Fluctuations in Miocene climate and sea levels along the southwestern South African coast: inferences from biogeochemistry, palynology and sedimentology

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The largest deposit of excellently preserved Mio-Pliocene vertebrate faunas in South Africa is located at the well known Langebaanweg (LBW) site along South Africa's west coast in the southwestern Cape. This research deals with a core (BH2) drilled at Langebaanweg 'E' Quarry, which captures Miocene fluvial deposits that unconformably underlie the Mio-Pliocene deposits. The aim of the study is to constrain fluctuations in climate and ecosystems in the region during the Miocene, using a combination of sedimentology, palynology, and biogeochemistry. The latter is a novel application and includes bulk C and N isotope measurements as well as branched GDGT (glycerol dialkyl glycerol tetraether) membrane lipid compositions of the organic fraction of the core. The mean annual temperature (MAT) at time of deposition and ambient pH values were calculated using the Methylation index of Branched Tetraethers (MBT) and Cyclization ratio of Branched Tetraethers (CBT) proxies. The core samples analysed spanned a depth interval of 17–33 m and had MATs ranging between 12.4°C and 26.6°C and pH range from 4.4 to 6.4. Furthermore, samples showed a low variance and light δ^{13}C distribution (from –25.52‰ to –24.27‰) and overall low C/N ratios. Palynological investigation supplemented earlier similar studies, reaffirming alternating sequences of tropical and subtropical elements including wetland taxa and complementing calculated MAT results. Pollen results from the lowermost subsection indicated a species-rich tropical/subtropical *Podocarpus* dominated forest with MATs at 16.2°C. The pollen-bearing middle subsection shows initially subhumid conditions, with MATs between 15.4 and 26.6°C, similar to the bottom of the section with high Podocarpaceae percentages, low Restionaceae and aquatics. Above this, local taxa, e.g. Restionaceae and algae increase, and a marine influence is indicated by abundant dinoflagellates. The uppermost subsection from a depth of 7.20–7.60 m shows similar conditions (with little marine influence) to that of the lower part of the middle subsection II. Biochemical and sedimentological data are not available for the uppermost subsection. The results suggest that regional Miocene climate showed high amplitude fluctuations (possibly driven by orbital forcing as seen in marine cores), underscoring the potential of biogeochemistry for unravelling past climates and ecosystems.

Keywords: Miocene, palms, Podocarpaceae, glycerol dialkyl glycerol tetraethers (GDGTs), South Africa.

INTRODUCTION

Setting

The LBW research area, a National Heritage Site, is an internationally known deposit which was originally a large-scale phosphate mine and now forms the West Coast Fossil Park (Fig. 1; Hendey 1982; Roberts *et al.* 2011). It lies on the coastal platform with the Great Escarpment to its landward aspect. The platform originated in the Mesozoic break-up of Gondwana and was influenced by several marine transgressions (Roberts *et al.* 2011). This study deals with a core drilled at the West Coast Fossil Park at Langebaanweg 'E' Quarry which captures the Elandsfontyn (Miocene) and Varswater (Mio-Pliocene) Formations of the Sandveld Group (Fig. 1).

Southern African perspective of Miocene climate change

The opening of oceanic gateways in the southern hemisphere in the Late Oligocene–early Miocene resulted in the thermal isolation of Antarctica by the middle Miocene, and established new atmospheric and oeanic circulation systems (Miller & Fairbanks 1985; Holbourn *et al.* 2005; Miller *et al.* 2005; Potter & Szatmari 2009). These changes were influential in reducing temperatures following the Middle Miocene climatic optimum (MMCO) in the southern hemisphere, where temperatures reached up to 6°C warmer than present (Flower & Kennett 1994; Holbourn *et al.* 2007; Donders *et al.* 2009; Shunk *et al.* 2009; Tong *et al.* 2009; Majewski & Bohaty 2010).

On the southwest coast of South Africa, increased global temperatures and subsequent sea-level changes are
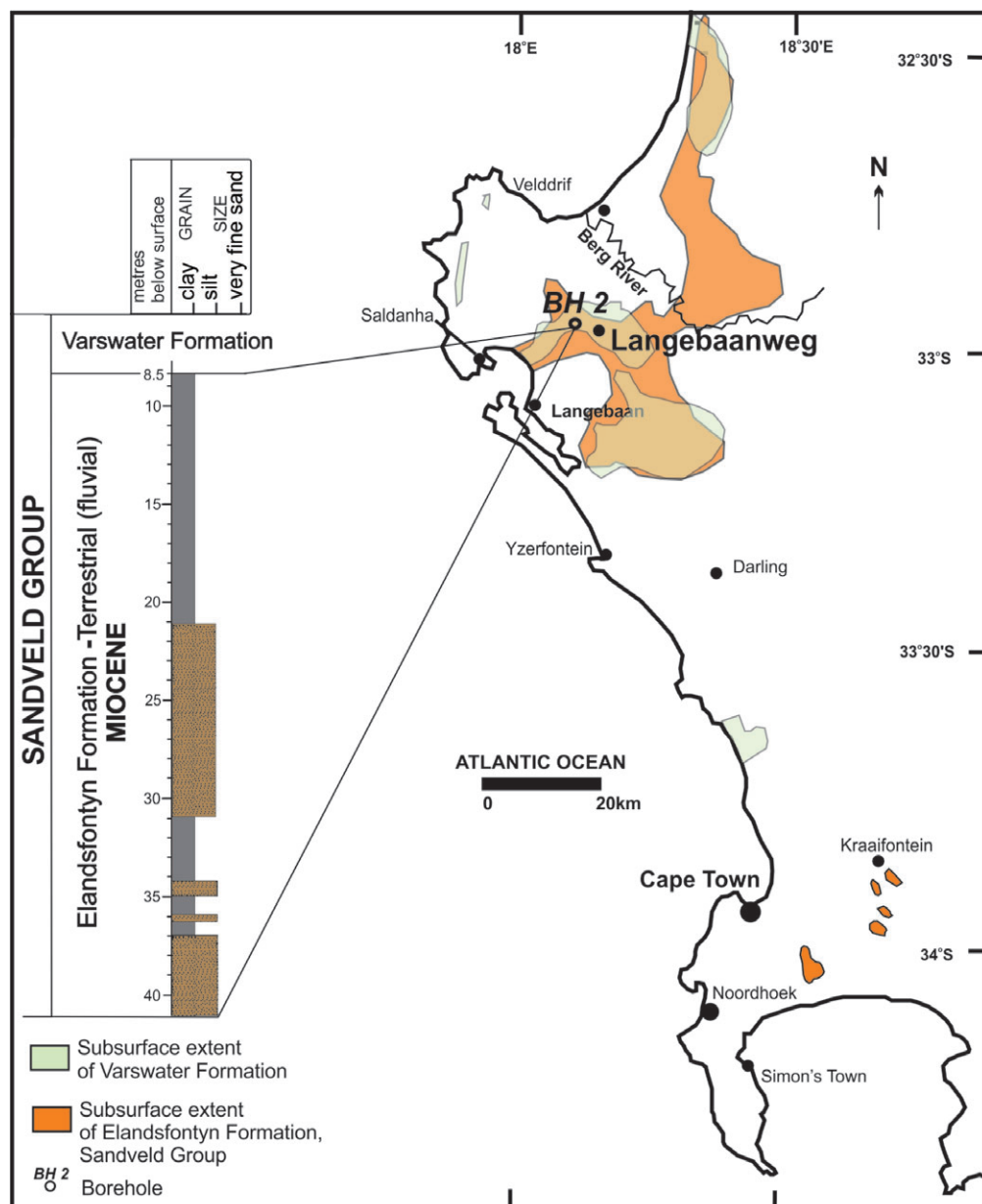
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Figure 1. The map indicates the subsurface extent of the Elandsfontyn Formation (orange), Sandveld Group, which unconformable overlies Precambrian basement, southwestern Cape, South Africa (adapted from Roberts 2006a), in relation to the study site at Langebaanweg (LBW) and sampled core (BH2; 32°57′50.9″S, 18°06′45.7″E) with its alternating organic clays and silts (inserted log).

recorded by a series of sea-level high stands (Hendey 1982; Clarke & Crame 1992; Wigley & Compton 2006). A MMCO age of 19 Ma–14 Ma has been assumed by previous Miocene terrestrial palynological studies in South Africa (Coetzee 1978). This age range is relative, and largely based on the presence of palm-dominated subtropical-tropical vegetation (Coetzee 1978) indicating increased regional temperatures and reduced seasonality. Other palynological studies and reviews of the Elandsfontyn and Varswater Formations by Coetzee & Rogers (1982), van Zinderen Bakker & Mercer (1986) and Scott *et al.* (1997) illustrate a change in palaeovegetation, from riparian tropical and temperate forest to the expanding Pliocene grasslands, and winter rainfall fynbos biomes. However, certain fynbos elements, e.g. Restionaceae, Ericaceae and Proteaceae, already contributed to the vegetation in southwestern Africa during the Palaeogene (Scholtz 1985; Linder 2003). Grasslands are thought to

have broadened globally during the late Miocene–early Pliocene, accompanying summer-rainfall-dominated savanna biomes, which are now firmly established in southern and East Africa (Cerling *et al.* 1997; Franz-Odendaal *et al.* 2002). The same time period saw the major radiation of fynbos elements (Goldblatt & Manning 2000; Cowling & Pressey 2001; Linder 2003).

A major uncertainty in understanding Cenozoic climate fluctuations, in general, is the scarcity of terrestrial counterparts for the long and detailed oeanographic/climate records contained in marine cores, which have generally well-constrained chronologies. Our study focuses on the organic-rich fluvio-wetland succession (Elandsfontyn Formation) at LBW. Here we use sedimentology alongside preliminary palynological results and organic biogeochemistry to reconstruct the sequential palaeoecology of the Miocene Elandsfontyn Formation deposit at LBW. Climate fluctuations, through time and space, are provi-
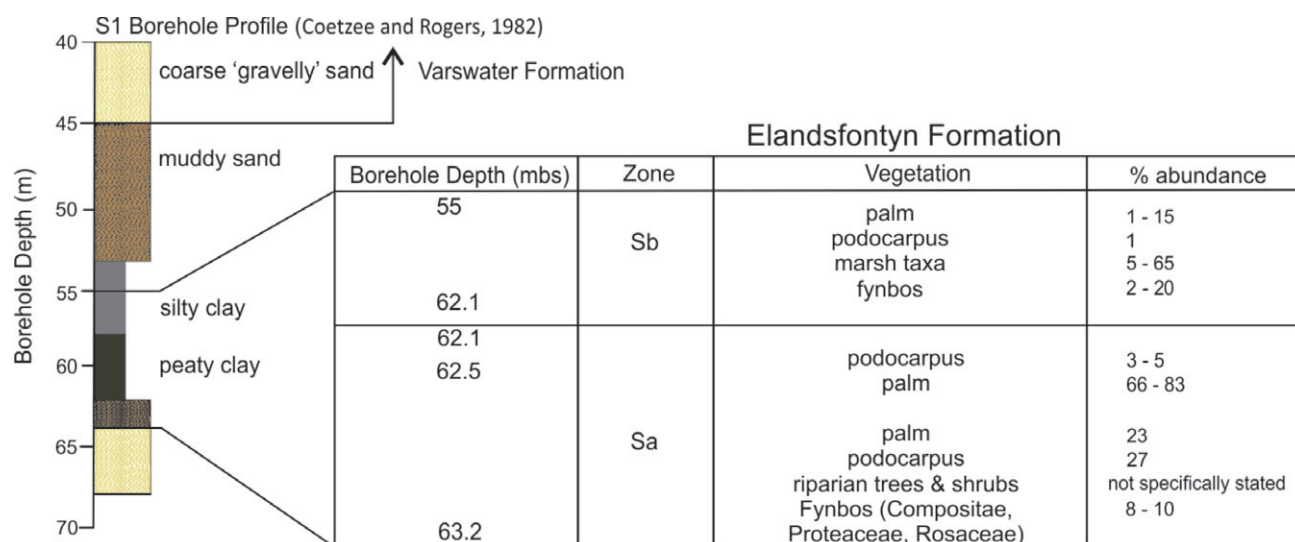
Figure 2. S1 borehole log of Coetzee & Roger's (1982) study of the Elandsfontyn Formation vegetation highlighting the sampled peaty- and silty-clay units. Inserted table provides percentage abundances for the zones (Sa and Sb) described by Coetzee & Rogers (1982). The uppermost part of Zone Sa is characterized by high abundances of palm pollen (66–83%) and dramatic reduction of *Podocarpus* pollen (3–5%) in comparison to underlying abundances and the overlying Zone Sb abundances.

sionally mapped. Biogeochemical proxies are potentially highly informative but have thus far been little used in southern Africa (Carr *et al.* 2010, compare Roberts *et al.* 2013). It is also a useful method if oxidation processes have destroyed palynomorphs so that only amorphous organic matter (AOM) is preserved. Biogeochemistry includes quantifying the amount and composition of branched bacterial tetraether lipids (branched GDGTs), which are used as proxies for the determination of palaeo-mean annual air temperatures (MATs) and soil pH. Bulk carbon and nitrogen isotope analyses were also applied for the determination of organic matter source and character.

Regional geological and palaeontological setting

The sediments at LBW have undergone various nomenclature changes since their first informal names were assigned. Roberts (2006a,b,c,d) revised the coastal Cenozoic lithostratigraphy of the southwestern Cape and established the Sandveld Group, comprising the Elandsfontyn, Prospect Hill, Varswater, Langebaan and Velddrif Formations.

Dating of the Elandsfontyn Formation at LBW

Biostratigraphic dating has been conducted on the mammalians of LBW, especially of the upper Varswater Formation, because of their large numbers and variety; little focus has been placed on the flora of the underlying Elandsfontyn Formation (Fig. 1; Coetzee 1978; Tankard & Rogers 1978; Hendey 1974, 1981, 1982, 1983; Coetzee & Rogers 1982; Franz-Odendaal *et al.* 2002; Matthews *et al.* 2006, 2007). A biostratigraphic age of middle Miocene has been assigned to the Elandsfontyn Formation, late Miocene (10 Ma) age to the lower Varswater Formation (Konings Vlei Gravel Member) and Mio-Pliocene age to the upper Varswater (Dingle *et al.* 1979; Hendey 1974, 1981, 1982, 1983; Coetzee & Rogers 1982). The latter age, on the basis of palaeomagnetic data, was honed to 5.1 Ma (Roberts *et al.* 2011). These ages are reaffirmed by the correspondence of Cenozoic Western Cape sea-level fluc-

tuation to global Cenozoic sea-level trends (Tankard 1976; Haq *et al.* 1987; Shackleton 1995; Franz-Odendaal *et al.* 2002).

LBW palaeoflora

Coetzee's (1978) pilot study of the organic-rich unit (uppermost Elandsfontyn Formation), below the main fossil bed at LBW, gave the first clues to the subtropical floral assemblages present. This was later supplemented by a more detailed study of the peaty-clay and organic silty-clay horizons from the S1 borehole (33°58.20'S and 18°6.97'E; Fig. 1) on Langeberg 188 Farm (Coetzee & Rogers 1982). Compared to the work of Tankard & Rogers (1978) of the Varswater Formation floras, Coetzee & Rogers' (1982) study provides a comprehensive look at the underlying LBW floras of the Elandsfontyn Formation. It established two pollen zones within the uppermost Elandsfontyn Formation of the S1 borehole (Fig. 2). However, the pollen results are only shown in composite form and percentages of pollen taxa are occasionally mentioned in the text.

The lowermost pollen zone (Sa) between 62.1 m and 63.2 m borehole depth (2.95 m to 4.0 m above sea level) is dominated by palms (21–23%), *Podocarpus* (26–27%) as well as tropical forest and riparian plants such as Oleaceae, Celastraceae, *Celtis*, Casuarinaceae and others no longer occurring in the southwestern Cape (Fig. 2; Coetzee & Rogers 1982). Fynbos elements are present throughout the zone, e.g. Asteraceae (1–2%), Proteaceae (1–4%), and Rosaceae (1–9%). In the upper part of the Sa-zone, between 62.1 and 62.5 m depth below surface, there is a reduction of forest taxa with a dominance of palm pollen (66–87%) and lesser amounts of *Podocarpus* pollen (3–5%; Fig. 2; Coetzee & Rogers 1982). Generally, this zone shows a decrease in humidity over time. The upper zone (Sb) is between 55.0 m and 61.2 m depth below surface, and shows a dramatic reduction in palm pollen abundance (1–15%; Coetzee & Rogers 1982). Forest elements are still present, but this may be as a result of their

long-distance dispersal capability, as is the case with *Podocarpus* pollen (±1%; Coetzee & Rogers 1982). This zone is considered to have been deposited in a freshwater environment, and contained high proportions of marsh taxa, which include local elements like *Sparganiaceae pollenites* and algae (Coetzee & Rogers 1982).

These studies exhibit the interplay of alternating wet and dry period floras (Coetzee & Rogers 1982). However, Coetzee & Rogers (1982) largely account for the change in vegetation to have been due to the meander and lateral migration of the river channel over time with the influence of sea-level variation. Thus, a change from river channel with significant representation of riparian forest (*Podocarpus* and other arboreal taxa) was developed proximal to it and earlier in the succession. Later there developed proximal and distal floodplains, where palm and swamp vegetation dominated on poorly drained soils (Coetzee & Rogers 1982). Other deposits in the Cape region (Sakriver and along the Olifantsriver), dated to the middle Miocene, contain wood of tropical trees and indicate a humid forest (Bamford 1999). Therefore, the shift from a rather tropical environment to vegetation with fynbos characteriztics together with a decrease in palms might also be interpreted as a change from a warm, more mesic climate, to a cooler more seasonal climate (Linder 2003).

METHODS AND MATERIALS

Sample collection

For this preliminary study 49 samples from core BH2 were processed for palynomorphs using a standard procedure, which includes the addition of HCl, HF, KOH and heavy liquid separation within a saturated $ZnCl_2$ solution. The pollen samples were mounted in glycerol jelly (solid medium), so that these slides can be stored for longer periods. *Lycopodium* spore tablets were added to calculate pollen and charred particle concentrations, which were expressed as the number of grains and fragments per gram sediment (Stockmarr 1971). Identification and counting was performed with a light microscope under ×400 and ×1000 magnification. Pollen, algae, fungi and spores were identified using the pollen reference collection at the University of the Free State, Bloemfontein, and the Bernard Price Institute for Palaeontology at the University of the Witwatersrand, Johannesburg, and references (amongst others Coetzee 1978, 1981; Coetzee & Rogers 1982; Coetzee & Praglowski 1988; de Villiers 1997; de Villiers & Cadman 1997; Scott 1982; Stuchlik *et al.* 2001, 2002, 2009). Approximately 250 terrestrial pollen grains were counted per sample (see results, Table 2).

Lipid extraction and analysis

For this initial study, approximately 1–2 g of five of the 49 samples were extracted with an accelerated solvent extractor (ASE Dionex 2000) using a dichloromethane (DCM): MeOH 9:1 (v/v) solvent mixture in three cycles under conditions of 100°C and 1000 Psi (5 min static, 100% flush, 90 s purge). The extracted solution, after rotary evaporation, was then separated over an activated Al_2O_3 column into an apolar fraction (eluent hexane/DCM, 9:1, v/v), a pure DCM-fraction, a tetraether fraction (eluent DCM/MeOH, 95:5, v/v) and a polar fraction (eluent MeOH/DCM, 1:1, v/v). The apolar and polar fractions of each sample were dried under N_2-gas. The polar fraction, which contained the branched GDGTs, was dissolved in hexane: propanol 99:1 (v/v) mixture to produce a 2 mg/ml solution, and placed in an ultrasonic bath. The mixture was then filtered through a 0.45 µm PTFE filter (Alltech) in preparation for high performance liquid chromatography mass spectrometry (HPLC/MS) analysis. An internal standard C_{46} GDGT was added to all samples (Huguet *et al.* 2006). For separation, an Alltech Prevail Cyano column (150 mm × 2.1 mm; 3 µm) was used. Samples were run with a hexane:propanol (99:1, v/v) eluent. An Agilent 1100/1100MSD series instrument, equipped with auto-injector and HP Chemstation software was used. Following Hopmans *et al.* (2004), after the samples were run through HPLC/MS, quantification of branched glycerol dialkyl glycerol tetraethers (GDGTs) was achieved through the comparison and integration of the area of the protonated molecular ion $[M+H]^+$ peaks with an isoprenoid GDGT crenarchaeol internal standard of known quantity (Huguet *et al.* 2006).

BIT, MBT and CBT proxies

Branched GDGTs were described and identified, using nuclear magnetic resonance (NMR) spectroscopy, by Sinninghe Damsté *et al.* (2000) when studying a northern hemisphere Holocene peat deposit (branched GDGTs I–III). They are typically found in lake sediments, soils and peat bogs, and have also been detected in, and used to represent the terrestrial sedimentary inputs to coastal sediments (Hopmans *et al.* 2004; Weijers *et al.* 2006). Hopmans *et al.* (2004) used the occurrence of branched GDGTs in marine sediment to construct the Branched and Isoprenoid Tetraether (BIT) index which measures the ratio of relative abundance of terrestrially produced branched GDGTs and isoprenoid marine/lacustrine GDGTs. Weijers *et al.* (2006) then established that the various branched GDGT lipids differed by two main features: (i) number of methyl branches (attached to alkyl chains) and (ii) number of cyclopentyl moieties. Weijers *et al.* (2007c) relationship between the number of cyclopentyl moieties and methyl groups of the tetraether membrane lipids is related to the soil pH and mean annual temperature (MAT), at the time of deposition, and allow for two indices to be described. These indices are: (i) methylation index of branched tetraethers (MBT), and (ii) cyclization ratio of branched tetraethers (CBT) (Weijers *et al.* 2007a,b,c). The CBT index allows for the calculation of soil pH and the MBT index of MAT and soil pH. The analytical reproducibility of these indices (0.1 units) allows for a temperature estimate reproducibility of 0.9°C (Weijers *et al.* 2007b,c).

Stable isotope preparation

Carbon and nitrogen isotope ratios were measured against the international standard ($\delta^{13}C$ nylon = –26.2 ± 0.2‰ and $\delta^{15}N$ nylon = –2.0 ± 0.4‰, n = 130) at the Department of Earth Sciences, Durham University, using protocols outlined in Palmqvist *et al.* (2003).

Table 1. Comparison between the two broad groupings of LBW samples based on measured bulk organic parameters and CBT/MBT proxy values. MAT– mean annual air temperature.

	'High' MAT group 17–18.40 m	'Low' MAT group 19.04–33 m
MAT	High; 21.5–26.6°C av. 24 ± 4°C	Moderate to low; 12.4– 21°C av. 16 ± 3°C
pH	Near neutral 6.2–6.4, av. 6.0 ± 0.11	Mild/moderately acidic 4.4–6.3, av. 5.0 ± 0.8
BIT	0.74–0.78	0.98–1.0
$\delta^{13}C$ (‰ vs PDB)	av. – 24.5 ± 0.3‰	av. –25.1 ± 0.6‰
$\delta^{15}N$ (‰ vs AIR)	3.5 ± 0.3‰	2.8 ± 0.5‰
TOC	av. 3.6 ± 1.3	av. 5.75 ± 3.5
Corg/Norg	42.5 ± 12.7	37.6 ± 6.4

RESULTS

The samples of this study form part of a fine-grained assemblage which caps an upward fining sequence of the fluvial Elandsfontyn Formation. With reference to the lithological log (Fig. 1) samples are largely homogeneous organic-rich clays which become silty with increasing stratigraphic depth. The size, uniformity and homogeneity of grains of all samples indicate deposition from suspension. Descriptive sedimentology of the samples showed no significant changes in colour, grain size or sorting between samples. The five LBW samples provided for analysis are homogeneous argillaceous (~80% clay) sediments, with the presence of organic matter implied by the dark colouration (brown to black).

Bulk geochemistry, isotopes, BIT index and CBT/MBT proxy

Total organic carbon (TOC) contents of the five selected samples vary between 2.3% and 10.7%, the latter value recorded at 19.50 m. By contrast, total nitrogen (TN) values are low, ranging from 0.09% (at 18.40 m) to 0.22% (at 19.50 m). TOC/TN ratios vary from 23 to 48.6 with the lowest value recorded at 33 m depth (Fig. 3). Characteristic in all profiles is the minimum values of TOC, TN and TOC/TN ratio at 18.40 m that increase progressively to highest TOC, TN and TOC/TN ratio at 19.50 m (Fig. 3). The C_{org}/N_{org} ratio ranges from 27 (at 33 m depth) to as high as 57.4 (at 19.5 m).

The $\delta^{13}C$ record ranges from –25.5‰ to –24.3‰ with the lightest value obtained at 19.50 m depth (Fig. 4). A peak in both TOC and $\delta^{13}C$ values is reached at 19.50 m with values of 49% and –25.5‰, respectively (Figs 3 & 4). These elevated values at 19.50 m then decline to 33 m depth with a TOC and $\delta^{13}C$ values of 23% and –25.5‰, respectively. The $\delta^{15}N$ values also show a narrow range between 2.4‰ (19.14 m) and 3.7‰ (17 m) with the stratigraphic relationship between $\delta^{15}N$ and $\delta^{13}C$ being broadly antithetic (Fig. 4). Such $\delta^{15}N$ values fall within the accepted norm of soils (between +2‰ and +5‰; Meyers 2003; Sharp 2007). All five samples taken for bacterial tetraether lipid analyses showed a positive result (Fig. 4: C). BIT indices range from 0.74 to 1.0 (Fig. 4: C); the lowest BIT values were recorded from samples at 18.4 m and 17 m (0.74 and 0.78, respectively), whilst all other samples exhibit high BIT indices above 0.98.

BIT index and CBT/MBT proxy

The MBT/CBT proxy was used to calculate mean annual temperatures (MAT) and pH of the environment at time of deposition (Fig. 4: D, E). The LBW samples have MATs which range from 12.4°C to 26.6°C and a pH range from 4.4 to 6.4 over the 16 metres studied (Fig. 4: D, E). There is a broadly positive linear correlation, with low R^2 values, relationship between MAT, pH and stable carbon isotope $\delta^{13}C$ values of samples. Interestingly, the calculated MATs also show two distinct groupings relative to the present day MAT of the Western Cape:

'Low' (<16°C) temperatures (samples between 19.04 and 33 m depth).

'High' (>21.5°C) temperatures (samples between 17 and 18.40 m) (Table 1).

Such a distinction can be further supported on the basis of additional parameters such as the BIT index as well as $\delta^{15}N$ values (Table 1).

Palynological results

Forty out of 49 samples, from three sections of the profile (III. 30.91–33.02 m, II. 16.47–20.32 m, I. 7, 20–7.61 m; Figs 5 & 6), contained palynomorphs in reasonable abundance and condition. By contrast, all samples between a depth of 7.61 m and 16.475 m were either barren or had low quantities of palynomorphs, but amorphous organic matter was abundant (Figs 5 & 6). The interval between subsection II and III was not subsampled and is not available for this study. Palynomorph diversity is similar to that recorded in Coetzee & Rogers (1982). Nomenclature of all fossil genera and species mentioned in this manuscript and/or the pollen histograms is given in Table 2, which also includes suggestions for the botanical affinities and references. Few palynomorph types, e.g. *Cliffortia* sp., are provisionally assigned to recent taxa (compare Coetzee & Rogers 1982). Many different types of palynomorphs were recognized (119), among them pollen, spores, algae, acritarchs, microforaminiferal linings and dinoflagellates. The latter three groups are only represented by one genus. A single silicoflagellate, phytoliths and few unidentified diatoms apparently survived the HF digestion of the sediments. We have to assume that they are actually much more common in the sediments. Forest taxa were diverse and *Podocarpidites* and *Cupressacidites* especially abundant.
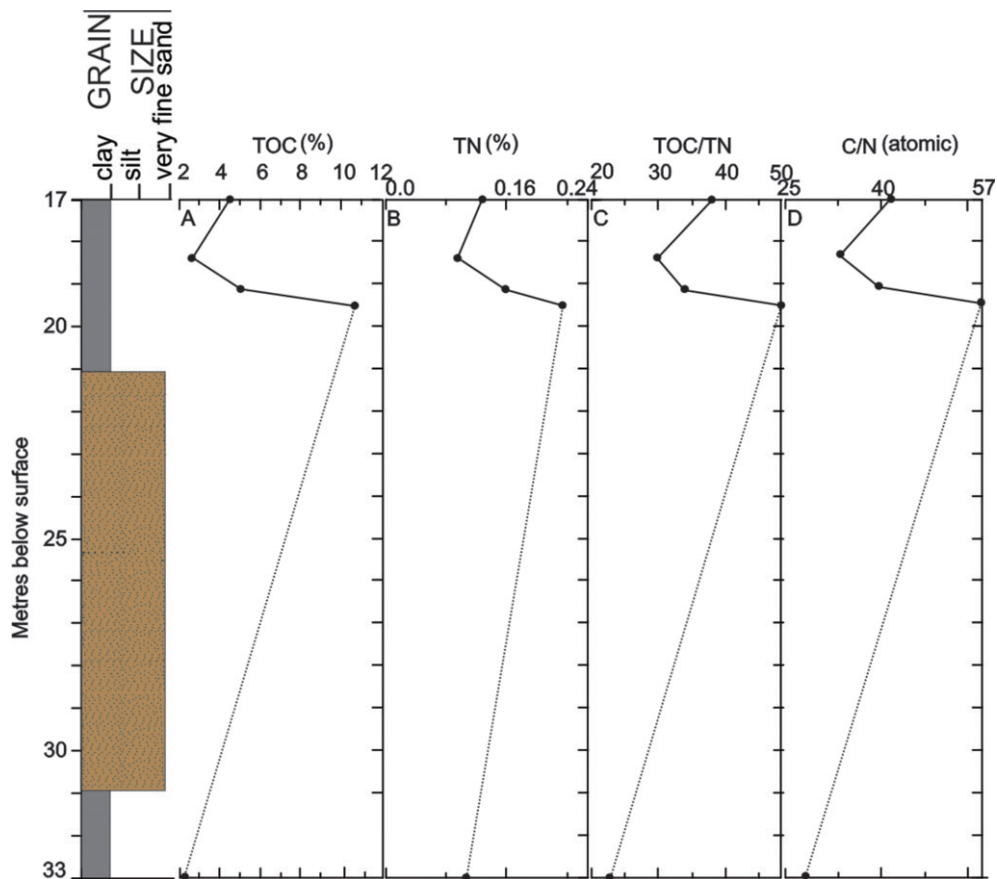
Figure 3. Stratigraphic down-core variability of Langebaanweg organic clay samples in core BH2 (A) total organic carbon (TOC) content (%), (B) total nitrogen (TN) content (%), (C) TOC/TN ratio and (D) the C/N (atomic) ratio. There is no strong relationship between TOC and δ^{15}N (‰ vs AIR) and TN and δ^{15}N (‰ vs AIR) (correlation coefficients of 0.24 and 0.078, respectively).
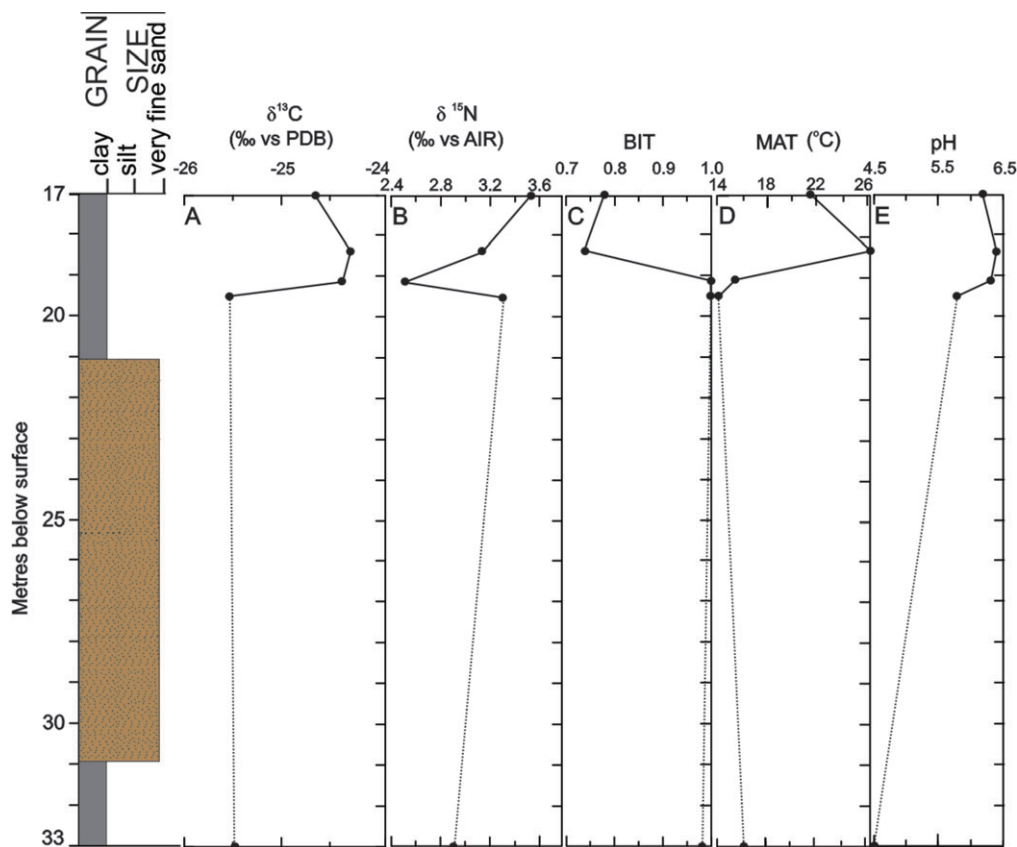

Figure 4. Stratigraphic down-core variability of the (A) stable carbon and (B) nitrogen isotope values; (C) Branched and Isoprenoid Tetraether index (BIT), (D) mean annual air temperature (MAT – calculated from the CBT/MBT proxy), and (E) pH for the Langebaanweg samples in core BH2 between 17 m and 33 m depth.
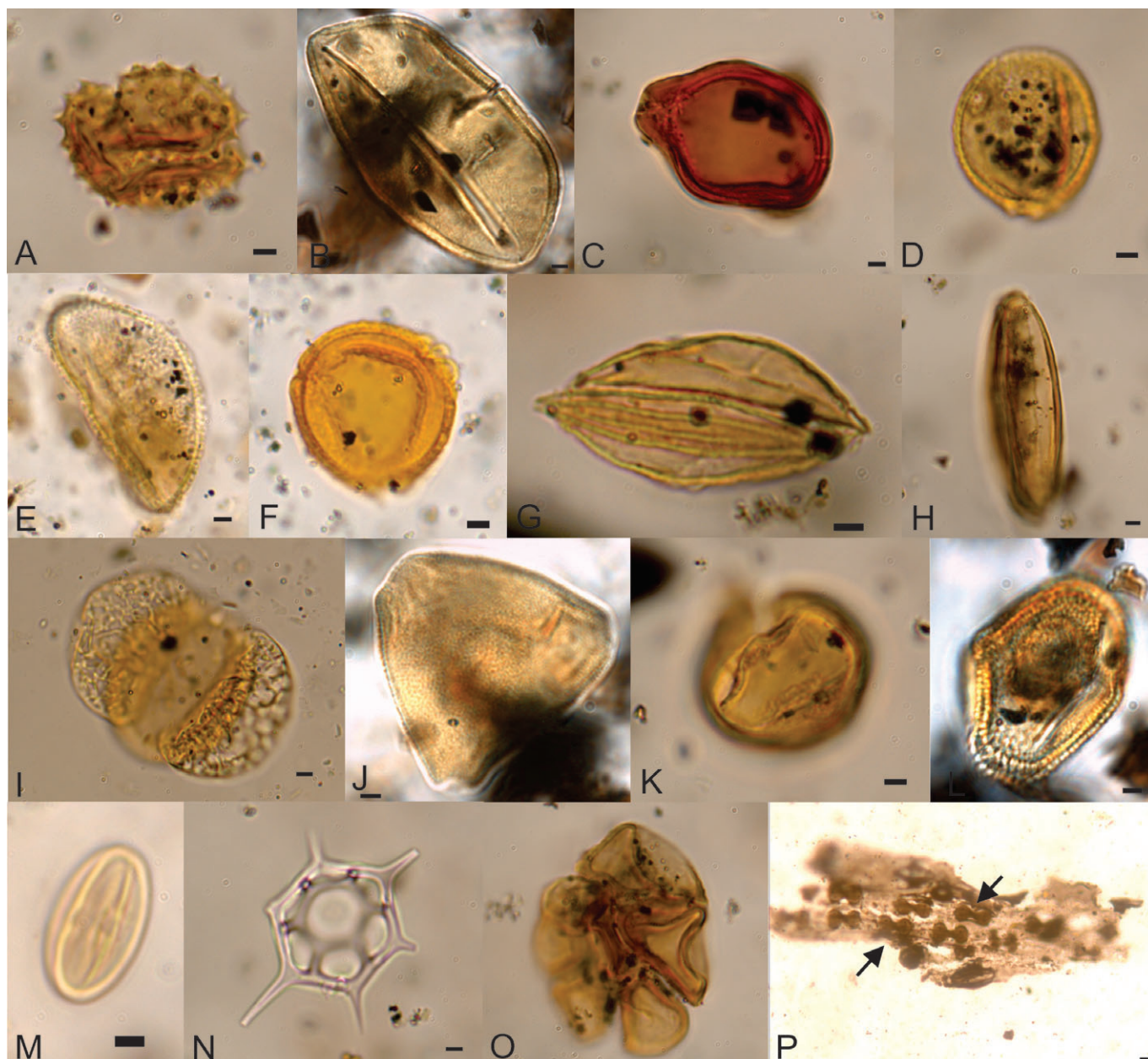
Plate 1. Common palynomorphs and microfossils from Langebaanweg. **A**, Asteraceae; **B**, *Arecipites* sp.; **C**, *Casuariniidites* sp.; **D**, *Celtipollenites* sp.; **E**, *Clavatipollenites* sp.; **F**, *Cliffortia* sp.; **G**, *Cupressacites* sp.; **H**, *Distachyapites* comb. nov.; **I**, *Podocarpidites* sp.; **J**, *Proteacidites* sp.; **K**, *Psilatricolporites operculatus*; **L**, *Rhoipites pilatus*; **M**, *Tricolporopollenites* sp. C, **N**, *Distephanus* sp.; **O**, microforaminiferal lining; **P**, grass phytoliths, bilobate short cell (arrows). Scale bars = 3 μm.

Fynbos elements (Proteaceae, Restionaceae) are present throughout the profile, but Ericaceae (*Ericipites* sp.) are absent. Pollen of Asteraceae including *Mutisiapollis* (compare Zavada & De Villiers 2000; Scott *et al.* 2006), *Pentzia*-type and *Stoebe*-type, appear regularly (max. 5%) but without fluctuations. Poaceae and *Tricolporopollenites* spp. C & D (botanical affinity: Crassulaceae, maybe Cornaceae, Scholtz 1985) are common herbs, *Sparaganiaceaepollenites* is the most widespread plant characterizing wet habitats. Poaceae are also evidenced by phytoliths (Plate 1, bilobate short cells, compare Rossouw *et al.* 2009). Restionaceae, considered by Coetzee & Rogers (1982) as a marsh element, are added to the land pollen sum in the current study. Modern Restionaceae are frost-tolerant, have a strong winter-rainfall affiliation and spread when either there is strong water-logging or drainage (Cowling *et al.*

2003). Due to the discontinuous character of the pollen profile a pollen zonation was not undertaken and palynological results were shown as histograms (Figs 5 & 6). A large number of palynomorphs were corroded and/or crumpled and therefore listed as 'varia'. Common palynomorphs and microfossils are shown in Plate 1.

DISCUSSION

Depositional setting

The sampled sediments have been ascribed to deposits of the palaeo-Berg River (Roberts 2006b) and fall into the Elandsfontyn Formation (Sandveld Group). Descriptive sedimentology of the samples showed no significant changes in colour, grain size or sorting between samples. They are likely to represent the overbank fines, formed in

Table 2. Nomenclature of pollen types (form genus respectively form species, in alphabetic order) and botanical affinities including references. Only pollen types discussed in text and shown in pollen histograms are given.

| Name (form genus, form species)	Reference	Botanical affinity	Reference	Plate 1
Arecipites sp.	Krutzsch 1970, Nichols *et al.* 1973, Jansonius	Palmae	Nichols *et al.* 1973 & Hills 1976	B
Caryapollenites sp.	Raatz 1973	Juglandaceae, *Carya*	Stuchlik *et al.* 2009	–
Casuarinidites sp.	Cookson & Pike 1954	Casuarinaceae, *Casuarina*	Cookson & Pike 1954 , compare Coetzee & Praglowski 1981	C
Celtipollenites sp.	Nagy 1969	Ulmaceae, *Celtis*	Nagy 1969, see Stuchlik *et al.* 2009	D
Celtipollenites sp. 1 (cf. *Trema*)	Stuchlik *et al.* 2009	Ulmaceae, *Trema*, *Celtis*	Stuchlik *et al.* 2009	–
Clavatipollenites sp.	Couper 1958	Chloranthaceae	Couper 1958, compare Kemp & Harris 1975	E
Crotonipollis sp.	Baksi, Deb & Siddhanta 1979	Euphorbiaceae, *Croton*	Baksi, Deb & Siddhanta 1979	–
Cupanieidites sp.	Cookson & Pike 1954 (M. major), Jansonius & Hills 1976	Sapindaceae, Cupanieae	Cookson & Pike 1954	–
Cupressacites sp.	Bolchovitina 1956, Krutzsch 1971	Cupressaceae	Krutzsch 1971	G
Distachyapites comb. nov.	Krutsch 1961	Gnetales, *Ephedra*	Krutzsch 1961, see Stuchlik *et al.* 2002	H
Ilexpollenites sp.	Thiergart 1938	*Ilex*, Aquifoliaceae	Thiergart 1938	–
Liliacidites sp.	Couper 1953	Liliaceae, Araceae, Amaryllidaceae	Pocknall 1982	–
Margocolporites rauvolfii	Salard-Cheboldaeff 1977	Apocynaceae, *Rauvolfia*	Coetzee & Rogers 1982	–
Microcachrydites sp.	Cookson 1947, Kemp & Harris 1977	*Microcachrys*	Cookson 1947, Kemp & Harris 1977	–
Mutisiapollis sp.	Macphail & Hill 1994	Asteraceae, tribe Mutisiae, *Dicoma*, *Mutisia*	Macphail & Hill 1994, Scott *et al.* 2006, compare Coetzee & Rogers 1982	–
Myricipites sp.	Wodehouse 1933	Myricaceae, *Myrica*	Wodehouse 1933	–
Myrtaceidites sp.	Jansonius & Hills 1976	Myrtaceae	Jansonius & Hills 1976	–
Palmae-trichotomosulcate	Jansonius & Hills 1976, Traverse 1988	*Phormium*, other Palmae	Jansonius & Hills 1976	–
Podocarpidites sp.	Couper 1953	*Podocarpus*, Podocarpaceae	Couper 1953	I
Proteacidites sp.	Cookson 1947, Couper 1953	Proteaceae	Cookson 1947, Couper 1953	J
Psilatricolporites operculatus	Gonzales-Guzman 1967	Euphorbiaceae, *Alchornea*	Coetzee & Rogers 1982	K
Rhoipites sp.	Wodehouse 1933	diverse, Rutaceae (?)	Pocknall & Crosbie 1982, Erdtman 1952	L
Rhuspollenites sp.	Thiele-Pfeiffer 1980	*Rhus*	Thiele-Pfeiffer 1980	–
Sparganiaceaepollenites sp.	Raatz 1937, Thiergart 1938	Sparganiaceae, *Sparganium*	Krutzsch 1970	–
Tetracolporopollenites sapotopides	Thomson & Pflug 1953	Sapotaceae, *Mimusops* (own observation)	Thomson & Pflug 1953	–
Thymelipollis sp.	Krutzsch 1966	Thymeleaceae	Jansonius & Hills 1976	–
Tricolporopollenites arnottensis	Scholtz 1985	Rubiaceae, *Anthospermum*	Scholtz 1985	–
Tricolporopollenites reticulatus	Cookson 1947, Couper 1953	*Olea*, Oleaceae	Couper 1953	–
Tricolporopollenites spp. C & D	Scholtz 1986	Cornaceae, Crassulaceae	Scholtz 1986	M
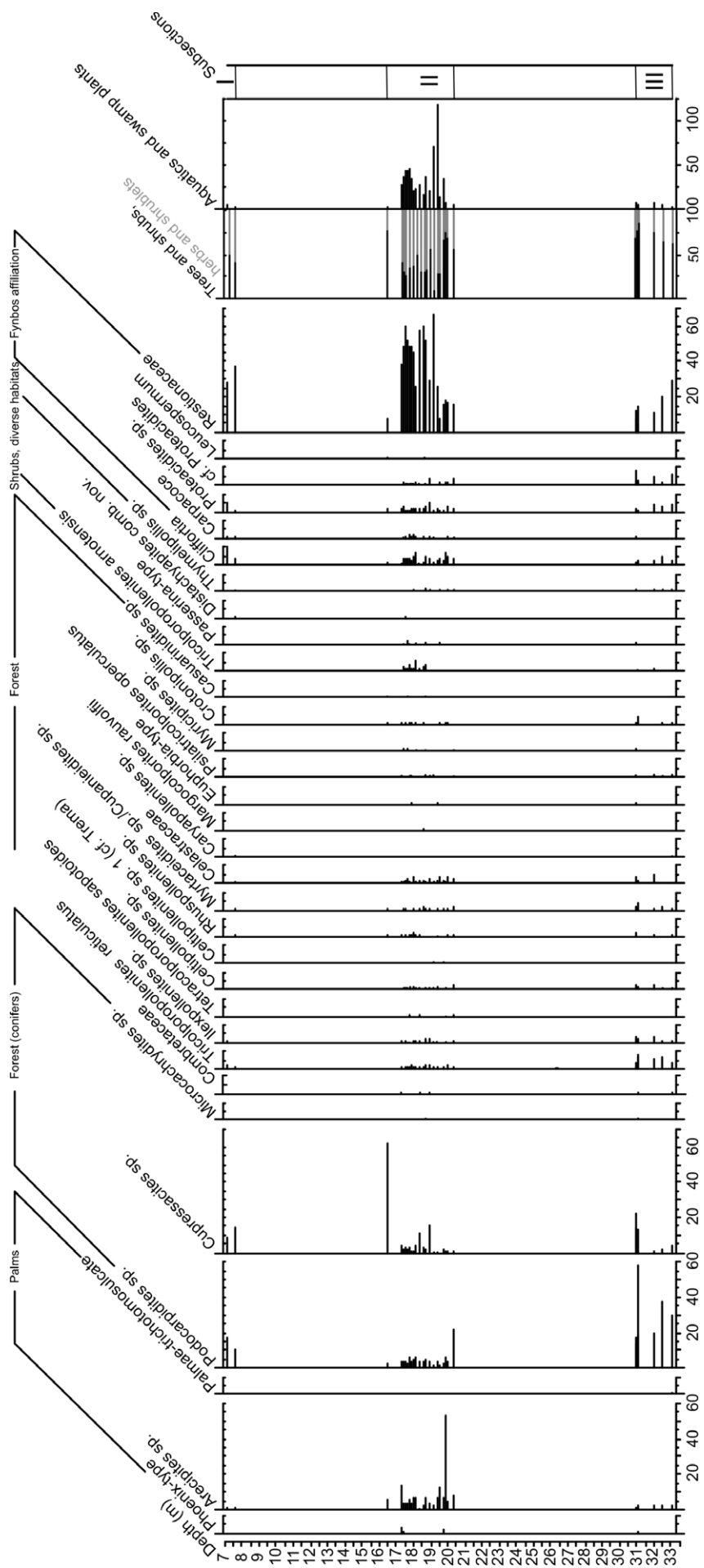
Figure 5. Histogram of condensed and selected palynomorphs of core BH 2, Langebaanweg: Palms – fynbos affiliation.
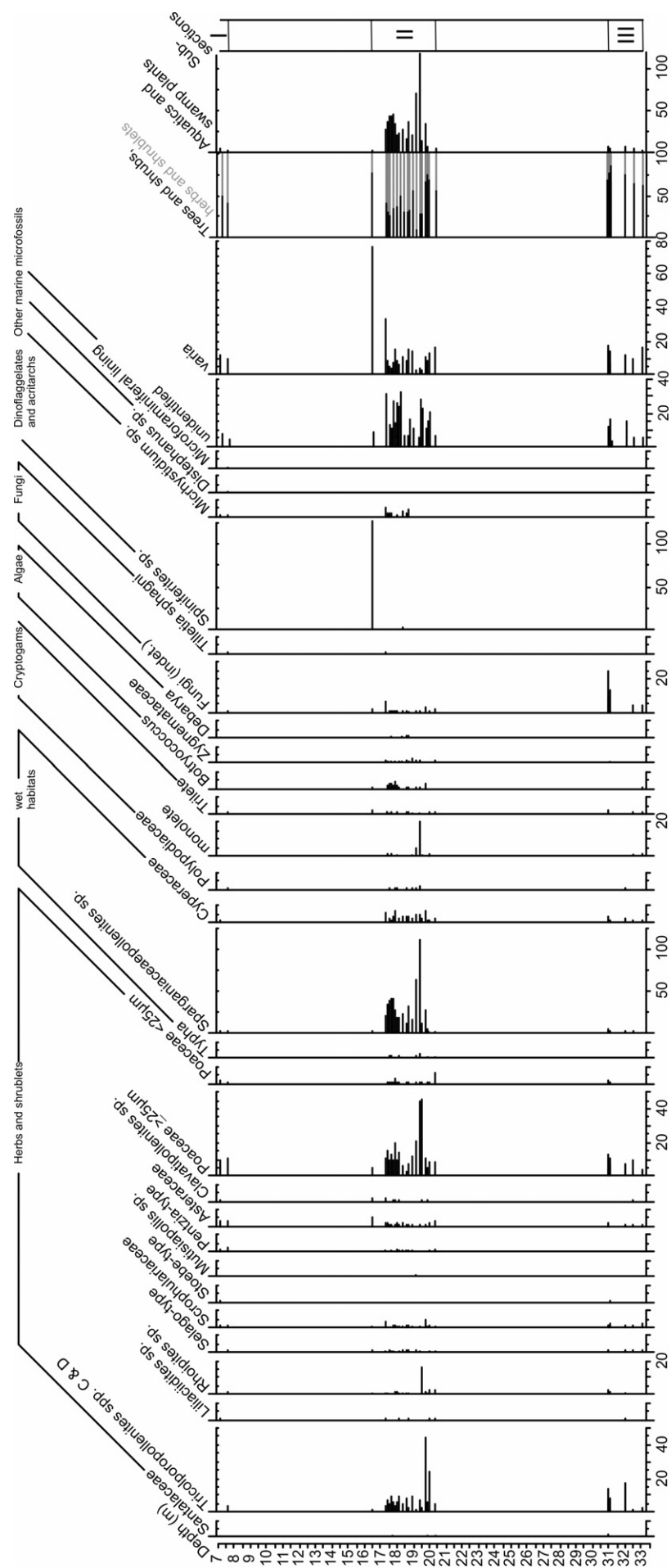
Figure 6. Histogram of condensed and selected palynomorphs of core BH 2, Langebaanweg: Herbs and shrublets – other marine microfossils.

the low-lying alluvial plain of the coastal platform, owing to their clay-rich nature and the presence of organic material (pollen, charcoal) which is terrestrially derived (Fig. 1; Miall 1985). The preliminary palynomorph studies presented here, show that the sediments studied contain pollen and charcoal (not analysed in the current study).

Vegetation history

Oxidation is likely to have destroyed the majority of palynomorphs between subsections I and II. Between subsection I and II only few corroded palynomorphs and AOM have survived.

Miocene Elandsfontyn Formation, subsection III: 30.915– 33.025 m depth

The lithology comprises organic-rich, clayey sediments. Pollen results (Figs 5 & 6) show evidence that *Podocarpidites* sp. (20–60%; botanical affiliation: *Podocarpus*, Plate 1: I) *was* widespread in the southern hemisphere and dominated forests with *Tricolporopollenites reticulatus* (see Table 2), *Microcachrydites* sp. (botanical affinity: *Microcachrydites* Podocarpaceae; Cookson 1947; Kemp & Harris 1977, compare De Villiers 1997). Taxa such as Celastraceae including *Maytenus* (not shown in Fig. 5), high *Celtipollenites* sp. percentages (Plate 1: D), rare *Ilexpollenites* sp., *Myricipites* sp., and *Proteacidites* sp. (botanical affinity: Proteaceae, Cookson 1947; Couper 1953; Martin & Harris 1974, Plate 1: J) are more common than in the upper sections. Neogene palm elements (*Arecipites* sp., Plate 1: B, Palmae-trichotomosulcate) and *Phoenix*-type are rather low. Palms are currently absent from the Cape region and are not recorded in the Pleistocene. Their presence, together with typical summer rain arboreal and herbal elements, e.g. *Psilatricolporites operculatus* (similar to recent pollen of *Alchornea* sp. which is widespread in northeastern South Africa, see Table 2, Coates-Palgrave 2002; Plate 1: K) and *Dissotis*-type, shows the occurrence of rain during the summer. Those conditions (maybe inter-annual or winter rain with a strong summer rain component) are prevalent throughout the pollen profile. Restionaceae are relatively high (30%), but decrease towards the middle of the subsection whereas podocarps and *Tricolporopollenites* spp. C & D are increasing (Fig. 5, *Tricoloporopollenites* sp. C; Plate 1: M). The top of the subsection is characterized by higher Cupressaceae (*Cupressacites* sp., possibly *Widdringtonia*, Plate 1: G), a decrease of podocarps and a slight increase of Restionaceae. The overall decrease of Restionaceae might show the spread of the podocarp forest or may be due to a lowering of the groundwater level since aquatics and swamp plants, e.g. *Sparganiaceae-pollenites* sp. and algae, were scarce. Poaceae, as noted before by Coetzee & Rogers (1982), are already well established in the area and achieve percentages between 5 and 15%.

Miocene Elandsfontyn Formation, subsection II: 16.47–20.32 m depth

The subsection initially shows subhumid conditions similar to the bottom of the whole section with high podocarp percentages (*c.* 25%), low Restionaceae (*c.* 20%)

and *Sparganiaceaepollenites* sp. (Figs 5 & 6) with *Tricolporopollenites* spp. C & D pollen, possibly indicating dry habitats on the increase. *Cupressacites* sp. pollen percentages decline, replicating other cases shown in the pollen diagram when podocarps are predominant. *Clavatipollenites* sp. (botanical affinity: Chloranthaceae, Couper 1958) and *Rhoipites* sp. (botanical affinity: diverse, maybe Rutaceae, Pocknall & Crosbie 1982; Erdtman 1952) are more common than in subsection III. *Rhoipites pilatus* (Pocknall & Crosbie 1982, Plate 1: L) is abundant in the profile. The middle part of subsection II shows a slight increase of palm pollen (e.g. *Arecipites* sp., representing Tertiary palms) with a single sample showing exceptionally high *Arecipites* sp. percentages, a strong spread of Restionaceae, Poaceae and *Sparganiaceaepollenites* sp., pointing to a high groundwater level. The pollen type resembles *Sparaganiaceaepollenites barungensis*-type as identified by Coetzee & Rogers (1982). Locally wet conditions are also signalled by ferns and algae, e.g. Zygnemataceae including *Debarya* and *Botryococcus*. *Cliffortia* sp., commonly found in damp places (Levyns 1966, Plate 1: F), is more prominent while *Typha,* which is not recorded in subsection III, appears here. Shrubby elements, e.g. *Tricolporopollenites arnotiensis* (botanical affinity: *Anthospermum,* Scholtz 1985), *Passerina*-type and *Distachypites* comb. nov. (Krutzsch 1961, botanical affinity: Gnetales, Table 2, Plate 1: H) are common. Tropical and subtropical forest elements (compare Table 3.1 in Sciscio 2011) are abundant, e.g. *Crotonipollis* sp. (botanical affinity: *Croton*, see also Scholtz 1985), *Celtipollenites* sp., *Celtipollenites* sp. 1 (*cf. Trema*, compare Stuchlik *et al.* 2009*), Tricolporopollenites microreticulatus/T. oleoides, Margocolporites rauvolfii* (botanical affinity: *Rauvolfia* sp. Coetzee & Rogers 1982), Celastraceae and Myrtaceae (*Myrtaceidites*). *Rauvolfia caffra* and *Trema orientalis*, nowadays common in eastern South Africa, are associated with groundwater or grow near water courses (Coates-Palgrave 2002). Recent pollen of both taxa resemble those found in the current study and attributed to *Margocolporites rauvolfii* and *Celtipollenites* sp. 1, respectively (own observations). Pollen of *Tetracolporopollenites sapotoides* resemble pollen of the recent tree *Mimusops* sp., another element which is now absent from the Cape region but present in the eastern half of the subcontinent (Coates-Palgrave 2002). All recent species of the genus *Mimusops* are typical of riverine vegetation, swamps or coastal regions. *Cupaniedites* sp. (*Myrtaceidites* and *Cupanieidites* of similar morphology, e.g. Frederiksen (1980), were not differentiated in the current study), and *Casuarinidites* sp. pollen likely suggests that Gondwana elements were widespread (Coetzee & Praglowski 1981, Plate 1: C). Cupanieae currently occur in America, Madagascar and Australasia, but are missing in continental Africa (see discussion in Coetzee & Muller 1984).

The top of the subsection, only represented by a single sample, is characterized by a decrease in aquatics, especially *Sparganiaceaepollenites* sp., Restionaceae and freshwater algae (e.g. Zygnemataceae, *Debarya*), whereas marine dinoflagellates (probably *Spiniferites* sp.) increase.

These observations might point to a stronger marine

influence in the upper part of the fluvial Elandsfontyn Formation, maybe even a limited ingression. More samples are needed in this subsection to identify the start of this event. *Cupressacites* sp. appear strongly (*c.* 65%) and Restionaceae pollen decrease whereas most forest taxa are reduced, which might signal a change in the forest composition. The palynological interpretation of this level is hampered by a high number of varia and unidentified types (Fig. 6). Less podocarp pollen and an increase in palm pollen supports the assumption of a warmer, more subtropical climate which was locally wetter than during the deposition of subsection III. Typical trees of riverine fringe thicket, e.g. *Rauvolfia* and *Trema*, probably occurred. Several samples above subsection II bear few pollen or are barren.

Miocene Elandsfontyn Formation, subsection I: c. 7.20–7.61 m depth

Two samples from the top of the core section show similar subhumid environmental conditions to those during the lower and middle part of subsection II. Podocarps and *Cupressacites* sp. percentages are moderate. Restionaceae are relatively high but algae, cryptograms and *Spargania-ceapollenites* sp. are low. This might indicate relatively low groundwater levels although more samples from this core section are necessary. The isolated appearance of micro-forameniferal linings (Plate 1: O) and silicoflagellates (probably *Distephanus* sp., e.g. McCartney *et al.* 1995, Plate 1: N) points to a limited influence of marine or estuarine conditions. The preliminary results point to a strong contribution of forms that presently occur in subtropical and tropical conditions in the summer rainfall region in eastern South Africa, e.g. *Crotonipollis* sp. Palmae including *Phoenix*-type, *Margocolporites rauvolfii*, or *Psilatricolporites operculatus* (botanical affinity *Alchornea*, see Table 1, Plate 1: K). Additional elements that are currently present on other Gondwana continents, e.g. South America and Australia, but are extinct in Africa are found, for example *Cupanieidites* sp., *Microcachrydites* sp., *Clavatipollenites* sp. (see Table 2, Plate 1: E). The arboreal elements indicate that the climate was altogether wetter than present-day which is in agreement with pollen data given in Coetzee & Rogers (1982) as well as Coetzee & Muller (1984). The absence of Ericaceae and the occurrence of only two *Stoebe*-type pollen, together with palms and subtropical trees, might support the indication of warm climatic conditions. Notably Ericaceae were also not identified in LBW sediments by Coetzee & Rogers (1982).

The climate might be either inter-annual (comparable to the contemporary southern Cape although palms are missing there today) or characterized by a greater contribution of summer rain. We also cannot exclude the possibility that winter rain was already established as suggested by Roussow *et al.* (2009), but higher rainfall throughout the year might have created favourable conditions for certain trees and palms. All pollen fluctuations can be explained by sea and groundwater level fluctuations (perhaps also changes in precipitation), that may have been induced by changes of the Berg River as suggested by Coetzee & Rogers (1982), including limited

marine transgressions during the deposition of the Elandsfontyn Formation in the Miocene. Strong thermal changes are not clearly evident from the pollen record so far.

Biogeochemical constraints

The relatively low bulk TOC data (2.7–10.7 wt%) indicate a substantial dilution of organic matter by inorganic clastic input during deposition, and/or significant degradation after deposition (Meyers 2003; Mielnik *et al.* 2009). High TOC/TN ratios (mean = 35; STD = 3.01) and C_{org}/N_{org} (mean 41; STD = 10.52) suggest that the bulk organic matter was not derived from algae in a lacustrinal setting (itself typified by C_{org}/N_{org} ratios of 13–14), but rather represents largely terrestrial, higher plant-derived material such as lignin and cellulose (Figs 3 & 4; Meyers & Lallier-Vergés 1999; Meyers 2003). Cellulose, however, will not be preserved in the sediments in significant amounts as it is relatively labile. Lignin is not especially recalcitrant unless conditions are anoxic. The high C/N ratio may represent some preferential degradation of N-containing organic matter and also preferential preservation of high C-containing OM – e.g. aliphatics and aromatics, some of which may originally be plant-derived (i.e. fatty acids, cutin), and/or some of which may be modified with diagenesis (Carr *et al.* 2011).

The TOC maximum at 19.50 m corresponds to a BIT index of 1.0 which signifies a purely terrestrial source of the organic matter during sediment deposition in this instance (Fig. 4; Hopmans *et al.* 2004). By contrast, the stratigraphically uppermost samples (18.40–17.0 m) exhibit relatively lower BIT, TOC, lower TOC/TN and higher $\delta^{13}C$ values, suggesting a likely lacustrine origin of the organic matter, which may or may not carry an additional coastal marine contribution (Hopmans *et al.* 2004; Weijers *et al.* 2006b).

The meandering nature of the palaeo-Berg River means that channels may have been periodically abandoned forming ephemeral (ox-bow) lakes (Miall 1985; Nichols & Fisher 2007). Furthermore, transitional lakes may have formed in the distal regions of a distributary system on the alluvial plain, during periods of increased water supply (Nichols & Fisher 2007). The changes in the proxy signals as documented here may also be related to the lateral migration of the river system and the concurrent changes in facies: the overbank fine deposits are likely to contain branched GDGTs within a developed soil profile, which are less likely to have been affected by riverine branched GDGT production. By contrast, the MBT-CBT index of channel-fills is likely to be influenced by the *in situ* biological production of branched GDGTs within the river system. Facies changes are thus likely to reflect corresponding changes on the source, production and/or preservation of branched GDGTs, and therefore the interpretation of the respective proxy signals.

Loomis *et al.* (2011) have cautioned against using the temperature calibration from soils if there is any *in situ* branched GDGT production in a lake. The range of reconstructed temperatures seems rather high and that while this may imply major climatic variability, but the these

Table 3. Comparison of modern Saldanha Bay mean annual air temperatures, sea surface temperatures, carbon dioxide concentration and precipitation levels in comparison to measured and predicated Miocene values.

Simulation	Mean annual temperature (°C)	Southern ocean mean annual surface temperature (°C)	Atmospheric carbon dioxide (ppmv)	Mean annual rainfall (mm)
Saldanha Bay (present day)	17°C Adelana *et al.* 2010	16.3°C	387 ppmv	256 mm
	Mean annual summer air temp. range: 18.4–27.5°C	Ranges from 10–16°C within the bay	Tripati *et al.* (2009)	
	Mean annual winter temp. range 7.1–14.9 °C (Hanekom *et al.* 2009)			
Middle Miocene	Globally 3–6°C higher	~27°C at 14Ma	Globally 300–600 ppmv (Kurschner *et al.* 2008)	Globally higher than present by at 200 mm (Tong *et al.* 2009)
	12.4–26.6°C (this study)	Decreasing to 18°C at 5 Ma (Dupont *et. al.* 2009)		
	But also MATs lower by 1–5°C than modern Cape MAT			

data are not always straight forward to interpret, especially given that the source, *in situ* terrestrial or lacustrine, of the branched GDGTs within the sediments is not known (Blaga *et al.* 2010).

On the basis of basement configuration studies of the LBW area by Roberts *et al.* (2013), the Langebaan embayment at the mouth of Saldanha Bay, is likely to have periodically held an expanse of water into which palaeo-Berg River sediments (Elandsfontyn Formation) were deposited. The location of the core in this study is within the Langebaan embayment, thus making this assumption valid (Fig. 1). Ultimately, with lateral migration of the river system, the inter-fingering of channel, overbank fines and lacustrine facies will develop as seen in the LBW core stratigraphically and geochemically.

Given the altitude of the site above modern sea level (~50 m), and taking into consideration a Miocene sea-level high stand potentially of 150 m (Haq *et al.* 1987; Kominz *et al.* 1998) it is possible that the LBW coastal platform would have been periodically inundated. This would have been so even if a pulse of uplift in the early-middle Miocene of 150 m is considered (Partridge & Maud 1987). Ultimately, this could account for the higher crenarchaeol abundance within the BIT indices noted for samples at 17 m and 18.40 m depth, as well as for the relatively lowered TOC and δ^{13}C values, reflecting periodic declines in productivity with estuarine development/salt water incursion, and/or poorer preservation within the wetland of the overbank fines (Meyers 2003). The high C_{org}/N_{org} ratio of these uppermost 'high' MAT samples, a good indicator of the presence of lignin and cellulose of higher plants, points to a higher plant-derived depositional setting (Sharp 2007) that overwhelmed organic matter supply by lacustrine species (Meyers 2003; Sharp 2007).

The samples with low MATs corresponded to more acidic sediment pH values (Fig. 4). Modern soil acidity is related to rainfall, with acidic soil profiles linked to increases in precipitation (Weijers *et al.* 2007a). It can therefore be inferred that sediments deposited below 19.50 m depth, experienced consistently high rainfall, and that plant matter deposition and decomposition facilitated the maintenance of low-pH sediments (Killops & Killops 2005). It is generally known, however, that lowered MAT

(as shown by the samples between 19.50 and 33 m) are likely to decrease the rate of organic matter degradation, and decrease the sediment pH, as lesser amounts of humic substances are produced (Coûteaux *et al.* 1995; Killops & Killops 2005). Accordingly, the high percentages of *Podocarpidites* pollen below *c.* 20m might also be due to edaphic reasons, since *Podocarpus* is more adapted to acidic soils (compare Sciscio 2011).

By contrast, the uppermost samples at 17 m and 18.40 m depth below sea level have high MAT and near neutral pH. These samples are associated with warm temperatures and are likely to also represent a period of high rainfall and a sea-level high stand during the late middle Miocene (Roberts 2006d) as indicated by marine microfossils (tentatively identified as *Spiniferites* sp.) found in the current study. Although the above is at odds with increased rainfall which should lower the acidity of the sediment, if these uppermost samples are indeed lacustrine or estuarine in origin it is likely that their pH was buffered by the overlying water column.

Regional climate and forcing mechanisms

On a regional scale, pollen evidence from sediments of the uppermost Elandsfontyn Formation indicates periodic flooding and drought within the Saldanha environs during the late middle Miocene (Hendey 1981; Coetzee & Rogers 1982). Globally, the middle Miocene is considered a time of reduced seasonality and relatively stable climate with elevated temperatures, high humidity and rainfall. At LBW rainfall variability, with strongly developed wet (i.e. implied from palynomorph taxa) and dry (charcoal; Roberts *et al.* 2013) year distinctions, are likely the result of changes in the strength and position of the westerlies, cloud bands and SSTs (Coetzee & Rogers 1982; Scott 1994; Vogel 2003). These forcing parameters of continental climate are strongly linked to Southern Ocean conditions (Gallagher *et al.* 2001; Romero *et al.* 2005), and ultimately coupled to Milankovitch cycles. The 'low' mean annual temperatures (15.8 ±3.3°C) from stratigraphically lower (older) samples are comparable with modern winter mean annual temperatures for the Cape region today (average ±12.5°C; Table 3; Cowling *et al.* 2009), but are more closely associated with the modern MATs (17°C).

These low MATs, for the locality of the sample site and in the knowledge that the Miocene Epoch is generally typified by raised global temperatures, is therefore unusual.

The samples with recorded 'high' MATs (24 ± 3.6°C) were deposited most probably during the late middle Miocene marine transgression due to the low BIT indices, high MATs and sedimentology (Table 1). Study of the palaeo-Atlantic through the offshore Cape Basin sediments from ODP Site 1085 indicate declining Miocene SSTs during this time from >27°C to 18°C (late Miocene; Table 3; Westerhold *et al.* 2005, Dupont *et al.* 2009, Rommerskirchern *et al.* 2011). These high SSTs at the end of the MMCO would complement the high MATs obtained for samples at 18.40 m (27°C) and 17 m (22°C) depth when it is considered that theocean is a strong forcing mechanism of continental air temperatures (Dommenget 2009). However, the palynomorph content does not reflect those temperature shifts and rather shows hydric fluctuations and possible marine ingressions as pointed out above.

Furthermore, during the deposition of the sampled sediments it is likely that rainfall was elevated. In southern Africa, higher rainfall would be expected with elevated SSTs due to fluctuations in parameters of the Southern Oscillation such as evaporation and wind (Table 3; Lutjeharms *et al.* 2001). Elevated SSTs give rise to the thermal expansion of ocean water and a concurrent rise in sea level, supporting the late Miocene transgression reported on the west coast (Kennett 1977). Additionally, warm oceans have a lowered capacity for the storage of CO_2, further increasing the amount of this greenhouse gas in the atmosphere and the raising of global temperatures (Kürschner *et al.* 2008).

The middle to late Miocene global cooling trend, as known from co-varying planktonic and benthic foraminifers stable oxygen isotope studies, is punctuated by several glaciation events (Mi-events) (Westerhold *et al.* 2005). The cooling of SSTs (Table 3) has been related to glacio-eustatic sea-level fall (due to the WAIS growth), and bottom-water cooling associated with increased upwelling (Westerhold *et al.* 2005; Dupont *et al.* 2009). There is an intimate connection between marine carbon draw-down and SST cooling and continental climate, with short- and long-term oscillations of Mio-Pliocene terrestrial climate (of the northern hemisphere) seen in the marine isotope record (Mosbrugger *et al.* 2005). This may also be the case with the western coast LBW site; however, more detailed sampling would be required stratigraphically higher in the section following the high MATs of the uppermost samples at 18.40 m and 17 m.

CONCLUSION AND FUTURE RESEARCH

The analysis of branched tetraether membrane lipids derived from organic-rich clays at LBW yielded credible results, and underpins the usefulness of the method in palaeo-environmental interpretation as long as the limitations are recognized. Employing the novel CBT/MBT proxy at LBW has complemented palaeoclimate interpretations from previous palynological studies and the present work by quantifying actual temperature fluctuations through time. The results encourage further work at Langebaanweg and at other similar Western Cape sites such as Noordhoek (Cape Town) or other sites in the region mentioned by Sciscio (2011) and Roberts *et al.* (2013).

The present pollen analysis confirmed and extended the conclusions drawn from previous studies. The BH2 core shows an upward transition from fluvial to estuarine conditions, which would indicate rising groundwater levels in the light of rising sea level, a notion reinforced by the presence of marine microfossils and lowered BIT indices. However, the driving force(s) behind the observed fluctuations in palynofloras, i.e. either climate change or a rise in the water table independently of climate (possibly caused by sea-level rise), can only be resolved by further data input.

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