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Diatom, phytolith, and pollen records from a $^{10}\text{Be}/^9\text{Be}$ dated lacustrine succession in the Chad basin: Insight on the Miocene–Pliocene paleoenvironmental changes in Central Africa



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ABSTRACT

A discontinuous 200 m-long borehole drilled in the Bol Archipelago (13°N, Lake Chad) provided 25 samples, which were dated using the $^{10}\text{Be}/^9\text{Be}$ method and analyzed for their micro-biological content. The dating provided ages ranging from 6.3 ± 0.1 to 2.6 ± 0.1 Ma, a period contemporaneous with the Pliocene fossil localities located in the current Djurab desert of Chad (16–17°N).

Well-preserved diatom assemblages first occurred at 4.7 ± 0.1 Ma and were dominated by the freshwater planktonic genera *Aulacoseira* and *Stephanodiscus* until the end of the Pliocene. This supports the recurrence of lacustrine conditions at Bol during all the Pliocene. The presence of pelite and argillaceous deposits in the core before 4.7 ± 0.1 Ma, however, suggests that the lake settled earlier, at least since 6.3 ± 0.1 Ma. The abundance of Afromontane pollen taxa at 4.2 ± 0.1 Ma and the occurrence of trapeziform polylobate phytoliths throughout the sequence suggest significant vegetation inputs from the southern highlands, while the importance of kaolinite in the clay sediments indicates a water supply predominantly from the south during the Pliocene.

Phytolith assemblages are all dominated by lobate grass silica short cells and by blocky and elongate types, which attest to the presence of herbaceous-dominated vegetation around Bol and/or in the southern drainage basin during the Pliocene. This result is also supported by the pollen assemblage described at 4.2 ± 0.1 Ma, which shows highest affinity for the savanna biome. Moreover, low values for the Xerophytic grass phytolith index indicate the presence of humid-loving (mesophytic and aquatic) grass communities in this vegetation.

At last, significant variations in the abundance of blocky and elongate phytoliths are indicative of local alternations of fully lacustrine and marshy conditions at Bol. Particularly between 3.6 and 2.7 Ma, the abundance of silicified bulliform cells combined with the absence of diatoms support a significant lacustrine reduction at Bol favoring the increasing of local marshy vegetation.

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1. Introduction

First hominins appeared during the late Neogene, when significant vegetation changes occurred in Africa (e.g. Bonnefille, 2010). These changes notably include the appearance and expansion of C_4 grasses (Cerling et al., 1997; Feakins et al., 2005; Edwards et al., 2010; Cerling

et al., 2011; Strömberg, 2011; Dupont et al., 2013; Feakins et al., 2013), which are currently part of tropical grasslands and savannas at low altitude (Sage, 2004). Moreover, marine records off the African coasts indicate high climate variability in tropical Africa during the late Neogene (Feakins and deMenocal, 2010). This environmental variability is revealed by a succession of humid and arid phases tuned with precessional cyclicity of the Earth's orbit (deMenocal, 2004).

Global vegetation and climate changes may have triggered hominin speciation and adaptive features (see review by Maslin et al., 2015). Particularly, the spread of C_4 -grass savannas may have favored the

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appearance of bipedalism in early hominins (e.g. Bonnefille, 2010; Cerling et al., 2011; Dominguez-Rodrigo, 2014). Among the factors that may have contributed to the increase of C_4 grasses in African landscapes, there is the decrease in the length but the intensification of the rainy season, or so called seasonality (Beerling and Osborne, 2006), and the subsequent enhancement of the fire activity (Archibald et al., 2009). Hydrological changes and the decrease of pCO_2 recorded at the end of the Miocene (Pagani et al., 1999) may also have been significant in the spread of C_4 plants in the African vegetation (Tippie and Pagani, 2010). As the hominin record is remarkably well documented in the East African rift system, much attention was given to this region for reconstructing late Neogene paleoenvironmental changes and especially C_4 -grass expansion in Africa (Bonnefille, 2010). Indeed, about ten early hominin species from this region were described between the Messinian and the early Pliocene, including three of the first hominins: *Orrorin tugenensis* (Kenya, ca 6 Ma, Senut et al., 2001), *Ardipithecus kaddaba*, and *Ardipithecus ramidus* (Ethiopia, ca 5.5–4.4 Ma, White et al., 1994; Haile-Selassie, 2001), and most of Australopithecines (Brown et al., 2013). The volcanic context of the East African rift system offers the advantage of providing datable ash layers useful for intersite paleoenvironmental comparisons. Biomarker data from the marine record DSDP231 off the East African coast indicate two major and distinct steps in the expansion of C_4 biomass in East Africa: one during the Tortonian (11–9 Ma) and one during the Pliocene (4.3–1.4 Ma) (Feakins et al., 2013). Yet the increase in C_4 biomass was not necessarily linked with an expansion of C_4 grasses but rather with a greater abundance of Amaranthaceae and Chenopodiaceae (xerophytic C_4 forbs) in the vegetation as precised by pollen data from the same record (Bonnefille, 2010).

In Chad (Central Africa), the hominin record is less documented than in the East African rift system but nevertheless significant. It includes *Sahelanthropus tchadensis* and *Australopithecus bahrelghazali*, which are the only two hominin species described in tropical Africa outside the East African rift valley (Brunet et al., 2002; Vignaud et al., 2002; Brunet et al., 2005). *S. tchadensis* is what's more the oldest hominin species described so far (Brunet et al., 2002). Fossil faunal assemblages associated with the Chadian hominins bear ages around 7 Ma for *S. tchadensis* and of 3.5 Ma for *A. bahrelghazali*, respectively (Brunet et al., 1995; Vignaud et al., 2002). The authigenic $^{10}Be/^{9}Be$ dating method applied to the fossiliferous areas of northern Chad produced after absolute ages that were consistent with the ages provided by faunas (Lebatard et al., 2008; Lebatard et al., 2010).

Paleoenvironmental studies at Chadian hominin sites (faunal composition and structure, dental mesowear and isotopes) concluded to the existence of a diversified vegetation cover, with a mosaic of forests, woodlands, grasslands, up to desert conditions, in close relationship with aquatic/lacustrine areas in northern Chad during the Messinian (Vignaud et al., 2002; Schuster et al., 2006; Jacques, 2007; LeFur et al., 2009; Schuster et al., 2009; Blondel et al., 2010; Otero et al., 2010), and open C_4 -vegetations later on during the Pliocene (Brunet et al., 1997; Zazzo et al., 2000; Geraads et al., 2001; Lee-Thorp et al., 2012), in association with large lake occurrences (Schuster et al., 2009). Paleovegetation modeling for the Messinian does not support the existence of mosaic vegetation in northern Chad, but rather a spread of the tropical savanna biome (Pound et al., 2012). For the same time period for comparison, it suggests the simultaneous presence of tropical forests, savannas, and grasslands in eastern Africa (Pound et al., 2012). The apparent inconsistency between paleontological data and simulated biomes in northern Chad during the Messinian could be however reconciled by considering that the recurring of large lake episodes must have triggered and sustained, at a local scale, dense and diverse riparian vegetation. Besides, the vegetation signal represented by fossil faunas from northern Chad may better reflect local environmental features related to lake littoral coasts and associated marsh/fluvial areas than regional biome. For the Pliocene period, vegetation modeling indicates that tropical savanna and/or xerophytic shrubland biomes were present in

northern Chad (Salzmann et al., 2008; Contoux et al., 2013), which is consistent with the occurrence of C_4 -dominated vegetation inferred from mammalian isotopic data (Zazzo et al., 2000; Lee-Thorp et al., 2012). Coupled climate-vegetation simulations also suggested that lacustrine episodes, even large, had little effect on biome distribution. During the mid-Pliocene, savanna and/or shrubland environments are therefore thought to result from regional climatic conditions rather than from lacustrine expansion (Contoux et al., 2013).

Paleobotanical data from the Lake Chad basin are restricted to the vertebrate localities of northern Chad. They consist of sparse silicified plant macro-remains (Coppens and Koeniguer, 1976) and abundant phytolith remains (Novello, 2012) preserved in paleosols, which are also full of fossilized root-systems and termite nests (Schuster et al., 2000; Durringer et al., 2006, 2007). The identification of this termite nest suggests the presence of dry grasslands or more probably wooded savannas in northern Chad during the Miocene-Pliocene (Durringer et al., 2006).

Previous paleoenvironmental investigations mostly focused on the northern part of the Lake Chad basin. More data from the southern part of the basin are therefore needed to provide direct evidences for the presence (or not) of the savanna biome inferred by modelling at the scale of the entire basin, at time crucial for hominin emergence. In order to better document paleovegetation and paleohydrological changes that occurred in this part of the basin, we undertook the analysis of an old archive obtained in 1973 from the drilling of a borehole, at the locality of Bol (13°N). The drilling was carried out by the BRGM ("Bureau de Recherches Géologiques et Minières", France) during a campaign engaged for groundwater prospection in the Lake Chad basin. The samples (cuttings), since then kept by one of the co-authors (Dr J. Maley), are now curated by l'Ecole et Observatoire des Sciences de la Terre (EOST, University of Strasbourg). We have first established the chronological background of this sedimentary archive by the determination of 25 new absolute ages using the authigenic $^{10}Be/^{9}Be$ dating method. We have then analyzed several micro-biological remains (phytoliths, pollen, and diatoms), which cover a time interval from about the Messinian (late Miocene) through the Piacenzian (late Pliocene). The results presented here are unique for both the time period and the place investigated.

2. Study area

The Lake Chad basin is a vast intracratonic basin of 2 500 000 km², which expands on Chad, Niger, Nigeria, and Cameroon. It is bordered by the Tibesti uplift in the north (>3 000 m), the Ennedi and the Ouaddaï Plateaus in the east, the Adamaoua and Jos Plateaus in the south, and the Aïr and the Hoggar in the west. The basin consists of Cretaceous to Neogene deposits resting on a Paleozoic sedimentary cover and Proterozoic basement (Genik, 1992; Kusnir and Moutaye, 1997). The Neogene continental deposits mainly outcrop in the northern part of the Chadian basin (Schneider, 1989), in the Djurab desert, and consist in sequences of eolian sands and perialacustrine clayish sandstones interbedded with lacustrine mudstones and diatomites, the whole testifying of recurring large lake expansions in the region since at least 7 Ma (Vignaud et al., 2002; Schuster et al., 2006; Schuster et al., 2009). Four main Neogene fossiliferous areas are described in the Djurab according to their respective ages, sedimentary features, and faunal assemblages: TM (7.3 ± 0.1 Ma), which yielded *Sahelanthropus tchadensis* fossil remains, KB (5.4 ± 0.6 Ma), KL (4.0 ± 0.1 Ma), and KT (3.6 ± 0.1 Ma, which yielded *Australopithecus bahrelghazali* fossil remains (Brunet et al., 1995, 1996, 1998; Brunet and M.P.F.T., 2000; Brunet et al., 2002; Vignaud et al., 2002; Brunet et al., 2005; Lebatard et al., 2010).

Present-day Lake Chad is supplied at 95% by the Chari-Logone river system, which drains a large basin of 620 000 km² in the humid tropics from Central African Republic and North Cameroon, and to a lesser extent by direct rainfall and inflow from other small tributaries (El Beid and Komadugu Yobe rivers) (Bader et al., 2011; Leblanc et al., 2011). It is set in the Sahelian domain where mean annual rainfall is

<400 mm/year, and dry season lasts from 9 to 11 months (Griffiths, 1972; New et al., 2002, CRU 10'x10' database). Average annual temperature is around 27–29 °C (New et al., 2002).

The borehole of Bol (13°28'N, 14°44'E) was drilled in the southern part of the Lake Chad basin, at the edge of the Saharan desert, about 400 km to the south of the Miocene–Pliocene hominin localities (Fig. 1). Bol is currently an archipelago surrounded by a large complex of small episodic islands of eolian sand, probably deposited in the region during the last glacial maximum (between 20 and 12 ka BP) (Servant, 1973). Present-day vegetation of Bol archipelago is essentially aquatic, with abundant herbaceous plants and few trees and shrubs (Gaston and Dulieu, 1976; Gaston, 1996; Olivry et al., 1996; César and Lebrun, 2003). Three vegetation types can be distinguished according to their distribution: (1) Aquatic areas are mostly occupied by hydrophytic Poaceae (grasses) (*Vossia cuspidata*, *Phragmites australis*, *Echinochloa* sp.), Cyperaceae (sedges) (*Cyperus papyrus*, *Cyperus articulatus*) and Typhaceae (*Typha domingensis*), which often constitute patches of vegetation drifting on water. The hydrophytic species *Aeschynomene elaphroxylon*, growing close to water courses and lake banks, may constitute dense vegetated areas when lake level is low (Gaston and Dulieu, 1976); (2) In areas that are regularly flooded and unflooded, the vegetation is mainly composed by marsh herbaceous taxa, such as

Cyperaceae, *Potamogeton*, and hydro-/helophytic Poaceae (e.g. *Leersia hexandra*); (3) In contrast, drought-adapted plants, such as the shrub *Cassia sieberdana*, and xerophytic Poaceae such as *Cenchrus biflorus*, *Dactyloctenium aegyptium*, and *Panicum laetum* occur in the dry and sandy areas on the islands. On the highest islands, sub-desertic trees of *Acacia* spp., *Balanites aegyptiaca*, *Calotropis procera*, *Leptadenia pyrotechnica*, and Arecaceae (palm) trees of *Hyphaene thebaica* are also present.

3. Material

The sedimentary archive consists of a discontinuous succession of samples that could be originally associated with three major lithological units: alternating sand and clay at the bottom (673–297.2 m), intermediate lacustrine pelite and diatomite (297.2–71.5 m), and eolian sand at the top (71.5–0 m) (Fig. 2a) (Moussa, 2010). However, only 114 samples from the intermediate section between 297.2 m and 71.5 m were preserved. Of these, 25 samples (sampled between 297.2 m and 90.8 m) were studied as a test. These 25 samples were analyzed for $^{10}\text{Be}/^9\text{Be}$ dating, phytolith, pollen, and diatom remains. A supplementary sample, BOL 74 (74 m), was considered for $^{10}\text{Be}/^9\text{Be}$ dating only.

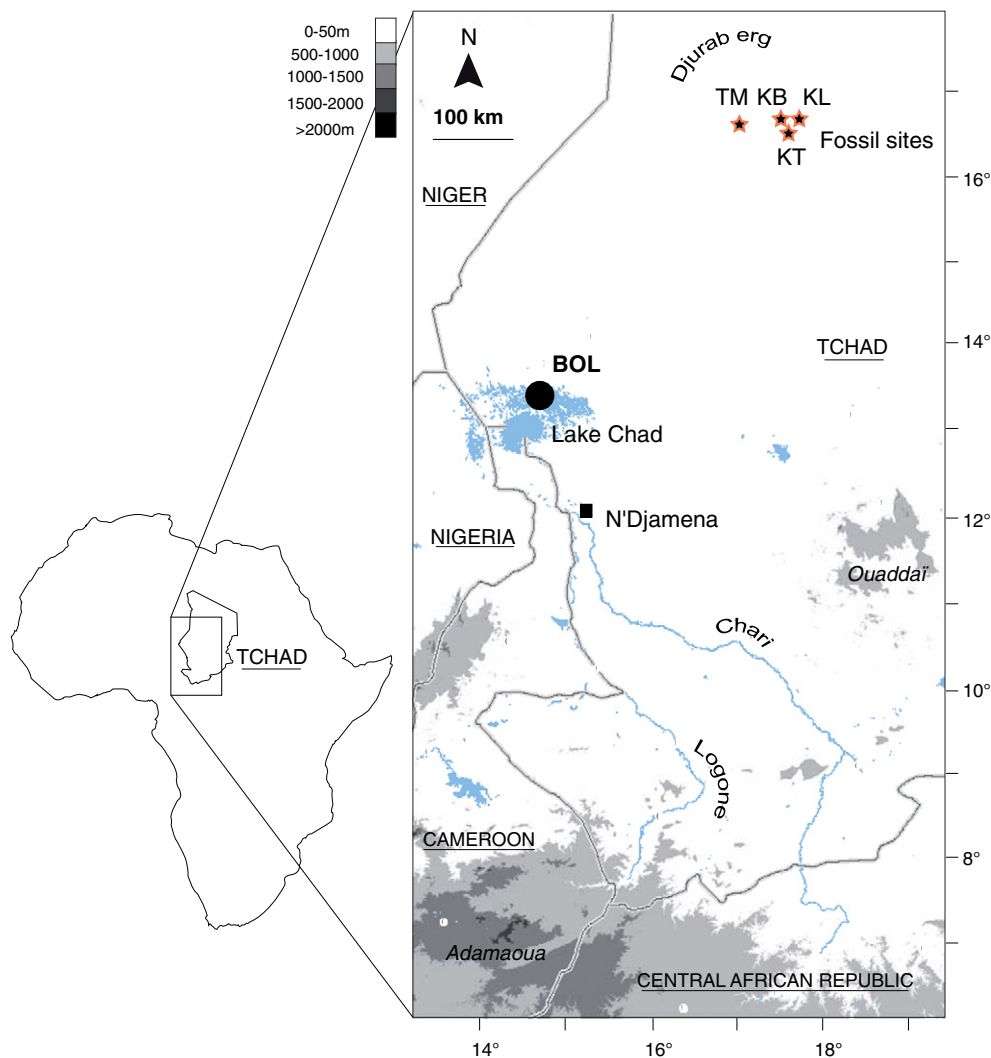


Fig. 1. Map locating the Bol borehole in the Lake Chad archipelago, about 400 km to the south of the Miocene–Pliocene fossil sites of the Djurab desert (stars). TM: 7.3 ± 0.1 Ma (type-locality of *Sahelanthropus tchadensis*), KB: 5.4 ± 0.6 Ma, KL: 4.0 ± 0.1 Ma, and KT: 3.6 ± 0.1 Ma (type-locality of *Australopithecus bahrelghazali*) (Lebatard et al., 2010).

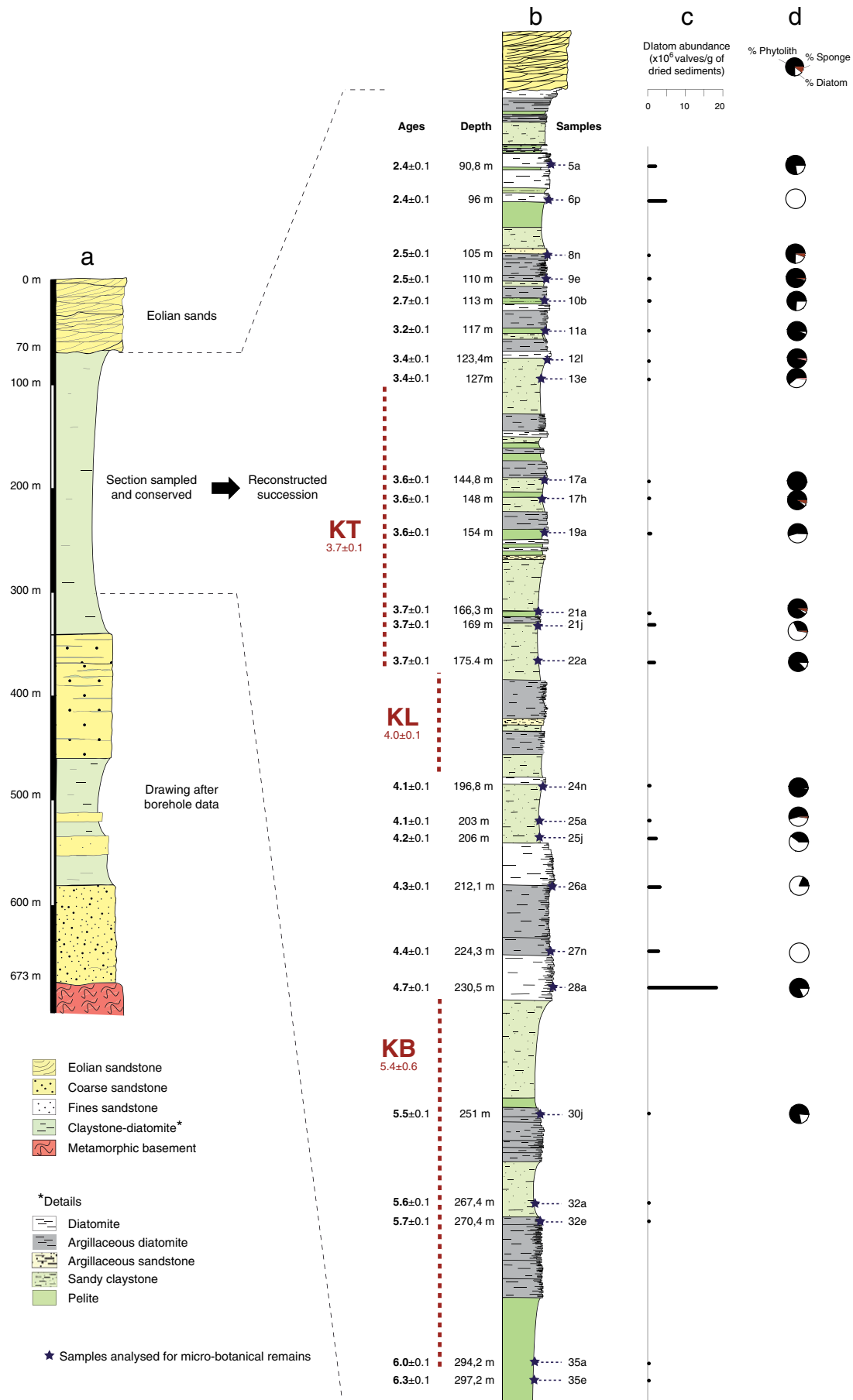


Table 1

Authigenic Be concentrations, $^{10}\text{Be}/^9\text{Be}$ ratios, and ages (initial ages and age model) for the 25 Bol samples. Initial ages re-estimated from age model are boxed. [§]Samples specifically measured at the AMS Tandemtron facility (Gif/Yvette, France). The original ages that were re-estimated from the age model are framed in bold.

Samples	Depth (m)	^{10}Be ($\cdot 10^7$ at.g $^{-1}$)	^9Be ($\cdot 10^{15}$ at.g $^{-1}$)	$^{10}\text{Be}/^9\text{Be}$ ($\cdot 10^{-9}$)	Age (Ma)	Age model (Ma)
BOL 74§	74.0	13.044 ± 0.449	5.163 ± 0.061	25.263 ± 0.919	–	–
BOL 5a	90.8	3.399 ± 0.065	4.911 ± 0.030	6.920 ± 0.139	2.61 ± 0.08	2.38 ± 0.07
BOL 6p	96.0	7.660 ± 0.073	10.028 ± 0.042	7.639 ± 0.080	2.41 ± 0.07	2.41 ± 0.07
BOL 8n	105.0	4.136 ± 0.064	5.174 ± 0.053	7.994 ± 0.149	2.32 ± 0.08	2.45 ± 0.08
BOL 9e	110.0	5.298 ± 0.075	7.192 ± 0.053	7.366 ± 0.117	2.48 ± 0.08	2.48 ± 0.08
BOL 10b	113.0	6.449 ± 0.082	9.916 ± 0.115	6.504 ± 0.112	2.73 ± 0.08	2.73 ± 0.08
BOL 11a	117.0	6.177 ± 0.093	12.034 ± 0.089	5.134 ± 0.086	3.20 ± 0.08	3.20 ± 0.08
BOL 12l	123.4	6.221 ± 0.067	11.375 ± 0.089	5.469 ± 0.073	3.08 ± 0.08	3.35 ± 0.08
BOL 13e	127.0	2.851 ± 0.050	6.219 ± 0.021	4.584 ± 0.082	3.43 ± 0.08	3.43 ± 0.08
BOL 17a	144.8	3.345 ± 0.052	8.006 ± 0.066	4.178 ± 0.074	3.62 ± 0.08	3.62 ± 0.08
BOL 17h§	147.0	2.001 ± 0.098	6.202 ± 0.066	3.226 ± 0.161	4.13 ± 0.13	3.62 ± 0.08
BOL 19a	154.0	2.758 ± 0.049	6.678 ± 0.059	4.129 ± 0.082	3.64 ± 0.09	3.64 ± 0.09
BOL 21a	166.3	3.277 ± 0.057	8.036 ± 0.052	4.078 ± 0.076	3.66 ± 0.08	3.66 ± 0.08
BOL 21j§	169.3	2.481 ± 0.119	7.013 ± 0.021	3.538 ± 0.170	3.95 ± 0.12	3.69 ± 0.09
BOL 22a	175.5	2.494 ± 0.053	6.359 ± 0.036	3.922 ± 0.086	3.74 ± 0.09	3.74 ± 0.09
BOL 24n	197.0	–	–	–	–	4.09 ± 0.09
BOL 25a§	203.0	1.595 ± 0.082	6.362 ± 0.035	2.507 ± 0.129	4.64 ± 0.13	4.19 ± 0.09
BOL 25j	206.0	2.391 ± 0.049	7.828 ± 0.087	3.054 ± 0.071	4.24 ± 0.09	4.24 ± 0.09
BOL 26a	212.1	2.193 ± 0.068	6.479 ± 0.052	3.385 ± 0.109	4.04 ± 0.10	4.31 ± 0.09
BOL 27n§	224.3	1.829 ± 0.087	8.943 ± 0.042	2.045 ± 0.097	5.04 ± 0.13	4.44 ± 0.10
BOL 28a	230.5	1.531 ± 0.045	6.175 ± 0.079	2.479 ± 0.080	4.66 ± 0.10	4.66 ± 0.10
BOL 30j	251.0	1.432 ± 0.040	8.866 ± 0.026	1.615 ± 0.045	5.52 ± 0.10	5.52 ± 0.10
BOL 32a§	269.0	1.803 ± 0.095	17.644 ± 0.114	1.022 ± 0.054	6.43 ± 0.14	5.63 ± 0.10
BOL 32e	270.4	2.172 ± 0.047	14.430 ± 0.027	1.506 ± 0.033	5.66 ± 0.10	5.66 ± 0.10
BOL 35a	294.2	0.664 ± 0.019	5.218 ± 0.028	1.273 ± 0.037	5.99 ± 0.10	5.99 ± 0.10
BOL 35e	297.2	1.379 ± 0.044	12.946 ± 0.030	1.065 ± 0.034	6.35 ± 0.11	6.35 ± 0.11

4. Methods

4.1. Dating

It has been demonstrated (Lebatard et al., 2008) that the environmental context prevailing since at least 8 Ma in the Chad basin is favorable for the use of the dating method based on the authigenic $^{10}\text{Be}/^9\text{Be}$ ratio (authigenic ^9Be isotope being used as a normalizing factor to overcome environmental effects, see Bourlès et al., 1989). This method offers the opportunity to date this 230 m long sedimentary record.

The dating method applied here exactly follows the methodology used in Lebatard et al. (2008, 2010) for the Chadian hominin sites dating. Authigenic beryllium isotopes were selectively extracted from dried and crushed sediments using a 0.04 M $\text{NH}_2\text{OH}-\text{HCl}$ in a 25% acetic acid leaching solution (Bourlès et al., 1989). A 2 ml aliquot is removed from the leachate for the ^9Be measurements. For the ^{10}Be measurements, the remaining leachate was spiked with 300 μl of a 10^{-3} g.g $^{-1}$ ^9Be solution (Merck 1000 mg/l Be standard), purified by solvent extractions of Be acetylacetonate in presence of EDTA followed by precipitations of $\text{Be}(\text{OH})_2$ at pH 8.5 and rinsing. The final precipitate, dissolved in 100 μl of HNO_3 , was dried and heated at 800 °C to obtain BeO . This chemical process and the measuring protocols were described in full details by Lebatard and coll. (Lebatard et al., 2010 and references therein).

The study was conducted in two stages. The first six samples (BOL 74, 17 h, 21j, 25a, 27n, and 32a; Table 1) ^{10}Be concentrations were measured at the now closed AMS Tandemtron facility (Gif/Yvette, France), the ^9Be concentrations being measured on an Hitachi Z-8200 Atomic Absorption Spectrophotometer using Zeeman Effect background correction at CEREGE (Aix-en-Provence, France) as fully described in Ménabréaz et al. (2011). The 20 additional samples (Table 1) ^{10}Be concentrations were measured at the French AMS national facility ASTER installed

since 2006 at CEREGE, the ^9Be concentrations being measured on a Thermo Scientific ICE 340 Atomic Absorption Spectrophotometer using Zeeman Effect background correction at CEREGE (Aix-en-Provence, France) as fully described in Lebatard et al. (2008, 2010). The ^{10}Be concentrations are normalized to $^{10}\text{Be}/^9\text{Be}$ SRM 4325 NIST reference material with an assigned value of $(2.79 \pm 0.03) \cdot 10^{-11}$ (Nishiizumi et al., 2007). This standardization is equivalent to 07KNSTD within rounding uncertainty. The ^{10}Be half-life of $(1.39 \pm 0.01) \cdot 10^6$ years (Chmeleff et al., 2010; Korschinek et al., 2010) was used for the age calculations, and the age uncertainties (1σ) result from the propagation of the ^{10}Be (linked to the number of ^{10}Be events detected coupled to a 3% analytical uncertainty deduced from the reproducibility through the measurement sequences), ^9Be and half-life uncertainties ultimately. ^{10}Be and ^9Be concentrations, the $^{10}\text{Be}/^9\text{Be}$ ratio and $^{10}\text{Be}/^9\text{Be}$ age of each individual sample are reported in Table 1.

4.2. Diatom, phytolith and pollen analyses

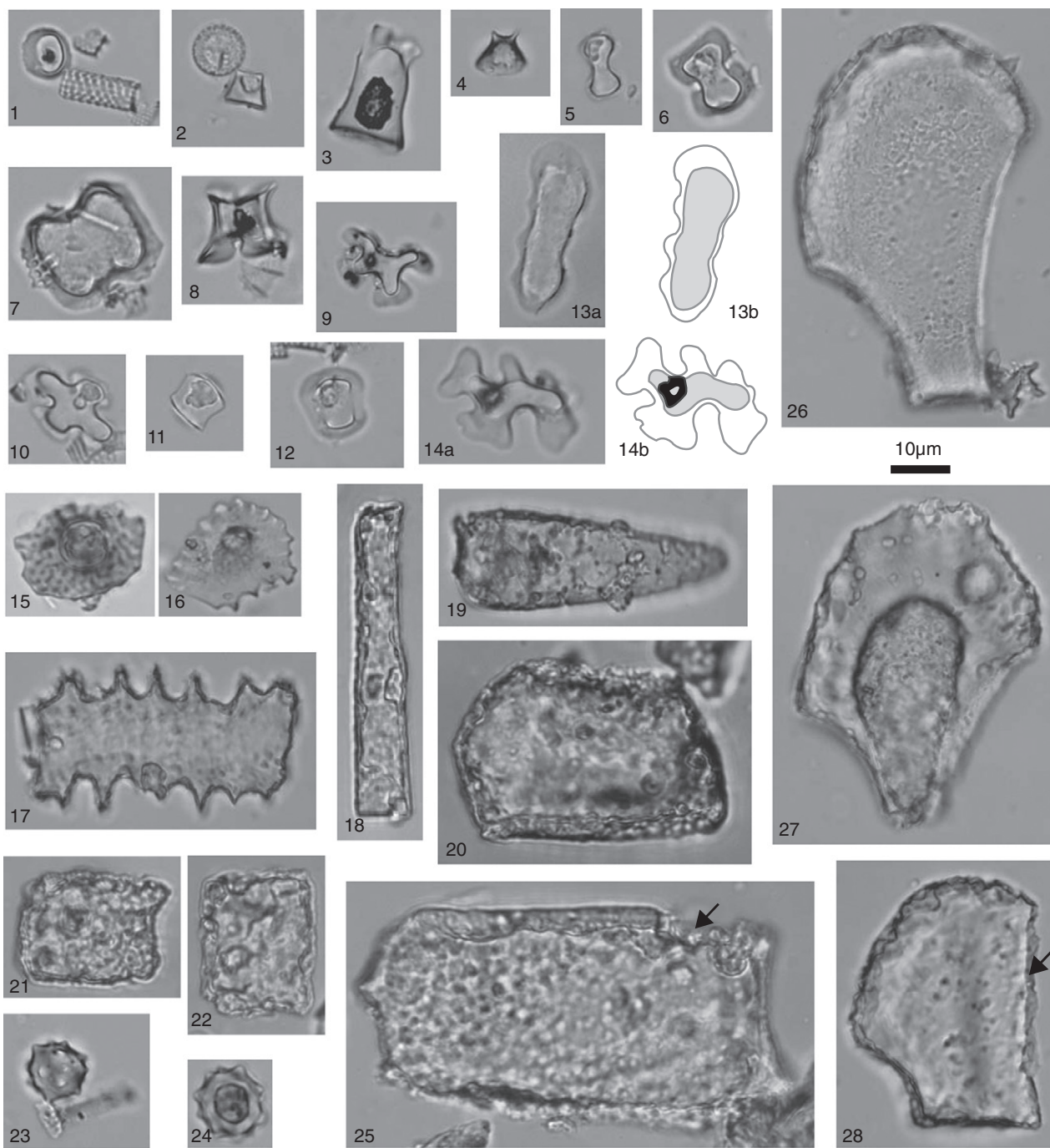
4.2.1. Laboratory procedures

Diatom analyses were conducted on 0–5 g samples and treated through standard procedures (1:1 mixture of $\text{H}_2\text{O}_2/\text{water}$, 1:1 mixture of HCl/water , and repeatedly rinsed in distilled water; slides were made using Naphrax high resolution mounting) (Battarbee et al., 2001). For each sample, at least 400 diatom valves were identified and counted using a Nikon Eclipse 80i light microscope (differential interference contrast optics, $\times 1000$ magnification, N.A. = 1.25). Specimens were identified to their lowest taxonomic level (i.e. variety) following the species concept used by Krammer and Lange-Bertalot (1986, 1988, 1991). However, many genera and species outlined have been re-assessed using the classification by Round et al. (1990) and synonyms are introduced. We quantified the number of diatom valves for

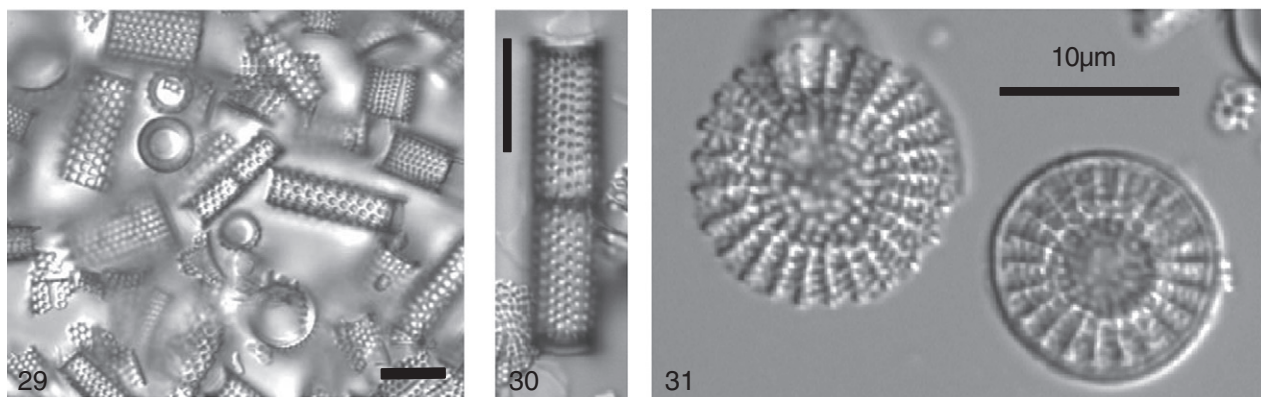
Fig. 2. Sedimentology of Bol borehole and stratigraphical positioning of the 25 studied samples (a. general synthesis; b. detailed sequence reconstructed from preserved cuttings), c. diatom concentration in the samples (number of valves/g of dry sediments), and d. quantification of phytoliths, diatoms, and sponge spicules recovered in the silicified assemblages. The time range of the Miocene-Pliocene fossil sites from the Djurab is given in parallel to the chronology of Bol borehole.

Table 2 (continued)

[illegible]



➤ Examples of prints interpreted as indices of dissolution or fragmentation



each sample using an inverted microscope Olympus CKX41 following the procedure described by Fröhlich and Servant-Vildary (1989) (Table 2, Fig. 2c). The concentration of diatoms was calculated using the dry weight of the sample and its dilution as well as the amount of dilution from which the diatom valve are numerated.

Phytolith analyses were conducted on 15 g samples. Chemical treatment included carbonate dissolution with hydrochloric acid (HCl 37%), oxidation of organic matter with hydrogen peroxide (H₂O₂ 33%, at 90 °C), iron removal (with C₆H₅Na₃O₇ and Na₂O₄S₂, H₂O₂), clay removal by decantation, and densimetric separation with zinc bromide heavy liquid (ZnBr₂) set at $d = 2.3$. The recovered silicified particles include phytoliths, diatoms, and sponge spicules. Glycerin was used as mounting medium to allow three-dimensional observations (WoldeGabriel et al., 2009). Silicified particles were counted at $\times 400$ and $\times 1000$ magnification. A minimum of 200 grass silica short cell phytoliths (GSSCs) was counted in each sediment sample, except for samples 30j, 26a, 21a, and 9e for which the extracted material was exhausted before reaching the count. In these cases, only 143 to 192 GSSCs could be counted (Table S1). We estimated the relative abundances (%) of phytoliths, diatoms, and sponge spicules in each sample through counting of the silicified particles under the microscope (Table 2, Fig. 2d). Phytoliths were classified following the classification criteria established by Novello et al. (2012) based on modern plants and soils from Chad. Photographs of phytoliths (Fig. 3) were realized using Canada balsam as mounting medium because its refraction index (1.55) gives a better contrast than glycerin.

Pollen extraction was done using 5 g of material and following the standard method of acid/base treatments. Carbonate-rich samples were treated with HCl (37%) before silicate removal using HF (40%, overnight), and humic acid removal using KOH (20%). Slides were mounted with glycerin. Pollen identifications were performed at $\times 1000$ magnification. Only one (sample 25j) out of 25 samples contained enough pollen material for carrying out a suitable counting (total pollen sum of 803 grains, Table S2).

4.2.2. Modern phytolith reference data and analytical approaches

A modern soil/mud phytolith dataset of 57 surface samples from Chad (between 8°55'–13°56'N and 15°33'–20°06'E) was used as a reference for interpreting the phytolith signal of fossil sediments from Bol. The modern dataset includes samples from the marshes of current Lake Chad (10) and from various Sahelian (18) and Sudanian (29) environments and savanna types (Novello, 2012; Novello et al., 2012). Sixteen phytolith soil samples from the Guineo-Congolian forests (Runge, 1999) were also included in our modern referential where grass-dominated environments are obviously over-represented.

The D/P° index (Bremond et al., 2008), which is the ratio of tree and shrub phytoliths (i.e. globular granulate types) over grass (Poaceae) phytoliths (grass silica short cells exclusively) was used as a proxy for the openness of the environment (Fig. 4). Globular echinate phytoliths (not included in the D/P° index) were used to infer the presence of the Arecaceae family, i.e. palms (Barboni et al., 2010; Albert et al., in press). The relative abundance of blocky and elongate phytoliths was used to trace lacustrine marshes, following our observations of modern surface soil samples (Fig. 4). Indeed, although blocky and elongate phytoliths may have various origins (trees, shrubs or herbaceous taxa)

(e.g. Strömberg, 2003; Mercader et al., 2009), we observed that they are more abundant in the surface samples of current Lake Chad marshes and of Sahelian temporary swamps than in any other surface soil samples of the sub-Saharan region (Novello et al., 2012). Among the blocky types, we gave particular attention to the cuneiform (fan-shaped) phytoliths. These are produced in the bulliform cells, of grasses and sedges (Novello et al., 2012). Heavy silicification of bulliform cells occurs when these plants undergo high evapotranspiration rate while their root system is submerged (Sangster and Parry, 1969). Indeed, silicified bulliform cells happen to be particularly abundant in mud samples from current sub-Saharan marshes (Fig. 4) (Novello, 2012). We used the Aquatic and Xerophytic grass phytolith indices (hereafter Iaq and Ixe) (Novello et al., 2012) now calibrated on an extended modern plant dataset of 98 grass species (Novello, 2012) as proxies for the environmental affinity of grass communities (Fig. 4). Based on the new calibration, Iaq and Ixe phytolith indices are defined as follows,

$$I_{aq} = \frac{Ro2 + Ro6 + Bi7 + Bi11 + Bi12 + Cr3 + Cr5 + Cr6 + Cr7 + Bi14 + Poly1}{N_{totGSSC}} \times 100$$

$$I_{xe} = \frac{Ro1 + Ro5 + Bi2 + Bi3 + Bi4 + Bi15 + S1 + S2 + S3}{N_{totGSSC}} \times 100.$$

where $N_{totGSSC}$ is the total of GSSC types counted for each sample, Ro: Rondel, Bi: Bilobate, Cr: Cross, S: Saddle (see Novello, 2012 for morphological details of the phytolith types used in the indices).

When applied to modern soil/mud surface assemblages, the newly calibrated grass phytolith indices provide a better identification of herbaceous wetland marshes, and of grasslands of the Sahelian and Sudanian domains than the indices calibrated previously on a smaller grass dataset (Novello et al., 2012) (Fig. S1). The Ixe index was here preferred to the Iph index (Bremond et al., 2005b) because it intends to exclude the phytolith signal of local aquatic grasses. The abundance of aquatic grasses on the lake margins may indeed conceal the phytolith signal of extra-local (i.e. non lacustrine) grass communities more representative of the regional vegetation.

Decision trees (rpart function, R.13.0) based on modern phytolith assemblages were used to produce threshold values of the Iaq and of the Ixe indices, below and above which a fossil sample can be related to the Sudanian, Sahelian, or Lake Chad (aquatic) grass signal. Above an Iaq index value of 34%, most phytolith assemblages are related to Lake Chad marshes (8/9 assemblages, 89% of accurate classifications). Below this threshold value of 34% for Iaq, phytolith assemblages are unlikely to originate from aquatic environments but rather from surrounding grass communities of the Sahelian or Sudanian domains (Figs. 4, 7). A threshold value of 46% for the Ixe index allows discriminating Sudanian phytolith assemblages from Sahelian phytolith assemblages (Figs. 4, 7). As a result, most modern Sahelian phytolith assemblages (15/18 assemblages, 83% of accurate classifications) have Ixe index values $\geq 46\%$. We are aware that the robustness of these thresholds and indices is criticized, as it depends on the number of phytoliths taken in account for calculating the grass phytolith indices Iaq and Ixe (here ranging from 33 to 114 phytoliths) (Strömberg, 2009). These thresholds were therefore used as a support to reflect only the main tendencies recorded in the fossil phytolith assemblages.

Fig. 3. Selected phytolith types and diatom species observed in the fossil sediments of Bol. Phytolith IDs follow the classification of Novello et al. (2012). 1–2. Rondel, conical, top truncated: Ro1 (1: in top view, 2: in side view) (near diatoms); 3. Rondel, conical, top truncated, tall ($h > 15 \mu m$); Ro7; 4. Rondel, conical, top keeled: Ro2; 5. Bilobate, short and tabular, round lobes: Bi2; 6. Bilobate, short and trapeziform, length > height: Bi11; 7. Bilobate, short and tabular, missing shank: Bi9; 8. Bilobate, short and trapeziform, length = height: Bi12; 9. Bilobate, trapeziform, concave lobes: Bi14; 10. Cross, trapeziform, 4-lobed, cross top: Cr5; 11. Saddle, tabular long: S3; 12. Saddle, trapeziform, base round and vaguely constricted: S4; 13a–b. Polylobate, trapeziform, base regularly sinuate: Poly4*; 14a–b. Polylobate, trapeziform, base irregularly sinuate with closed concavities/convexities: Poly3*; 15. Plate, regularly scrobiculate with a short depressed cone in the center: Pla4; 16. Plate, sinuous shape, knobby apex: Pla5; 17. Elongate, tabular, echinate margins: El3; 18. Elongate, tabular, smooth margins: El3; 19. Acicular, psilate, pitting of dissolution/alteration: Ac1; 20–22, 25. Blocky, cubic to parallelepipedal, prints of dissolution/alteration on surface: Blo3, –6 or –7; 23–24. Globular, echinate: Glo1; 26–28. Blocky, cuneiform with prints of dissolution/alteration on surface (28: broken cuneiform): Blo4; 29. Diatomite dominated by *Aulacoseira granulata*, sample 10b; 30. *Aulacoseira granulata*; 31. *Stephanodiscus carconensis* (var. *pusilla*). *New types.

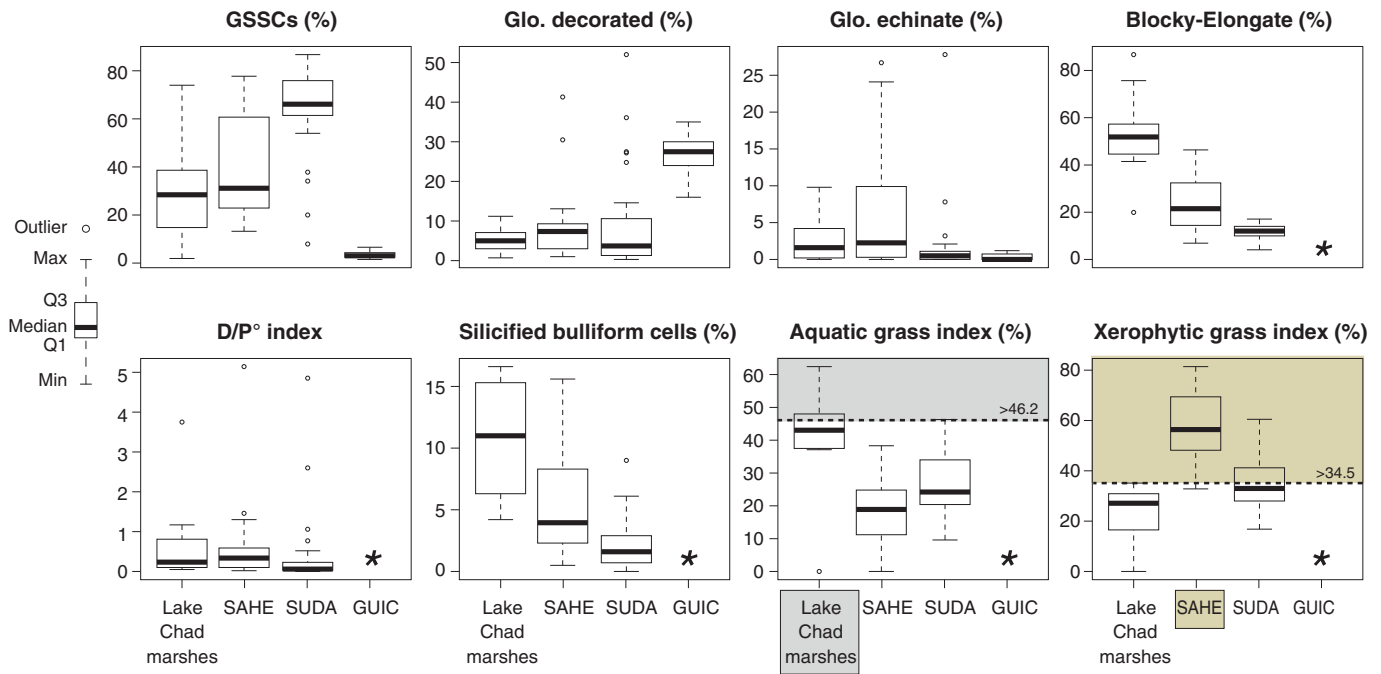


Fig. 4. Box-plots showing the abundance patterns of the main phytolith indicators and indices considered in this study according to modern sub-Saharan environments: Lake Chad marshes, Sahelian, and Sudanian vegetation, and Guineo–Congolian vegetations (from Runge, 1999). Gray areas indicate the range of Aquatic (Iaq) and Xerophytic (Ixe) grass indices characterizing a phytolith signal from present-day Lake Chad marshy grasses and from Sahelian grasses.

4.2.3. Modern pollen reference data and analytical approaches

The biomization method was applied to the pollen sample 25j (Jolly et al., 1998). Biome scores (%) were calculated for the fossil pollen data of Bol, and compared to scores calculated from modern pollen data of Lake Chad (Maley, 1972, 1981; Amaral et al., 2013). Score values were obtained from pollen abundance data (square-root transformed to lower the weight of over-represented taxa) and from known taxa ecological and phytogeographical affinities (Fig. 8). Assignment of pollen taxa to pant functional types (PFTs) and of PFTs to biomes follow Lézine et al. (2009) (Appendix 2).

5. Results

5.1. $^{10}\text{Be}/^9\text{Be}$ ages

Twenty-five authigenic $^{10}\text{Be}/^9\text{Be}$ ratios were measured. Using the initial ratio N_0 of $(2.54 \pm 0.09) \cdot 10^{-8}$ determined from Holocene diatomites of Megalake Chad (~7 to 4.4 ka BP) (Schuster et al., 2005; Lebatard et al., 2008) and the ^{10}Be decay period of $(1.387 \pm 0.012) \cdot 10^6$ years (Chmieleff et al., 2010; Korschinek et al., 2010), the resulting authigenic $^{10}\text{Be}/^9\text{Be}$ ages range from 6.35 ± 0.11 Ma at the bottom (at 297.2 m depth) to 2.61 ± 0.08 Ma for the top sample (at 90.8 m depth). The Bol sequence therefore represents a time interval from the Messinian (late Miocene) to the Piacenzian (late Pliocene) (Table 1). A sedimentation rate of about $53.0 \text{ m} \cdot \text{Ma}^{-1}$ was estimated from linear regression performed with initial age data (Fig. 5). The upper sample BOL 74 (Table 1) corresponding to 74 m of depth yields an authigenic $^{10}\text{Be}/^9\text{Be}$ ratio similar to the initial ratio N_0 of $(2.54 \pm 0.09) \cdot 10^{-8}$ (Lebatard et al., 2008). This initial ratio is used here, because it was established using specifically selected diatomites deposited during the last Holocene expansion of Megalake Chad (~7 to 4.4 ka BP). Lacustrine diatoms built their silica skeleton made from dissolved nutrients and thus are an ideal substrate for registering the dissolved isotopic ratio of $^{10}\text{Be}/^9\text{Be}$ at the time they were alive (Lebatard et al., 2008).

We obtained eight date inversions along the chronological sequence. An age-model was therefore calculated in order to represent the fossil

abundances along a time scale (Fig. 5). This model excludes the upper sample BOL 74 related to late Pleistocene times. Each modeled age value was estimated from the linear equation defined by the age values of its closest neighboring samples. This method was also used to provide an absolute age for the non-dated sample 24n (Table 1). A linear regression performed from age model values provides a new estimated sedimentation rate value of about $54.5 \text{ m} \cdot \text{Ma}^{-1}$ (Fig. 5). We consider our age model reliable because it barely changes the sedimentation rate estimated from initial data ($53.0 \text{ m} \cdot \text{Ma}^{-1}$).

5.2. Diatom, phytolith, pollen remains

5.2.1. Occurrences, preservation, and relative abundance

Of the 25 analyzed samples, four are sub-sterile in diatoms and phytoliths, five only sub-sterile in diatoms, and two sub-sterile in phytoliths (Table 2). Sub-sterile samples 35e, 35a, 32e, and 32a are part of the oldest section of the record and dated to >5.6 Ma. They include $<5 \mu\text{m}$ non-describable bio-silica fragments, as well as sparse and very fragmented diatom frustules. Other samples sub-sterile in diatoms are dated to 5.5 ± 0.1 Ma (30j) and between 3.6 ± 0.1 and 2.5 ± 0.1 Ma (17a, 13e, 11a, and 8n). Samples sub-sterile in phytoliths are dated to 4.4 ± 0.1 Ma (27n) and 2.4 ± 0.1 Ma (6p). We note that fragmentation (and dissolution?) of phytoliths and/or diatoms is common also in the samples rich in bio-silica particles (Fig. 3).

The number of diatom valves in the sediment is highly variable as concentrations range from 1.10^7 valves/g of dry sediments to $>18.10^6$ valves/g of dry sediments (Table 2, Fig. 2c). The relative abundance of phytoliths (versus diatoms and sponge spicules) ranges from 0% to 100% (Fig. 2d). Out of 21 samples, 17 are dominated by phytoliths. Particularly during the 3.6 ± 0.1 to 2.4 ± 0.1 Ma time interval, phytoliths account for $>61\%$ of the siliceous particles.

5.2.2. Diatom assemblages

The diatom-rich samples are dominated by two freshwater planktonic genera: *Aulacoseira* and *Stephanodiscus* (Table 2). *Aulacoseira granulata* (Ehrenberg) Simonsen is the dominant species, accounting for $>50\%$ in all samples, and up to 90% in samples 27n, 22a, 21a, 10b,

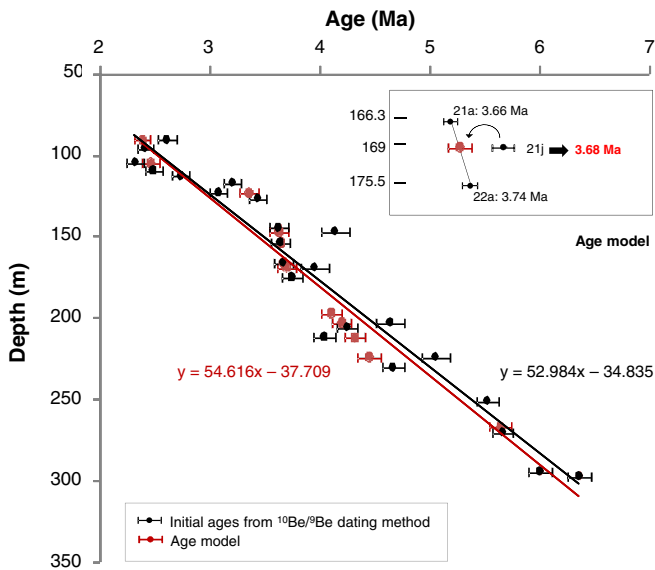


Fig. 5. Age model estimated from the $^{10}\text{Be}/^9\text{Be}$ initial age values. Calculation method for modeling ages is illustrated for sample 21j as an example.

and 6p. The occurrence of *Stephanodiscus carconensis* Grunow and its varieties is limited to the time interval between 4.3 ± 0.1 and 2.5 ± 0.1 Ma (samples 26a to 9e). *S. carconensis* accounts for up to 37% of the total diatom assemblages with two major peaks of abundance in samples 26a (4.3 ± 0.1 Ma) and 17 h (3.6 ± 0.1 Ma) (Table 2, Fig. 6).

Other diatom species account for 1% to 30% and include a variety of both planktonic (such as *Cyclotella* spp., *Coscinodiscus* spp.) and epiphytic species (such as *Navicula confervacea* (Kützinger) Grunow, *Nitzschia compressa* (Bailey) C. S. Boyer, and *Rhopalodia gibba* (her.) O. Müller) (Table S1). Epiphytic diatoms are rare: they occur in 11 out of 16 samples and represent <5% (Table 2). Sample 28a (4.7 ± 0.1 Ma), which has the highest diatom valves concentration, also exhibits the highest diatom diversity with a total of 23 species, against an average of nine species in other samples (Table 2). In this sample, epiphytic diatoms account for 4%.

5.2.3. Phytolith assemblages

We observed 40 different types of grass silica short cells (GSSCs) and 25 non-GSSC types (Table S1). Among these, two types of trapeziform polylobate phytoliths (Poly3 and Poly4) were not observed in modern soil/sediment assemblages of Chad (Novello et al., 2012). Poly3 type is characterized by an irregular and narrow multi-lobate base topped by a rectangular to keeled top that can reach up to 20 μm in height (Fig. 3, 13). Poly4, on the contrary, has a regular multi-lobate base topped by a well-defined rectangular top that does not exceed 8 μm in height (Fig. 3, 14).

GSSCs, blocky bodies and elongates are the most abundant phytolith categories present in the fossil samples. GSSCs (Poaceae) account for 18–82% of the total assemblages, while blocky and elongate types (ubiquists) account for 2–55% and 4–29%, respectively (Table 2). When present, globular phytoliths (mostly related to trees, shrubs, and palms) account for <14%, Cyperaceae papillae and Commelinaceae polyhedral phytoliths for $\leq 1\%$ of the total phytolith assemblages (Table 2). Among the GSSCs (Poaceae), the lobate phytolith types (bilobates, crosses, and tabular polylobates) are the most abundant and represent 61% to 94% of the total GSSC assemblages. Rondel and saddle phytolith types are less represented as they account for <28% and <13% of the GSSC assemblages, respectively. Polylobate trapeziform GSSCs (Poly3, Poly4) are observed in 7 out of 19 samples but represent <2% of the total GSSC assemblages (Table 2).

The D/P^o (“Dicot versus Poaceae phytoliths”) index is strictly <0.1 for all 25 samples between 5.5 ± 0.1 and 2.4 ± 0.1 Ma (Table 2, Fig. 6). Such low D/P^o values are observed for present-day grass-dominated

environments of the Sudanian and Sahelian domains. They are strikingly different from the D/P^o values of >2.5 obtained for the climatic forests of the Guineo-Congolian domain (Fig. 4). Arecaceae (palms) phytoliths are rare in all samples except 30j (5.5 ± 0.1 Ma) where they reach 13% of the total phytolith assemblage (Fig. 6). The (ubiquist) blocky and elongate phytoliths are present in remarkable amounts (29–38%) between 3.6 ± 0.1 and 3.2 ± 0.1 Ma (Fig. 6). Such abundances are observed in modern sediment samples from current marshes near Lake Chad (Fig. 4). Peaks of abundance of silicified bulliform cells are recorded at 3.6 ± 0.1 , 3.4 ± 0.1 , and 3.2 ± 0.1 Ma (38%, 17%, and 29%, respectively). Abundances of silicified bulliform cells >17% are associated, in our modern referential (Novello et al., 2012), with samples from Lake Chad marshes only (Fig. 4). The lxe index strictly <41% suggests humid-loving grass communities in the regional vegetation (Figs. 6, 7). For seven samples (28a, 25j, 24n, 21a, 13e, 11a, and 9e), the laq index signal is >38% and therefore close to that of aquatic grasses that may have been present in the local vegetation associated with the lake (Figs. 6, 7). For other samples the laq is $\leq 35\%$, and therefore is close to that of regional mesophytic grass communities (Figs. 6, 7).

5.2.4. Pollen assemblage (sample 25j)

The only productive sample for pollen analysis (25j, 4.2 ± 0.1 Ma) is characterized by abundant Poaceae (68%) and Cyperaceae (15%) elements, and the marked presence of about a total of 7% of tree taxa leaving today only in the southern reach of the Chad basin (Table S2). These southern pollen taxa belong to Guineo-Congolian elements (<1%) with *Tetrorchidium*, to Sudano-Guinean elements (3%) with *Alchornea* (3%), *Acalypha*, *Hymenocardia*, and *Syzygium guineense*, and to Afrotropical elements (4%) with *Olea capensis* (4%), *Maesa*, *Myrica*, and *Rapanea melanophloeos* (Table S2).

Biome scores calculated for sample 25j are highest for SAVA (savanna) and STEP (steppe) (Fig. 8). Moreover, they are higher for the WAMF (warm afrotropical forest), TSFO (tropical seasonal forest), and TRFO (tropical rain forest) biomes in comparison with present-day and Holocene biome scores also obtained from Lake Chad area (after Maley, 1972, 1981; Amaral et al., 2013) (Fig. 8).

6. Discussion

More than 100 samples from the Bol borehole were available to us but just the 25 samples for which we obtained $^{10}\text{Be}/^9\text{Be}$ authigenic dating were chosen for micro-botanical analyses. The resulting dataset is therefore a partial contribution that could be further improved to document paleoenvironmental changes that occurred in the Lake Chad Basin during the Miocene–Pliocene. Nevertheless, our study provides the first paleoenvironmental data set that is directly comparable with other coeval records in continental Africa because each sample has an absolute age (Salzmann et al., 2008; Bonnefille, 2010; Feakins et al., 2013) thus providing direct evidences of paleovegetation and paleohydrological conditions in Central Africa Chad basin between 6 and 2 Ma.

6.1. Diatom biostratigraphy and absolute $^{10}\text{Be}/^9\text{Be}$ ages

The Bol diatom record is consistent with the known biostratigraphy of *Stephanodiscus carconensis* Grunow in the fossil lacustrine records from Africa (Gasse, 1980) (Fig. 6). *S. carconensis* seems to have appeared during the late Miocene of Africa, diversified during the Pliocene, and finally disappeared during the early Pleistocene (Gasse and Fourtanier, 1991). As no diatoms were preserved in the Bol record before 4.7 Ma, the first appearance of *S. carconensis* recorded for the late Miocene cannot be confirmed here. Higher abundance of *S. carconensis* and higher number of varieties between 4.7 ± 0.1 and 3.6 ± 0.1 Ma support a Pliocene diversification. After 2.7 ± 0.1 Ma, *S. carconensis* only accounts for <1%, while other species as *Stephanodiscus medius* Håkansson and *Stephanodiscus astraea* (Ehrenberg) Grunow make their apparition and diversify (Table S1). This pattern may be interpreted as a sign of

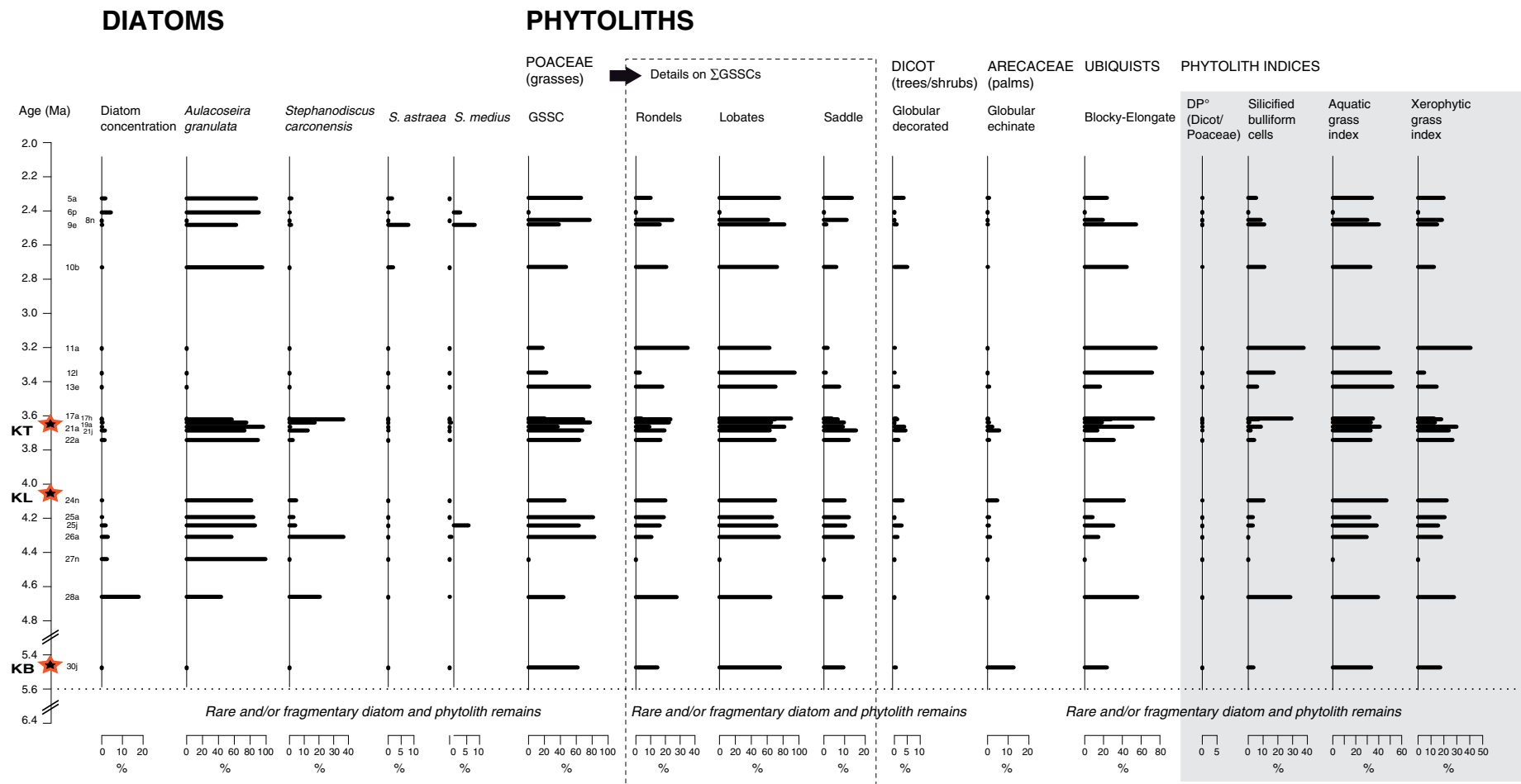


Fig. 6. Abundance diagrams of main diatom species, and phytolith types and indices (in gray) observed in the fossil assemblages of Bol. A quantitative estimation of the abundance of diatoms in the sediments is also provided. Stars indicate fossil localities of the Djurab (northern Chad) on the time scale: KB (5.4 ± 0.6 Ma), KL (4.0 ± 0.1 Ma), and KT (3.6 ± 0.1 Ma).

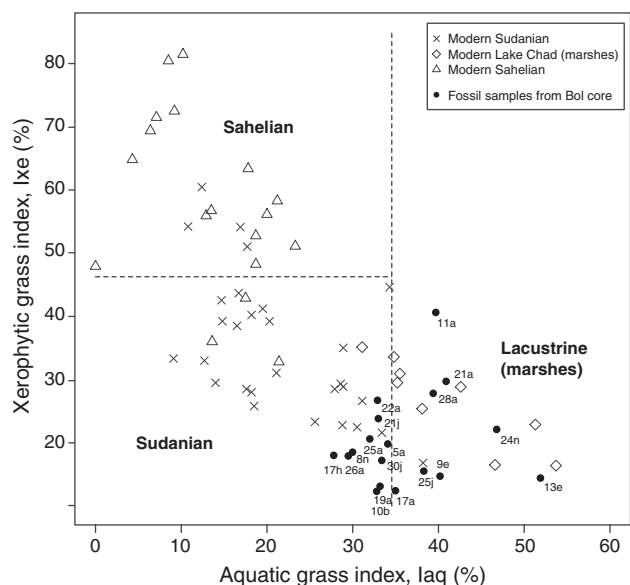


Fig. 7. Plot comparing the distribution of the 25 fossil samples with modern samples from Chad according to the laq and lxe grass phytolith indices. Threshold values (dotted lines) were estimated using the *rpart* function (R.13.0), as also indicated on modern box-plots (Fig. 4).

S. carconensis disappearance at the beginning of the Pleistocene. The agreement between the *S. carconensis* occurrence time span and the calculated $^{10}\text{Be}/^9\text{Be}$ ages along the core section, in which the species is present, confirms the reliability of the dating method based on meteoric ^{10}Be (Lebatard et al., 2008; Lebatard et al., 2010). *S. carconensis* varieties occurred in Africa during the Messinian and reached their optimal development in large lakes during the Piacenzian (Gasse and Fourtanier, 1991). At that time, they were widespread throughout the world (Gasse and Fourtanier, 1991). In the Afar region (Ethiopia), for instance, where at least three successive lacustrine phases occurred between 2.5 and 0.75 Ma, they predominate before 1.5 Ma (Gasse, 1980). They disappeared from the African continent during the early Pleistocene, but still survive today in some great temperate lakes of the Northern Hemisphere. Their disappearance from the African continent is assumed to result from the establishment of arid conditions which prevailed during the middle Pleistocene, and which led to desiccation of the northern African large lakes (Servant-Vildary, 1978; Gasse and Fourtanier, 1991).

6.2. The paleo-lake Chad: evidence and variations during the Miocene–Pliocene

Diatom assemblages dominated by *Aulacoseira granulata* and *Stephanodiscus carconensis* support freshwater lacustrine conditions at Bol (even interrupted) since at least 4.7 ± 0.1 Ma. *Aulacoseira* and *Stephanodiscus* are considered as freshwater planktonic species. They were widespread during the Piacenzian and early Pleistocene in large and deep lakes from East Africa (Gasse, 1980, 1986), the Chad basin (Servant, 1973), and the Hoggar (Rognon, 1967). *A. granulata* still dominates the present-day diatom flora, notably in Lake Chad (Rirongarti, 2014). The core lithology between 297 m and 230 m, and below 297 m depth, however, suggests that lacustrine sedimentation most likely occurred earlier than the first diatom evidence recorded at 4.7 ± 0.1 Ma, probably as early as 6.3 ± 0.1 Ma (Fig. 2). Indeed, before 4.7 ± 0.1 Ma, the Bol sequence includes pelites (35a, 35e), sandy claystones (32a), and apparently argillaceous “diatomite” (32e, 30j) according to the macroscopic aspect of sediments. Microscopic investigations of the sediments dated older than 5.5 ± 0.1 Ma, however, revealed no preserved diatoms (or phytoliths) but $<5 \mu\text{m}$ non-describable bio-silica fragments (35e,

35a, 32e, and 32a). These fragments may be related to undetermined diagenetic processes had affected the preservation of bio-silica material (including both diatoms and phytoliths) in this part of the core. It is not possible, therefore, to document paleoenvironmental changes and lake variations at Bol before 5.5 Ma although the nature of sediments attests to the presence of a lake.

Sample 30j, dated to 5.5 ± 0.1 Ma, provides the oldest biogenic evidence. It has no diatoms but is rich in phytoliths of obligate terrestrial plants such as palm trees. In this sample, palm phytolith abundance (about 13%) is comparable to that observed in modern soil samples from the Sudanian and Sahelian domains (Fig. 4). Elsewhere in the Bol sequence, the abundance of palm phytoliths never exceeds 5%. This suggests that terrestrial vegetation was present in the close surroundings of Bol around 5.5 Ma, just at the end of the Messinian, and that peri-lacustrine condition may therefore have prevailed at Bol during that time. During the Pliocene, the dominance of planktonic *Aulacoseira* species indicates the existence of expanded water areas at Bol. At 4.7 ± 0.1 Ma, the diatom flora is remarkably diversified and abundant (Table 2), as it can be observed during a lacustrine expansion. After 4.7 ± 0.1 Ma and then until 2.4 ± 0.1 Ma, variations in the abundance of diatom frustules (per gram of sediments) then suggest recurring phases of expansion-regression of the lake, or otherwise *in-situ* modifications of the lake properties that would have prevented the optimal development of diatoms. Extreme water turbidity as well as low silica availability, which is most unlikely in the Lake Chad basin (Olivry et al., 1996), may impede an optimal development of diatoms. But as the aquatic grass phytolith signal also varies temporally during that period (Fig. 6), it is likely that alternating marshy and fully lacustrine conditions occurred at Bol during the Pliocene as it did during the Holocene (Maley, 1977; Maley, 1980, 2010). At 4.4 ± 0.1 Ma (27n) and 2.4 ± 0.1 Ma (6p), no phytoliths and pollen, but only diatoms are observed. This may indicate fully lacustrine conditions at Bol, with lake margins located too far to allow any phytolith input from plants. It is unlikely that post-depositional dissolution of phytoliths occurred, given the great preservation of diatoms in these sediments. Phytolith preservation may have been altered during transport from the (local and regional) sources to the lake, but it would then be surprising that they all suffered similarly from total dissolution at all these different original places and/or during transport along their different journeys.

After 3.6 ± 0.1 Ma (17 h), the quasi-absence of diatoms likely suggests (episodic or continuous?) reduced lacustrine conditions at Bol until at least 2.7 ± 0.1 Ma (10b), which are consistent with high abundances of blocky-elongate phytoliths (including silicified bulliform cells) recorded in the same sediments (Fig. 6, Table 2). In modern sub-Saharan environments of Chad, blocky-elongate phytoliths, and notably silicified bulliform cells, are most abundant in wetland sediments from marshes where aquatic herbaceous vegetation is dominant (Novello et al., 2012). High relative abundances of silicified bulliform cells only may be related to plants with submerged root system undergoing high evapotranspiration rates (Novello et al., 2012). Combined evidences of marshy vegetation and high evapotranspiration rates at the same time interval suggest a temporary period of increased aridity in the Lake Chad basin. This hypothesis is consistent with a general trend toward aridification in the basin from ~ 7 Ma to 3 Ma inferred from oxygen isotopes on fish remains from northern Chad (Otero et al., 2011). Lake variations recorded at Bol are also in agreement with repeated lacustrine transgressive events documented in northern Chad ($>16^\circ\text{N}$) during the Pliocene (Schuster et al., 2009).

6.3. A lake supplied from the south

Several botanical and sedimentary evidences suggest that major water input to paleolake Chad came from its southern drainage basin. Fluvial contributions from the south are first inferred from the pollen assemblage at 4.2 ± 0.1 Ma. Maley (1972, 1981) previously showed that the presence of taxa living in the humid south (Sudano-Guinean, Guineo-

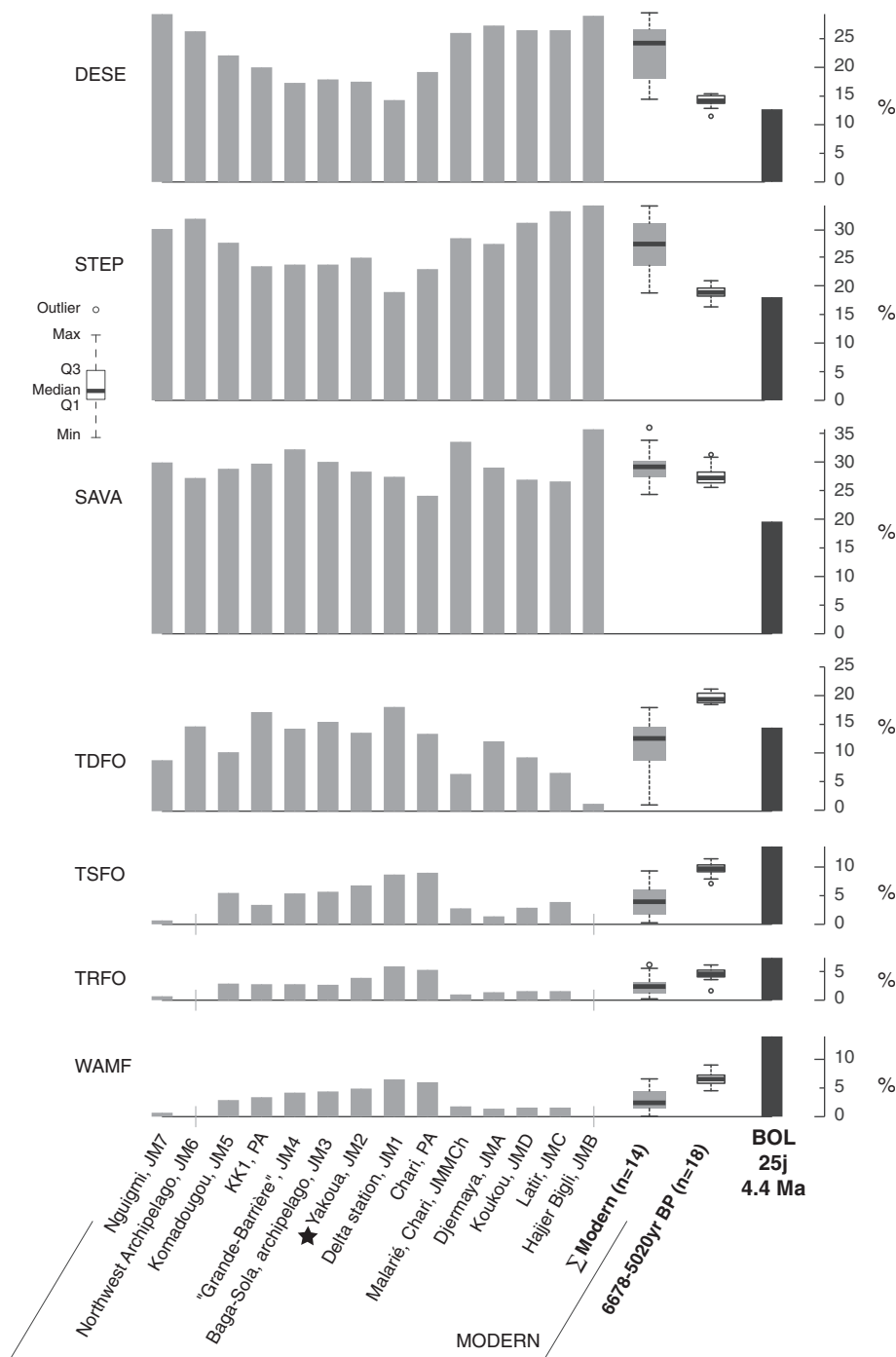


Fig. 8. Biome scores inferred from the fossil pollen data of Bol sample 25j (dated 4.2 Ma) compared to biome scores calculated for modern (Maley, 1972, 1981; Amaral et al., 2013) and Holocene (6678–5020 yr BP) data (Amaral et al., 2013) from Lake Chad. Biome classification follows Lézine et al. (2009): DESE: desert; STEP: steppe; SAVA: savanna; TDFO: tropical dry forest; TSFO: tropical seasonal forest; TRFO: tropical rain forest; WAMF: warm mixed-forest.

Congolian zones) in modern pollen assemblages of Lake Chad, clearly indicates that these pollens were brought by fluvial influx. Such a pattern was also suggested for Holocene sediments originated from the Kanem region, in the north-east of Lake Chad (Maley, 2004). Moreover, the Pliocene pollen assemblage is also rich in southern Afrotropical pollen taxa such as *Olea capensis*, *Rapanea melanophloeos*, *Maesa*, and *Myrica*. These taxa only occur nowadays in the Adamawa and Jos plateaus, which are located at mid-elevation and close to the southern reach of the Chad Basin (Letouzey, 1968). For instance, similar pollen taxa were determined for a middle Holocene pollen spectrum obtained on a core collected in

Lake Mbalang (7°19'N, 13°44'E, 1110 m a.s.l.), located in the eastern Adamawa plateau (Vincens et al., 2010). Conversely, taxa from the northern Saharan uplifts such as the olive tree *Olea europaea* (Letouzey, 1968) are totally absent in the Pliocene pollen assemblage. Water supply from the south is also consistent with the presence of kaolinite in the clay fraction (Moussa, 2010). Kaolinite is the main clay mineral developed in lateritic weathering profiles of the tropical zone. It has been identified as the preponderant mineral in the suspended matter presently exported from their southern catchment by the Logone and Chari rivers to the lake (Gac, 1980; Gac and Tardy, 1980; Olivry et al., 1996).

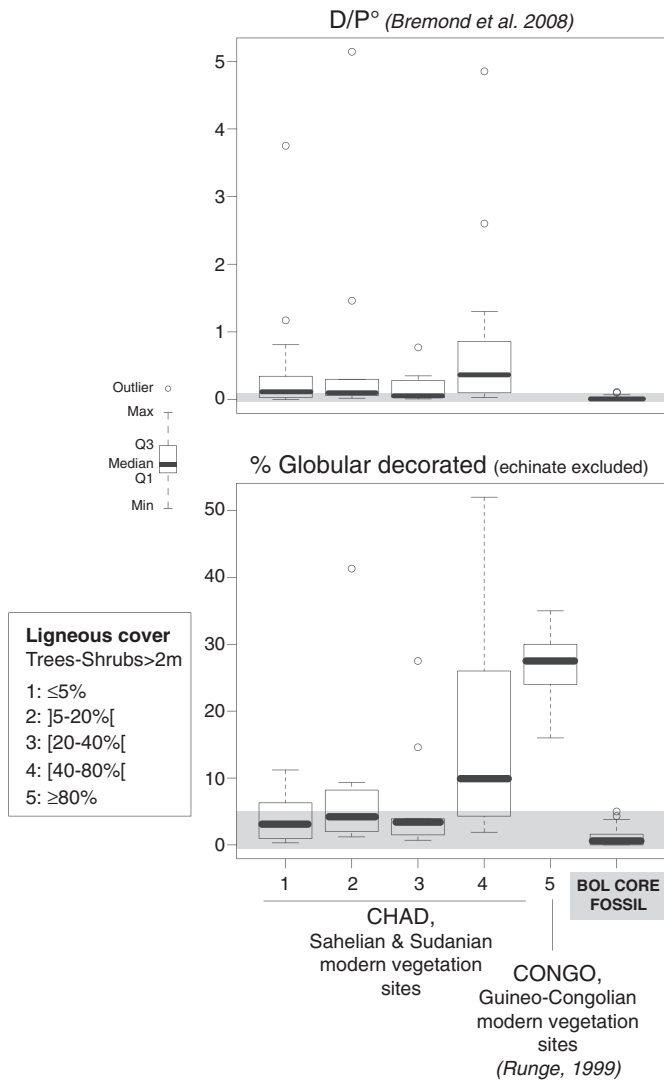


Fig. 9. Box-plots of the values of D/P° phytolith index values and of the percentages of globular decorated phytoliths (echinate excluded) in the Bol fossil assemblages, and compared with those observed from non-forested soils of Chad, and from tropical evergreen forest soils of the Democratic Republic of Congo, DRC (Runge, 1999). Modern environments are classified according to their percentages of ligneous cover directly estimated in the field for Chadian environments, and estimated by satellite for forested environments from DRC (see Barboni et al., 2007). Gray areas draw the ranges of D/P° values and of the globular decorated phytolith percentages obtained for the fossil samples of Bol.

Compared to biome scores calculated for modern and Holocene pollen assemblages from sediments of Lake Chad (Maley, 1972, 1981; Amaral et al., 2013), the WAMF (warm afro-montane forest) biome score is significantly higher at 4.2 ± 0.1 Ma. This may also suggest a more intense drainage of the southern highlands than observed nowadays. The occurrence of trapeziform polylobate phytoliths related to C_3 -Pooideae grasses (Barboni and Bremond, 2009; Rossouw, 2009) throughout the Bol sequence also further supports this hypothesis. Indeed, these types are currently absent in modern phytolith assemblages of the Lake Chad basin (Novello et al., 2012).

6.4. Tropical savanna as the dominant biome in the basin

Both phytolith and pollen data from the Bol sequence are congruent with the presence of the savanna biome during the Miocene–Pliocene in the Lake Chad basin. Low D/P° index values ($<<1$) and high relative abundance of grass pollen (68%) are indeed typical for savannas and grasslands. The Ixe phytolith index suggests that humid-loving but

likely non-lacustrine grasses, i.e. mesophytic grasses, occurred in the Lake Chad basin during all the Pliocene. The Iaq phytolith index indicates that local aquatic grasses also contributed to the grass phytolith signal recorded at Bol. It is noteworthy that no significant xerophytic grass signal is indicated by the Ixe phytolith index (Figs. 6, 7), even during the period of increased aridity that we hypothesized to have occurred between 3.6 and 2.7 ± 0.1 Ma. The absence of xerophytic grass signal in the phytolith record is even more surprising considering the high steppe biome score that we obtained for the pollen assemblage at 4.2 ± 0.1 Ma (Fig. 8). Thus we expect that the grass phytolith inputs to the lake during the Pliocene were quasi-exclusively originated from fluvial contributions from the humid south and from the local grasslands associated with the lake.

Climate and hydrological simulations for the mid-Pliocene (with fixed CO_2 at 405 ppm) show three potential biomes in the Lake Chad basin: tropical xerophytic shrubland, tropical savanna, and tropical deciduous forest/woodland (Salzmann et al., 2008; Contoux et al., 2013). The most forested biome (tropical deciduous forest/woodland) is simulated when the ITCZ reaches a northerly position (Contoux et al., 2013). The phytolith and pollen grass signals support the presence of the tropical savanna biome in the Lake Chad basin as far as the Bol locality throughout the Pliocene, and therefore give poor support for the existence of tropical xerophytic shrubland for latitudes $<13^\circ N$. As the savanna biome was present, the monsoon regime probably reached a more northern position during the summer rainfall season than observed nowadays in Central Africa, but not far enough to allow forest biome to expand until the location of Bol. Forest tropical biomes, therefore, were most likely present in the southern part of the drainage basin of paleolake Chad during the Pliocene, as suggested by higher biome scores for tropical seasonal and rain forests at 4.2 ± 0.1 Ma than for modern and Holocene samples (Fig. 8).

The presence of the tropical savanna biome at $13^\circ N$ does not exclude abundant ligneous taxa in the paleovegetation, as it can be currently observed for some modern savannas of West-Central Africa (“woodland savanna”, “tree savanna”, Boughey, 1957). Phytoliths are good indicators to trace the existence of true tropical forest environments in tropical Africa (Fig. 4) (Bremond et al., 2005a; Aleman et al., 2011) but, conversely, they still fail to reconstruct the ligneous cover in tropical savanna environments (Neumann et al., 2009; Novello et al., 2012). As a result, the Bol phytolith data cannot provide much information on the presence/abundance of the ligneous cover in these paleo-savannas. The abundance of globular decorated phytoliths, as well as the values of the D/P° phytolith index observed for fossil samples, could be only indicative of a large range of >2 m-ligneous cover values, from 0% to 80% according to the calibration performed between our modern phytolith data and ligneous cover percentages directly estimated at the modern Chadian sites we sampled (Fig. 9). Then the existence of savannas with high ligneous component cannot be totally excluded since Aleman et al. (2014) also have demonstrated that the “super-local” grassy vegetation associated with a lake may hide the signal from the surroundings and regional vegetation.

6.5. Lesson for the Pliocene grass expansion in Central Africa

Evidences of grass-dominated vegetation are observed in northern Chad ($16^\circ N$) during the Messinian (Blondel et al., 2010), but it is likely that the spread of grass-dominated environments occurred by the Zanclean–Piacenzian boundary, around 3.6 Ma (Zazzo et al., 2000; Geraads et al., 2001). Our data cannot document the history of grasses in Central Africa before ~ 5.5 Ma as no micro-botanical remains were preserved in the Bol core beyond this period. However, our data suggest that the expansion of grasses in Central Africa may have occurred earlier than the Zanclean as grass silica short cell phytoliths already account for 61% in Bol sediments at ~ 5.5 Ma. This hypothesis is consistent with the increase of grass cuticles and pollen grains recorded in a core off the Niger Delta (5 – $6^\circ N$) during the Messinian, just before the Miocene–Pliocene transition (Morley and Richards, 1993). Carbonate isotopes on mammalian teeth (including the hominin

A. bahrelghazali) indicate that C₄ vegetation was dominant in northern Chad (16°N) around 3.6 Ma (Zazzo et al., 2000; Lee-Thorp et al., 2012). At Bol, the fossil assemblages are dominated by bilobate, cross (i.e. lobate), and saddle phytoliths (Table 2; Fig. 6), which are most likely related to C₄ grasses according to previous phytolith studies (Fredlund and Tieszen, 1994; Barboni et al., 1999; Bremond et al., 2005b, 2008).

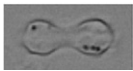
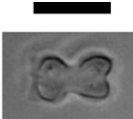
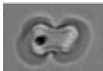
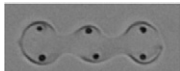
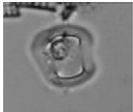

Recent studies, however, have demonstrated that some lobate and saddle phytoliths may also be produced by C₃ aquatic grass species. This was notably observed for the C₃ aquatic species *Phragmites australis* (Arundinoideae), which can produce about 68% of saddle phytoliths in its leaves, and for the C₃ aquatic species *Leersia hexandra* (Ehrhartoideae), which can produce 21% of lobate (bilobate, cross, and tabular polylobate) phytoliths in its leaves (Novello et al., 2012). Today, these aquatic grass species are part of the aquatic herbaceous vegetation present in the Lake Chad marshes, and therefore are expected to largely contribute to the phytolith assemblages of modern sediments deposited in the lake. However, C₃ aquatic grasses cannot presently be properly identified using phytoliths. *Phragmites*, for instance, produces trapeziform saddles that are not identified to the genus-level (Novello et al., 2012) because they can be confused with trapeziform saddles also produced by some tropical C₄ grass species of the Chloridoideae subfamily such as *Cynodon dactylon* and *Eragrostis pilosa* (Fig. 9), and even of the Panicoideae subfamily such as *Andropogon gayanus* (Fig. 9) currently occurring in the Lake Chad basin (Gaston, 1996). A similar ambiguity applies to *hexandraLeersia*. This species produces indeed various types of lobate phytolith types (Bi2, Bi8, Bi11, and Poly11) that can be also produced by tropical C₄ grass species of the Aristidoideae subfamily such

as *Aristida stipoides* (Fig. 9), and of the Panicoideae subfamily such as *Andropogon pseudapricus*, *Diheteropogon amplexans*, and *Panicum laetum* (Fig. 9) currently occurring in the Lake Chad basin (Gaston, 1996). Some authors (Neumann et al., 2009; Prasad et al., 2011), however, asserted to have identified a specific bilobate type – “the bilobate type with scoped ends” – which they described as diagnostic of the Ehrhartoideae grass subfamily. The description criteria and classification used in this study were not designed to specifically isolate this bilobate type that may have thus be counted as part of the bilobate types with notched ends (Bi8 type) (see Table 3 for Bi-8 type photograph), which in fact probably include a larger morphological variability than the bilobate type with scoped ends itself. Although a new counting of the fossil samples of Bol should be done to investigate the presence of this diagnostic type in the record, further investigations are nevertheless needed to allow discriminating C₃ aquatic grasses from C₄ grasses using phytoliths. The respective presence/abundance of C₃ and C₄ grasses in the vegetation of the Bol area during the Pliocene thus remains uncertain.

6.6. New insight on climate in Central Africa during the Pliocene

The Pliocene is known to include long periods of warmer and cooler climate than today (Feakins and deMenocal, 2010). The mid-Pliocene is particularly characterized by higher global mean annual temperatures (+2–3 °C) (Ravelo et al., 2004) compared to present times (Feakins and deMenocal, 2010; Salzmann et al., 2011; Contoux et al., 2012). At Bol, the mid-Pliocene global warm event may be expressed by an aridification of the climate as suggested by the paucity of diatom

Table 3
Occurrences of the bilobate, cross, and tabular polylobate phytolith types produced by the aquatic C₃ grass species *Leersia hexandra* on the one hand, and of the saddle phytolith types produced by the aquatic C₃ grass species *Phragmites australis* on the other hand, compared with the occurrences of the same phytolith types in 27 tropical C₄ grass species largely distributed presently in the Lake Chad basin. Original data are from Novello et al., 2012. Scale bar: 10 µm.

	<i>Leersia hexandra</i>				<i>Phragmites australis</i>	
	Bi2	Bi8	Bi11	Poly1	S4	S5
C ₄ grass species (from Novello et al., 2012)						
Aristidoideae						
<i>Aristida stipoides</i> Lam.	x	x	x			
Chloridoideae						
<i>Ctenium elegans</i> Kunth		x				
<i>Cynodon dactylon</i> (L.) Pers.		x				x
<i>Dactyloctenium aegyptium</i> (L.) Willd.						x
<i>Eragrostis pilosa</i> (L.) P. Beauv.		x				x
<i>Eragrostis tremula</i> Steud.						x
<i>Sporobolus cordofanus</i> (Steud.) Coss.					x	x
Panicoideae						
<i>Andropogon gayanus</i> Kunth	x	x				x
<i>Andropogon pseudapricus</i> Stapf	x	x		x		
<i>Brachiaria xantholeuca</i> (Schinz) Stapf		x	x	x		
<i>Cenchrus biflorus</i> Roxb.	x	x				
<i>Diheteropogon amplexans</i> (Nees) Clayton	x	x	x	x		
<i>Echinochloa pyramidalis</i> (Lam.) Hitchc. & Chase	x	x		x		
<i>Hyparrhenia bagirmica</i> (Stapf) Stapf		x	x	x		
<i>Hyparrhenia barteri</i> (Hack.) Stapf		x		x		
<i>Hyparrhenia dissoluta</i> (Nees et Steud.) Clayton		x	x			
<i>Loudetia simplex</i> (Nees) Hubb.		x				
<i>Loudetia togoensis</i> (Pilg.) Hubb.	x	x				
<i>Panicum anabaptistum</i> Steud.	x	x		x		
<i>Panicum fluvicola</i> Steud.	x		x	x		
<i>Panicum laetum</i> Kunth	x	x	x	x		
<i>Panicum subalpinum</i> Kunth	x	x	x			
<i>Panicum turgidum</i> Forssk.		x	x			
<i>Pennisetum pedicellatum</i> Trin.	x	x	x			
<i>Setaria sphacelata</i> (Schumacher.) Moss				x		
<i>Vetiveria nigriflora</i> (Benth.) Stapf						x
<i>Vossia cuspidata</i> (Roxb.) Grill.		x	x			

remains, and the abundance of marsh phytolith indicators (blocky and elongate phytoliths), and high evapotranspiration rates (inferred by high abundances of silicified bulliform cells). Besides it is likely that North, West, and Central African regions all experienced aridity during the mid-Pliocene. Indeed, pollen records indicate aridification of the northwestern African climate between ~3.5 and ~3.2 Ma, and at ~2.8 Ma (Leroy and Dupont, 1994, 1997), while dust records from marine core indicate increased wind strength at ~2.8 Ma in West Africa (deMenocal, 2004). Mid-Pliocene aridity may be related to significant elevated pCO_2 (330–425 ppm) compared to present times (200–260 ppm) (Pagani et al., 2010), and/or to direct effects of Northern Hemisphere Glaciation such as ocean circulation and trade wind modification (Feakins and deMenocal, 2010; Salzmann et al., 2011).

6.7. Pliocene environment and hominin occurrences in tropical Africa

The core sediments provide no micro-biological remains during the time spanning *Sahelanthropus tchadensis* occurrence. During Abel's occurrence (~3.6 Ma) major changes in the micro-biological record are observed at Bol, which we have interpreted as a lake regression under higher arid climate conditions. For the same period in the Djurab (further north), evidences of C_4 vegetation (Lee-Thorp et al., 2012), open environments (Zazzo et al., 2000; Geraads et al., 2001), and trend toward aridity (Otero et al., 2011) were observed. In East Africa, the early occurrence of the species *Australopithecus afarensis* is dated from ~3.9–3.6 Ma (Upper Laetoli Beds, Tanzania) (Harrison, 2011). Phytoliths from Upper Laetoli beds indicate a shift to xeric environmental conditions followed by an increase of C_4 grasses in the vegetation (Rossouw and Scott, 2011). A trend toward aridity, in relationship with a Chenopodiaceae and Amaranthaceae pollen increase, has also been recorded at the regional scale of East Africa at ~3.6 Ma (Bonnefille, 2010). Yet it seems that arid conditions have not impeded the presence of hominins in both Central and East Africa during the Zanclean–Piacenzian boundary. Such arid conditions may have notably triggered the evolution of hominins in Africa through bipedalism, which may allow them consuming less energy under high-temperature conditions (Wheeler, 1991a, 1991b).

7. Conclusion

The study of the Bol borehole provides new evidences on the paleoenvironment that prevailed in the Lake Chad basin during the Miocene–Pliocene, between ca 6 and 2 Ma. The nature of the sedimentation indicates that Paleo-Lake Chad existed at least since 6.3 ± 0.1 Ma, but probably settled earlier during the Miocene (Moussa, 2010). Siliceous remains are however too poorly preserved to document the lacustrine conditions and the surrounding vegetation that existed at Bol before 5.5 ± 0.1 Ma. Both sedimentology (for instance, kaolinite in the sediments) and plant remains (presence of *Afromontane* pollen taxa, and occurrence of C_3 –*Pooideae* phytolith types) are consistent with the hypothesis that water supply to the lake during the Pliocene originated from the southern regions, including the southern highlands. Preserved diatom floras and phytolith assemblages attest for fully lacustrine conditions in the southern Lake Chad Basin during the beginning of the Zanclean (4.7 – 4.1 ± 0.1 Ma) and at the end of the Piacenzian (2.7 – 2.4 ± 0.1 Ma), and peri-lacustrine conditions at the end of the Messinian (at 5.5 ± 0.1 Ma) and during the beginning of the Piacenzian (3.6 – 2.7 ± 0.1 Ma). Phytolith assemblages and pollen elements indicate the presence of the tropical savanna biome in the Lake Chad basin during the Pliocene, in agreement with previous simulations by models (Salzmann et al., 2008; Salzmann et al., 2011; Coutoux, 2013). Another important result only observed between 3.6 and 2.7 ± 0.1 Ma, is that sediment diatom concentration exhibits a significant decrease, concomitantly with the increase of blocky and elongate phytoliths, including the silicified bulliform cells, in the phytolith assemblages. This evidence supports the aridification recorded in northern Chad and elsewhere in North

and West Africa during the Zanclean–Piacenzian transition (Leroy and Dupont, 1994, 1997; Otero et al., 2011).

The first question that arises concerns the paleoenvironmental conditions that existed at Bol before the end of the Messinian. The second question concerns the respective role of C_3 and C_4 grasses in the region in relationship with aquatic vegetated areas. Only the analyses of additional fossil samples of the Bol record as well as a deep and continuous lacustrine record combined with advanced micro-biological remains calibrations on modern material would enable to complete this dataset. Moreover, further paleontological excavations regarding the Piacenzian deposits of northern Chad would confirm the hypothesis that the event recorded at Bol may have impacted the evolution of vertebrate faunas in the region, and particularly hominins.

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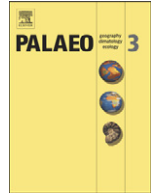
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Update

Palaeogeography, Palaeoclimatology, Palaeoecology

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Corrigendum

Corrigendum to “Diatom, phytolith, and pollen records from a $^{10}\text{Be}/^9\text{Be}$ dated lacustrine succession in the Chad basin: Insight on the Miocene–Pliocene paleoenvironmental changes in Central Africa” [PALAEO: 430 (2015) 85–103]



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The authors regret for the correction and wish that Fig. 4 is replaced with the figure below.

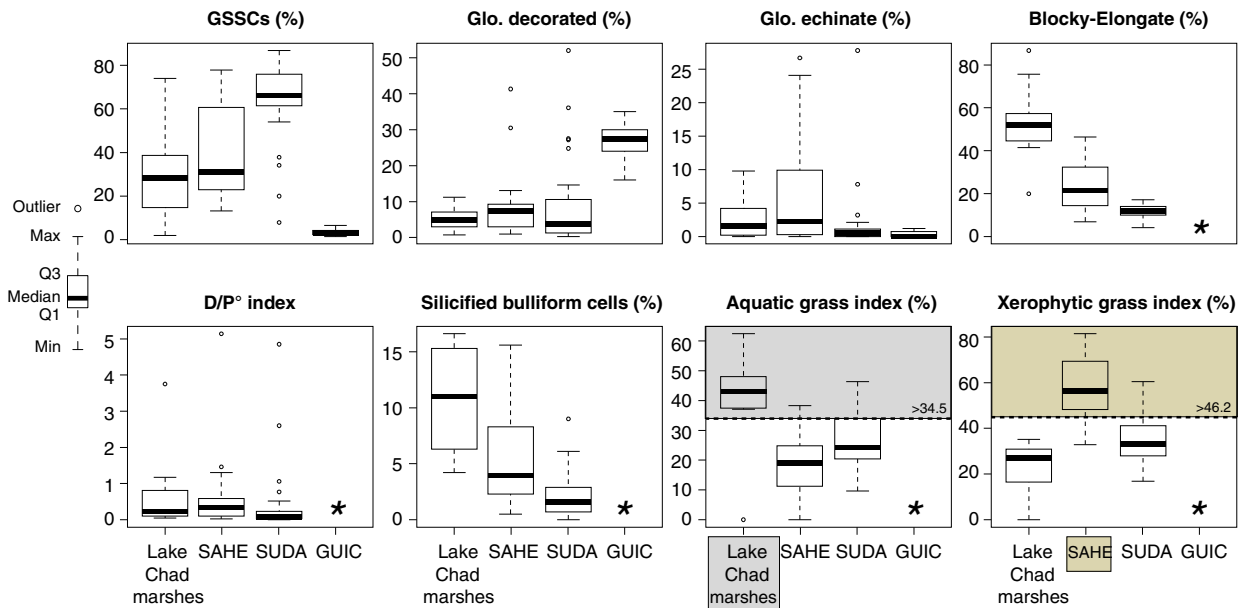


Fig. 4. Box-plots showing the abundance patterns of the main phytolith indicators and indices considered in this study according to modern sub-Saharan environments: Lake Chad marshes, Sahelian, and Sudanian vegetation, and Guineo-Congolian vegetations (from Runge, 1999). Gray areas indicate the range of Aquatic (Iaq) and Xerophytic (Ixe) grass indices characterizing a phytolith signal from present-day Lake Chad marshy grasses and from Sahelian grasses.

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