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Climate of the Past

Past environmental and climatic changes during the last 7200 cal yr BP in Adamawa plateau (Northern-Cameroun) based on fossil diatoms and sedimentary carbon isotopic records from Lake Mbalang

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Abstract. Past limnological conditions of Lake Mbalang (7°19′ N, 13°44′ E, altitude: 1130 m) and vegetation type were reconstructed from diatoms and sedimentary stable carbon isotope records ($\delta^{13}C$) since 7200 cal yr BP. The data showed that before 3600 cal yr BP, the water column was dominantly stable except around 5000–5300 cal yr BP where diatoms evidenced a mixed upper water layer and $\delta^{13}C$ data suggest more forested vegetation in the landscape. These stable conditions can be explained by a strong monsoon flux and relatively northern position of the ITCZ that entailed high or low rainfall well distributed over the year, allowing the development of mountainous forest taxa. The decreasing trend of the monsoon flux towards the mid-Holocene was affected by several abrupt centennial to millennial-scale weakening at 6700, 5800–6000, 5000–5300, 4500 and 3600 cal yr BP. However, their impact on the vegetation is not visible, probably because rainfall distribution was favourable to forest maintenance or extension. After 3600 cal yr BP, the water column became very mixed as a result of more intense NE trade winds (Harmattan) that led at ∼3000 cal yr BP to the establishment of savannah in the vegetation landscape. At that time, rainfall was probably reduced following the southward shift of the ITCZ, and the distribution of yearly rainfall was not favourable anymore to forest development. A strong seasonality with a marked dry season was established, conditions that maintained the savannah vegetation until today. Diatom data suggest the lake did not dry up during the last 7200 cal yr BP; however, a low lake level observed at 2400–2100 cal yr BP is contemporaneous to a climatic event evidenced in several areas of tropical Africa and could correspond to the southernmost position of the ITCZ. Other low lake levels are observed at 1800 and 1400 cal yr BP, after which the lake rose to its present level.

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1 Introduction

Climatic changes during the Holocene in Western Africa have been mostly studied in the subequatorial forest and Sahelian/arid regions. The two regions are submitted to the atmospheric monsoon flux from the tropical Atlantic that reaches its northern maximum extension during the northern summer (July–August) in the present. It is present over the year in the northern subequatorial regions except during a 3-month dry season centered in January. At these latitudes, this monsoon flux is characterised by a deep atmospheric convection; however, a relative stability of the atmosphere at low levels at the base of the monsoon flux is observed in July–August when the Intertropical Convergence Zone (ITCZ) is farthest north. Convective rainfalls are almost suppressed during this period of the year at the northern border of the Guinean Gulf.

During the Holocene, the monsoon flux penetrated more or less deeply inside the Saharan region, entailing an alternation of wet and dry phases (e.g. Servant and Servant-Vildary, 1980; Gasse, 2000) superimposed on a general trend of monsoon weakening in response to decreasing summer insolation of the Northern Hemisphere (Kutzbach and Street-Perrot, 1985). Modifications in the intensity of the monsoon were also suggested by changes of precipitation *minus* evaporation balance at subequatorial latitudes (Talbot and Delibrias, 1980; Nguetsop et al., 2004).

Concordant data from low and high altitudes in western Cameroon (Maley and Brenac, 1998; Reynaud-Farrera et al., 1996; Nguetsop et al., 1998; Stager and Anfang-Sutter, 1999; Vincens et al., 1999; Ngomanda et al., 2007, 2009b; Kossoni and Giresse, 2009) suggest that climatic changes were also controlled by modifications in the vertical structure of the atmosphere (Nguetsop et al., 2004). The present stable air layer situated at the base of the monsoon flux in July–August could have extended on the western Cameroon lowlands and mid altitude areas during the greatest part of the year, entailing the almost suppression of convective rains before 3000 cal yr BP. After that date, the influence of the stable air layer was strongly reduced and convective rainfall reappeared. If this is true, one can expect different climate evolutions between lowlands south of the Adamawa plateau, mid altitude regions such as Adamawa (1000–1100 m), and western Cameroon highlands (>2000 m).

Available paleoclimatic records of the last 3000 yr in the tropical zones of Africa, close to the Atlantic coast of Gabon, West-Cameroon and South-Congo (Ngomanda et al., 2009a; Nguetsop et al., 2004; Vincens et al., 1999), suggest significant modifications in abundance and/or seasonal distribution of rainfall in response to north south shift of the Intertropical Convergence zone (ITCZ). Thus, climatic changes affected in the past water resources that impacted on human population and vegetation landscape of central and north tropical Africa. Paleoenvironmental studies showed that the rain forest belt was reduced and persisted only in refuge zones during the Last Glacial maximum (e.g. Maley, 1987). Between ~2500–2000 cal yr BP, the rain forest was strongly disturbed or was replaced by savannas, depending on the sensibility to climate change of each site in central Atlantic Africa (Vincens et al., 1999). The present day "hot spots" of biodiversity (Tchouto et al., 2006) and the spatial heterogeneity of the rain forest are probably inherited from past climate changes. The question is how the Adamawa plateau located between the dry zones in the north and wet areas in the south responded to theses major climatic changes.

Organic components in lake sediments are supplied by allochtonous organisms and riverine, terrestrial and atmospheric inputs. They are biomarkers of biological production, source organisms, and paleolimnological changes in the drainage basin. Here we study paleolimnological changes inferred from a multi-proxy data set of microfossils and carbon stable isotope ratios of sediment of the core M4 retrieved from Lake Mbalang in the Adamaoua in Cameroon, along with sedimentary facies (Ngos and Giresse, 2011) and AMS carbon-14 datings (Tandetron Accelerator Mass Spectrometry). Specifically, past limnological conditions will be accessed through the analysis of diatom ecological groups; variations in trade wind (Harmattan and monsoon) intensity will be reconstructed from allochtonous diatom taxa or species that characterise stable water table. The evolution of sedimentary δ^{13}C will be compared to published palynological data, showing that Lake Mbalang area was only made up of patches of forest surrounded by savannas (with fluctuations of their respective areas) and from ca. 2500 cal yr BP, the region was completely covered by savannas (Vincens et al., 2010). These phenomena are discussed in relation to the Monsoon African System and environmental changes for the last 7000 yr.

2 The site

2.1 Location and general characteristics of the studied lake

Lake Mbalang (7°19′ N, 13°44′ E, altitude: 1130 m) lies on the Adamawa plateau that belongs to the Cameroonian volcanic line (Fig. 1). This high topographic unit (850–1200 m) extends between latitudes 6° and 8° north and between longitudes 11°30′ to 15°45′ E. The plateau is limited in the north by the relatively lowlands of the Benue plain (800–300 m) and in the south by the sub-Cameroonian plateau (800–500 m). Crystalline and foliated metamorphic rocks make up the substratum of this unit which is largely covered today by ancient volcanic basaltic flows differently altered from one region to another (Humbel, 1967). According to Gèze (1943), in the Adamawa as well as in the whole volcanic line of Cameroon, three volcanic series can be encountered: the lower black series to which Mbalang region belongs dated from upper Cretaceous to upper Eocene (Bachelier, 1957), the medium white series (end of Neogene) and the upper black series (Quaternary). These series are respectively composed of basalts and andesites conserved as yellowish clays, trachyte and phonolithe lavas, and basaltic volcanic deposits. Volcanic and ferralitic materials in form of dome and outcrops are encountered at the vicinity of the lake. Soils are mostly ferralitic, rich in aluminium and iron oxides with frequent neoformation of halloysite and kaolinite. Other clay minerals present include gibbsite and siderite.

The lake, with a surface area of 50 ha and a narrow watershed (~90 ha), is a volcanic maar described as an asymmetric bowl with steep slopes. Lake Mbalang water maximum depth is about 52 m and is characterized by the absence of a present day river inlet. The euphotic zone is 3.45 m deep (Kling, 1987). According to Kling (1987), Lake Mbalang is "moderately stable"; however a relatively cool epilimnion is subjected to period of surface warming in times of low wind stress. The water column is then affected by yearly modifications of mixing depth that could be attributed primarily to higher temperatures as well as intensity of storms or maximum wind speed. The lake is fed only by rainfall and runoff

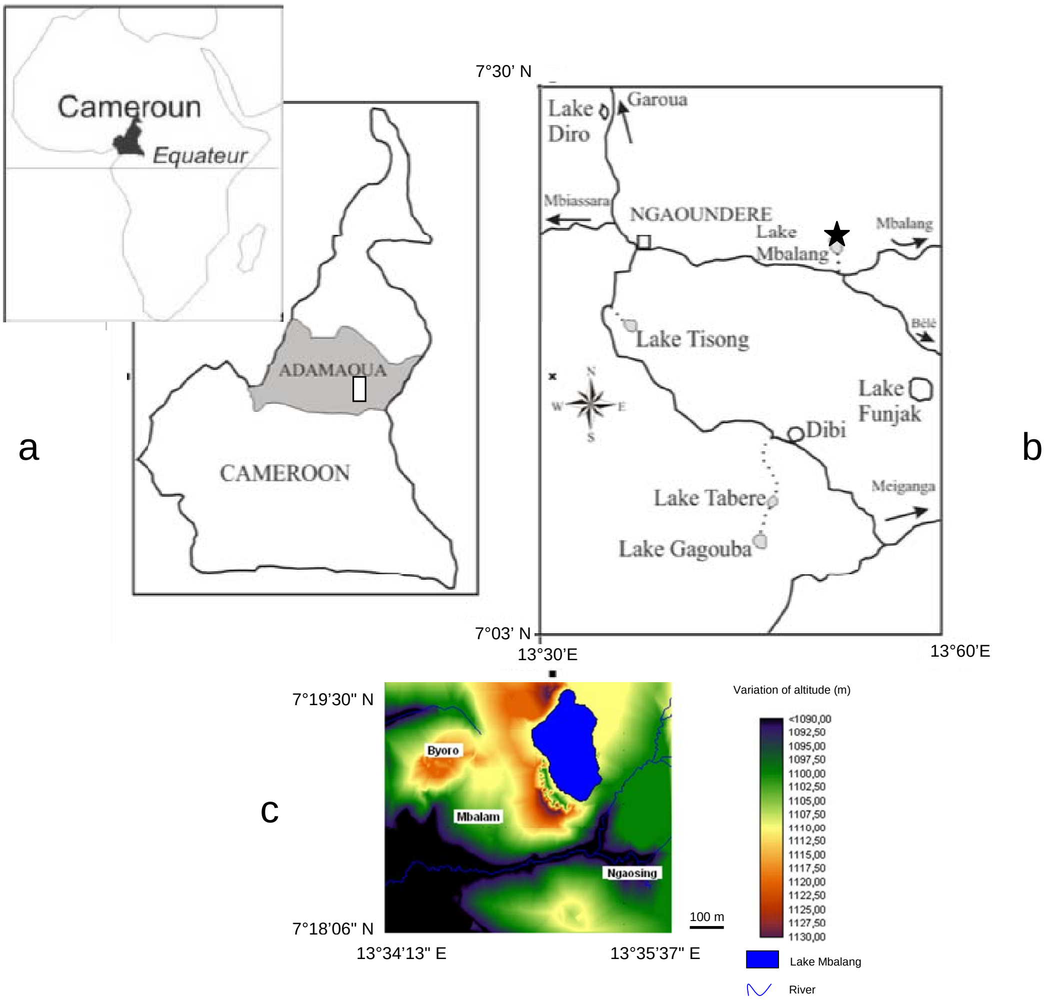

Fig. 1. Location of Lake Mbalang in the Adamawa plateau; morphometric features and area of the lake. The location of the lake is shown with a black star in **(b)**.

from the catchments, water losses occur through evaporation; however a surface outlet is present at the southeastern part of the lake but functions only during very high lake levels over the year. The ^{210}Pb profiles along the first 80 cm of the sediment in the lake suggested regular sediment supply from smooth erosion of the surrounding catchments, hence fossil sediments of the lake can be presumably suitable for paleoenvironmental studies (Pourchet et al., 1987).

2.2 Vegetation

The Adamawa region is occupied by tree or shrub savannas characterized by *Daniellia oliveri* (*Caesalpiniaceae*) and *Lophira lanceolata* (*Ochnaceae*); these savannas are strongly altered in some areas due to their permanent use as grazing land. Highest altitudes areas are occupied by soudano-guinean vegetation dominated by *Hymenodyction floribundum* (Letouzey, 1968, 1985). The edges of the lake are more forested with taxa such as *Croton macrostachyus*, *Sterculia tragacantha*, *Polyscias fulva*, *Rauvolfia vomitoria*, *Pittosporum mannii*, *Ficus capensis*, etc... Typical savanna trees encountered were *Annona senegalensis*, *Allophilus africanus*, *Cussonia barteri*, *Piliostigma thonningi*, *Terminalia glaucescens* and *Harungana madagascariensis*.

2.3 Climate

The region is under the influence of the altitudinal tropical climate that shows two distinct seasons: the dry season that last from November to March and the rainy season from April to October with rainfall maxima in July–September. The mean annual rainfall is 1500 mm; mean annual temperature varies from 23 to 26 °C (Suchel, 1988). In a classic picture, seasonal changes are explained by the displacement during the year of the intertropical convergence zone (ITCZ) in direction of the most heated hemisphere. During the dry

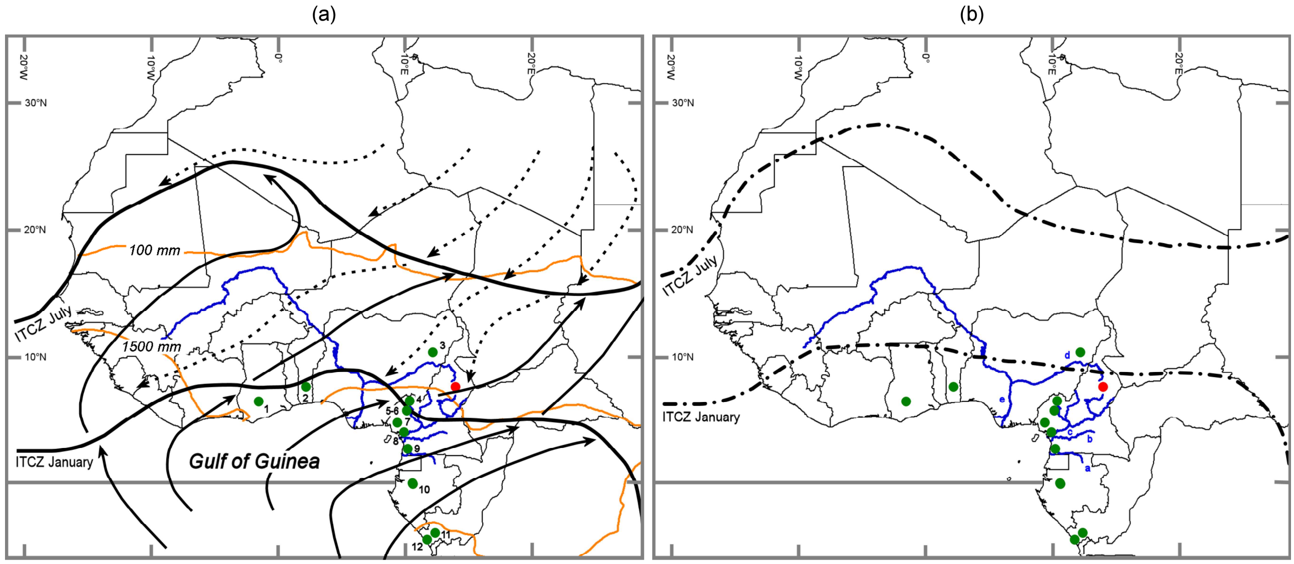

Fig. 2. Map showing the **(a)** modern positions of Intertropical Convergence Zone (ITCZ) during the northern summer (ITCZ July) and northern winter (ITCZ January). The solid arrows represent the monsoon flux while dotted arrows represent the NE trade winds (Harmattan) (Leroux, 2001). Orange full lines represent isohyetal lines 1500 mm and 100 mm (New et al., 2000). Colored dots correspond to sites were paleorecords (green dots) are available: 1 – Bosumtwi, 2 – Sele, 3 – Tilla, 4 – Djupi, 5 – Shum Laka, 6 – Bambili, 7 – Barombi Mbo, 8 – Ossa, 9 – Nyabessan (Ntem River), 10 – Nguène, 11 – Sinnda, 12 – Kitina and Mbalang (red dot). **(b)** Possible position of ITCZ before 3600 cal yr BP inferred from diatom and δ^{13}C isotopic data. Rivers of the Gulf of Guinea: Ntem **(a)**, Nyong **(b)**, Sanaga **(c)**, Benoué **(d)** and Niger **(e).**

season (boreal winter), the ITCZ is located south of the Adamawa plateau, the zone is then under the influence of the dry north-eastern trade winds (Harmattan). It moves northwards during the rainy season (boreal summer), the zone is then under the influence of humid south-western air masses (monsoon flux) that bring precipitation (Fig. 2). However, the African easterly waves may strongly modulate the spatial organisation of rainfall over West Africa (Nicholson, 2009).

3 Material and methods

3.1 Description of the core

The core was collected in March 1998 at the centre part of the lake (44 m deep) with a Mackereth air-compressed corer by Ecofit program team. Lithology and sedimentology of the core M4 has been described by Ngos et al. (2008) and Ngos and Giresse (2011). The lithology of the 6 m long core showed globally a dark clayey organic mud with clearer/darker laminas at certain levels (Fig. 3a). Coarser sandy laminas (up to 10 % sand in some levels) are observed at the base of the core between 560 and 580 cm (Ngos et al., 2008; Ngos and Giresse, 2011). Thin-section examinations and XRD analysis of Lake Mbalang core show variable siderite quantities (Ngos et al., 2008). The siderite is ubiquitous but only at small amounts. In few places, nebula-like masses of very small (<5 µm) siderites prisms were observed (see Ngos et al., 2008).

Preliminary observations of thin-sections showed that biogenic particles composed of spongiae spicules and diatoms are present throughout the core. Phytoliths and spicules were observed and counted during diatom counting under the light microscope, but not identified to generic or specific levels (Fig. 3b and c). Minerals such as siderite, quartz, feldspars and augite could also be observed in the form of layers or scattered in the sediment (Ngos et al., 2008; Ngos and Giresse, 2011; Fig. 3d).

Spicules were more abundant at the base of the core (587–225 cm), the ratio spicules/diatoms (Fig. 3c) counted was relatively high ($>20\times10^{-2}$). The most important peaks appeared at 535 cm (3664×10^{-2}), at 557 cm (363×10^{-2}), at 508 cm (649×10^{2}) and between 391 and 379 cm (403–500×10^{-2}). At the upper part of the core, the ratio was generally low ($<10\times10^{-2}$), the only relatively high values were observed at 182 cm (89×10^{-2}) and 67 cm (35×10^{-2}). The ratio phytoliths/diatoms followed broadly the same pattern of variation as spicules/diatoms but values are lower (Fig. 3b), peaks were evidenced at 557 cm (1208×10^{-2}), 535 cm (2545×10^{-2}) and at 508 cm (1607×10^{-2}). Relatively lower ratios were observed between 587–585 cm (17–31×10^{-2}), at 544 cm (141×10^{-2}) and at 526 (290×10^{-2}). A decreasing trend was observed towards the top of the core, the ratio reaches values close to 2×10^{-2} between 39–26 cm.

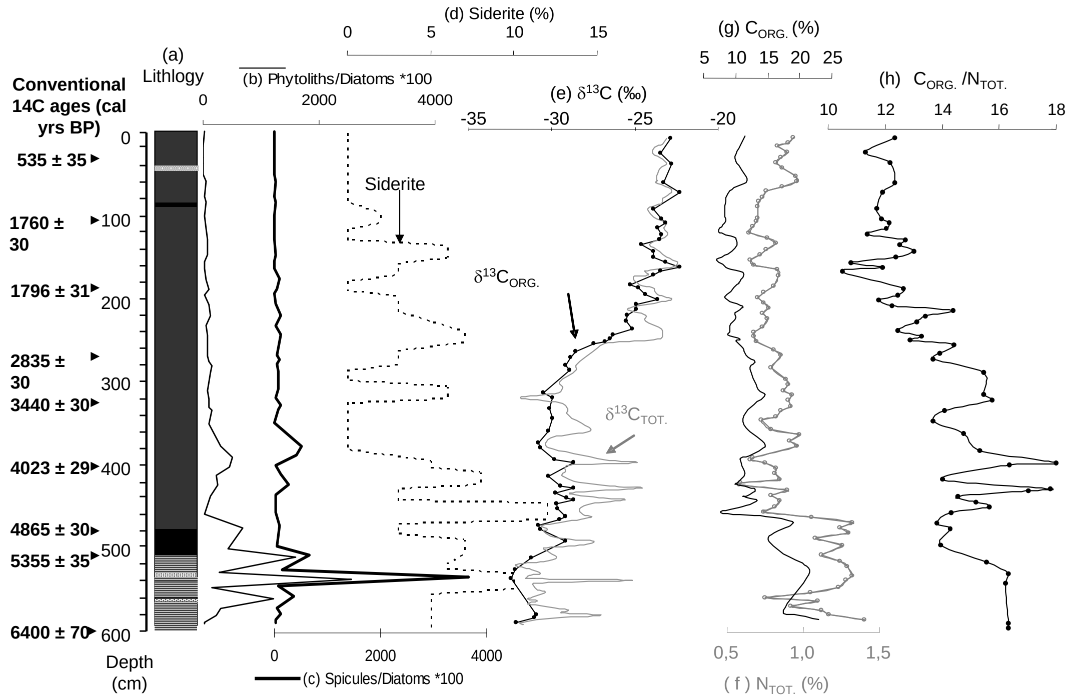

Fig. 3. Lithology of the core M4 modified after Ngos et al. (2008, 2011) radiocarbon ages **(a)** Phytoliths/Diatoms and Spicules/Diatoms ratios (**b** and **c**); siderite occurrence (% total crystals), **(d)**; total δ^{13}C ($\delta^{13}C_{TOT,}$ ‰) organic δ^{13}C ($\delta^{13}C_{ORG,}$ ‰) **(g)**, total nitrogen ($N_{TOT,}$ %, h) total organic carbon (C_{ORG}, %, g) variations and Corg/N_{TOT} **(h)** changes along the core.

3.2 Radiocarbon dates

The chronological control is based on nine AMS radiocarbon dates performed on total organic matter (Table 1). Four of the nine dates (indicated by stars) were already published and discussed in previous articles (Ngos et al., 2008; Vincens et al., 2010). The other five radiocarbon dates were processed at the "Laboratoire de Mesure du Carbone 14 (Salclay, France)" with the ARTEMIS AMS facility. The calibration of ^{14}C yr BP into cal yr BP was performed using the radiocarbon calibration program Calib Rev 6.0 (Stuiver and Reimer, 1993). Eight of the nine dates showed a good internal consistency as function of depth while one performed at 102 cm appeared older than expected (1760 $\pm$ 30 yr BP). From Ngos and Giresse (2011) recent study, we know that the volcanic activity of Lake Mbalang was insignificant and that of Lake Tizong located at 15 km west of Lake Mbalang has been of small radius suggesting CO_2 volcanic gases have not affected Lake Mbalang datings. Hence we suggest that older age at 102 cm cannot be attributed to low radiocarbon activity of volcanic CO_2. The older age at 102 cm may indicates an increase in sedimentation rate as it is observed in Lake Assom (Ngos et al., 2003) and possibly in Lake Tizong in the southern part of Adamawa between 1300 and 2800 yr BP. The lithology of the core did not show any particular unit that could indicate the changes of sedimentation, nevertheless the ratio quartz and plagioclase over kaolonite and gibbsite revealed an elevation of coarse elements in the core at 80–100 cm (Ngos et al., 2008; Ngos and Giresse, 2011) but the time resolution is not good enough to confirm the change. Here we consider the date older as a result of allochtonous or reworked organic material and consequently the date was excluded in constructing the age model. Assuming that no radiocarbon reservoir age affected the organic carbon of Lake Mbalang the remaining eight dates allowed a construction of a polynomial depth-age model (Fig. 4) in order to calculate by extrapolation the estimated age of each studied sample. This age model was also applied to recent periods (between 535 yr BP and the present) because ^{210}Pb analyses are not yet available for accurate calculation of sedimentation rate for that period of time. The polynomial regression intercept near the surface indicates an age of approximately 150 cal yr BP suggesting that reservoir age and volcanic CO_2 have little impact on the proposed chronology.

3.3 Diatom analyses

Diatom slides were prepared from ~0.5 g of dry sediment by gently heating in 30 % hydrogen peroxide (Battarbee, 1986) followed by several washings with distilled water. Few drops (0.2 ml) of the resulting residue suspended in distilled water were evaporated onto a coverslip, which was subsequently mounted on a glass slide with Naphrax^{TM}. At least 600 diatom valves were counted per sample or approximately

Table 1. Radiocarbon dates from the core M4.

Laboratory codes	Level (cm)	Material	Conventional 14C ages (cal yr BP)	Calibrated 14C dates (cal yr BP)	2-sigma calibrated 14C ages range (cal yr BP)	Relative area (probability)
Unknown	35	TOM	535 ± 35*	546	509–562	0.69859
					594–635	0.30141
SacA 18586	102	TOM	1760 ± 30**	1664	1567–1739	0.970552
					1757–1780	0.025613
					1803–1806	0.003835
Unknown	185	TOM	1796 ± 31*	1729	1922–1671	0.174361
					1688–1820	0.825639
SacA 18587	276	TOM	2835 ± 30	2939	2860–3007	0.949524
					3012–3036	0.035758
					3050–3061	0.0014719
SacA 18588	321	TOM	3440 ± 30	3698	3631–3780	0.826694
					3787–3828	0.173306
Unknown	407	TOM	4023 ± 29*	4481	4421–4536	0.949341
					4542–4549	0.015457
					4555–4568	0.035202
SacA 18589	481	TOM	4865 ± 30	5605	5490–5501	0.02962
					5583–5654	0.97038
SacA 18590	506	TOM	5355 ± 35	6139	6002–6084	0.310786
					6095–6218	0.555934
					6235–6274	0.13328
Beta 143097	600	TOM	6400 ± 70*	7333	7173–7222	0.06448
					7234–7432	0.93552

* Dates already published (Ngos et al., 2008; Vincens et al., 2010). ** Date not used in the age model.

200–300 valves when the diatom concentrations were too low. Counts were done at magnification 1000× with oil immersion objective ($na = 1.32$) using an Olympus BHT light microscope equipped with Nomarski optics. Diatom preservation was good throughout the core.

Identification and taxonomy of diatoms were based principally on Krammer and Lange-Bertalot (1986–1991), Gasse (1980, 1986), Germain (1981), Schoeman (1973), Simonsen (1987).

Ecological interpretations were essentially based on the modern data of Lake Ossa area (Nguetsop, 1997; Nguetsop et al., 2010) coupled with previously documented taxa preferences in other regions of Africa (Gasse et al., 1995; Servant-Vildary, 1978), for most including taxa counts and water-chemistry characteristics at sampling sites.

3.4 Stable carbon isotope analyses

For measurement of carbon stable-isotope content and C/N ratios, 106 samples were taken along the core at intervals varying between 2 and 16 cm with an average interval sampling of 6 cm. According to Ngos et al. (2008) inorganic carbon (IC) in M4 core is less than 1 % and X-ray diffraction (XRD) siderite (iron carbonate $FeCO_3$) estimates show variable relative amount along the core. Bahrig (1988) reported $\delta^{13}C$ values of siderite ranging from ∼+2 to +15 ‰ for the siderite $\delta^{13}C$ though more recent experimental studies of Jimenez-Lopez and Romanek (2004) show that isotopic fractionation of the siderite is close but lower than isotope fractionation factors for the calcite CO_2(gaz). Enriched $\delta^{13}C$ of carbonates may increase the bulk sedimentary $\delta^{13}C$ and preclude the interpretation of $\delta^{13}C$ in terms of organic bulk material. Hence, one should remove siderite by sample pretreatment with acid. However acid treatment methods affect the reliability of organic carbon (C) and nitrogen (N), and

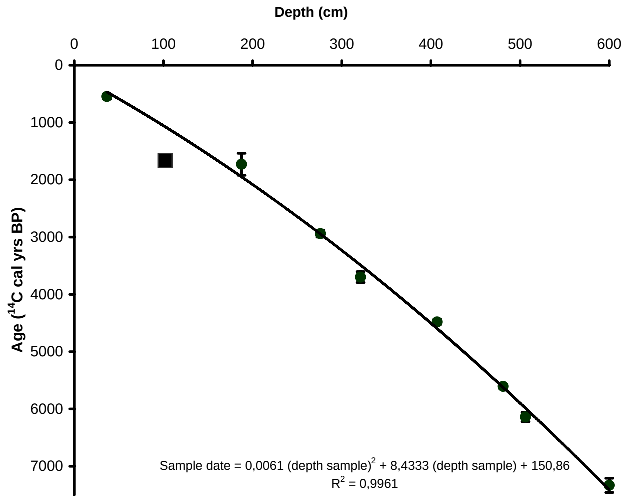

Fig. 4. Calibrated ^{14}C yr BP versus depth in the core M4. The black square represents the measure date that was excluded in the age model.

δ^{13}C values (Brodie et al., 2011). Hence, we proceeded to the analysis of δ^{13}C of 106 bulk sediment samples of the M4 core without treatment ($\delta^{13}C_{TOT.}$) and 63 among the 106 samples ($\delta^{13}C_{ORG.}$) were sub-sampled for acid-wash treatment with HCl (0.6N) to remove carbonates (referred as rinse method according to Brodie et al., 2011). Following the recommendations of these authors we included centrifugation steps to minimize the loss of fine colloidal components of the sample material (e.g. fine organic fragments and clays) and reduce potential biasing towards coarser grained fractions. The acid washed samples were rinsed three times with de-ionized water and centrifuged. Bulk sediment and acid-washed samples were dried at 50 °C for 48 h. About 1 cm^3 subsamples were ground using a mortar pilar and sieved through a 60 µm mesh. About 0.5 mg bulk sediment powder is weighed and introduced in tin capsules prior to elemental and isotope analysis. Nitrogen content results refer only to the bulk material.

Elemental C and N contents (%) and carbon isotope values of sediment were measured by dry combustion on a Euro Vector 3000 Elemental Analyzer coupled with a Micromass Optima Isotope Ratio Mass Spectrometer at ISEM laboratory (Montpellier). Elemental analysis of total carbon (C_{TOT}.), total organic carbon (C_{ORG}.) and total nitrogen (N_{TOT}.) and therefore C/N ratios were measured using the C and N contents of the alanine standard (C % = 40, N % = 16). The analytical precision of the N % and C % is about 1 %. The δ^{13}C results are expressed in delta (δ) notation where: δ (‰) $= [(R_{\mathrm{sample}}/R_{\mathrm{standard}}) - 1] \times 1000$ where R_{sample} and R_{standard} refer to the ^{13}C/^{12}C ratios of sample and standard, respectively. δ^{13}C values are reported in parts per thousand (‰) relative to the Vienna Pee Dee Belemnite (VPDB) standard. Precision for isotope measurements of chemical standards (Nist-8541 graphite international standards and alanine) within sample runs were better than 0.2 ‰.

4 Results and interpretations

A total of 98 species and varieties of diatoms were identified in the 48 studied samples of the core M4. Figure 5 shows the evolution of the most represented species (≥5 % in at least one sample). The ecological preferences of diatoms allowed the individualization of 2 major phases (with 6 sub-phases) in the paleohydrological evolution of the lake. Planktonic and tychoplanktonic diatoms were present throughout the core indicating that Lake Mbalang has never dried up (Figs. 6b, c, d and 7a). This assumption is reinforced by the fact that benthic, epiphytic and aerophilic diatoms remained consistently low along the core (Figs. 6e, f, g and 7a). Some diatoms that were recognised to be allochtonous were excluded from the diatom percentage calculations and sum (Fig. 5).

We measured $\delta^{13}C_{TOT.}$, $C_{TOT.}$ and $N_{TOT.}$ contents and $C_{TOT.}/N_{TOT.}$ ratios of 106 bulk samples and $\delta^{13}C_{ORG.}$, $C_{ORG.}$ and $C_{ORG.}/N_{TOT.}$ of 63 carbonate free samples of M4 core (Table 2 and Fig. 3e, g and h). Both curves of $\delta^{13}C_{TOT}$ and $\delta^{13}C_{ORG}$ show the same trend. However, between 400 and 600 cm, and 220 and 240 cm the $\delta^{13}C_{ORG}$ show more depleted values compare to $\delta^{13}C_{TOT}$. The isotopic difference between untreated and treated samples may reach 4 ‰ (426 cm) likely due to siderite reaching 10–12 % of mineral crystals at those depths (Fig. 3d).

According to Ngos and Giresse (2011) the organic carbon content fluctuates between 15 and 20 % in the lower two thirds portion of the sedimentary column before dropping, to 10 % in the upper part i.e. during the last 2500 cal yr BP. Our new data of $C_{TOT.}$ (6.5 to 21.5 %) and $C_{ORG.}$ (7.1 to 22.9 %) and N_{TOT} (0.6 to 1.4 %) confirm this change (Fig. 3g).

Lake sedimentary organic matter may results from a complex combination of sources: autochtonous organisms (freshwater food chain) and/or allochtonous material (terrestrial riverine and atmospheric inputs). Hence the isotopic composition of the lake organic matter may reflect a mixture of these diverse sources. Generally autochtonous lacustrine organic matter is characterised by relatively low C/N ratios, typically <10 (Meybeck, 1982) and terrestrial plants have high C/N ratios (>20 and may be >200; Hedges et al., 1986). Talbot et al. (1992) reported C/N ratios for plants collected near the Lake Bosumtwi varying between 26.4 and 156.3.

The M4 core C/N ratios of untreated and acid washed samples range from 9.7 to 17.9 and 10.5 to 18.0 respectively (Table 2) showing that $C_{ORG.}/N_{TOT.}$ ratios are slightly higher than $C_{TOT}./N_{TOT}$. In average, the $C_{ORG.}$ is 1.3 ± 06 % higher than $C_{TOT.}$ taking into account the precision of 1 %, we can assume that the difference is not significant. However, according to Brodie et al. (2011) the acid treatment using the rinse method can artificially elevate C values due to the loss of fine-grained materials (e.g. such as clays) and the removal of inorganic carbon. The highest C/N values are observed between 600 and 300 cm with a mean value of 14.0 ± 1.2 (untreated samples) and 15.6 ± 1.2 (treated samples). Between 300 cm and the top of M4 core, the mean

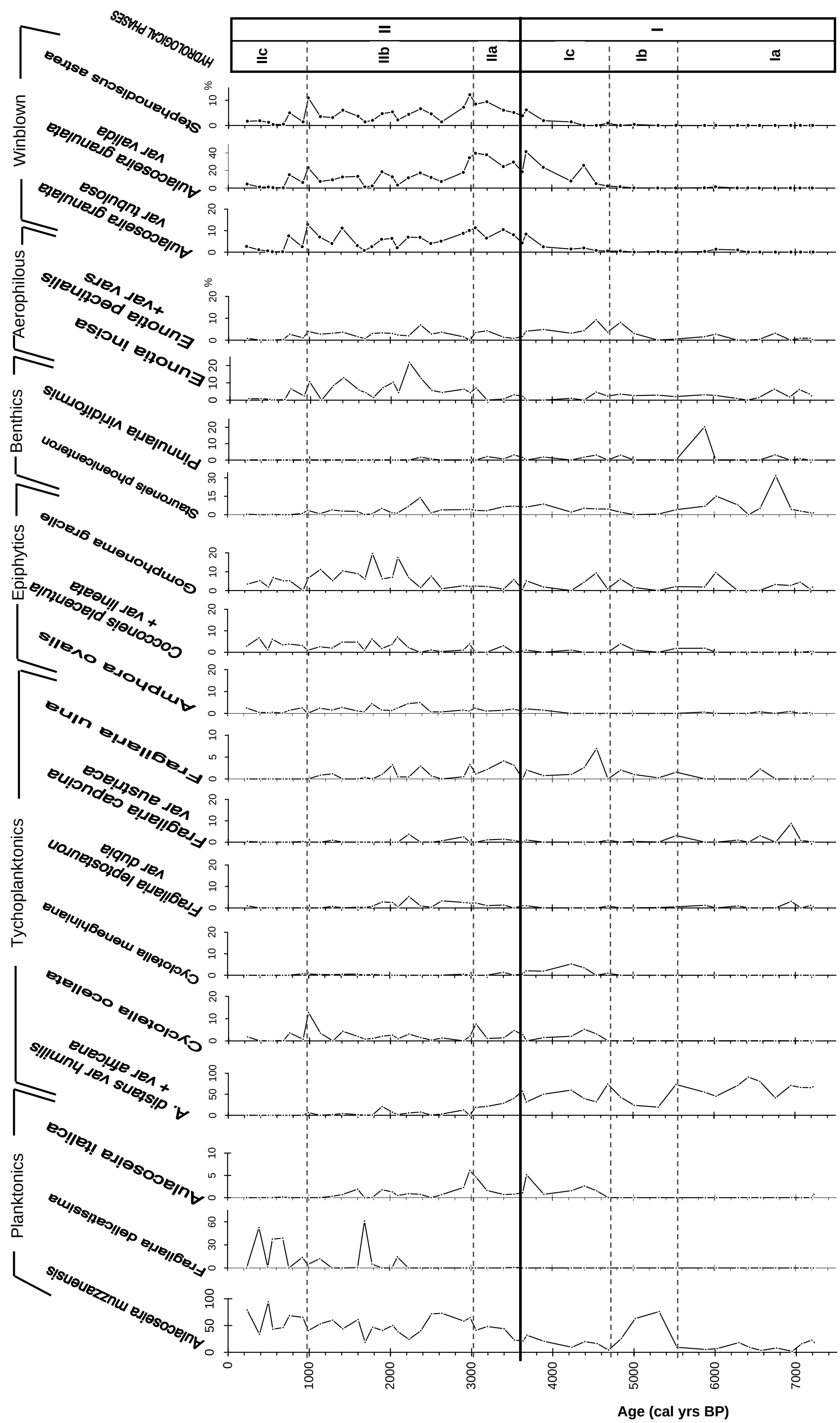

Fig. 5. Variation in abundances of the most dominant taxa (>5 % in at least one sample) belonging to different habitat groups and windblown diatoms over the core. Hydrological phases corresponding to diatom zones are indicated.

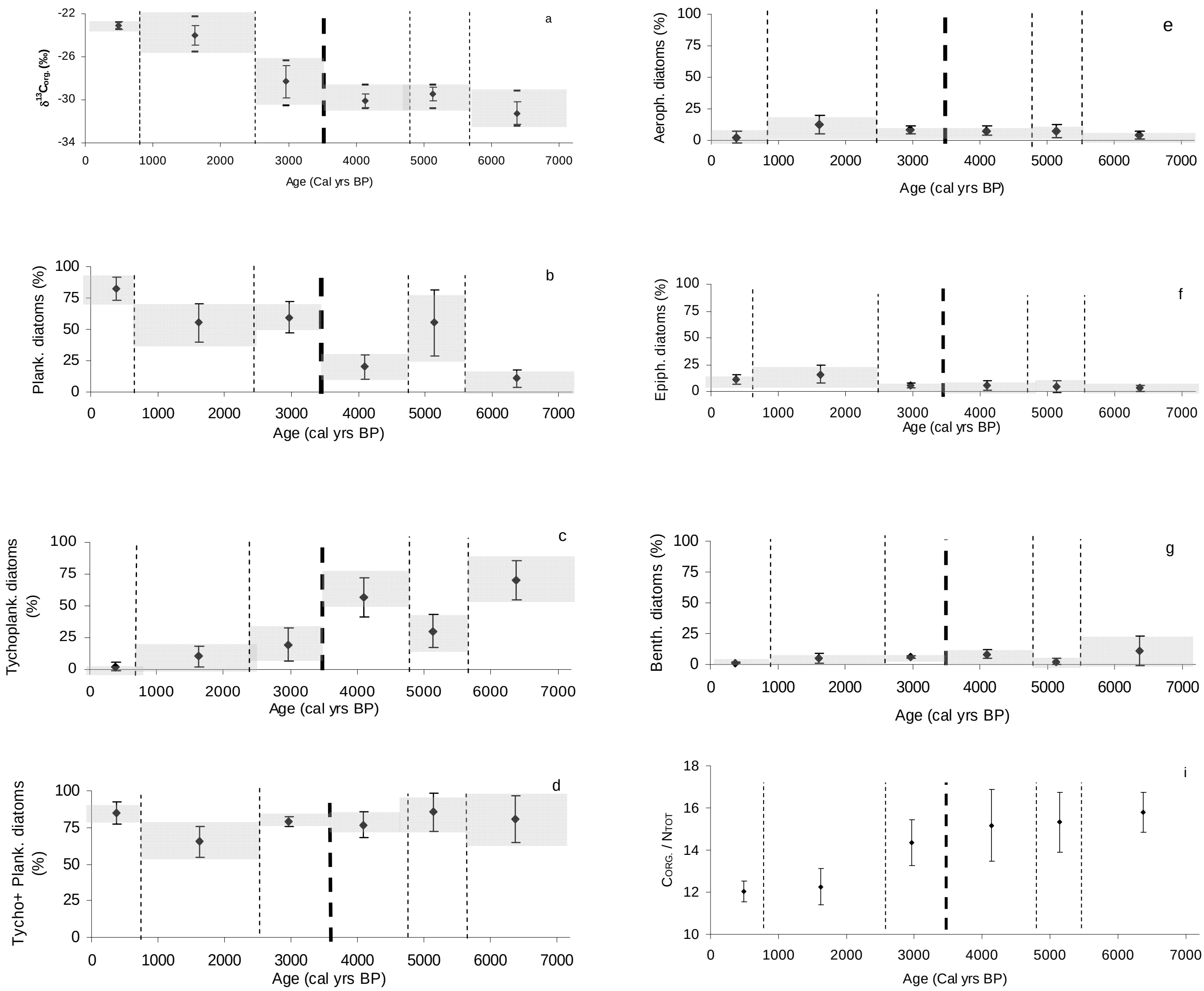

Fig. 6. Mean values and standard deviations of each diatom (**b** to **g**) habitat group (%) and δ^{13}C **(a)**. δ^{13}C extreme values are also reported. The dashed lines highlight the phase thresholds defined by diatom stratigraphy along the core.

C/N value decreases (11.0 ± 0.8 and 12.6 ± 1.1 for untreated and treated samples respectively). Do these values reflect a change in the proportion of terrestrial versus aquatic organic carbon inputs to Lake Mbalang? One can suggest that values of M4 core (from ∼12 to ∼15) may reflect a mixture of aquatic and terrestrial organic carbon and hence that the aquatic contribution tend to increase from the base to the top of M4 core. However, according to unpublished C/N data , a different interpretation for relatively low C/N ratios can be proposed. Modern soil of Central Africa in savanna area have been analyzed with a multi-proxy mean approach (see Fig. 2, Aleman et al., 2011). Measured C/N ratios vary from 10 to 19.3. These relatively low C/N ratios of terrestrial material show intermediate values of C/N are not necessarily linked to a mixture of aquatic and terrestrial organic carbon and that more study and data collection are necessary to understand the processes behind the C/N variations in plants, soils and lacustrine algae of Africa. Finally, erosional exposition and/or long distance transport of terrestrial organic matter can also increase the C/N ratios.

The analyses of $\delta^{13}C_{ORG.}$ along the core show two main phases and a transitional phase (3400 to 2500 cal yr BP): depleted values varying between −32.4 and −28.7 ‰ are observed between ∼7200 and ∼3400 cal yr BP (mean value of -30.2 ± 1.1 ‰) and enriched $\delta^{13}C_{TOC}$ values (∼−26.3 to −22.3 ‰) concomitant with Poaceae pollen increase (Vincens et al., 2010) for the last ∼2500 cal yr BP (Figs. 6a and 7d).

According to Roberts et al. (2001), high lacustrine primary production (mostly algae) may increase the lacustrine isotopic carbon composition values by up 15 ‰ due to the high CO_2 biological demand during photosynthetic

Table 2. Stable carbon and elemental analysis of M4 sedimentary core taken in Lake Mbalang.Total δ^{13}C ($\delta^{13}C_{TOT.}$, ‰) and total carbon (% $C_{TOT.}$) and nitrogen (% $N_{TOT.}$) refer to analysis made on the total sedimentary material (without acid treatment). $\delta^{13}C_{ORG.}$ and $C_{ORG.}$ (%) refer to analysis of the organic carbon fraction (after HCl treatment). C/N ratios are calculated for both total (C_{TOT}/N_{TOT}) and organic carbon ($C_{ORG.}/N_{TOT.}$).

Sample Number	Depth (cm)	Age (Cal Yr BP)	$\delta^{13}C_{TOT.}$	$C_{TOT.}$ (%)	$N_{TOT.}$ (%)	$C_{TOT.}/N_{TOT.}$	$\delta^{13}C_{ORG.}$	$C_{ORG.}$ (%)	$C_{ORG.}/N_{TOT.}$
1	10	234	−23.1	10.2	0.9	10.9	−22.84	11.50	12.3
4	16	289	−23.4	9.4	0.9	10.4			
6	20	321	−23.8	8.8	0.8	10.6			
9	27	387	−24.0	9.5	0.9	10.6	−23.44	10.13	11.3
12	34	440	−23.3	8.9	0.9	10.4			
15	40	495	−23.1	8.2	0.8	10.0	−22.80	9.98	12.2
18	46	552	−23.5	8.8	0.9	10.2			
21	56	644	−23.7	9.6	1.0	10.1			
23	60	679	−23.6						
24	62	700	−23.7	10.2	1.0	10.6	−23.28	11.82	12.3
27	68	755	−23.0	9.5	0.9	11.0			
30	74	811	−22.8	7.7	0.8	10.2	−22.29	9.02	11.9
34	81	873	−23.3	7.5	0.7	10.2			
37	86	921	−23.4	7.4	0.7	10.5			
40	93	988	−24.3	7.4	0.7	10.5	−23.91	8.26	11.7
45	106	1110	−23.7	7.5	0.7	10.7	−23.40	8.36	11.9
47	110	1155	−23.1	7.4	0.7	10.8	−23.18	8.40	12.2
50	116	1214	−23.3	6.9	0.7	10.2	−23.65	8.07	12.0
53	124	1287	−23.0	6.5	0.6	10.1	−23.39	7.35	11.4
56	130	1346	−23.5	8.6	0.8	11.2	−23.53	9.72	12.7
59	136	1413	−24.1	9.1	0.8	11.1	−24.64	10.27	12.5
62	144	1489	−23.5	8.6	0.8	11.4	−23.90	9.82	13.0
65	150	1557	−23.1	7.6	0.7	10.7	−23.91	8.83	12.4
68	157	1622	−22.5	6.6	0.7	10.1	−23.13	7.08	10.8
71	163	1688	−22.8	6.8	0.7	10.1	−22.31	8.04	11.9
73	167	1730	−24.4	8.0	0.8	9.7	−23.48	8.71	10.5
75	172	1780	−24.2				−23.89	11.18	
76	175	1811	−25.1	9.5	0.8	11.3			
78	183	1901	−24.7				−25.27	10.61	
80	187	1944	−24.0	9.2	0.8	11.4	−24.78	10.25	12.7
83	195	2027	−23.7	8.2	0.7	11.0	−24.36	9.24	12.4
86	201	2093	−22.8	7.4	0.7	10.6	−23.65	8.28	11.8
89	207	2159	−24.2	8.1	0.7	11.0	−24.91	9.07	12.3
92	213	2227	−25.5	8.6	0.8	11.2	−24.93	11.09	14.4
95	220	2301	−24.4	8.9	0.7	12.1	−25.45	9.87	13.4
98	227	2374	−24.4	8.9	0.8	11.5	−25.52	10.08	13.1
101	236	2482	−23.4	8.5	0.7	11.5	−25.13	9.17	12.4
104	243	2560	−23.4	8.0	0.7	11.8	−26.33	9.02	13.3
106	247	2611	−23.5	7.8	0.7	11.5	−26.47	8.71	12.9
108	252	2663	−25.6				−26.79	10.63	
109	254	2686	−26.8	8.4	0.7	12.0	−27.45	10.05	14.4
112	263	2793	−27.5	9.7	0.8	12.5	−28.56	11.18	13.9
115	271	2878	−28.0	10.6	0.9	12.4	−28.90	11.67	13.7
119	279	2982	−28.6				−29.19	13.07	
120	286	3056	−28.6	10.1	0.8	12.7	−28.94	12.20	15.5
123	299	3219	−29.5	11.7	0.9	13.2			
126	306	3297	−29.4	11.8	0.9	13.0			
129	312	3375	−29.6	11.7	0.9	14.0	−30.51	13.47	15.5
132	318	3451	−31.9	12.8	0.9	14.3	−29.96	14.64	15.8
135	324	3527	−29.5	12.3	0.9	14.2			
138	331	3607	−29.1	11.5	0.9	12.5	−30.15	12.93	14.1
141	337	3681	−29.1	11.2	0.9	13.0			

Table 2. Continued.

Sample Number	Depth (cm)	Age (Cal Yr BP)	$\delta^{13}C_{TOT.}$	$C_{TOT.}$ (%)	$N_{TOT.}$ (%)	$C_{TOT.}/N_{TOT.}$	$\delta^{13}C_{ORG.}$	$C_{ORG.}$ (%)	$C_{ORG.}/N_{TOT.}$
144	343	3764	−28.8	10.6	0.8	12.9	−29.99	11.29	13.7
146	348	3824	−28.2	9.2	0.7	12.8			
149	358	3954	−27.7	10.7	0.8	13.5	−30.20	11.67	14.7
152	365	4047	−29.9	11.3	1.0	11.6			
155	372	4135	−30.4	13.0	0.9	14.9	−30.82		
158	378	4215	−30.2	13.5	1.0	14.9	−30.69	14.66	15.3
162	392	4394	−26.9	10.2	0.7	15.9	−29.82	12.03	18.0
163	395	4432	−24.9	10.2	0.7	15.3	−28.67	10.65	16.4
165	399	4490	−29.1	10.6	0.8	14.8			
168	406	4574	−29.2	11.1	0.8	13.5			
171	412	4654	−29.3	11.0	0.8	13.5	−30.19	11.44	14.0
174	418	4745	−28.4	11.6	0.8	13.7			
176	423	4810	−26.5	10	0.6	17.9	−29.44	10.42	17.8
177	426	4845	−24.6	9.3	0.6	17.2	−28.66	9.84	17.0
180	432	4930	−28.9	12.1	0.9	13.5	−29.75	13.00	14.6
183	438	5015	−26.9	10.9	0.8	13.9	−29.08	11.95	15.2
184	440	5047	−25.6				−28.68	11.14	
186	444	5101	−27.6	11.7	0.8	13.8	−29.72	13.29	15.7
189	451	5192	−28.3	10.9	0.8	13.4	−29.67	11.67	14.3
192	457	5282	−28.1	10.1	0.7	13.7			
193	459	5310	−27.6				−29.19	7.71	
195	464	5372	−28.2	13.07	1.1	12.3	−29.56	14.66	13.8
198	470	5466	−30.6	17.11	1.3	13.0	−30.81	18.86	14.3
200	475	5530	−30.3	17			−30.66	18.46	
201	477	5559	−30.5	15.74	1.2	12.7			
204	483	5647	−30.1	16.33	1.3	12.6			
206	487	5705	−28.9						
207	489	5733	−27.5	13.97	1.1	12.9	−29.18	15.05	13.9
208	493	5791	−28.1						
210	497	5849	−30.5	17.04	1.3	13.5			
211	500	5885	−30.4						
215	509	6024	−29.7	14.85	1.1	13.3	−31.24	17.44	15.6
217	514	6090	−30.5						
218	516	6132	−32.2	17.13	1.2	13.8			
221	522	6221	−31.2	17.96	1.3	13.9	−32.20	21.06	16.3
223	527	6292	−30.7						
224	529	6323	−31.3						
226	534	6386	−31.5	17.93	1.3	13.6	−32.45	21.42	16.2
227	536	6424	−25.1						
228	539	6467	−31.1	17.69	1.3	13.8			
230	545	6565	−31.0						
231	547	6594	−30.9	17.82	1.2	14.5			
232	550	6627	−30.4						
234	554	6694	−31.0	16.1	1.1	15.3			
236	558	6757	−31.5						
237	560	6790	−30.8	10.61	0.8	14.1			
239	564	6851	−31.4	16.7	1.1	15.2			
242	571	6949	−29.0	13.19	0.9	14.3			
244	575	7009	−29.4	17.1	1.1	15.3			
245	576	7035	−27.0	16.4			−30.94	17.49	
246	578	7068	−27.1						
247	581	7102	−29.6	17.2	1.2	14.7	−31.07	19.11	16.3
250	586	7190	−32.0	21.5	1.4	15.3	−32.14	22.88	16.3
251	588	7221	−31.7						

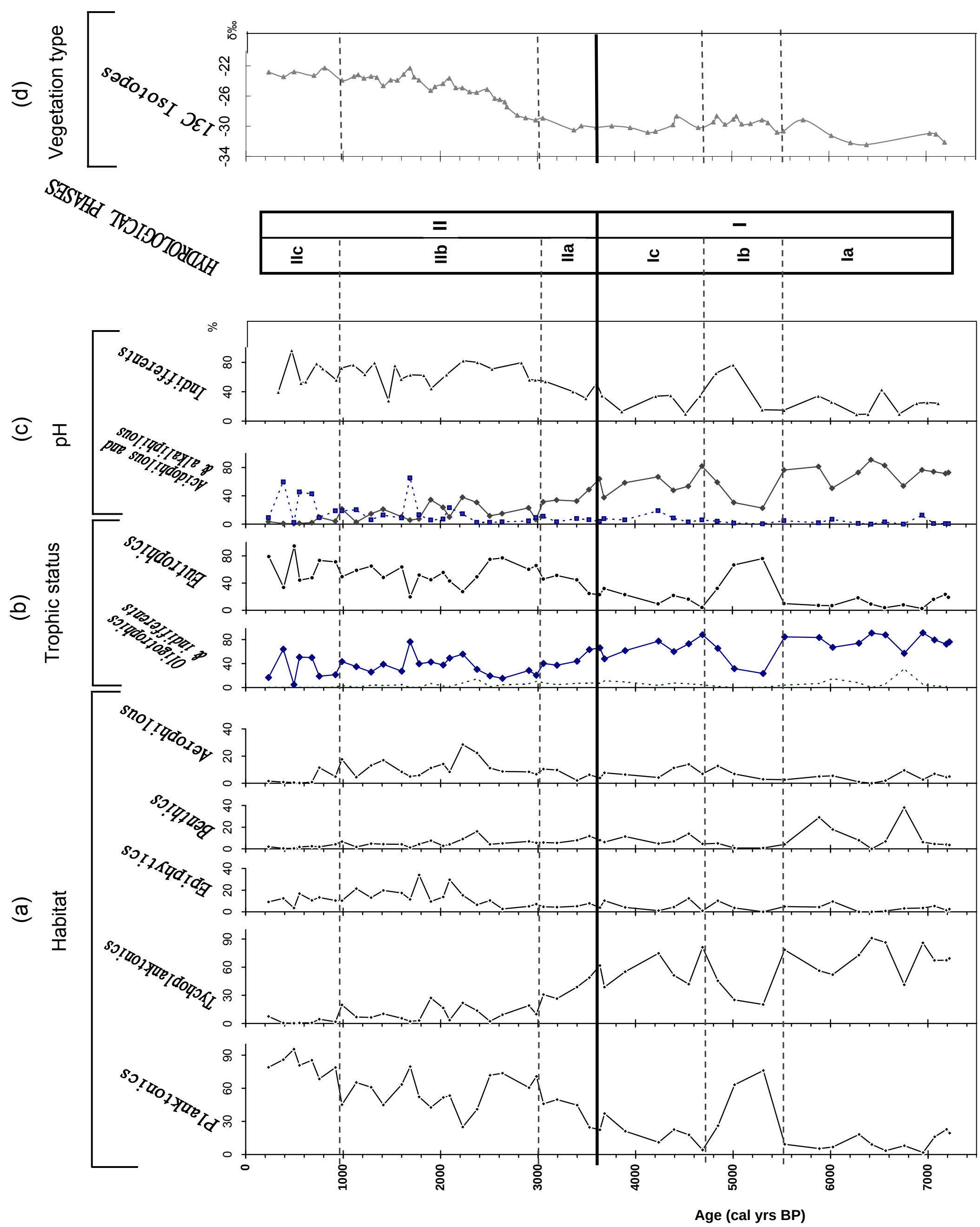

Fig. 7. Variations of habitat **(a)**, trophic status **(b)** and pH **(c)** groups over the core. Habitat: Planktonics, Tychoplanktonics, Epiphytics, Benthics and Aerophilous. Trophic status: Oligotrophics strong line, indifferent (dotted line) and eutrophic taxa. pH: Acidophilic (strong line), Alkaliphilic (dotted line) and pH-indifferent taxa. Modifications of vegetation type over the core **(d)**.

processes leading to isotopic disequilibrium of the lacustrine carbonate system. However, the organic carbon content shift observed by Ngos and Giresse (2011) and our C_{ORG} and C_{TOT} data do not support the hypothesis of a greater lacustrine productivity though eutrophic and planktonic algae show and increase. According to the analysis of literature results examining the complete range of δ^{13}C of benthic and planktonic algae on a global basis, France (1995) shows that freshwater benthic algae exhibit δ^{13}C values of -26 ± 3 ‰ and phytoplankton of -32 ± 3 ‰, an average difference of about 6 ‰. Thus assuming that this is true for diatoms of Lake Mbalang we compared the peaks of abundance of benthic and planktonic diatoms (Fig. 6g and b) of M4 core to the $\delta^{13}C_{ORG}$. We did not find covariations with benthic diatoms.

The two main peaks of benthic diatom abundances (>30 % at 557 cm and 498 cm) observed at the base of M4 core are not correlated to enriched carbon isotope ratios (<−28 ‰, Table 2) and along the M4 core planktonic diatoms increase (decrease) when $\delta^{13}C_{ORG}$ increases (decreases). These inverse covariations suggest that planktonic and benthic diatoms are not the main sedimentary carbon source.

We cannot exclude that enriched values from 3400 cal yr BP onwards may also reflect a mixture of terrestrial and freshwater organic matter. However we suggest that $C_{ORG.}/N_{TOT}$ ratios and carbon stable isotope ratios of M4 core may well be indicators of vegetation cover of the Lake Mbalang watershed. Abundance of the tychoplanktonic diatoms showed also an inverse covariation with the $\delta^{13}C_{ORG}$ (Fig. 6a and c) though the two proxies show differences when compared in details. $C_{ORG.}/N_{TOT}$ and $\delta^{13}C_{ORG}$ are negatively correlated along the M4 core (Fig. 3h). This covariation is not a sign of post depositional changes of the original isotopic characteristics of the primary organic matter but we will show thanks to this multi-proxy study that the concomitant increase of the $\delta^{13}C_{ORG}$ and Poaceae pollen evolution (Vincens et al., 2010) together with a decrease of the $C_{ORG.}/N_{TOT}$ ratios can be interpreted in terms of paleolimnological and paleovegetation variations forced by climatic changes.

4.1 Phase I: between 7200 and 3500 cal yr BP

The diatom flora of the lake was dominated by the oligotrophic, acidophilous tychoplanktonic diatoms represented essentially by *Aulacoseira distans* var. *humilis* and *A. distans* var. *Africana*. These taxa were reported in several tropical swamps and swampy lakes from East Africa as Lake Kioga (Uganda) and in the swamps of Bangweulu (Zambia). In Lake Kioga (altitude 1036 m) they can represent up to 75 % of the plankton samples (Gasse, 1986). From the study of the modern diatom and associated water characteristics of Saharan/Sahelian waterbodies, they were encountered in cold stratified water conditions (Gasse, 1987) although they generally prefer warm water conditions (Gasse, 1986). High percentages (40–80 %) of these taxa are encountered in the modern data set of Adamawa in bottom mud of lake borders occupied by aquatic vegetation. In swampy locations, dominated by sedges and Poaceae, their abundance was relatively high (28 %) (Kom, 2010). We thus inferred that these species are characteristics of low to high water depth and stable water table that can be occupied by aquatic vegetation or not. They also indicate oligotrophic and acidophilous waters.

From 7200 to 5500 cal yr BP (subphase Ia): high abundance of tychoplanktonic species *Aulacoseira distans* var. *humilis* and *A. distans* var. *africana* (41–91 %) suggest a generally acidic, oligotrophic and relatively stable or less mixed water table. Alkaliphilous tychoplanktonic taxa (*Cyclotella ocellata* and *C. meneghiniana*) remained consistently low except at the end of the subphase. Planktonic diatoms represented mainly by *A. muzzanensis* were also present but exhibit relatively low abundance; their highest abundance in this sub-phase is observed between 7200 and 6300 cal yr BP (18–23 %). *A muzzanensis* is considered as an eutrophic (Hustedt, 1927–1966; Cholnoky, 1968), planktonic taxa (Shoeman, 1973), encountered in the plankton of lakes and great rivers (Hustedt, 1930; Krammer and Lange-Bertalot, 1991) but it can also occur in some lakes in shallow turbid waters. Their presence can thus be interpreted in this sub-phase as a result of relatively high water depth and more mixed eutrophic water table, conditions that can be observed during the dry season when the north-eastern dry winds are preponderant. Hence, we can infer from the two previous groups during this subphase, a generally moderate to high lake level that can be mixed episodically. Benthic diatoms represented mainly by *Stauroneis phoenicenteron*, *S. anceps* var. *gracilis* and *Pinnularia viridiformis* were more important in this subphase, they peaked between 6900 and 6600 cal yr BP (up to 38 %) and at 6300–5900 cal yr BP (up to 29 %); their high abundance suggests periods of clearer water column or at least episodic lowering of lake level. The hypothesis of lake level lowering is also suggested by the presence of sand in the lowermost part of the core along with abundant phytoliths and spicules but the low abundance of epiphytic diatom taxa excluded a very low lake level where the lake basin could have been occupied by dense macrophytic vegetation. $\delta^{13}C_{ORG.}$ values are low with a mean value of -31.2 ± 1.1 ‰ consistent with a C3-dominated terrestrial flora. Some very low $\delta^{13}C_{ORG.}$ values (−32.5 ‰) may also be due to the presence of plant material influenced by the isotopic effects of a dense, closed canopy forest that developed at that time. This phase is also characterized by $\delta^{13}C_{TOT}$ shifts towards higher values: One peak (−27.1‰) covaries with one of the major sand layer evidence described at 580–560 cm and the two other peaks at 508 and 534–535 cm (∼ −25 ‰ and −27 ‰, respectively) covarying with the Phytoliths/Diatoms and Spicules/Diatoms ratios (Fig. 3). After acid treatment, the peaks disappeared suggesting siderite influences the total sedimentary carbon isotopic ratios. These results and the absence of covariation with the benthic diatoms reinforced the hypothesis of episodic lowering of lake level and the presence or the vicinity of the aquatic vegetation and important terrestrial organic matter input as supported by Ngos et al. (2008) and Ngos and Giresse (2011). Though epiphytic diatom abundance (*Amphora ovalis*, *Cocconeis placentula* and varieties and *Gomphonema gracile*) remained consistently low, the hypothesis is nevertheless supported also by high values of total organic carbon (Ngos et al., 2008). The mean $C_{ORG.}/N_{TOT.}$ ratio is 15.8 ± 1.0.

From 5500 to 4800 cal yr BP (subphase Ib): planktonic diatoms represented mainly by *A. muzzanensis* increased markedly and reached 63–76 % abundance while tychoplanktonic diatoms decreased. This suggests an increase of lake level and/or a well mixed water table. Benthic and

Table 3. C/N ratios and δ^{13}C of modern soils sampled in the Ombella-Mpoko and Lobaye provinces of Central African Republic (see the sampling field design in Aleman et al., 2011). These data were obtained in the frame of the ANR - ERA Net BIODIVERSA COFORCHANGE Project. The region is a rainforest progressively giving way to savannas, on ferralsols under a typical tropical climate consisting in an alternation of a dry season from November to February followed by a 8-months long wet season, with about 1500 mm annual precipitation (Bangui weather station, FAOCLIM, 2005).

Sample Number	$\delta^{13}C_{Bulk}$	$C_{Tot.}/N_{Tot.}$
1	−18.1	16.0
2	−18.2	16.6
3	−17.7	15.7
4	−18.6	15.2
5	−15.8	14.7
6	−17.8	17.5
7	−18.1	17.0
8	−18.6	18.1
9	−25.7	14.2
10	−24.7	14.0
11	−19.0	19.3
12	−18.1	
13	−16.8	
14	−23.0	15.3
15	−26.7	12.1
16	−27.2	12.2
17	−27.6	10.1
18	−28.3	

epiphytic diatoms nearly disappeared; aerophilous taxa (*Eunotia incisa* and *E. pectinalis)* exhibit very low abundance (∼3 %). The $\delta^{13}C_{ORG.}$ values vary from −30.8 to −28.7 ‰ between 470 cm and 423 cm. The mean $\delta^{13}C_{ORG}$ value for this period is −29.5 ± 0.7 ‰ and the mean $C_{ORG.}/N_{TOT.}$ ratio is 15.3 ± 1.4. At those levels, we also noticed the effect of carbonate siderite on the stable isotopic composition of the total carbon. This episode of slight increase of $\delta^{13}C_{ORG}$. compared with the previous period may be related to a slightly higher photosynthetic activity of eutrophic algae (Hollander and McKenzie, 1991; Law et al., 1995) and/or Poaceae taxa (Vincens et al., 2010). We suggest that during this time, the lake level was generally high, nevertheless, episodes of wind stress comparable to present day's dry season were longer or more severe than before. Consequently the lake experienced low level episodically, but benthic and epiphytic taxa could not develop due probably to a mixed, turbid water column. The high lake level can be explained by high and probably well distributed rainfall over the year that allowed the maintenance of forest vegetation as shown by $\delta^{13}C_{ORG.}$ data and the presence of savanna patches.

From 4800 to 3500 cal yr BP (subphase Ic): planktonic diatoms decreased significantly, tychoplanktonic species rose (up to 75–81 %) then showed a decreasing trend with short (centennial) spells of very low abundance towards the end of the sub-phase. This may indicate a slight lowering of lake level and probably a clearer, less turbid water column also evidenced here by the increase of both benthic, epiphytic taxa (12 %) at 4500 cal yr BP and aerophilous taxa (7–13 %) at 4800–4500 cal yr BP. This is also attested by a slight increase of spicules and phytoliths in samples which confirm the development of aquatic vegetation closer to the coring site. During this period, $\delta^{13}C_{ORG}$. background signature remained consistently low (−30.8 to −28.7 ‰). Mean $\delta^{13}C_{ORG}$ and $C_{ORG.}/N_{TOT.}$ ratios are −30.1 ± 0.7 ‰ and 15.2 ± 1.5 respectively suggesting similar environmental conditions as during the previous period. However, this subphase is also characterized by the appearance of *A. granulata* var. *valida, A. granulata* var. *tubulosa* and *Stephanodiscus astraea.* Although these taxa are typical planktonic species, they should be interpreted with caution because in Lake Ossa area in southern Cameroon (3°50′ N, 9°36′ E), it was shown based on their bad state of conservation, their distribution and their abundance in the lake modern sediment samples and in the uppermost layer of soils under the forest surrounding the lake, that they are originated from the Saharan diatomite deposits (Nguetsop et al., 2004). Moreover, recent analyses of modern sediments from a dried wetland (Ndjombi Swamp, near Kika SE Cameroon) revealed the presence of a comparable assemblage of taxa while other diatoms where completely absent. Hence their abundance in lake sediments was interpreted as an intensification of NE trade winds that are preponderant in Adamawa during the boreal winter rather than water depth or water trophic status changes. We can hypothesised that, the appearance of these taxa in Lake Mbalang marked as in Ossa area an intensification at least episodically of the NE trade winds.

4.2 Phase II: between 3500 and 0 cal yr BP

Planktonic diatoms, dominated by eutrophic *A. muzzanensis* indicated a high lake level and well mixed water. Tychoplanktonics declined significantly during this period and nearly disappeared. Windblown diatoms were consistently present, even if their abundance showed important fluctuations. $\delta^{13}C_{ORG.}$ values (ranging between −30.5 and −22.3 ‰) increase from 3500 to 0 cal yr BP. This increase is concomitant with the increasing proportion of C4 plant vegetation in the landscape cover and consequently its $C_{ORG.}$ contribution to the lake sediment.

From 3500 to 2800 cal yr BP (Subphase IIa): eutrophic diatoms characteristic of well mixed layer increased and reached about 70 % at 2600–2500 cal yr BP. *A. distans* var. *humilis* and *A. distans* var. *africana* which are indicators of the stability of water table decreased markedly. Mean $\delta^{13}C_{ORG.}$ values remained relatively low (−29.5 ± 0.7 ‰)

and the mean $C_{ORG.}/N_{TOT.}$ ratio does not change significantly (15.1 ± 0.9) suggesting that sources and proportions of organic matter have not changed yet $C_{ORG.}$ content (mean value 11.5 %) is lower compared to previous periods. The persistence of windblown diatoms showed an intensification of windiness on the lake environment. This phase marked an unequivocal change of climatic conditions in the area; from relatively more stable or less mixed water table reflecting probably the stability of the air at low layers of the atmosphere to more mixed water table linked to a reinforced seasonality.

From 2800 to 800 cal yr BP (Subphase IIb): This phase is marked by high fluctuations in abundances of planktonic taxa at plurisecular timescale. Although *A. muzzanensis* dominates throughout the sub-phase, *Fragilaria delicatissima* became more important and peaked at 2100 cal yr BP (14 %), between 1800 and 1700 cal yr BP (5–61 %) and between 1100 and 900 cal yr BP (5–14 %). Contrarily to *A. muzzanensis*, *F. delicatissima* is considered as an oligotrophic to mesotrophic taxa (Kammer and Lange-Bertalot, 1991). Lowest abundances of planktonics are observed at 2400–2200 cal yr BP, 1900, 1400 and 1000 cal yr BP. In these levels, the epiphytic (*Gomphonema. gracile, Amphora ovalis, Cocconeis placentula* and its variety *lineata)* and aerophilous (*Eunotia incisa* and *E. pectinalis* var. *minor*) diatoms increased, indicating a lowering of the lake level at least at seasonal or interannual timescales. The relatively high abundance of windblown diatoms indicated the maintenance of the influence of the north-eastern trade winds in the Lake Mbalang environment. The development of *F. delicatissima* when windblown diatoms are low indicated probably a less mixed water table and/or a slight increase in lake level. This idea is reinforced by the fact that epiphytic, benthic and aerophilous taxa are very low. The sub-phase represents probably the period of time during which short time maximum climate variability occurred. This variability is roughly reflected on $\delta^{13}C_{ORG.}$ with fluctuation of ∼2 ‰ amplitude and even smaller after 1500 cal yr BP (−24.6 to −22.3 ‰). These relatively high $\delta^{13}C_{ORG.}$ values and $C_{ORG.}/N_{TOT.}$ ratios (varying between 10.5 and 14.4; mean value 12.4 ± 0.9) suggest yet terrestrial organic matter input suggesting the maintenance of C4 plants in Lake Mbalang environment. Very high biological activity of eutrophic algae may be responsible of such enriched $\delta^{13}C_{ORG.}$ however, the decreasing $C_{ORG.}$ trend along the core does not support this hypothesis.

From 800 to 0 cal yr BP (Subphase IIc): high abundance of planktonics indicates a persistence of high lake level. The two main planktonic species alternated at this level, the change from *Aulocoseira* to *Fragilaria* dominated assemblage in diatom community is interpreted as the changing to more clear water column or shallowing, reduced mixing when P:E is low (Stager and Anfang-Sutter, 1999). The substantial decrease of windblown taxa supports the inference for more stable water column. This may also indicates important changes in water trophic status. Among other taxa, only *Gomphonema gracile, Cocconeis placentula* and its variety *lineata* remained present with percentages close to those of the precedent zone. The $\delta^{13}C_{ORG.}$ values and $C_{ORG.}/N_{TOT}$ ratios were similar to the end of the previous sub-phase suggesting yet the maintenance of C4 plants.

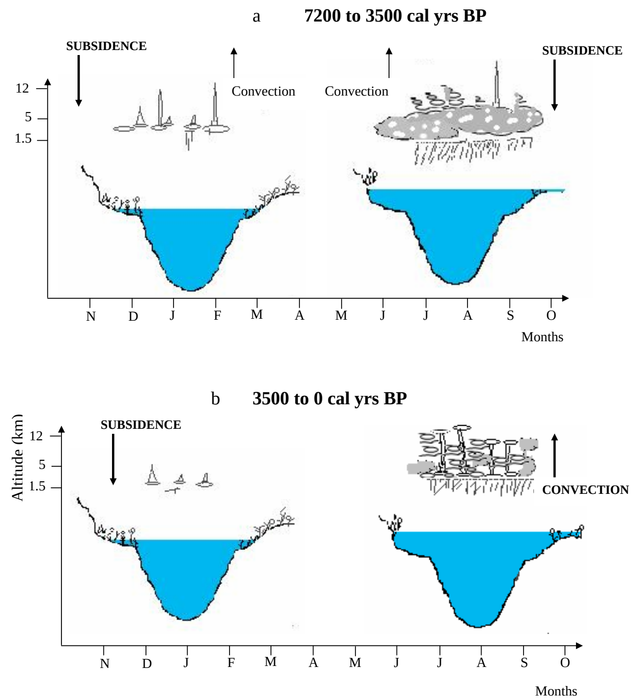

Fig. 8. Sketch of atmospheric features (clouds cover and air movement) and relative modifications of Lake Mbalang level, in the dry season (January) and rainy season (August) before 3500 cal yr BP **(a)** and afterwards **(b)**. Before 3500 cal yr BP, stratiform cloud cover were abundant, convective cloud are dominant after 3600 cal yr BP.

5 Discussion

The variations of the abundances of planktonic and tychoplanktonics can be considered as indicators of lake level changes (Fig. 9a), although the curve should be interpreted with caution because these organisms can also thrive in large free water surface. Acidophilous oligotrophic and tychoplanktonic *Aulacoseira distans* var. *humilis*, *A. distans* var. *africana* and planktonic taxa *Fragilaria delicatissima* are characteristic of stable or less mixed water table, which presupposes also a relatively stable air layer over the lake. During the boreal summer, a deep atmospheric convection (zone C of the cross-section of the troposphere over tropical Africa (after Leroux, 1970, 2001), entails heavy rains that directly cool waters of the surface. In these conditions also characterised by heavy clouds and subsequent reduced solar radiation inputs, thermal differences between epilimnion and hypolimnion are reduced and finally mixing occurred.

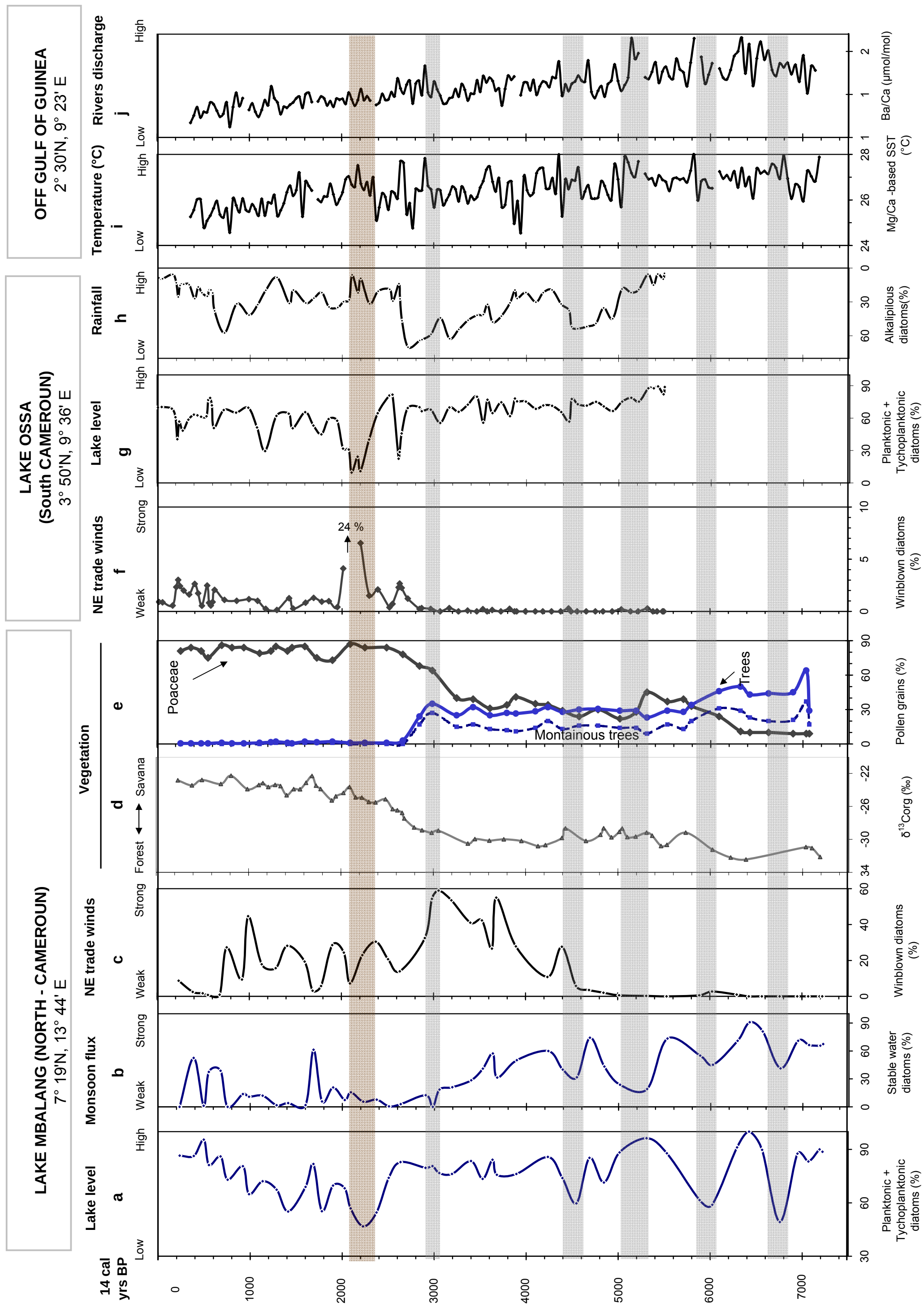

Fig. 9. Comparisons between Lake Mbalang (North-Cameroon), Lake Ossa (South-West Cameroon) and Gulf of Guinea. Lake Mbalang level variations evidenced by relative abundance of Planktonics + Tychoplanktonics **(a)**, Monsoon flux intensity reflected by stable water diatoms, higher percentages correspond to more intense monsoon flux **(b)**, NE trade winds (Harmattan) intensity, higher allochtonous diatom abundance indicates more intense Harmattan **(c)**. Changes from C3 to C4 dominant plants in vegetation is evidenced by δ^{13}C of sedimentary organic matter **(d)**, also shown by palynological data **(e)** (Vincens et al., 2010). Variations in NE trade winds (Harmattan) **(f)** and lake level **(g)** are shown in Lake Ossa as well as relative change in rainfall evidenced from alkaliphilous diatoms **(h)** (Nguetsop et al., 2004). Variations in temperature off Gulf of Guinea is shown from Mg/Ca based SST **(i)**, Rivers discharge based on ration Ba/Ca is also shown **(j)** (Weldeab et al., 2005, 2007).

However, if in the past the convective zone moved farther north than today, the Adamawa plateau could have been subjected to a climate that is described by Leroux in zone D where subsiding air masses present at mid-levels of the atmosphere generate stability at low levels. Consequently, the weather is cloudy and rainfall strongly reduced in the form of light rain and drizzle. In these conditions, evaporative heat loss may be suppressed or reduced, and surface warming during this period of low wind stress is likely to cause more stability in the water column (Kling, 1987). Hence, high abundance of the two species in the past can suggest conditions close to those observed in the boreal summer when the ITCZ is farther north, which entail the stability of the water column and/or the development of aquatic vegetation. Conversely, the planktonic *Aulacoseira muzzanensis* and *Aulacoseira granulata* thrive better in well mixed water tables that are associated to high temperatures, intense storms and windiness. These conditions are observed nowadays, mostly during the boreal winter in the Adamawa plateau and entail a deeper and unique thermocline in the water table (Kling, 1987). Such large diatoms have also been used as indicators of water table mixing in east African lakes (Stager et al., 1997). The variations in the intensity of the NE trade winds are inferred, as in Lake Ossa, from relative abundance of windblown diatoms (Fig. 9c). We suggest that the mixing is mostly due to the intensification of the north-eastern trade winds (Harmattan) during the year, although crater lakes of the Cameroon volcanic line show high volume/surface ratios and are relatively sheltered from winds.

Paleoclimatic data suggest that tropical Africa experienced during the Holocene important paleoclimatic changes that are now well dated (Servant et Servant-Vildary, 1980; Gasse, 2000). The base of the core M4 (7200 cal yr BP) belongs to the African humid phase that is documented in several continental sites (e.g. Gasse, 2000; Talbot and Johanessen, 1992; Stager et al., 1997) and marine sites offshore Africa.

5.1 Middle to late Holocene: from 7200 to 3600 cal yr BP

Diatoms data of Lake Mbalang inferred a stable water table that may indicate a stronger monsoon flow. These data are consistent with appearance of mountain forest taxa pollens in the palynological spectrum. The two most abundant taxa *Olea capensis* and *Podocarpus sp* were probably developed on nearby mountains that are today covered by shrubby savannas dominated by *Hymenodictyon floribundus* (Vincens et al., 2010). The nearest modern ecological niche of these two taxa according to Letouzey (1968, 1985) is located at Mount Ngan-Ha (1923 m), some 35 km east of Lake Mbalang. These species are also present some 300 km north of the lake at Mount Poli (7°50′ N; 2049 m) and at Tchabal Mbabo highlands (7°18′ N, 2460 m) located 165 km west of Ngaoundere on the Cameroon volcanic line. In fossil records, *O. capensis* and/or *Podocarpus* sp. occurrences in several locations in the northern subtropics and subequatorial areas of Africa (Salzmann et al., 2002) and especially during the Last Glacial Maximum were interpreted as indicative of cooler air conditions during a longer period of the year linked to stratiform cloud cover that are observed today only during the boreal summer when upwelling system is reinforced off the Gulf of Guinea (Maley and Brenac, 1998). But this hypothesis is less likely during the Holocene because marine isotopic data off the Gulf of Guinea showed no evidence of past strong upwellings system at that area (Weldeab et al., 2005, 2007). Another alternative is to consider episodic cold air mass advections of middle and high latitudes that can also contribute to such air conditions, but the weakness of this hypothesis is shown by the absence of such occurrences in the Saharan/Sahelian regions during this period (Servant and Servant-Vildary, 1980). If the climatic determinism is the same as today, their abundance in Adamawa fossil spectra should imply a northward displacement of ecological boundaries as shown by palynological data (Watrin et al., 2009; Lezine, 2009) and reproduced by vegetation models (Hély et al., 2009). Diatoms in Lake Mbalang inferred a moderate to high lake level which can correspond to precipitation lower than today in a context of low evaporation because of the far northern position of the ITCZ, but precipitation distribution remained favourable for forest development as shown by palynological and $\delta^{13}C_{ORG.}$ data. Although sponge spicules and phytopliths were relatively abundant, low epiphytic and benthic diatoms abundance showed that water level was not strongly reduced. It is possible that these phytoliths were from a more important belt than today of ligneous tree fringing the lake (*Alchornea* sp.) during this period of relative low evaporation and high water content in soils as is observed in other sites of central Africa (Ngomanda et al., 2009b).

From 7200 cal yr BP onwards, the decreasing trend of diatoms characterising the stable water column is punctuated by several abrupt low abundance at 6700, 5800–6000, 5000–5300, 4500 and 3600 cal yr BP (Fig. 9b) corresponding probably to episodes of weaker monsoon flux superimposed on the general trend, showing the complexity of climate change towards late the Holocene drier conditions. This pattern is reflected in water balance and vegetation landscape in several areas of tropical Africa and was largely discussed to underline the timing and magnitude of climate change from one region to another and associated climatic mechanisms (Gasse, 2000). The abrupt dryness of the climate corresponding to the end of the African Humid Period (AHP) shown by marine data off Mauritania (de Menocal et al., 2000; Adkins et al., 2006) is close to 5000–5300 cal yr BP (420–470 cm) low spell of tychoplanktonics observed in Lake Mbalang (Fig. 9b). This event is interpreted in Lake Mbalang as a period of increased mixing of the water table forced probably by north-eastern trade winds of the dry season. Therefore, the more development of Poaceae in Lake Mbalang area at that time can be explained by increased seasonality with a longer dry season compared to the previous period rather than an absolute decrease of rainfall as suggested by Vincens et al. (2010). Lake Ossa located south of Adamawa (3°50′ N) experienced convective rainfall in agreement with our model (Fig. 9h).

The spell dated at 4500–4000 cal yr BP corresponds probably to the most documented climatic phase throughout Africa; it was recorded at several sites of both southern and northern tropics (Servant and Servant-Vildary, 1980; Gasse, 2000). Drier conditions are also registered both by palynological, limnological and/or sedimentological data in subtropical latitudes of western Africa in Biu plateau (12°32′ N) and around lake Sele (7°9′ N) after ∼3800 yr BP (Salzmann et al., 2002, 2005) with the opening of the Dahomey Gap in the rain forest belt. In sub-equatorial regions this period was marked in Lake Bosumtwi by a low lake level at about 4000 yr BP (Talbot and Delibrias, 1980) although recent data did not confirmed this low stand (Russell et al., 2003). In central African subequatorial regions, proxy data inferred important disturbances in the periphery of the equatorial rain forest belt with possible appearance of included savannas (Ngomanda et al., 2009a, b) and complete dryness of lakes as Lake Sinnda in south Congo by 4400 yr BP (Vincens et al., 1994; Bertaux et al., 2000). In inner forest block, lakes were less affected by this climatic change (Vincens et al., 1999; Ngomanda et al., 2007; Kossoni et al., 2009). This period is characterised in Lake Mbalang by the maintenance of indicators of stable water table in agreement with the palynological and $\delta^{13}C_{ORG.}$ data, and thus to a stronger monsoon. But the appearance of windblown diatoms (∼4400 cal yr BP) attests probably the beginning of the aridification of the Sahara and/or the intensification of the NE trade winds (Fig. 9c).

Despite the scarcity of paleoclimatic records on highlands, the Bambili (western Cameroon) core provided a 24 000 yr time series that highlighted the comprehension of paleoclimatic evolution around the Gulf of Guinea. Contrarily to lowlands, Lake Bambili registered a dramatic low lake level from 10 000 to 7000 cal yr BP, then fluctuated around this low value afterwards (Stager and Anfang-Sutter, 1999) while other sites of tropical Africa underwent the so called “African humid period”. In Lake Njupi located north of Bambili at 1020 m altitude, *Olea. capensis* and *Podocarpus* sp. were present until around 3000 yr BP, suggesting a comparable evolution as the Adamawa plateau. Thus highlands as Bambili (2264 m altitude) may probably have evolved differently during greater part of the Holocene in term of water balance as suggested by Stager and Anfang-Sutter (1999), however synchronous evolutions between lowlands and highlands seems to have started at 3000 cal yr BP. Lake Mbalang evolved like lowlands in term of the pattern of change even though the palynological and hydrological signals seem to have been also controlled by altitudinal and meridian variations of climatic factors. It is thus possible that lowlands and highlands below 1200 m altitude like the Adamawa plateau were under conditions characterised by an important cloud cover during a greater part of the year while highlands such as Bambili were submitted to drier climate over the year.

5.2 The Late Holocene (last 3600 cal yr BP)

After 3600 cal yr BP, diatoms and other proxies of Lake Mbalang inferred significant changes of the climatic conditions. High abundance of *A. muzzanensis* and *A. granulata* suggest a more mixed water layer and a deeper thermocline. These conditions prevail today during the boreal winter. The lake level remained relatively high after 3000 cal yr and decreased between 2400–2100 cal yr BP. The other relative lowstands are dated at 1800 and 1400 cal yr BP, time after which the lake started its evolution towards present day’s high level (Fig. 9a). The windblown diatoms remained relatively important consistent with a significant influence of the NE trade winds during the year responsible of a well mixed water table. Nevertheless, the diatom derived lake depth reflects limnological variations and consequently water balance at centennial to millennial timescales. The relatively higher abundance of epiphytic, benthic and aerophilous mixed with planktonic and tychoplanktonic diatoms in individual samples reflects the lowering of lake level at the interval of time represented by one sample (∼6 yr) or could reflect seasonal variability. In that case, one can hypothesise in such climatic conditions the development of planktonic diatoms during the rainy season high lake level and development of littoral forms during the dry season at the lake borders on Cyperaceae (sedges) that fringe the lake today. But this short term variability did not strongly affect the vegetation cover: among minor changes we noticed a depletion of the $\delta^{13}C_{ORG.}$ values (Fig. 9d), concomitant to a slight decrease of the Poaceae at 1800 and 1400 cal yr BP (fig 9e). Palynological data in Lake Mbalang showed the expansion of Poaceae at 3000 cal yr BP, they remained the most abundant than any other groups of plants until the present days. Sedges also developed and reached their highest abundance suggesting the lowering of lake level at a short timescale. Montane forest regrowth (Fig. 9e), and arboreal savannas taxa abundance became very low. These modifications in the vegetation landscape implied a more dry and contrasted climate (Vincens et al., 2010) as also suggested by diatom habitat groups and windblown diatoms (Fig. 9c). The 2400–2100 cal yr BP event is also well marked in other sites of the subequatorial regions of central Africa (Vincens et al., 1999). The data confirmed a more dry climate in southern Congo, but at the latitude of Lake Ossa, woody pioneer heliophilous taxa appear in the rain forest (Reynaud-Farrera et al., 1996), probably as a result of stormy rainfall rather than absolute low precipitation (Nguetsop et al., 2004) as well as in Nyabessan located 200 km south of Lake Ossa (Ngomanda et al., 2009b). In Lake Bosumtwi (6°30′ N; 1°25′ E), sedimentological records showed an evolution towards aridity and more seasonality at about 3000 yr BP (Russell et al., 2003;

Talbot et Johannessen, 1992). The reduction of the mixing at 1700, 700–600 and at 400 cal yr BP is marked by a slight decrease of Poaceae and the increase of Cyperaceae, $\delta^{13}C_{ORG.}$ values decreased also slightly. This last event shows the sensitivity of vegetation and hydrology to recent centennial climate variability as it was demonstrated by Ngomanda et al. (2007, 2009b).

5.3 Paleoclimatic interpretation

Diatoms data suggest a decreasing trend of the monsoon flux in Adamawa area from mid-Holocene (7200 cal yr BP) to mid-late Holocene, consistently with the decreasing summer insolation in the Northern Hemisphere and correlatively reducing land-ocean contrast linked to orbital changes. Although orbital changes account for a greater part in explaining the hydrological changes (Kutzbach and Street-Perrot, 1985), they induced regional atmospheric factors that may be useful in understanding the response of the local hydrological system. The better comprehension of climatic changes in central Africa regions around the Gulf of Guinea should integrate the structure of the atmosphere during the wet season when the monsoon flux overrides the NE trade winds in the northern summer. According to Leroux (1970, 2001), five climatic zones can be individualized in the meridian structure of the troposphere at this period of the year, they have been used in interpreting past climatic conditions by several authors (e.g. Nguetsop et al., 2004; Ngomanda et al., 2009b). The compression and dilatation of these climatic zones over the year can explain a series of climatic conditions that are encountered yearly today between 20° N and 5° S. One can then hypothesize that, if in the past the rain belt moved northwards and entailed rainfall at Saharan region at around 6000 yr BP as shown by paleoclimatic data (e.g. Gasse, 2000) and reproduced by paleoclimatic models (Kutzbach and Street-Perrot, 1985; Kutzbach and Guetter, 1986) it is likely that all the climatic zones that are linked to the strengthening of the monsoon, and not only the convection area, were more extended than today during the boreal summer. This hypothesis is reinforced by the fact that cloud cover and low evaporation that are limited today between 5° S and 4° N are also reproduced by climatic model in higher latitude at 6000 yr BP (Kutzbach and Guetter, 1986).

From 7200–3600 cal yr BP, the lake level was mostly moderate to high as evidenced by planktonic diatoms and the water column generally stable. We suggest that the ITCZ mean position at that time was north of the Adamawa plateau (Fig. 2b) in agreement with paleoclimatic data (Gasse and Van Campo, 1994); this position entailed at the latitude of the studied lake, stratiform cloud cover and low precipitation (Fig. 8a). Temperatures were consequently relatively low due primarily to these atmospheric features, but also, to the relatively high altitude of the Adamawa plateau (1100–1200 m). These conditions were favourable for the development of the mountain forest taxa in the vegetation landscape and the regrowth at the forest borders (Vincens et al., 2010). This period was characterized by very low mixing except between 5000–5300 cal yr BP; the Harmattan was probably very weak until 4500 cal yr BP.

From 7200–6900 cal yr BP, diatoms data suggest a relatively deep and stable lake. Despite the age uncertainties offset and the different time resolution in published data, this subphase could correspond to the wet episode that is well known in Saharan and Sahelian regions (Servant et Servant-Vildary, 1980; Gasse, 2000), the African Humid period. The high monsoon inflow suggested by diatoms at 6400, 5500, 4600 and 4200 cal yr BP and characterized by relatively high lake level in Adamawa plateau (Fig. 9a) appeared at certain periods of time to be uncorrelated with data of sea surface temperatures (Fig. 9i) and rivers discharges (Fig. 9j) off the Gulf of Guinea (Weldeab et al., 2005, 2007). This can be explained if the variability of the mean position of the ITCZ is considered at multi-secular to millennial timescale as it is observed today over the year. Consistent northernmost mean position of the ITCZ may have favoured rainfall at the northern part of the catchments of Niger River while the southern part and probably a great part of the Sanaga and Ntem may have been under stable air layers (Fig. 2b). Conversely, the southernmost mean position of ITCZ may have favoured high rainfall around the Gulf of Guinea in the greatest part of the Sanaga and Ntem river catchments, and drier conditions in the upper part of the Niger River. These rivers discharge off the Gulf of Guinea may thus be very variable depending on SSTs and the position of ITCZ which strongly influences precipitation and evaporation on the continent. Lake Ossa high lake levels are observed in the context of low rainfall between 4800 and 4400 cal yr BP suggesting a climate with low evaporation and low rainfall consistent with the northernmost position of the ITCZ. Hence, the apparent discrepancies observed between rainfall on the continent, SSTs and rivers discharges off the Gulf of Guinea during Middle to Late Holocene (Weldeab et al., 2005, 2007) could be explained by these meridian changes of the structure of the lower levels of the atmosphere at centennial to millennial timescales.

The low monsoon inflows at 6700, 6000–5800 and 4600–4400 cal yr BP are also characterised by relatively low lake levels in Lake Mbalang. SSTs are low to medium except at 6700 cal yr BP where they are high. The general conditions suggest a position of the ITCZ further north than today, but brief low lake levels could indicate at least the ITCZ episodic southward displacements entailing either low rainfall (if the ITCZ moves south of Adamawa) or high rainfall (if the Adamawa region is included in the convection zone) and high evaporation in the two cases (Fig. 8b). The appearance of windblown diatoms in Adamawa at 4500 cal yr BP corresponds probably to the desiccation of the Sahara or intensification of the NE trade winds. At 5300–5000, lake levels are high in Adamawa, SSTs and river discharges are high, rainfall is high in low latitude (Lake Ossa), suggesting the displacement towards the south of the ITCZ at a position favourable to convective rainfall in the two regions. This phase is probably contemporaneous of the onset of a dry episode in Sahara; paleolakes retreated around 5800 yr BP (Vernet, 1995; Servant and Servant-Vildary, 1980) consistently with the termination of the African humid period (de Menocal, 2001). At that time, high mixing as observed at the upper part of the studied core (after 3600 cal yr BP) shows that position of the ITCZ was closed to its modern position (Fig. 2a). This hypothesis is reinforced by the development of savannah in the vegetation landscape indicating as today a more contrasted climate. Between 3600 and 3000 cal yr BP, SSTs off the Gulf of Guinea alternate between moderate and low values, and river discharges were relatively low or moderate in good agreement with low rainfall in Lake Ossa but high lake level inferred in Lake Mbalang may indicated higher rainfall in Adamawa plateau linked to the displacement towards the north of the convective rain belt.

Between 3000 and 0 cal yr BP, diatom data suggest the significant reduction of the monsoon flux. The lake level remained broadly high except between 2400–2100, 1800 and 1400 cal yr BP. Although the lake did not decline dramatically, indicating that rainfall remained relatively important, the increase of savannah taxa and their maintenance until today attest a seasonality change of the rainfall distribution. The influence of the NE trade winds during the year is shown by the persistence of windblown diatoms. A low lake level registered both in Lake Mbalang and in Lake Ossa between 2400 and 2100 cal yr BP and in others subequatorial regions of Africa coincided with higher rainfall in Ossa region and important fluctuations of SSTs off the Gulf of Guinea while the river discharges decreased gradually. It revealed the unstable position of the ITCZ and consequently the rainfall belt modifications during this southwards shift. In agreement with our model, this episode corresponds to the southernmost position of the ITCZ, at least episodically. Consequently, it entailed more arid conditions northwards as shown by intense windblown diatoms, indicating the strengthening of NE trade winds in Ossa region, stormy rainfall around the Gulf of Guinea with subsequent disturbances inside the forest block. After 2000 cal yr BP, the evolution towards present days is observed. These new conditions are roughly characterized by relatively high lake level in Lake Ossa and in the Adamawa, high rainfall in Ossa region suggesting a sharp northwards shift of the mean position of the ITCZ. Meanwhile, both river discharges and SSTs showed a decreasing trend. Brief highstand at 2000–1900 cal yr BP, lowstands at 1800 and 1400 cal yr BP attested the more intense or weakening of the monsoon inflow respectively. The reduction of the mixing at 700–600 and at 400 cal yr BP marked a slight intensification of the monsoon which is well recorded both by rainfall regime and lake level in Ossa.

Holocene short climatic events were evidenced in several sites of the monsoon domain both in Africa and Asia, the forcing factors is primarily the modifications of insolation that is modulated by sea surface temperatures and land surfaces feedback mechanisms (Gasse et Van Campo, 1994; de Menocal et al., 2000).

6 Conclusions

Planktonic and tychoplanktonic diatoms variation suggested that Lake Mbalang did not dry during the last 7200 cal yr BP as relative fluctuations of water level are observed. A low lake level recorded at 2400–2100 cal yr BP is contemporaneous to a climatic event evidenced in several areas of tropical Africa, and other low lake levels are observed at 1800 and 1400 cal yr BP, after which the lake rose to its present level. Nevertheless, diatom data showed that the lake evolved from oligotrophic stable water table before 3600 cal yr BP to mixed and eutrophic conditions afterwards, corresponding respectively to a strong monsoon flow before and a more intense north-eastern trade winds (Harmattan) after. The $\delta^{13}C_{ORG.}$ data indicated the development in the landscape of more forested vegetation, also confirmed by palynological data in good agreement with the inferred climate. However, the decreasing monsoon trend was punctuated by several abrupt weakenings at 6700, 5800–6000, 5000–5300, and 4500 cal yr BP. After 3000 cal yr BP, the savanna vegetation developed in the Adamawa area and has persisted until today. These climatic changes can be attributed to the modifications of the position of the Intertropical Convergence Zone (ITCZ), its northernmost position between 7200 and 3600 cal yr BP entailed at the level of the Adamawa plateau, a climate characterized by very low precipitation and also very low evaporation as is observed today during the boreal summer in the southwest of Cameroon. After 3600–3000 cal yr BP, the ITCZ moved southward and reached a position where convective rainfall became dominant, but its amount and/or its distribution were no longer favourable to forest development.

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