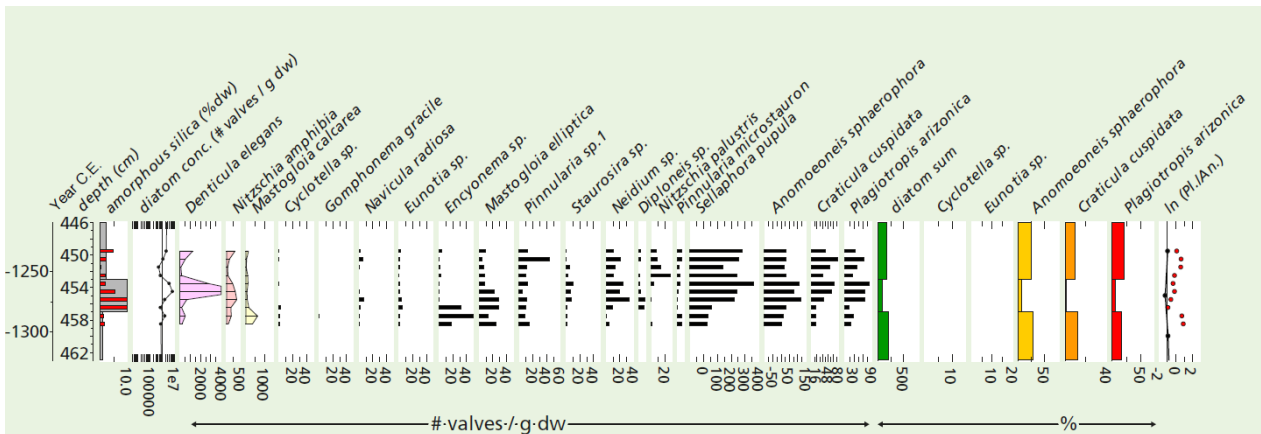


Fig. A6

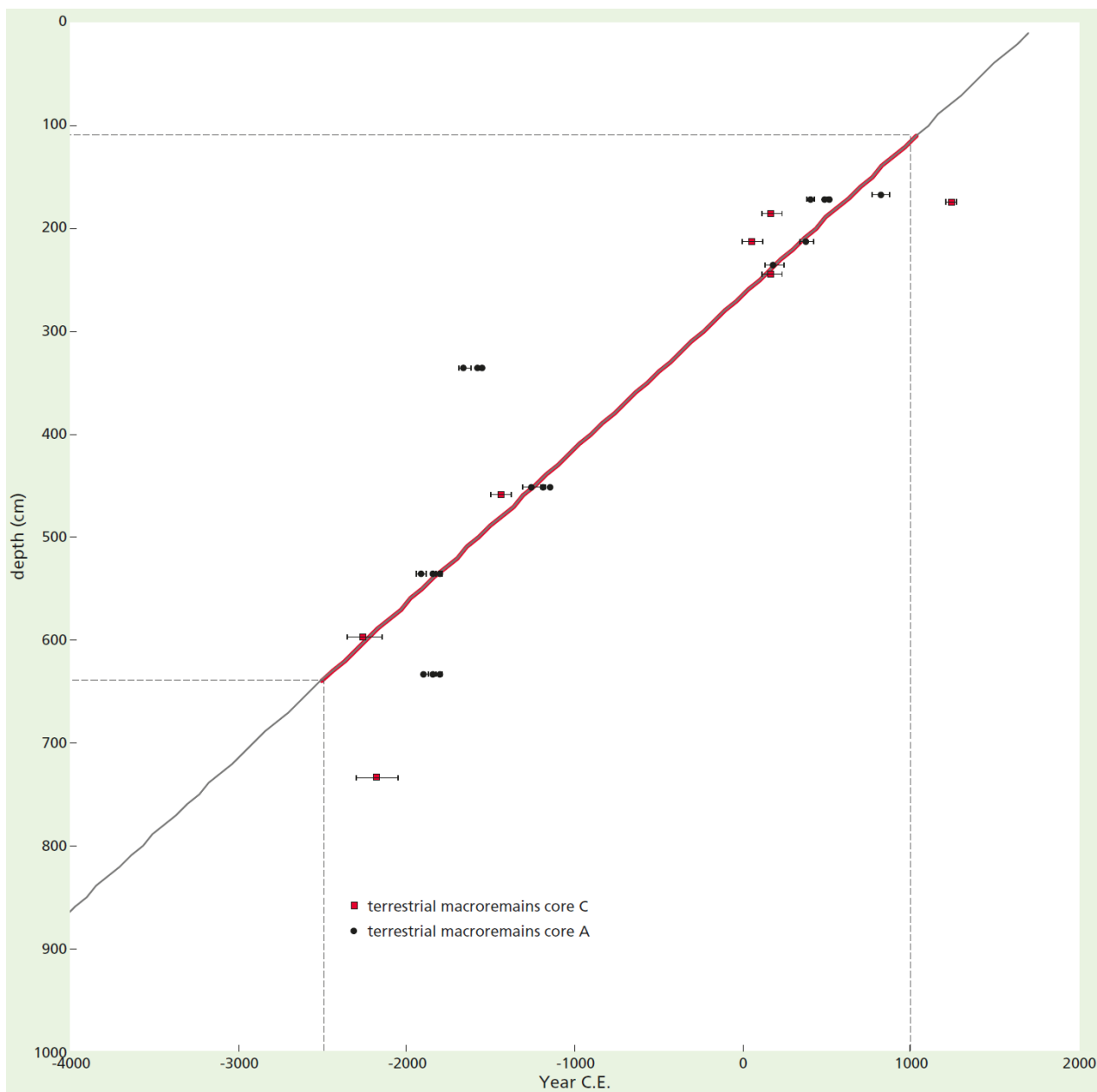


Fig. A7

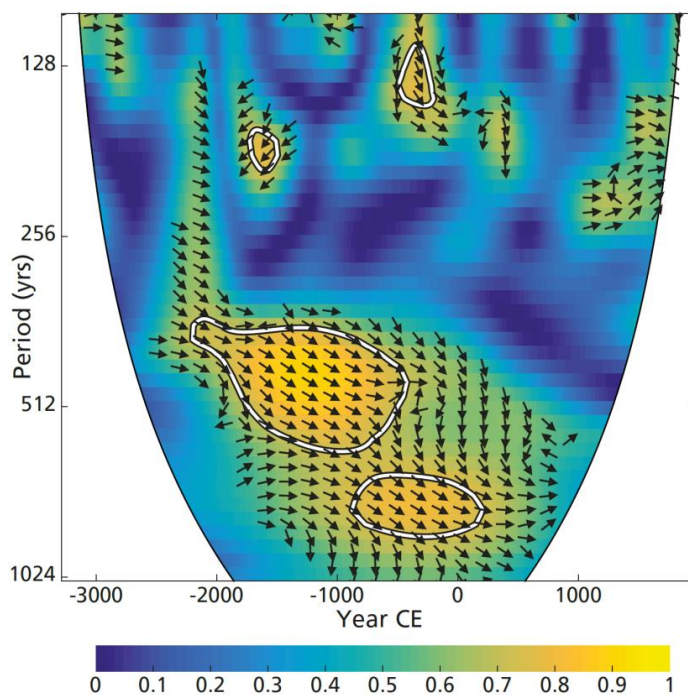


Fig. A8

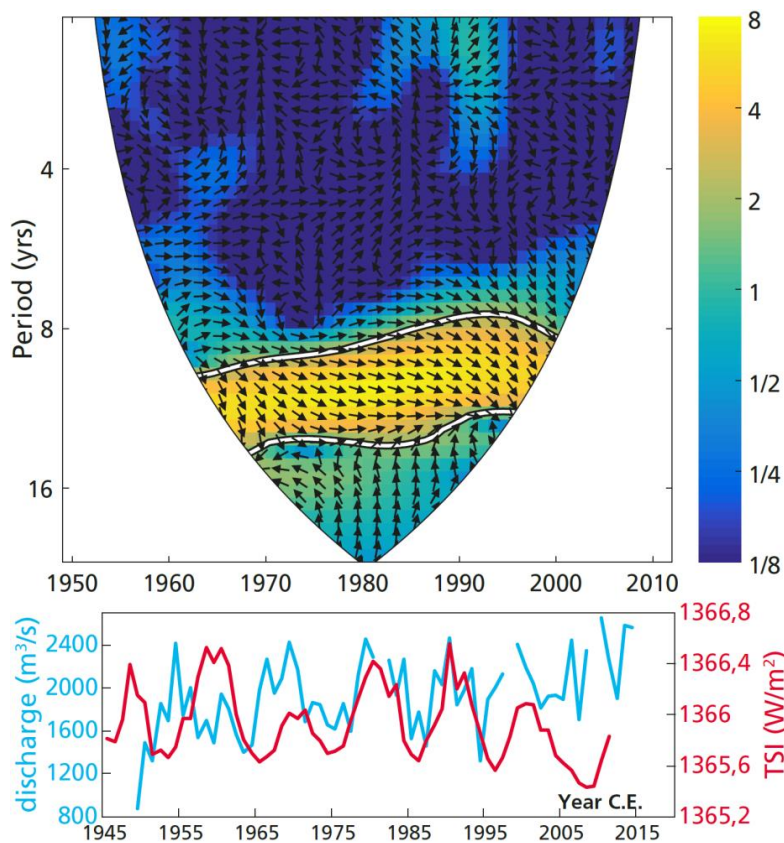


Fig. A9



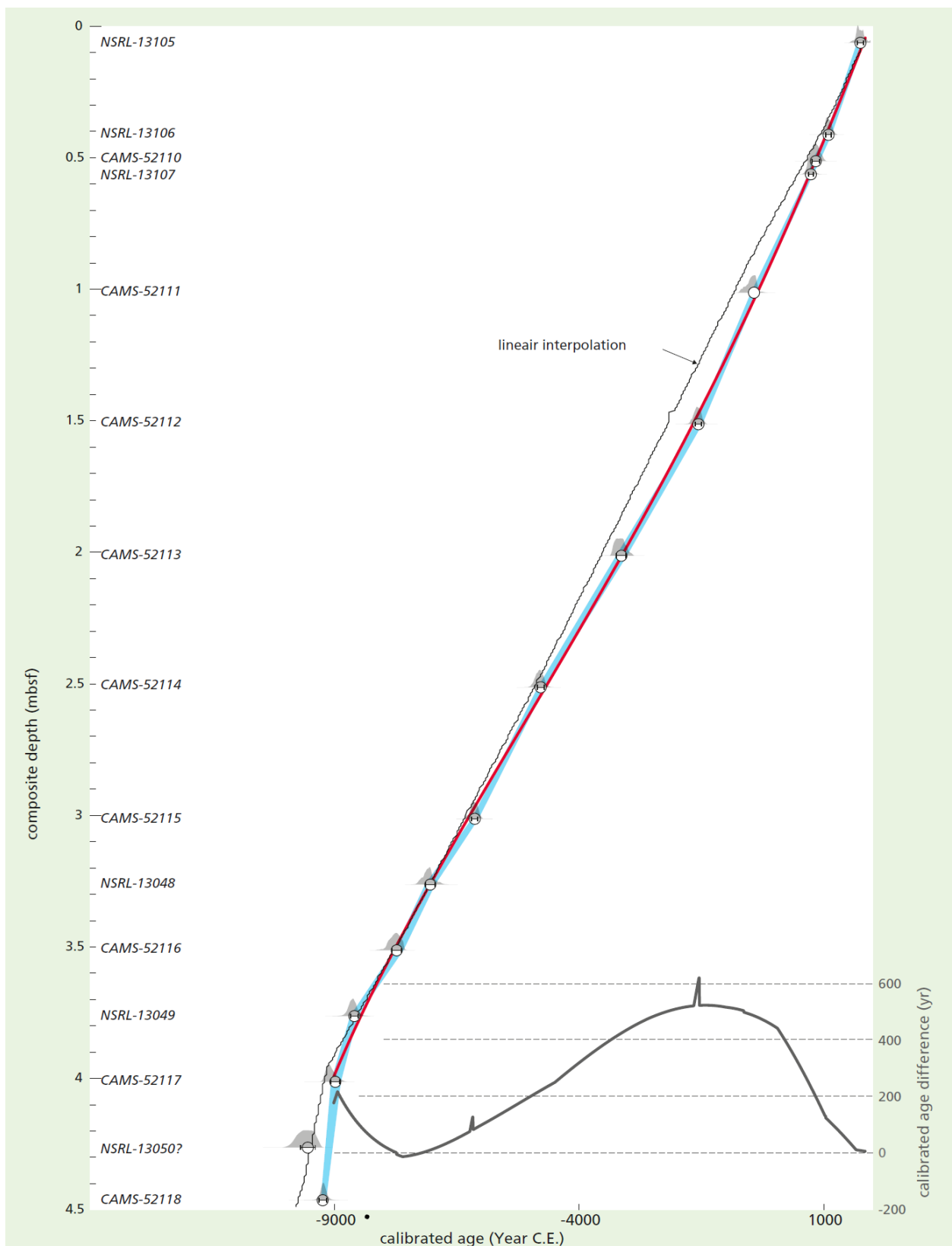


Fig. A10