Geochemical records of palaeoenvironmental controls on peat forming processes in the Mfabeni peatland, Kwazulu Natal, South Africa since the Late Pleistocene

A. Baker, Joyanto Routh, M. Blaauw and A.N. Roychoudhury

Linköping University Post Print

N.B.: When citing this work, cite the original article.

Original Publication:

A. Baker, Joyanto Routh, M. Blaauw and A.N. Roychoudhury, Geochemical records of palaeoenvironmental controls on peat forming processes in the Mfabeni peatland, Kwazulu Natal, South Africa since the Late Pleistocene, 2014, Palaeogeography, Palaeoclimatology, Palaeoecology, (395), , 95-106.

<http://dx.doi.org/10.1016/j.palaeo.2013.12.019>

Copyright: Elsevier

<http://www.elsevier.com/>

Postprint available at: Linköping University Electronic Press

<http://urn.kb.se/resolve?urn=urn:nbn:se:liu:diva-106296>

Geochemical records of palaeoenvironmental controls on peat forming processes in the Mfabeni peatland, Kwazulu Natal, South Africa since the Late Pleistocene

A. Baker^a, J. Routh^{b*}, M. Blaauw^c, A.N. Roychoudhury^a

^aDepartment of Earth Sciences, Stellenbosch University, Private Bag X1, Matieland, 7600, South Africa

^bDepartment of Water and Environmental Studies, Linköping University, 581 83 Linköping, Sweden

^cSchool of Geography, Archaeology and Palaeoecology, Queen's University Belfast, Belfast BT7 1NN, U.K.

Abstract

The Mfabeni peatland is the only known sub-tropical coastal fen that transcends the Last Glacial Maximum (LGM). This ca. 10 m thick peat sequence provides a continuous sedimentation record spanning from the late Pleistocene to present (basal age c. 47 kcal yr BP). We investigated the palaeoenvironmental controls on peat formation and organic matter source input at the Mfabeni fen by: 1) exploring geochemical records (mass accumulation rate, total organic carbon, carbon accumulation rate, $\delta^{13}\text{C}$, $\delta^{15}\text{N}$ and C/N ratio) to delineate primary production, organic matter source input, preservation and diagenetic processes, and 2) employ these geochemical signatures to reconstruct the palaeoenvironmental conditions and prevailing climate that drove carbon accumulation in the peatland. We established that the Mfabeni peat sediments have undergone minimal diagenetic alteration. The peat sequence was divided into 5 linear sedimentation rate (LSR) stages indicating distinct changes in climate and hydrological conditions: LSR stage 1 (c. 47 to c. 32.2 kcal yr BP): predominantly cool and wet climate with C4 plant assemblages, interrupted by two short warming events. LSR stage 2 (c. 32.2 to c. 27.6 kcal yr BP): dry and

*Corresponding author: Joyanto Routh, Department of Water and Environmental Studies, Linköping University, 581 83 Linköping, Sweden. Tel: +46 7049 31066. Email: joyanto.routh@liu.se

windy climate followed by a brief warm and wet period with increased C4 sedge swamp vegetation. LSR stage 3 (c. 27.6 to c. 20.3 kcal yr BP): initial cool and wet period with prevailing C4 sedge plant assemblage until c. 23 kcal yr BP; then an abrupt change to dry and cool glacial conditions and steady increases in C3 grasses. LSR stage 4 (c. 20.3 to c. 10.4 kcal yr BP): continuation of cool and dry conditions and strong C3 grassland signature until c. 15 kcal yr BP, after which precipitation increases. LSR stage 5 (c. 10.4 kcal yr BP to present): characterized by extreme fluctuations between pervasive wet and warm to cool interglacial conditions with intermittent abrupt millennial-scale cooling/drying events and oscillations between C3 and C4 plant assemblages. In this study we reconstructed a high-resolution record of local hydrology, bulk plant assemblage and inferred climate since the Late Pleistocene, which suggest an anti-phase link between Southern African and the Northern Hemisphere, most notably during Heinrich (5 to 2) and Younger Dryas events.

Key words: Palaeoenvironment; subtropical peatland; stable isotopes; elemental analyses; carbon accumulation.

1. Introduction

Peatlands play a pivotal role in the global carbon (C) cycle, serving as a direct link between the short-term (atmosphere, biosphere and hydrosphere) and long-term (geosphere) carbon reservoirs. Under sequestering conditions, peatlands serve as sinks for atmospheric CO₂, important sources of CH₄ and exporters of fluvial dissolved and particulate organic carbon to down-stream ecosystems (Worrall et al., 2003). The balance between ecosystem productivity and respiration, controlled primarily by precipitation, temperature, water table fluctuation and local topography, determines if a peatland acts as a C sink or source.

Contemporary global peatland C stocks have been estimated to be in excess of 450 Petagrams (1 Pg = 10^{15} g), equivalent to 75% of CO₂ stored in the atmosphere at any given time (Strack, 2008). The vast majority (~90%) of peatlands are found in the Northern Hemisphere temperate and boreal regions, with tropical and sub-tropical peatlands constituting the balance (Immirzi et al., 1992; Page et al., 2011). However, because of their relatively higher carbon accumulation rate (CAR; Chimner and Ewel, 2005; Strack, 2008), tropical peatlands are estimated to represent up to a quarter of the potential global peatland C stock (Strack, 2008; Page et al., 2011), and they are at a greater risk of degradation and overexploitation due to their proximity to populated areas and climate change (Rieley et al., 1996). Currently, there is limited scientific understanding of the processes that regulate C cycling and accumulation in tropical peatlands (Chimner and Ewel, 2005), and how changing climate and increasing anthropogenic pressures will affect low latitude peatland systems and their ability to sequester C.

Southern Africa is situated at the interface of tropical and temperate climate systems. The region is influenced by the largest asymmetrical cross-continental tropical convection (Stokes et al., 1997) as a consequence of seasonal fluctuations in the Inter Tropical Convergence Zone (ITCZ), and large temperature gradients between the warm Agulhas and cold Benguela oceanic currents (Preston-Whyte and Tyson, 1998; Tyson and Preston-Whyte, 2000). Due to the topography and semi-arid climate of Southern Africa, archives are not commonly preserved, and few continuous palaeoenvironment records exist (Chase and Meadows, 2007), despite strong evidence of climate variability from Antarctic ice cores and low latitude African hydrological investigations (Bard et al., 1997; Blunier et al., 1998; Gasse, 2000; Stocker, 2000; Stenni et al., 2001). A few detailed limnology studies have been

undertaken in regional lakes (Meadows et al., 1996; Meadows and Baxter, 1999; Partridge, 2002; Kristen et al., 2010), but due to the fact that most of southern Africa is water scarce, freshwater lakes are uncommon. Speleothem archives from South African caves have yielded high-resolution climate records (Talma and Vogel, 1992; Lee-Thorp et al., 2001 Holmgren et al., 2003; Holzkämper et al., 2009), but these records either do not span the Last Glacial Maximum (LGM) or are incomplete with at least one or more hiatuses of between 2.5 and 10 kyr. Several palynology studies have been undertaken in regional coastal peatlands and hyrax midden deposits (Finch and Hill, 2008; Neumann et al., 2008, 2010; Walther et al., 2011; Valsecchi et al., 2013) however, with the exception of the Mfabeni peatland study (Finch and Hill, 2008) the palynology records are limited to the deglacial and Holocene periods. To our knowledge, the only geochemical palaeoclimatic study undertaken on the sub-tropical Braamhoek peatland in the austral summer rainfall region of South Africa is by Norström et al. (2009). This inland wetland is located on the Eastern escarpment at 1700 m a.s.l. and lies over 290 km from the nearest coastline with a palaeorecord extending only as far back as c. 16 kcal yr BP.

The lack of high-resolution terrestrial palaeoclimate records on the African sub-continent continues to hinder the understanding of past climate forcing factors and their environmental impacts (Chase and Meadows, 2007; Gasse et al., 2008). Additional high resolution multi-proxy and multi-archive studies are therefore needed to elucidate past climate fluctuations and modelling of the ensuing environmental responses to these changes in the future. In this context, organic matter (OM) rich peat deposits are ideally suited for palaeoenvironmental studies as they are well preserved archives that are subject to mainly autochthonous depositional regimes that are largely regulated by climate (Strack,

2008). The aim of this research is to investigate climatic and environmental conditions that have prevailed in the southern African region since the Late Pleistocene. Our objective is to reconstruct the palaeoenvironmental controls on past C accumulation and OM source input at the sub-tropical coastal Mfabeni fen by: 1) delineating primary production, OM source input, OM preservation and diagenetic processes which affected the formation of these peat deposits, and 2) using multiple geochemical proxy signatures such as, mass accumulation rate (MAR), total organic carbon (TOC), carbon accumulation rate (CAR), $\delta^{13}\text{C}$, $\delta^{15}\text{N}$ and C/N, to reconstruct the palaeoenvironmental conditions and prevailing climate in the peatland over the last c. 47 kcal yr BP.

2. Methods

2.1. Site description

St Lucia is one of the largest estuarine systems on the African continent (Vrdoljak and Hart, 2007). It falls within the UNESCO World Heritage iSimangaliso Wetland Park, situated on the northern shores of Kwazulu-Natal province, South Africa (Fig. 1). Lake St Lucia, the dominant north-south aligned water body, has an extent of 350 km² and an average depth of only 90 cm. At its northern end, the lake is fed by four regional rivers with a combined catchment of approximately 6085 km², namely the uMkhuze, Nyalazi, Mzinene and Hluhluwe rivers. To the south, a narrow 22 km long waterway sporadically links the lake to the Indian Ocean, and to the east the lake is fed with fresh water by the large Maputoland unconfined aquifer (Kelbe et al., 1995; Taylor et al., 2006a).

The Mfabeni fen lies within an interdunal valley (Botha and Porat, 2007) on the eastern shores of Lake St Lucia, running parallel to the coastline and measuring ca. 10 x 3 km

(Clulow et al., 2012; Grundling et al., 2013), and thickness of up to ca. 11 m (Grundling, 2001; Grundling et al., 2013). The hydrological regime of the fen is controlled by local precipitation and circum-neutral Ca^{2+} and HCO_3^- dominated groundwater emanating from the Maputaland aquifer (Venter, 2003; Taylor et al., 2006b; Grundling et al., 2013). The region lies within a humid, sub-tropical climate which experiences primarily austral summer rainfall of between 900 and 1200 mm/yr (Grundling, 2001; Taylor et al., 2006a). However, modern rainfall data indicate distinctive wet and dry cyclical events which can effectively halve the annual precipitation for extended periods (Bate and Taylor, 2008). The surface waters on the eastern part of the minerotrophic Mfabeni peatland drain northwards into the southern part of Lake Bhangasi, while the western sections drain southwards into Lake St Lucia (Grundling, 2001; Clulow et al., 2012).

The Mfabeni fen forms part of the greater Natal Mire Complex (NMC; Fig. 1) that extends from southern Mozambique to the south of Richards Bay, Kwazulu-Natal (Smuts, 1992). The NMC falls within the Maputaland group of coastal Cenozoic deposits, constrained in the west by the Lebombo monocline, uPongola and uMkhuze River valleys (Botha and Porat, 2007). The Mfabeni peatland accumulated by valley infilling on a lacustrine or intertidal non-permeable clay layer (Grundling et al., 2013) within the reworked late Pleistocene KwaMbonambi Formation coastal dune depression (Smuts, 1992), as a consequence of blockage of the Nkazana palaeo-channel and sustained groundwater input from the Maputaland aquifer (Grundling et al., 2013).

The iSimangaleso wetland park encompasses several heterogeneous habitats, mainly as a response to topography, hydrology and historical land use (Vrdoljak and Hart, 2007).

Mucina et al. (2006) broadly categorised the wetland habitats as Maputaland wooded

grassland, coastal belt and sub-tropical freshwater wetlands surrounded by northern coastal forests. The Mfabeni fen vegetation is largely represented by herbaceous reed sedge vegetation (Finch, 2005), and is dominated by *Rhynchospora holoschoenoide*, *Fimbristylis bivalve*, *Panicum glandulopaniculatum* and *Ischaemum fasciculatum* (Lubke et al., 1992; Vaeret and Sokolic, 2008).

2.2. Sampling techniques

A 810 cm deep sediment core, SL6, was extracted from the middle of the peatland (28.15021°S; 32.52508°E) using a 5 cm diameter x 50 cm length Russian peat corer in June 2011. The core was logged in the field and described and sectioned into 1–2 cm increments in the laboratory. All samples were weighed before freeze-drying, then again afterwards to calculate bulk density and porosity of each segment.

2.3. Radiocarbon dating / age model

Evenly spaced samples were sent for ¹⁴C dating at the Poznań Radiocarbon Laboratory, Poland. The sediments were chemically pre-treated as described by Brock et al. (2010) with the exception of using 0.25M instead of 1M HCl; samples SL1 31-32 and SL4 89-90 were not treated with NaOH due to their low carbon content. The samples were combusted with CuO and Ag wool at 900°C for 10 hrs and the CO₂ was reduced to pure graphite in a vacuum line as described by Czernik and Goslar (2001). Coal or IAEA C1 Carrara Marble and international modern Oxalic Acid II standards were subjected to the same pre-treatment and combustion procedure. The ¹⁴C content of the samples were measured on a Compact Carbon AMS (National Electrostatics Corporation, USA) as described by Goslar et al. (2004). The conventional ¹⁴C age was calculated using a correction factor for isotopic fractionation as per Stuiver and Polach (1977).

Since the ages of some of the ^{14}C dates lie beyond the limit of the southern hemisphere calibration curve (McCormac et al., 2004), dates were calibrated using the Northern hemisphere terrestrial calibration curve IntCal09 (Reimer et al., 2006) while applying a southern hemisphere offset of 40 ± 20 ^{14}C years. Post-bomb ages were calibrated using the southern hemisphere post-bomb curve of Hua and Barbetti (2004). All ages were adjusted within the Bayesian framework, using age-depth modelling software Bacon (Blaauw and Christen, 2011). This method divides a core into sections and models the accumulation rate for each of these sections. Accumulation rates were constrained by prior information (here a gamma distribution with mean 50 yr/cm and shape 1.5), and the variability of accumulation rate from one depth to the next was constrained by a 'memory' parameter (here using the default beta distribution with mean 0.7 and strength 4). Stable runs were obtained using multiple Markov Chain Monte Carlo (MCMC) iterations.

2.4. Elemental and stable isotope analyses

Selection of acid treated and raw samples were analysed for C and N stable isotopes composition and elemental ratios at the Department of Archaeology, University of Cape Town. Peat samples were combusted in a Thermo Scientific Flash 2000 organic elemental analyser, coupled to a Thermo Scientific Delta V Plus isotope ratio mass spectrometer via a Thermo Scientific ConFlo IV gas control unit (detection limit $5\mu\text{g}$). The sand samples (low C and N wt%) were combusted in a Thermo Finnigan Flash EA 1112 series elemental analyzer, coupled to a Thermo electron Delta Plus XP isotope ratio mass spectrometer via a Thermo Finnigan ConFlo III gas control unit (detection limit $15\mu\text{g}$). In-house standards used were: chocolate/egg mixture, Australian National University sucrose, Merck proteinaceous gel and dried lentils, calibrated against International Atomic Energy Agency standards (N1, N2, NBS

18, 19, SMOW and SLAP). The precision for both analytical systems was 0.05 and 0.08‰ for N and C, respectively. Nitrogen (N) isotopic composition is expressed relative to atmospheric N, whereas C is expressed relative to Pee-Dee Belemnite.

2.5. C3/C4 plant mass balance

To broadly gauge the relative proportions of C input from C4 and C3 plants during peat accumulation in Mfabeni, a mass balance equation was used as per Gillson et al. (2004) and Boutton et al. (1998):

$$\delta^{13}\text{C}_{\text{SOM}} = (\delta^{13}\text{C}_{\text{C4}_{\text{plants}}}) (x) + (\delta^{13}\text{C}_{\text{C3}_{\text{plants}}}) (1-x) \quad (\text{Equation 1})$$

Where:

- $\delta^{13}\text{C}_{\text{SOM}}$ bulk stable carbon isotope composition of C_{org} in sample
- $\delta^{13}\text{C}_{\text{C4}_{\text{plants}}}$ and $\delta^{13}\text{C}_{\text{C3}_{\text{plants}}}$ are the average stable C isotope values from regional C4 (-13.2‰) and C3 (-27.5‰) vegetation (Muzuka, 1999).
- x is the proportion of C from C4 plant sources and $(1-x)$ the proportion of C from C3 plants.

3. Results

3.1. Core description

Core SL6 consists of 7 different sediment types (Fig. 2), dominated by peat with occasional sandy lenses of varying thicknesses: black fine-grained amorphous peat (810-610 cm), dark brown fine-grained peat with grey sand mottled zones (610 – 535 cm), black fine-grained peat with grey sand mottled zones (535 – 440 cm), black fine-grained amorphous peat with increasing sandy texture with depth (440 - 340 cm), black fine-grained amorphous peat with minimal rootlets (340 - 110 cm), black fine-grained amorphous peat, with extensive

rootlets (110 - 61 cm), and dark brown “fibrous” peat transitioning to fine grained black amorphous peat sediments with depth (61 - 0cm).

The peat sediments show an average bulk density of 0.29, 0.34, 0.28, 0.28, 0.24 g.cm⁻³ for linear sedimentation rate (LSR, see section 3.2 for explanation) stages 1 to 5, respectively (Table 1). The average core porosity for the dominant peat sediment was calculated to be 0.7.

3.2. Age model

The SL6 basal age recorded at 805 cm is c. 47.0 kcal yr BP (Table 2; Fig. 3). Peat sediments dominate (Fig. 2) from the base of the core up to the age of c. 31.9 kcal yr BP (609 cm). Above this a succession of sand lenses cross cut the core, with relatively low C% up to c. 28.8 kcal yr BP (540 cm). A fining up transition occurs from “sandy” peat to peaty sediments until a modelled age of c. 14.3 kcal yr BP (350 cm), after which amorphous peat sediments with increasing rootlet content dominate (340 - 61 cm), ending in the top “fibrous” surface peat layer.

The linear sedimentation rates (LSR) calculated for core SL6 suggests several changes in sedimentation regimes (Fig. 4; Table 1). When a trendline is fitted to grouped data points, the lower third of the core, spanning c. 47.0 to c. 32.4 kcal yr BP, displays an average LSR of 0.13 mm.yr⁻¹ (LSR Stage 1). The average LSR increases to 0.22 mm.yr⁻¹ between c. 32.1 and c. 27.9 kcal yr BP (LSR Stage 2), declining to 0.14 mm.yr⁻¹ (c. 27.6 to c. 20.3 kcal yr BP; LSR Stage 3) before dropping to its lowest average rate of 0.10 mm.yr⁻¹ between c. 19.8 and c. 10.4 kcal yr BP (LSR Stage 4). Thereafter, a marked increase in average LSR to 0.29 mm.yr⁻¹ occurs during the Holocene (LSR Stage 5).

3.3. Past sedimentation and C measurements

3.3.1. Mass accumulation rate (MAR)

MAR in core SL6 fluctuates between a minimum of 20.8 (c. 14.3 kcal yr BP) and a maximum of 103 g.m⁻².yr⁻¹ (c. 3.5 kcal yr BP) with a total core MAR average of 57.5 g.m⁻².yr⁻¹ (Fig. 4 and Table 1). LSR stages 1, 3 and 4 display overall lower than core average MAR values of 37.6, 41.4 and 28.1 g.m⁻².yr⁻¹, respectively, punctuated by abrupt shifts to above core average MAR values for LSR stages 2 and 5 (76.4 and 70.9 g.m⁻².yr⁻¹, respectively). LSR stage 5 shows the least consistency with overall elevated MAR values, and short intermittent excursions to below average MAR values at c. 7.1, c. 5.3, c. 4.7 and c. 1.4 kcal yr BP.

3.3.2. Total organic carbon (TOC)

The sediment TOC of core SL6, for the most part, trends similarly but opposite to MAR records, fluctuating between a minimum of 9.65 g C.m⁻² (c. 30.4 kcal yr BP) and a maximum of 1600 g C.m⁻² (c. 44.6 kcal yr BP) with a total core average of 981 g C.m⁻² (Fig. 4; Table 1). During LSR stage 1, TOC fluctuates dynamically between the core maximum and 327 g C.m⁻² at c. 35.6 kcal yr BP. LSR stage 2 displays the lowest TOC stage average (553 g C.m⁻²) with the exception of an abrupt increase of 584 g C.m⁻² towards the latter part of the stage (28.5 kcal yr BP). LSR stage 3 fluctuates around the core average until a sharp decrease occurs at c. 22.3 kcal yr BP followed by a relatively low TOC values till c. 14.3 (350 cm) kcal yr BP. Thereafter TOC increases in the lead up to and during the Holocene which exhibits the highest stage TOC average (1108 g C.m⁻²; LSR stage 5), , despite sporadic sharp declines associated with increases in MAR at c. 5.3 kcal yr BP and c. 7.7 kcal yr BP.

3.3.3. Carbon accumulation rate (CAR)

From the bottom of core SL6 up until c. 10.2 kcal yr BP, the CAR fluctuates below the total

core average of $22.05 \text{ g C.m}^{-2}.\text{yr}^{-1}$ (stages 1 – 4 core average = $11.9 \text{ g C.m}^{-2}.\text{yr}^{-1}$), with the exception of between c. 28.6 and c. 27.9 kcal yr BP where the CAR maximises at just below $28 \text{ g C.m}^{-2}.\text{yr}^{-1}$ (Fig. 4; Table 1). The lowest CAR values occur between c. 30.4 and c. 30.0 kcal yr BP (LSR stage 2), coinciding with an increase in sand dominated sediments. The top quarter of core SL6 (LSR stage 5) displays an elevated average CAR of $32.3 \text{ C.m}^{-2}.\text{yr}^{-1}$, notwithstanding three sharp declines at c. 7.1, between c. 5.5 and c. 5.3 and c. 1.4 and c. 1.3 kcal yr BP.

3.4. Elemental and stable isotopes measurements

3.4.1 C/N ratio

The atomic C/N ratio for core SL6 displays a range of between 16.9 (c. 1.4 kcal yr BP) and 71.1 (c. 24.2 kcal yr BP) with a core average of 40.5 (Fig. 5). The C/N ratio at the base starts out below 30.0, but steadily increases to 57.3 at c. 38.7 kcal yr BP before sharply declining to 37.6 at c. 37.5 kcal yr BP. The C/N signal then stabilises to between 36 and 52 until c. 24.5 kcal yr BP, where a significant positive shift occurs to a maximum of 71.1, followed by a return to a below core average of 29.9 at c. 17.1 kcal yr BP. Thereafter, the C/N signal gradually increases until c. 1.9 kcal yr BP, where it sharply declines from 57.7 to the core minimum value of 16.9 at c. 1.4 kcal yr BP.

3.4.2. Stable N isotopes

$\delta^{15}\text{N}$ values for core SL6 range between 1.9 (c. 41.5 kcal yr BP) and -2.9‰ (c. 0.44 kcal yr BP) and show an overall depletion in ^{15}N up core (Fig. 5). The $\delta^{15}\text{N}$ signal shifts to enriched ^{15}N values from the base of the core till the core maximum at c. 41.5 kcal yr BP. Thereafter, the $\delta^{15}\text{N}$ signal steadily trends to depleted ^{15}N values, with the exception of a period of

enrichment in ^{15}N between c. 30.0 and c. 24.5 kcal yr BP, and a sharp positive incursion to enriched ^{15}N from -1.5 to 0.9‰ at c. 1.4 kcal yr BP.

3.4.3. Stable C isotopes

$\delta^{13}\text{C}$ signal fluctuates between a maximum of -15.5‰ (c. 24.2 kcal yr BP) and minimum of -25.3 (c. 9.4 kcal yr BP), with no discernible overall trend (Fig. 5). With the exception of the base interval, the $\delta^{13}\text{C}$ signal becomes enriched in ^{13}C up core from -22.9 to -17.4‰ between c. 46.6 and c. 38.3 kcal yr BP before sharply decreasing to -22.3‰. Thereafter, the $\delta^{13}\text{C}$ signal increases steadily to the core maximum value at c. 24.2 kcal yr BP, and then declines again to the core minimum at c. 9.4 kcal yr BP. For the remaining part of the upper core, the $\delta^{13}\text{C}$ signal becomes enriched in ^{13}C through a succession of fluctuating cycles displaying pronounced enrichment shifts in ^{13}C of 4.5‰ (c. 8.8 to c. 8.4 kcal yr BP), 8.2‰ (c. 6.5 to c. 4.7 kcal yr BP) and 4.5‰ (c. 1.2 to c. 0.15 kcal yr BP).

4. Discussion

4.1. Sedimentation and carbon accumulation

Core SL6 returned a basal ^{14}C age of c. 47.0 kcal yr BP (805 cm), spanning the late Pleistocene and Holocene, positioning it as one of the oldest continuous coastal peatland records globally. According to Strack (2008), low lying coastal peat deposits in the (sub) tropics tend to accumulate faster than their temperate / boreal counterparts, however, most of the extensive coastal peatlands surveyed in SE Asia (Anderson and Muller, 1975; Staub and Esterle, 1994) only originated in the middle to late Holocene after the last sea level transgression. The Mfabeni peatland, therefore, is unique and owes its longevity to the protection against sea level fluctuations, and enhanced groundwater transmissibility

(Grundling et al., 2013) of the adjacent coastal dune corridor (c. 55 kcal yrs BP; Porat and Botha; 2008).

The varying LSR, MAR and CAR throughout core SL6 (Fig. 4) are indicative of changes in sedimentation regimes which are ultimately controlled by local climate. The total core LSR of between 0.10 and 0.29 mm.yr⁻¹ calculated for SL6 compares favourably with the average C accumulation rate reported by Grundling (2001, 2013) in the Mfabeni peatland. Peat accumulates when Net Primary Production (NPP) outstrips decomposition (Chimner and Ewel, 2005). In sub-arctic and boreal peatlands, the low to sub-zero temperatures retard microbial decomposition causing peat to accumulate (Francez and Vasander, 1995), notwithstanding the short growing seasons. However in the tropics, peatlands are subject to consistent hot and often humid conditions that are associated with rapid rates of decomposition. Although elevated temperatures facilitate microbial decomposition and rapid turnover of OM in tropical regions, it also increases NPP due to longer growing seasons and relatively higher precipitation, with plant roots mooted as the primary source of peat accumulation in tropical regions (Chen and Twilley, 1999; Chimner et al., 2002). The overriding dominant control on (sub) tropical peat formation is waterlogging, which enables prevailing anaerobic depositional conditions that ultimately retard the rate of decomposition and permits OM rich peat sediments to accumulate (Rieley et al., 1996). Waterlogging in peatlands results from an in-balance between moisture input and evapotranspiration, facilitated by local geology/ topography, impermeable mineral base layers, basin geomorphology and groundwater input (Cameron et al., 1989).

The SL6 CAR profile (Fig. 4) can be divided into two distinct segments, the top quarter of the core, spanning the Holocene, averages 32 g C.m⁻².yr⁻¹, with the rest of the core averaging

only 12 g C.m⁻².yr⁻¹. Estimates for Holocene CAR for the northern hemisphere peatlands have been published (Gorham, 1991; Turunen et al., 2002; Charman et al., 2013), with some authors using a statistical model developed by Clymo et al. (1992), to convert long-term apparent C accumulation (LARCA) rates to contemporary true rate of carbon accumulation (TRACA), with limited success (Gorham, 1991; Clymo et al., 1998). The LARCA average rate calculated for a northern Swedish fen was reported to be 25 g C.m⁻².yr⁻¹ (Oldfield et al., 1997) and 24 g C.m⁻².yr⁻¹ for a proximal fen 70 km away (Nilsson et al., 2008). Gorham (1991), on the other hand, published a global LARCA estimate of 29 g C.m⁻².yr⁻¹, which is more representative of the CAR (32 g C.m⁻².yr⁻¹) in the Mfabeni peatland during the Holocene.

4.2. Elemental and isotopic proxies

Atomic C/N ratios are often used as a proxy to delineate OM sources in lakes and marine environments (Meyers, 1994, 2003; Gälman et al, 2008). In peatlands, the C/N proxy has been employed in conjunction with C and N stable isotope signatures to elucidate OM preservation, redox depositional conditions and biogeochemical processes related to C and N cycling of sedimentary OM (Skrzypek et al., 2008; Jones et al., 2010; Andersson et al., 2012).

Palaeoresearchers commonly employ the $\delta^{15}\text{N}$ proxy as an indicator for bulk OM source input into lake / marine sediments and changes in palaeoproductivity (Meyers and Ishiwatari, 1993; Meyers, 1997, 2003; Routh et al., 2004; Choudhary et al., 2009). In peatlands, higher plants form the overwhelmingly dominant source of OM input; resulting in up to 95% of N originating from the degraded plant OM (Andersson et al., 2012). The premise for the N isotope bulk OM source proxy is based on the isotopically different sources of inorganic N available to plants. Plants that receive their N predominantly from

soil N fixers conventionally display $\delta^{15}\text{N}$ values within a small range (approximately -2 to +2‰) analogous to atmospheric $\delta^{15}\text{N}$ ($\pm 0\text{‰}$). In contrast, plants sourcing the majority of their N through microbially catalysed OM decomposition display characteristically more negative $\delta^{15}\text{N}$ ranges (ca. -2 to -8‰; Fogel and Cifuentes, 1993; Skrzypek et al., 2008). Bodelier and Laanbroek (2004) suggested that during increased rates of methanogenesis, and by that token increased temperatures and waterlogging in tropical peatlands, the high N demanding methanogenic bacteria tend to consume large quantities of ^{14}N to produce a more isotopically light N source for plants, which ultimately causes the peat N isotopic signal to become depleted in ^{15}N .

Stable carbon isotope signatures, in conjunction with atomic C/N ratios, have also proven to be robust indicators of OM sources, diagenetic alteration and primary production (Meyers and Ishiwatari, 1993; Meyers, 2003). C4 and C3 plants fractionate against ^{13}C differently, which results in significantly more positive bulk $\delta^{13}\text{C}$ values for C4 plant biomass in comparison with that of C3 plants. Therefore, the $\delta^{13}\text{C}$ signal in peat sediments can serve as an archive for changes in the relative contribution of C4 and C3 plants, which can be related to changes in environmental and climatic conditions (Skrzypek et al., 2008, 2010). The bulk $\delta^{13}\text{C}$ isotopic values in core SL6 demonstrates a strong correlation with concordant isotopic values of the leaf wax *n*-alkanes ($r=0.88$, $P=0.01$, $n=27$; in Baker et al., in preparation), suggesting the major OM source in the Mfabeni sediments are of terrestrial plant origin. A detailed study of how individual plant species identified in Mfabeni could potentially react to changes in moisture and temperature is beyond the scope of this research. We assume, therefore, that these environmental factors have a negligible effect on the $\delta^{13}\text{C}$ value of bulk OM in comparison with shifts in C3/C4 vegetation.

Finch and Hill (2008) undertook an extensive palynological study in the Mfabeni peatland and documented the changes in the pollen record. They observed a local vegetation taxa dominated by varying proportions of Poaceae and Cyperaceae since the late Pleistocene (c. 44 kcal yr BP to present). Even though the Poaceae and Cyperaceae plant families are comprised of both C3 and C4 species, Stock et al., (2004) and Vogel and Fuls (1978) observed a definite geographical distribution between the two photosynthetic pathways in southern Africa. The C4 grasses and sedges occur more abundantly in the north eastern austral summer rainfall areas, while the C3 species are more dominant in the winter rainfall areas and high altitude regions of the south western Cape and eastern Escarpment. Kotze and O'Connor (2000) did a study of contemporary vegetation within and around wetlands along an altitudinal gradient in the Kwazulu-Natal region. They documented the percentage variations of the two dominant species, namely C3/C4 grasses and sedges, in wetlands with varying hydrological conditions. The Mongolwane wetland, at 550 m elevation, recorded relative proportions of 49.2, 39.6, 9.5 and 1.7% for C4 sedges, C3 grasses, C3 sedges and C4 grasses, respectively, in permanently inundated sections, and 57.6, 21.0, 13.2 and 8.2% for C4 sedges, C4 grasses, C3 grasses and C3 sedges, respectively, in seasonally inundated areas. This represents a shift from ~50:50 split between C3 and C4 plant species in permanent wetlands to ~20:80 ratio in favour of C4 plant species in seasonal wetlands. Kotze and O'Connor (2000) also recorded an increase to near absolute C4 grasses dominance as a result of moisture reduction in low-lying wetlands.

Throughout core SL6, the C/N ratio does not fall below 20 (Fig. 5), besides at 54 cm (17; c. 1.4 kcal yr BP), reiterating the predominant OM input into the Mfabeni peatland being vascular land plants. The elemental C and N wt% values exhibit a strong correlation

throughout the core (total core: $r = 0.77$; $P=0.01$; $n = 196$; refer to Table 3 for individual LSR stages and Fig. 5), which represents a MAR, as opposed to diagenetic relationship for C and N concentrations with depth (Andersson et al., 2012). Kuhry and Vitt (1996) used the elemental C/N relationship to explore diagenetic effects in acrotelm and catotelm peat layers. They proposed that in the aerobic acrotelm layer, N is preferentially lost during OM breakdown, resulting in increased bulk C/N ratio with depth, whereas in the anaerobic catotelm layer, a decrease in C/N should be observed as a result of methanogenesis and N becoming immobile under anoxic conditions. The SL6 C/N ratio shows no overall trends down core, supporting the inference based on the wt% C and N relationship, that diagenetic alteration did not play a major role during overall peat formation in Mfabeni peatland.

The $\delta^{15}\text{N}$ signal in core SL6 displays a general increasing trend with depth, with values ranging between -2.9 and 1.9‰ suggesting the predominant source of bioavailable N during photosynthesis is N-fixing bacteria and not microbial reworking of OM. In contrast, the $\delta^{13}\text{C}$ profile fluctuates between -25.3 and -15.5‰, inferring an interchangeable source of C3 and C4 plant end members (Fig. 5 and 6). It has been shown that the lighter ^{12}C (Nichols et al., 2009; Jones et al., 2010) and ^{14}N isotopes (Amundson et al., 2005) are preferential removed during extensive microbial remineralisation, resulting in an enrichment in both the ^{13}C and ^{15}N isotopes in the remaining soil (and more ^{14}N available for plant uptake; Bodelier and Laanbroek, 2004). Therefore a significant statistical correlation can be expected between the $\delta^{13}\text{C}$, $\delta^{15}\text{N}$ and C/N signals in extensively decomposed soils (Engel et al., 2010), or conversely, an insignificant correlation in highly preserved soils where minor isotopic C and N fractionation occurred during diagenesis (Jedrysek and Skrzypek, 2005; Jones et al., 2010; Skrzypek et al., 2010). By comparing the isotopic and elemental signals for each of the five

LSR stages in core SL6, the statistical relationships can be used to infer the relative degree of SOM preservation / rates of decomposition for each LSR stages.

There is a general absence of any significant correlation between the $\delta^{15}\text{N}$ and $\delta^{13}\text{C}$ signals (Table 3) in all five of the LSR stages endorsing the Mfabeni sediments as highly preserved. When comparing the $\delta^{13}\text{C}$ and C/N signals, a significant correlation is exhibited in LSR stages 1 – 4, while the $\delta^{15}\text{N}$ and C/N comparison only shows a significant correlation during LSR stages 2 and 3. This unbalanced correlation between the isotopic and elemental combinations up core suggests the degree of OM preservation was influenced predominantly by environmental factors during deposition and diagenesis, as opposed to peat maturity (age), permitting us to gauge relative rates of decomposition during each LSR stage, and surmise the environmental factors affecting not only sedimentation but decomposition rates (Engle et al., 2010).

4.3. Palaeo reconstruction

4.3.1 LSR Stage 1 (c. 47.0 – c. 32.4 kcal yr BP)

LSR Stage 1 (average LSR = 0.13 mm.yr⁻¹) occurred in conjunction with relatively low MAR; high average TOC, below average C accumulation (Fig. 4), and relatively low rates of decomposition (Table 3) suggesting a cool and wet period, with extended phases of waterlogging, but slow peat accumulation probably as a result of lower NPP. The $\delta^{15}\text{N}$ values are comparatively more positive, ranging between -0.4 and +1.9‰, indicating a predominant atmospheric N fixing source while the $\delta^{13}\text{C}$ signal fluctuates between -22.9 and -15.8‰, representing a C4 plant dominated OM input (Fig. 5). Finch and Hill (2008) found evidence of high frequencies of forest tree pollen, with a locally dominant signal of grasses, and to a lesser extent sedges, between c. 44 and c. 33 kcal yr BP, which they interpreted to represent a cool and wet period. A cooler climate could have resulted in lower rates of

methanogenesis, and a reduction in bioavailable ^{14}N (Skrzype et al., 2008; Jones et al., 2010). For the $\delta^{13}\text{C}$ signal, however, the C isotope signature is only significantly influenced by the relative proportions of C3 and C4 plant OM input. Stock et al. (2004) concluded that C4 sedges endemic throughout South Africa appeared to have evolved more as a response to low atmospheric CO_2 , as opposed to limited nutrient and hydrologic adaptations, which are more typical of the regional C4 grasses. The shifts in $\delta^{13}\text{C}$ during this stage trends similarly, but opposite to the CO_2 records reported in the Byrd ice cores (Stocker, 2000) reinforcing the proposed evolutionary relationship between C4 plants and lower atmospheric $p\text{CO}_2$.

Notably, the core TOC maximum occurs in this stage at c. 44.6 kcal yr BP (1600 g C.m^{-2}), together with another abrupt increase in average TOC values between c. 37.9 and c. 35.9 kcal yr BP, concordant with the A2 and A1 warming event (and Heinrich 5 (H5) and Heinrich 4 (H4) cooling event in the Northern hemisphere) identified by $\delta^{18}\text{O}$ and $\delta^2\text{H}$ signals in the Byrd and Vostok ice cores from Antarctica (and GRIP ice core in Greenland; Blunier et al., 1998; Stocker, 2000). Bard et al. (1997) reported an overall 1.6°C gradual decline in alkenone proxy sea surface temperatures (SST) from a marine sediment core (MD79257) extracted at 20°S in the Mozambique Channel during this same period. However, included in this general decline, two prominent spikes were observed at c. 44.5 and c. 35.5 kcal yr BP, which correspond to the A2 and A1 warming events, respectively (Blunier et al., 1998; Stocker, 2000). Although the stable isotope signature does not definitively reflect the A2 warming event (or H5, 44.5 kcal yr BP), a shift of -4.1‰ to more negative $\delta^{13}\text{C}$ values and a decrease in C4 plant contribution (Fig. 6) is observed starting at c. 38 kcal yr BP (A2; H4). This $\delta^{13}\text{C}$ negative shift is followed closely by a negative shift in $\delta^{15}\text{N}$ values, which suggests

an increase in microbial sources of ^{14}N isotopes as a response to the A1 warming event. It is postulated that the prominent change in physical and geochemical parameters during the A1 (and A2) warming event was a result of permanent waterlogging and increased contributions from C3 grasses relative to C4 sedges during a period of permanent inundation (Kotze and O'Connor, 2000), accompanied by increased rates of methanogenesis towards the latter part of LSR Stage 1.

4.3.2. LSR stage 2 (c. 32.1 – c. 27.9 kcal yr BP)

The next LSR stage (Fig. 4) displays an increased average LSR of 0.22 mm.yr^{-1} , overall high MAR, low TOC and CAR, which was accompanied by a gradual overall positive trends in $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ values (Fig. 5) and an increase in decomposition rates compared to LSR stage 1 (Table 3). These parameters infer a period of minimal water logging at the site, and a shift to sand dominated sedimentation, most likely due to a shift to drier and windier conditions. Conversely, a sharp increase in TOC and C accumulation to above core average values, concordant with more negative $\delta^{15}\text{N}$ values occurred in the latter part of LSR stage 2 (c. 28.6 to c. 27.9 kcal yr BP), suggesting a short period of wet and warm conditions, in contrast to the general overall trend towards full glacial conditions. The $\delta^{13}\text{C}$ signal, however, continues along an overall positive trend, inferring a continued increase in C4 plant input which can be correlated to overall decreasing $p\text{CO}_2$ (compared to the Byrd Ice core; Stocker, 2000). This interpretation is supported by a general decline over the same timescale, punctuated by an abrupt increase of 1°C at c. 28 kcal yr BP in the MD79257 core derived alkenone SST (Bard et al., 1997), and a decline in forest pollen and switch towards sedge dominated swamp vegetation (Finch and Hill, 2008). Likewise, the Vostok ice core registered a noticeable return to more positive $\delta^2\text{H}$ values at c. 28 kcal yr BP (Stocker, 2000). Talma and Vogel (1992) calculated late Quaternary ambient temperatures from speleotherm $\delta^{18}\text{O}$ values in

the Cango Caves (22°E, 33°S), which is situated in the perennial rainfall region of the southern Cape, South Africa. They observed a slow overall decline in temperature leading up to the LGM (c. 19 to c. 17 kyr BP), punctuated by a ~1.5 °C temperature reversal at c. 28.5 kyr BP. It could also be argued that with the large abrupt increase in CAR observed at the conclusion of LSR stage 2, the Mfabeni peatland switched from being a temporary to a seasonally inundated fen resulting in an increase in C4 sedge dominance as reported by Kotze and O'Connor (2000). Partridge (2002) similarly recorded a peak in precipitation levels using a rainfall time series sediment proxy extracted from the Tswaing impact crater lake in north eastern South Africa around c. 28 kcal yr BP, before the onset of full glacial conditions.

4.3.3. LSR stage 3 (c. 27.6 – c. 20.3 kcal yr BP)

During LSR stage 3 (Fig. 4), the LSR drops to an average of 0.14 mm.yr⁻¹ accompanied by below average MAR, average to high TOC and near average CAR until c. 22.7 kcal yr BP, after which a sharp decline in TOC and to a lesser extent CAR is observed. A significant correlation between both $\delta^{15}\text{N}$ and $\delta^{13}\text{C}$ with C:N ratio suggests an overall elevated rate of decomposition during the LGM (Table 3). The $\delta^{15}\text{N}$ and $\delta^{13}\text{C}$ values remain relatively stable until c. 24.2 kcal yr BP. The N and C stable isotope records subsequently diverge (H2, Fig. 5), the $\delta^{15}\text{N}$ values become more negative, whereas the $\delta^{13}\text{C}$ and C/N ratio values abruptly increase, coincidental with elevated TOC and average CAR. These changing parameters can be interpreted as an initial shift to cooler temperatures and a peak in waterlogging (precipitation) and increased dominance of widespread C4 wetland sedges (Kotze and O'Conner, 2000), before a shift to cooler and drier conditions after c. 23 kcal yr BP. The sharp increase in C/N ratio at c. 24.2 kcal yr BP could possibly be due to an increase in peat accumulation, as a result of higher plant preservation during extensive waterlogging

suggested by the elevated TOC and more subtle increases in the LSR and CAR. Thereafter, the $\delta^{13}\text{C}$ signal steadily decreases, coinciding with a sharp drop in TOC and CAR, which we conclude as an overall increase in relative abundance of C3 grasses, which have an advantage over C4 grasses / sedges at lower growing season temperatures (Sage et al., 1999; Kotze and O' Conner, 2000; Finch and Hill, 2008; Fig. 6). The SST of core MD79257 displayed stable temperatures around 26 °C until c. 24 kcal yr BP (H2) after which the signal steadily declines towards the lowest core SST of 24.2 °C at c. 20 kcal yr BP (Bard et al., 1997). The Byrd Antarctic ice core $\delta^{18}\text{O}$ signal oscillates between -41.5 and -39.0‰ until c. 24 kcal yr BP, after which the signal declines to lowest levels for the next c. 3.5 kyr (Blunier et al., 1998). Finch and Hill (2008) observed an abrupt change from swamp sedge to grassland vegetation after c. 24 kcal BP which they interpreted as an abrupt shift to drier, cooler local conditions at the onset of the LGM (c. 24 kcal yr BP) indicated by a sudden increases in Poaceae pollen frequencies and a steady decline in Cyperaceae pollen. Likewise, the Cold Air cave stalagmite located in Makapansgat Valley (24°S, 29°E) in the summer rainfall area of South Africa recorded drier conditions in conjunction with lower temperatures between 23 – 21 kyr (Holmgren et al., 2003).

4.3.4. LSR Stage 4 (c. 19.8 – c. 10.4 kcal yr BP)

The lowest core average LSR (0.10 mm.yr⁻¹) occurs in LSR stage 4 (Fig. 4), corresponding to a period of low MAR, low but steadily increasing TOC and CAR dominated by an abrupt shift to negative $\delta^{13}\text{C}$ values. This trend is accompanied by fluctuating $\delta^{15}\text{N}$ values near the core average (Fig. 5) and relative reduction in decomposition rates (Table 3). Between c. 14.3 and c. 10.8 kcal yr BP, a sharp overall increase in TOC values occurs, whereas other physical and geochemical parameters remain relatively constant. These observations can be

interpreted as a continuation of dry and cool late glacial conditions up until c. 15 kcal yr BP, after which an abrupt increase in precipitation and waterlogging occurs. The sharply negative $\delta^{13}\text{C}$ values could have been as a result of increases in C3 grass input which is supported by the sharp decline in the proportion of C from C4 plants mass balance at the onset of LSR stage 4 (Fig. 6), and then a switch to C3 swamp forest vegetation after c. 15 kcal yr BP due to the abrupt increase in precipitation. Finch and Hill (2008) documented a continued dominance of grasslands over wetland sedges during this stage, which leads us to speculate that the predominant photosynthetic pathway employed by Poaceae grasses was C3 as a consequence of the better adaptation to lower growing season temperatures and general dry conditions compared to C4 plants (Sage et al., 1999) up to c. 15 kcal yr BP.

Norström et al. (2009) used palynology, and C and N stable isotope proxies to infer a cool and dry climate between 16 and 14.3 kyr, then a shift to more humid conditions culminating at 13.2 kyr before returning to drier conditions in the Braamhoek peatland (28°S, 29°E).

Talma and Vogel (1992) recorded a reversal of the generally declining ambient air temperatures in the Congo cave speleothem after c. 15.5 kyr BP, with the temperature steadily increasing up until stalagmite growth ceased at c. 13.8 kyr BP. The Cold Air cave stalagmite $\delta^{18}\text{O}$ signature inferred drier and cooler conditions around c. 19.5 - 17.5 and c. 15 - 13.5 kyr, with warming interludes (Holmgren et al., 2003), while biomarker and stable isotope proxies from Lake Tswaing (25°S, 28°E) suggested the period between c. 14 and c. 10 kcal yr BP experienced increased temperatures and moisture (Kristen et al., 2010).

Chase et al. (2011) recorded relative dry conditions between c. 19.5 and c. 17.5 kcal yr BP, followed by an increase in moisture up to the early Holocene in hyrax midden deposits located in the winter rainfall zone of the SW Cape, with the exception of a conspicuous

period of drier conditions concomitant with the Younger Dryas (YD). On the contrary, Schefus et al. (2011) observed large inputs of terrestrial sedimentary and plant input into the Zambezi catchment area (18° 33.9' S, 37° 22.8' E) during the H1 and YD northern hemisphere cold events, indicating significant increases in adjacent continental austral summer rainfall precipitation, while the proximal Mozambique channel MD79257 core alkenone SST recorded a warming trend after c. 15 kyr BP (H1; Bard et al., 1997). Stocker (2000), compared the Greenland GRIP ice core with the Vostok and Byrd Antarctic ice cores, and documented asynchronous responses to climate change between the two latitudinal hemispheres during the last deglaciation. The similarity between these trends in core SL6 and other regional climate records implies that the Mfabeni peat deposit faithfully recorded both regional and global late glacial climatic events and displays a comparable opposite phasing in trend and magnitude when compared to northern hemisphere Heinrich and YD climate events.

4.3.5. LSR stage 5 (c. 10.2 kcal yr BP – present)

The transition to interglacial conditions at the Pleistocene / Holocene boundary displays a prominent shift to highly fluctuating climatic conditions. Gasse (2000) concluded that dramatic Holocene hydrological changes documented in low latitudes on the African continent appear to rival fluctuations observed during the preceding glacial period. The LSR stage 5 is represented by elevated and highly variable sedimentation, displaying the highest LSR stage average (0.29 mm.yr⁻¹), accompanied by high MAR, overall high TOC and CAR (Fig. 4), with an ever decreasing $\delta^{15}\text{N}$ and fluctuating $\delta^{13}\text{C}$ signature (Fig. 5). The lack of any significant correlations between either of the stable isotopes and elemental ratio, suggests low rates of decomposition during the Holocene (Table 3). These parameters can be

interpreted as significant increases in humidity and temperatures typical of an interglacial period, punctuated by a collection of millennial-scale cooling events. Three abrupt excursions to low TOC and CAR values occur at c. 7.1, c. 5.3 and c. 1.4 kcal yr BP, which coincide with decreases in Mozambique Channel alkenone and Zambezi delta TEX₈₆-derived SST (Bard et al., 1997; Schefus et al., 2011), and signs of general aridity in the Kalahari Desert (Stokes et al., 1997). Norström et al. (2009) recorded similar large oscillations in the Braamhoek wetland proxies that are in general agreement with the Mfabeni peatland parameters, with the exception of between 7.5 and 2.5 ka which indicate a relatively dry and warm period, whereas core SL6 recorded overall elevated CAR values, excluding the two millennial-scale drying events at c. 7 and c. 5.3 kcal yr BP. Norström et al. (2009) however did stipulate that this time period can only be used to infer general palaeoenvironmental conditions due to the comparative low rate of peat accumulation and low chronological resolution in the Braamhoek record.

After c. 10.5 kcal yr BP, the CAR abruptly increases while the $\delta^{13}\text{C}$ record displays the minimum core value (-25.3‰; c. 9.4 kcal yr BP) and lowest C4 plant C contribution (Fig. 6) suggesting a significant increase in local precipitation and C3 plant abundance in and around the Mfabeni peatland. The palynology record showed rapidly increasing Podocarpus forest (C3) pollen after c. 11 kcal yr BP, which is interpreted to represent a period of cool and wet conditions (Finch and Hill, 2008). In contrast, Valsecchi et al. (2013) documented a shift from Protea-type pollen to arboreal pollen and attributed the change to warmer and wetter conditions in the winter rainfall fynbos area of the south Western Cape. The Cold Air cave stalagmite returned the lowest $\delta^{13}\text{C}$ values at c. 9 kyr, indicating a significant increase in C3 plant influence on the precipitating aragonite (Holmgren et al., 2003). At c. 9 kcal yr BP, the

562 $\delta^{13}\text{C}$ values reverse their decreasing trend and rapidly become more positive, peaking at c.
563 8.4 kcal yr BP, after which the signal returns to more negative values up until c. 7.2 kcal yr
564 BP (first of the millennium scale cooling events), emulated by similar trending TOC and CAR.
565 Finch and Hill (2008) documented a rapid increase in swamp forest vegetation, with
566 sustained levels of Podocarpus pollen, during the Holocene Altithermal (c. 8 to c. 6 kcal yr
567 BP), which they interpreted to signify a warming trend with high precipitation levels,
568 supported by a corresponding 0.6 °C increase in the Mozambique Channel core SST data at
569 c. 9 kcal yr BP (Bard et al., 1997). The enriched ^{13}C signal at c. 8.4 kcal yr BP, could possibly
570 be a result of a spike in C4 wetland sedge populations (Fig 6) contributing to the enrichment
571 of ^{13}C during a brief period of intensive waterlogging in the peatland, reinforced by a peak in
572 levels of concordant SL6 CAR (Fig. 4). The $\delta^{13}\text{C}$ signal returns steadily to more negative
573 values up until c. 7 kcal yr BP, suggesting a return to swamp forest (C3) dominance, and
574 humid and warm climatic conditions, supported by increases in TOC up until c. 7 kcal yr BP
575 cooling event. Pudsey and Evans (2001) studied changes in glacial till deposits from floating
576 ice sheets at the edge of the Antarctic Peninsula. They documented large deglaciation in the
577 northern James Ross Island, and proposed the full disappearance of the George VI Ice shelf
578 c. 6.5 kyr BP, signifying a warming trend in Antarctic. The Mfabeni $\delta^{13}\text{C}$ signal increases by
579 more than c. 5‰, coinciding with a decrease in $\delta^{15}\text{N}$ values, between c. 7 and c. 6 kcal yr BP
580 inferring a shift to C4 sedge dominated plant OM input (Kotze and O’Conner, 2000 and Fig.
581 6) and increased methanogenesis. These signatures indicate a short wet period before a
582 shift to cool dry conditions at the second millennium-scale c. 5.3 kcal yr BP cooling event,
583 delineated by the abrupt decrease in TOC and CAR. After c. 6 kcal yr BP, Finch and Hill
584 (2008) observed a steady decline in the abundance of Podocarpus forest pollen and an

585 increase in Poaceae and Cyperaceae frequency, indicating a move to grassland / savannah
586 dominance, and inferred a cooling and drying trend towards the middle Holocene.

587 Relatively more positive $\delta^{13}\text{C}$ values occur between c. 5 to c. 3 kcal yr BP, coinciding with
588 elevated TOC and CAR and continuance of wetland C4 sedge dominance (Finch and Hill,
589 2008; Fig. 6), signalling a return to warm and moist conditions. Anthropogenic farming
590 practices become apparent from the palynology record after c. 5 kcal yr BP, specifically
591 attributed by Finch and Hill (2008) to the decline in Podocarpus (C3) forests which could
592 have artificially elevated the $\delta^{13}\text{C}$ climate signal. As a consequence, bulk parameters need
593 to be interpreted with caution from c. 5 kcal yr BP as human influence on the
594 palaeoenvironment signal increased. An overall C4 peatland signature of the $\delta^{13}\text{C}$ trend and
595 a shift to more negative $\delta^{15}\text{N}$ up until c. 1.4 kcal yr BP suggests the continued dominance of
596 C4 sedge input (and perhaps anthropogenic forest thinning) due to extensive waterlogging.
597 Finch and Hill (2008) concluded an increase in savannah / grassland pollen to infer a drying
598 trend from c. 3 kcal yr BP to present. However, we interpret core SL6 data to suggest a
599 warming and moist trend over this period (with the exception of the 1.4 kcal yr BP cooling
600 event) as indicated by ever decreasing $\delta^{15}\text{N}$ and overall positive $\delta^{13}\text{C}$ signals during this
601 same period, signalling waterlogged conditions. The last of the abrupt cooling/drying events
602 (c. 1.4 kcal yr BP) occurs in conjunction with a prominent decrease in both TOC and CAR, a
603 positive 2.4‰ deviation in $\delta^{15}\text{N}$ and an initial increase, followed by a decrease in $\delta^{13}\text{C}$
604 values. The changes in the proxy parameters indicate that this abrupt cooling event
605 reduced the rate of methanogenesis, reducing the amount of bioavailable ^{14}N and an
606 increase in C4 grass input (Fig. 6) in response to sudden and brief period of dry conditions.
607 The swift return to the more elevated TOC and CAR and the lowest $\delta^{15}\text{N}$ core values,

accompanied by oscillating $\delta^{13}\text{C}$ signal infers a return to cyclical climate conditions and once again highlights the extreme climatic fluctuations characteristic of the Holocene, both regionally and globally (Gasse, 2000; Mayewski et al., 2004).

The Mfabeni record's overall opposite environmental response to better known climatic events in the Northern hemisphere suggests an anti-phase coupling to the southern hemisphere (Figs. 4 and 5). Core SL6 shows a trend in both physical and geochemical parameters towards increased C accumulation during cold Heinrich events (particularly H5 to H2 and YD), suggesting elevated precipitation, and arguably temperatures, which is in direct contrast to findings from the northern hemisphere records. Varying degrees of anti-phased interhemisphere coupling has also been observed by other authors (Bard et al., 1997; Blunier et al., 1998; Schmittner et al., 2003; Chase et al., 2011; Schefus et al., 2011) with several different mechanisms for climate forcing postulated; signalling a crucial requirement for additional regional palaeoenvironment studies to corroborate or challenge these hypotheses.

5. Conclusions

We employed several known geochemical indicators of peat forming processes and related them to changes in past primary production, OM preservation, OM sources and diagenetic alteration after deposition. By relating the bulk parameters to physical and biogeochemical processes, we were able to reconstruct the palaeoenvironment that controlled peat formation at the Mfabeni peatland and hypothesise the probable climate on the north east coast of South Africa since the late Pleistocene. We established that the Mfabeni peat sediments have undergone minimal diagenetic alteration, confirming the archive's high degree of preservation and accurate recording of palaeoenvironmental conditions. We

631 surmise the following sequence of palaeoenvironmental conditions and their climatic
632 controls from c. 47 kcal yr BP to present:-

633 LSR stage 1 (c. 47 – c. 32.2 kcal yr BP): we inferred a predominantly cool and wet climate
634 with extensive waterlogged but low NPP, and C4 plant dominant OM source assemblage,
635 punctuated by two short warming events, A2 (c. 44.5 kcal yr BP) and A1 (c. 37 kcal yr BP).
636 The latter events were delineated by elevated TOC attributed to increased NPP and
637 extensive waterlogging. LSR stage 2 (c. 32.2 to c. 27.6 kcal yr BP): a distinct period of sand
638 dominated deposition as a result of negligible waterlogging, suggesting dry and windy
639 conditions, followed by a brief period of warm and wet conditions (c. 28 kcal yr BP)
640 indicated by an abrupt increases in CAR and C4 sedge swamp vegetation abundance. LSR
641 stage 3 (c. 27.6 – c. 20.3 kcal yr BP): began with cool and wet conditions with a peak in
642 waterlogging at c. 24 kcal yr BP represented by high TOC and prevailing C4 sedge
643 assemblage. After c. 23 kcal yr BP (LGM), an abrupt change to dry and cool conditions
644 indicated by a sharp drop in TOC and steady increase in C3 grasses relative to C4 sedges as
645 waterlogging receded. LSR stage 4 (c. 20.3 – c. 10.4 kcal yr BP): continuance of cool and dry
646 conditions inferred by limited waterlogging and strong C3 grassland signature until c. 15 kcal
647 yr BP, after which, increases in waterlogging and precipitation during waning of the ACR and
648 lead up to the Pleistocene/Holocene boundary. This trend further accentuates the apparent
649 opposite climate phasing between the Northern and southern hemispheres. LSR stage 5 (c.
650 10.4 kcal yr BP – present): the Holocene epoch is characterized by fluctuations between
651 pervasive wet and warm to cool / dry interglacial conditions, with intermittent abrupt
652 millennial-scale cooling / dry events (c. 7.1, c. 5.3, c. 1.4 kcal yr BP). The MAR, TOC and CAR
653 values are distinctly elevated, with large abrupt magnitudes of variability accompanied by

frequent changes between C3 and C4 plant assemblages in comparison to the late glacial period.

In this study, we delineated the Mfabeni peat sequence to produce a high resolution record of past local hydrology and bulk plant assemblages, and by using these signals, we inferred regional climate variability since the Late Pleistocene. There has long been a quest amongst palaeoclimatologists to understand the interhemispheric climate coupling relationship. The Mfabeni archive suggests an anti-phase link between southern Africa and the Northern hemisphere, most notably during H5 to H2 and YD events. Many more analogous palaeoclimate studies need to be undertaken to firm up our understanding of the triggers and change mechanisms responsible for LGM climate variability in southern Africa.

6. Acknowledgments

Alistair Clulow helped facilitate field access and site identification. A Russian peat corer was loaned to the project by Piet-Louis Grundling. iSimangaliso Authority and Ezemvelo KZN Wildlife granted access and sampling permission. The following people assisted with laboratory set up, equipment, protocols and analysis: Esmé Spicer, Cynthia Sanchez-Garrido, Renata Smit, Ian Newton, Lena Lundman, Tomaz Gozlar, Eric Ward and Megan Hill. Nikolai Pedentchouk and Michael Meadows gave valuable manuscript input. The project was supported through a bilateral funding agreement by the Swedish Research Link-South Africa program. Student support was supplied by the National Research Foundation and Inkaba yeAfrica. This is an Inkaba ye Africa publication no. 78 and AEON publication no. 118.

674 **References**

- 675 Amundson, R., Austin, A.T., Schuur, E.A.G., Yoo, K., Matzek, V., Kendall, C., Uebersax, A.,
676 Brenner, D. & Baisden, W.T., 2003. Global patterns of the isotopic composition of soil
677 and plant nitrogen. *Global Biogeochemical Cycles* 17(1), 31-41.
- 678 Anderson, J.A.R., Muller, J., 1975. Palynological study of a Holocene peat and a Miocene coal
679 deposit from NW Borneo. *Review of Palaeobotany and Palynology* 19 (4), 291-317.
- 680 Andersson, R.A., Meyers, P., Hornibrook, E., Kuhry, P., Mörth, C., 2012. Elemental and
681 isotopic carbon and nitrogen records of organic matter accumulation in a Holocene
682 permafrost peat sequence in the east European Russian arctic. *Journal of Quaternary*
683 *Science* 27 (6), 545-552.
- 684 Bard, E., Rostek, F., Sonzogni, C., 1997. Interhemispheric synchrony of the last deglaciation
685 inferred from alkenone palaeothermometry. *Nature* 385 (6618), 707-710.
- 686 Bate, G.C., Taylor, R.H., 2008. Sediment salt-load in the St Lucia Estuary during the severe
687 drought of 2002-2006. *Environmental Geology* 55 (5), 1089-1098.
- 688 Blaauw, M., Christeny, J.A., 2011. Flexible paleoclimate age-depth models using an
689 autoregressive gamma process. *Bayesian Analysis* 6 (3), 457-474.
- 690 Blunier, T., Chappellaz, J., Schwander, J., Dällenbach, A., Stauffer, B., Stocker, T.F., Raynaudt,
691 D., Jouzel, J., Clausen, H.B., Hammer, C.U., Johnsen, S.J., 1998. Asynchrony of Antarctic
692 and Greenland climate change during the last glacial period. *Nature* 394 (6695), 739-
693 743.

694 Bodelier, P.L.E., Laanbroek, H.J., 2004. Nitrogen as a regulatory factor of methane oxidation
695 in soils and sediments. *FEMS Microbial Ecology* 47 (3), 265-277.

696 Botha, G., Porat, N., 2007. Soil chronosequence development in dunes on the southeast
697 African coastal plain, Maputaland, South Africa. *Quaternary International* 162-163,
698 111-132.

699 Boutton, T.W., Archer, S.R., Midwood, A.J., Zitzer, S.F., Bol, R., 1998. $\delta^{13}\text{C}$ values of soil
700 organic carbon and their use in documenting vegetation change in a subtropical
701 savanna ecosystem. *Geoderma* 82(1–3), 5-41.

702 Brock, F., Higham, T., Ditchfield, P., Ramsey, C.B., 2010. Current pretreatment methods for
703 AMS radiocarbon dating at the Oxford radiocarbon accelerator unit (ORAU). *Radiocarbon*
704 52 (1), 103-112.

705 Cameron, C.C., Esterle, J.S., Palmer, C.A., 1989. The geology, botany and chemistry of
706 selected peat-forming environments from temperate and tropical latitudes.
707 *International Journal of Coal Geology* 12 (1–4), 105-156.

708 Charman, D.J., Beilman, D.W., Blaauw, M., Booth, R.K., Brewer, S., Chambers, F.M., Christen,
709 J.A., Gallego-Sala, A., Harrison, S.P., Hughes, P.D.M., Jackson, S.T., Korhola, A.,
710 Mauquoy, D., Mitchell, F.J.G., Prentice, I.C., Van Der Linden, M., De Vleeschouwer, F.,
711 Yu, Z.C., Alm, J., Bauer, I.E., Corish, Y.M.C., Garneau, M., Hohl, V., Huang, Y., Karofeld,
712 E., Le Roux, G., Loisel, J., Moschen, R., Nichols, J.E., Nieminen, T.M., MacDonald, G.M.,
713 Phadtare, N.R., Rausch, N., Sillasoo, U., Swindles, G.T., Tuittila, E.-., Ukonmaanaho, L.,
714 Välranta, M., Van Bellen, S., Van Geel, B., Vitt, D.H., Zhao, Y., 2013. Climate-related

715 changes in peatland carbon accumulation during the last millennium. *Biogeosciences*
716 10 (2), 929-944.

717 Chase, B.M., Quick, L.J., Meadows, M.E., Scott, L., Thomas, D.S. & Reimer, P.J., 2011. Late
718 glacial interhemispheric climate dynamics revealed in South African hyrax middens.
719 *Geology* 39 (1), 19-22.

720 Chase, B.M., Meadows, M.E., 2007. Late Quaternary dynamics of southern Africa's winter
721 rainfall zone. *Earth-Science Reviews* 84 (3-4), 103-138.

722 Chen, R., Twilley, R.R., 1999. A simulation model of organic matter and nutrient
723 accumulation in mangrove wetland soils. *Biogeochemistry* 44 (1), 93-118.

724 Chimner, R.A., Ewel, K.C., 2005. A tropical freshwater wetland: II. Production,
725 decomposition, and peat formation. *Wetlands Ecology and Management* 13 (6), 671-
726 684.

727 Chimner, R.A., Cooper, D.J., Parton, W.J., 2002. Modeling carbon accumulation in Rocky
728 mountain fens. *Wetlands* 22 (1), 100-110.

729 Choudhary, P., Routh, J., Chakrapani, G.J., 2009. An environmental record of changes in
730 sedimentary organic matter from Lake Sattal in Kumaun Himalayas, India. *Science of*
731 *the Total Environment* 407 (8), 2783-2795.

732 Clulow, A.D., Everson, C.S., Mengistu, M.G., Jarman, C., Jewitt, G.P.W., Price, J.S., Grundling,
733 P., 2012. Measurement and modelling of evaporation from a coastal wetland in
734 Maputaland, South Africa. *Hydrology and Earth System Sciences* 16 (9), 3233-3247.

- 735 Clymo, R.S., 1992. Models of peat growth. *Suo* 43 (4-5), 127-136.
- 736 Clymo, R.S., Turunen, J., Tolonen, K., 1998. Carbon accumulation in peatland. *Oikos* 81 (2),
737 368-388.
- 738 Czernik, J., Goslar, T., 2001. Preparation of graphite targets in the Gliwice radiocarbon
739 laboratory for AMS¹⁴C dating. *Radiocarbon* 43 (2), 283-291.
- 740 Engel, Z., Skrzypek, G., Paul, D., Drzewicki, W., Nývlt, D., 2010. Sediment lithology and stable
741 isotope composition of organic matter in a core from a cirque in the Krkonoše
742 Mountains, Czech Republic. *Journal of Paleolimnology* 43 (4), 609-624.
- 743 Finch, J.M., 2005. Late Quaternary palaeoenvironments of the Mfabeni peatland, northern
744 KwaZulu Natal. MSc Thesis, University of KwaZulu-Natal, Pietermaritzburg.
- 745 Finch, J.M., Hill, T.R., 2008. A late Quaternary pollen sequence from Mfabeni Peatland,
746 South Africa: Reconstructing forest history in Maputaland. *Quaternary Research* 70 (3),
747 442-450.
- 748 Francez, A.J., Vasander, H., 1995. Peat accumulation and peat decomposition after human
749 disturbance in French and Finnish mires. *Acta Oecologica* 16 (5), 599-608.
- 750 Fogel, M.L., Cifuentes, L.A., 1993. Isotope fractionation during primary production. *Organic*
751 *geochemistry: Principles and applications*, 73-98.
- 752 Gälman, V., Rydberg, J., De-Luna, S.S., Bindler, R., Renberg, I., 2008. Carbon and nitrogen
753 loss rates during aging of lake sediment: Changes over 27 years studied in varved lake
754 sediment. *Limnology and Oceanography* 53 (3), 1076-1082.

755 Gasse, F., 2000. Hydrological changes in the African tropics since the Last Glacial Maximum.
756 Quaternary Science Reviews 19 (1-5), 189-211.

757 Gasse, F., Chalié, F., Vincens, A., Williams, M.A.J., Williamson, D., 2008. Climatic patterns in
758 equatorial and southern Africa from 30,000 to 10,000 years ago reconstructed from
759 terrestrial and near-shore proxy data. Quaternary Science Reviews 27 (25-26), 2316-
760 2340.

761 Gillson, L., Waldron, S., Willis, K.J., 2004. Interpretation of soil $\delta^{13}\text{C}$ as an indicator of
762 vegetation change in African savannas. Journal of Vegetation Science 15 (3), 339-350.

763 Gorham, E., 1991. Northern peatlands: role in the carbon cycle and probable responses to
764 climatic warming. Ecological Applications 1 (2), 182-195.

765 Goslar, T., Czernik, J., Goslar, E., 2004. Low-energy ^{14}C AMS in Poznań Radiocarbon
766 Laboratory, Poland. Nuclear Instruments and Methods in Physics Research Section B:
767 Beam Interactions with Materials and Atoms (223–224), 5-11.

768 Grundling, P.L., 2001. The Quaternary peat deposits of Maputaland, Northern Kwazulu-
769 Natal, South Africa: categorisation, chronology and utilisation. MSc Thesis, University of
770 Johannesburg.

771 Grundling, P., Grootjans, A.P., Price, J.S., Ellery, W.N., 2013. Development and persistence of
772 an African mire: How the oldest South African fen has survived in a marginal climate.
773 Catena 110, 176-183.

774 Hemming, S.R., 2004. Heinrich events: Massive late Pleistocene detritus layers of the North
 775 Atlantic and their global climate imprint. *Reviews of Geophysics* 42 (1), 1-43.

776 Holmgren, K., Lee-Thorp, J.A., Cooper, G.R.J., Lundblad, K., Partridge, T.C., Scott, L.,
 777 Sithaldeen, R., Siep Talma, A., Tyson, P.D., 2003. Persistent millennial-scale climatic
 778 variability over the past 25,000 years in southern Africa. *Quaternary Science Reviews*
 779 22 (21-22), 2311-2326.

780 Holzkämper, S., Holmgren, K., Lee-Thorp, J., Talma, S., Mangini, A., Partridge, T., 2009. Late
 781 Pleistocene stalagmite growth in Wolkberg Cave, South Africa. *Earth and Planetary*
 782 *Science Letters* 282 (1-4), 212-221.

783 Hua, Q., Barbetti, M., 2004. Review of tropospheric bomb C-14 data for carbon cycle
 784 modelling and age calibration purposes. *Radiocarbon*, 46 (3), 1273-1298.

785 Immirzi, C.P., Maltby, E., Clymo, R.S., 1992. The global status of peatlands and their role in
 786 carbon cycling: A report for the Friends of the Earth. Wetlands Ecosystems Research
 787 Group, Department of Geography, University of Exeter, London.

788 Jedrysek, M., Skrzypek, G., 2005. Hydrogen, carbon and sulphur isotope ratios in peat: The
 789 role of diagenesis and water regimes in reconstruction of past climates. *Environmental*
 790 *Chemistry Letters* 2 (4), 179-183.

791 Jones, M.C., Peteet, D.M., Sambrotto, R., 2010. Late-glacial and Holocene $\delta^{15}\text{N}$ and $\delta^{13}\text{C}$
 792 variation from a Kenai Peninsula, Alaska peatland. *Palaeogeography,*
 793 *Palaeoclimatology, Palaeoecology* 293 (1-2), 132-143.

794 Kelbe, B.E., Rawlins, B.K., Nomquphu, W., 1995. Geohydrological modelling of Lake St Lucia.
 795 Report, University of Zululand, KwaDlangezwa 3886 KwaZulu-Natal.

796 Kotze, D.C., O'Connor, T.G., 2000. Vegetation variation within and among palustrine
 797 wetlands along an altitudinal gradient in KwaZulu-Natal, South Africa. *Plant Ecology* 146
 798 (1), 77-96.

799 Kristen, I., Wilkes, H., Vieth, A., Zink, K., Plessen, B., Thorpe, J., Partridge, T.C., Oberhänsli,
 800 H., 2010. Biomarker and stable carbon isotope analyses of sedimentary organic matter
 801 from Lake Tswaing: Evidence for deglacial wetness and early Holocene drought from
 802 South Africa. *Journal of Paleolimnology* 44 (1), 143-160.

803 Kuhry, P., Vitt, D.H., 1996. Fossil carbon/nitrogen ratios as a measure of peat
 804 decomposition. *Ecology* 77 (1), 271-275.

805 Lee-Thorp, J., Holmgren, K., Lauritzen, S.E., Linge, H., Moberg, A., Partridge, T., Stevenson,
 806 C., Tyson, P., 2001. Rapid climate shifts in the southern African interior throughout the
 807 mid to late Holocene. *Geophysical Research Letters* 28 (23), 4507-4510.

808 Lubke, R.A., Moll, J.B., Avis, A.M., 1992. Rehabilitaion ecology in coastal and
 809 environmental services: eastern shores of Lake St Lucia. Kingsa/Tojan lease area special
 810 reports 1, biophysical environment, Council of Scientific and Industrial Research,
 811 Pretoria, South Africa.

812 Mayewski, P.A., Rohling, E.E., Stager, J.C., Karlén, W., Maasch, K.A., Meeker, L.D., Meyerson,
 813 E.A., Gasse, F., van Kreveld, S., Holmgren, K., Lee-Thorp, J., Rosqvist, G., Rack, F.,

814 Staubwasser, M., Schneider, R.R., Steig, E.J., 2004. Holocene climate variability.
 815 Quaternary Research 62 (3), 243-255.

816 McCormac, F.G., Hogg, A.G., Blackwell, P.G., Buck, C.E., Higham, T.F.G., Reimer, P.J., 2004.
 817 SHCal04 southern Hemisphere calibration, 0-11.0 cal kyr BP. Radiocarbon 46 (3), 1087-
 818 1092.

819 Meadows, M.E., Baxter, A.J., Parkington, J., 1996. Late Holocene environments at
 820 Verlorenvlei, Western Cape Province, South Africa. Quaternary International 33 (1), 81-
 821 95.

822 Meadows, M.E., Baxter, A.J., 1999. Late Quaternary palaeoenvironments of the
 823 southwestern Cape, South Africa: A regional synthesis. Quaternary International 57-58,
 824 193-206.

825 Meyers, P.A., 2003. Applications of organic geochemistry to paleolimnological
 826 reconstructions: A summary of examples from the Laurentian Great Lakes. Organic
 827 Geochemistry 34 (2), 261-289.

828 Meyers, P.A., 1997. Organic geochemical proxies of paleoceanographic, paleolimnologic,
 829 and paleoclimatic processes. Organic Geochemistry 27, (5-6), 213-250.

830 Meyers, P.A., 1994. Preservation of elemental and isotopic source identification of
 831 sedimentary organic matter. Chemical Geology 114 (3-4), 289-302.

832 Meyers, P.A., Ishiwatari, R., 1993. Lacustrine organic geochemistry-an overview of
 833 indicators of organic matter sources and diagenesis in lake sediments. *Organic*
 834 *Geochemistry* 20 (7), 867-900.

835 Mucina, L., Adams, J.B., Knevel, I.C., Rutherford, M.C., Powrie, L.W., Bolton, J.J., van der
 836 Merwe, J.H., Anderson, R.J., Bornman, T.G., le Roux, A., Janssen, J.A.M., 2006. Coastal
 837 Vegetation of South Africa, in: Mucina, L., Rutherford, M.C. (Eds.), *The vegetation of*
 838 *South Africa, Lesotho and Swaziland*. South African National Biodiversity Institute,
 839 Pretoria, pp. 658-696.

840 Muzuka, A.N.N., 1999. Isotopic compositions of tropical East African flora and their
 841 potential as source indicators of organic matter in coastal marine sediments. *Journal of*
 842 *African Earth Sciences* 28 (3), 757-766.

843 Neumann, F.H., Scott, L., Bousman, C.B., van As, L., 2010. A Holocene sequence of
 844 vegetation change at Lake Eteza, coastal KwaZulu-Natal, South Africa. *Review of*
 845 *Palaeobotany and Palynology* 162 (1), 39-53.

846 Neumann, F.H., Stager, J.C., Scott, L., Venter, H.J.T., Weyhenmeyer, C., 2008. Holocene
 847 vegetation and climate records from Lake Sibaya, KwaZulu-Natal (South Africa). *Review*
 848 *of Palaeobotany and Palynology* 152 (3-4), 113-128.

849 Nichols, J.E., Walcott, M., Bradley, R., Pilcher, J., Huang, Y., 2009. Quantitative assessment of
 850 precipitation seasonality and summer surface wetness using ombrotrophic sediments
 851 from an Arctic Norwegian peatland. *Quaternary Research* 72 (3), 443-451.

852 Nilsson, M., Sagerfors, J., Buffam, I., Laudon, H., Eriksson, T., Grelle, A., Klemedtsson, L.,
853 Weslien, P., Lindroth, A., 2008. Contemporary carbon accumulation in a boreal
854 oligotrophic minerogenic mire, a significant sink after accounting for all C-fluxes. *Global*
855 *Change Biology* 14 (10), 2317-2332.

856 Norström, E., Scott, L., Partridge, T.C., Risberg, J., Holmgren, K., 2009. Reconstruction of
857 environmental and climate changes at Braamhoek wetland, eastern escarpment South
858 Africa, during the last 16,000 years with emphasis on the Pleistocene-Holocene
859 transition. *Palaeogeography, Palaeoclimatology, Palaeoecology* 271 (3-4), 240-258.

860 Oldfield, F., Thompson, R., Crooks, P.R.J., Gedye, S.J., Hall, V.A., Harkness, D.D., Housley,
861 R.A., McCormac, F.G., Newton, A.J., Pilcher, J.R., Renberg, I., Richardson, N., 1997.
862 Radiocarbon dating of a recent high latitude peat profile: Stor Amyran, northern
863 Sweden. *Holocene* 7 (3), 283-290.

864 Page, S.E., Rieley, J.O., Banks, C.J., 2011. Global and regional importance of the tropical
865 peatland carbon pool. *Global Change Biology* 17 (2), 798-818.

866 Partridge, T.C., 2002. Were Heinrich events forced from the southern hemisphere? *South*
867 *African Journal of Science* 98 (1-2), 43-46.

868 Porat, N., Botha, G., 2008. The luminescence chronology of dune development on the
869 Maputaland coastal plain, southeast Africa. *Quaternary Science Reviews* 27 (9-10),
870 1024-1046.

871 Preston-Whyte, R.A., Tyson, P.D., 1998. *The atmosphere and weather of southern Africa.*
872 Oxford University Press, South Africa.

873 Pudsey, C.J., Evans, J., 2001. First survey of Antarctic sub-ice shelf sediments reveals mid-
874 Holocene ice shelf retreat. *Geology* 29 (9), 787-790.

875 Reimer, P.J., Baillie, M.G.L., Bard, E., Beck, J.W., Blackwell, P.G., Buck, C.E., Burr, G.S.,
876 Edwards, R.L., Friedrich, M., Guilderson, T.P., 2006. Comment on: Radiocarbon
877 calibration curve spanning 0 to 50,000 years BP based on paired $^{230}\text{Th}/^{234}\text{U}/^{238}\text{U}$
878 and ^{14}C dates on pristine corals by RG Fairbanks et al., 2005. *Quaternary Science*
879 *Reviews* 24, 1781-1796 and Extending the radiocarbon calibration beyond 26,000 years
880 before present using fossil corals by Chiu, T.C., 2005. *Quaternary Science Reviews* 24,
881 1797-1808. *Quaternary Science Reviews* 25, 855-862.

882 Rieley J.O., Ahmad-Shah A.A., Brady M.A., 1996. The extent and nature of tropical peat
883 swamps - Tropical Lowland Peatlands of Southeast Asia. In: Maltby E., Immirzi C.P.,
884 Safford R.J. (Eds.), *Tropical Lowland Peatlands of Southeast Asia -proceedings of a*
885 *workshop on integrated planning and management of tropical lowland peatlands:*
886 *workshop on integrated planning and management of tropical lowland peatlands.*
887 IUCN, Gland, Switzerland.

888 Routh, J., Meyers, P.A., Gustafsson, O., Baskaran, M., Hallberg, R., Scholdström, A., 2004.
889 Sedimentary geochemical record of human-induced environmental changes in the Lake
890 Brunnsviken watershed, Sweden. *Limnology and Oceanography* 49 (5), 1560-1569.

891 Sage, R.F., Wedin, D.A., Li, M., 1999. The Biogeography of C4 Photosynthesis: Patterns and
892 Controlling Factors in: Sage, R.F., Monson, R.K. (Eds.), *C4 Plant Biology*. Academic Press,
893 San Diego, pp. 313-373.

894 Schefusz, E., Kuhlmann, H., Mollenhauer, G., Prange, M. & Patzold, J., 2011. Forcing of wet
 895 phases in southeast Africa over the past 17,000 years. *Nature Geoscience* 480 (7378),
 896 509-512

897 Schmittner, A., Saenko, O.A., Weaver, A.J., 2003. Coupling of the hemispheres in
 898 observations and simulations of glacial climate change. *Quaternary Science Reviews* 22
 899 (5-7), 659-671.

900 Skrzypek, G., Jezierski, P., Szyrkiewicz, A., 2010. Preservation of primary stable isotope
 901 signatures of peat-forming plants during early decomposition - observation along an
 902 altitudinal transect. *Chemical Geology* 273 (3-4), 238-249.

903 Skrzypek, G., Paul, D., Wojtun, B., 2008. Stable isotope composition of plants and peat from
 904 Arctic mire and geothermal area in Iceland. *Polish Polar Research* 29 (4), 365-376.

905 Smuts, W.J., 1992. Peatlands of the Natal mire complex: geomorphology and
 906 characterization. *South African Journal of Science* 88 (9-10), 474-483.

907 Staub, J.R., Esterle, J.S., 1994. Peat-accumulating depositional systems of Sarawak, East
 908 Malaysia. *Sedimentary Geology* 89 (1-2), 91-106.

909 Stenni, B., Masson-Delmotte, V., Johnsen, S., Jouzel, J., Longinelli, A., Monnin, E.,
 910 Röthlisberger, R., Selmo, E., 2001. An oceanic cord reversal during the last deglaciation.
 911 *Science* 293 (5537), 2074-2077.

912 Stock, W.D., Chuba, D.K., Verboom, G.A., 2004. Distribution of South African C3 and C4
 913 species of Cyperaceae in relation to climate and phylogeny. *Austral Ecology* 29 (3), 313-
 914 319.

915 Stocker, T.F., 2000. Past and future reorganizations in the climate system. *Quaternary*
 916 *Science Reviews* 19 (1-5), 301-319.

917 Stokes, S., Thomas, D.S.G., Washington, R., 1997. Multiple episodes of aridity in southern
 918 Africa since the last interglacial period. *Nature* 388 (6638), 154-158.

919 Strack, M., 2008. Peatlands and climate change. International Peat Society, Jyväskylä,
 920 Finland.

921 Stuiver, M., Polach, H.A., 1977. Discussion: Reporting on ¹⁴C data. *Radiocarbon* 19 (3), 355-
 922 363.

923 Talma, A., Vogel, J.C., 1992. Late Quaternary paleotemperatures derived from a speleothem
 924 from Cango caves, Cape Province, South Africa. *Quaternary Research* 37 (2), 203-213.

925 Taylor, R., Kelbe, B., Haldorsen, S., Botha, G.A., Wejden, B., Varet, L., Simonsen, M.B., 2006a.
 926 Groundwater-dependent ecology of the shoreline of the subtropical Lake St Lucia
 927 estuary. *Environmental Geology* 49 (4), 586-560.

928 Taylor, R., Adams, J.B., Haldorsen, S., 2006b. Primary habitats of the St Lucia estuarine
 929 system, South Africa, and their responses to mouth management. *African Journal of*
 930 *Aquatic Science* 31 (1), 31-41.

- 931 Turunen, J., Tomppo, E., Tolonen, K., Reinikainen, A., 2002. Estimating carbon accumulation
932 rates of undrained mires in Finland - Application to boreal and subarctic regions.
933 *Holocene* 12 (1), 69-80.
- 934 Tyson, P.D., Preston-Whyte, R.A., 2000. The weather and climate of southern Africa, Oxford
935 University Press Incorporated, Cape Town, South Africa.
- 936 Vaeret, L., Sokolic, F., 2008. Methods for studying the distribution of groundwater-
937 dependent wetlands: a case study from Eastern Shores, St Lucia, South Africa, in:
938 Responses to global change and management actions in coastal groundwater
939 resources, Norwegian University of Life Sciences.
- 940 Valsecchi, V., Chase, B.M., Slingsby, J.A., Carr, A.S., Quick, L.J., Meadows, M.E., Cheddadi, R.
941 & Reimer, P.J., 2013. A high resolution 15,600-year pollen and microcharcoal record
942 from the Cederberg Mountains, South Africa. *Palaeogeography, Palaeoclimatology,*
943 *Palaeoecology* 387, 6-16.
- 944 Venter, C.E., 2003. The vegetation ecology of Mfabeni peat swamp, St Lucia, KwaZulu-Natal.
945 MSc Thesis, University of Pretoria.
- 946 Vogel, J.C., Fuls, A., 1978. The Geographical Distribution of Kranz Grasses in South Africa.
947 *South African Journal of Science* 74, 209-215.
- 948 Vrdoljak, S.M., Hart, R.C., 2007. Groundwater seeps as potentially important refugia for
949 freshwater fishes on the Eastern Shores of Lake St Lucia, KwaZulu-Natal, South Africa.
950 *African Journal of Aquatic Science* 32 (2), 125-132.

- 951 Walther, S.C., Neumann, F.H., 2011. Sedimentology, isotopes and palynology of late
952 Holocene cores from Lake Sibaya and the Kosi Bay system (KwaZulu-Natal, South
953 Africa). South African Geographical Journal 93 (2), 133-153.
- 954 Worrall, F., Reed, M., Warburton, J., Burt, T., 2003. Carbon budget for a British upland peat
955 catchment. Science of the Total Environment 312 (1-3), 133-146.
- 956

Tables

Table 1: Sediment properties with averaged accumulation data for the five LSR stages in core SL6

Core #	n	Sediment type	Porosity	Bulk Density g.cm ⁻³	LSR ^a mm.yr ⁻¹	MAR ^b g.m ⁻² .yr ⁻¹	TOC gC.m ⁻²	C Accumulation g C.m ⁻² .yr ⁻¹
Averages								
LSR 1	38	Peat	0.73	0.29	0.13	37.6	1068	13.6
LSR 2	20	Peaty sand	0.64	0.34	0.22	76.4	553	12.4
LSR 3	22	Peat	0.73	0.28	0.15	41.4	896	13.0
LSR 4	20	peat	0.72	0.28	0.10	28.1	711	7.4
LSR 5	98	peat	0.69	0.24	0.29	70.9	1108	32.3

^a LSR = Linear Sedimentation Rate

^b MAR = Mass Accumulation Rate

964 Table 2: Chronology of Mfabeni peatland core SL6 with uncalibrated AMS ^{14}C dates,
 965 corresponding percentage total organic carbon (TOC) and bulk stable carbon isotope values

966

Sample code	Depth (cm)	Material	uncalibrated date	% C	$\delta^{13}\text{C}_{\text{TOC}}$
SL6 10-11	10	Peat	109 ± 0.35 pMC	45.1	-21.8
SL6 109-110	109	Peat	3240 ± 30 BP	51.9	-16.8
SL6 209-210	209	Peat	6170 ± 30 BP	19.2	-24.5
SL6 309-310	309	Peat	9270 ± 70 BP	37.1	-21.5
SL6 405-406	405	Peat	16940 ± 80 BP	14.7	-19.1
SL6 510-511	510	Peat	22800 ± 130 BP	29.6	-18.6
SL6 609-610	609	Peat	27600 ± 190 BP	22.5	-19.0
SL6 709-710	709	Peat	>48000 BP	28.1	-17.6
SL6 805-806	805	Peat	49000 ± 2200 BP	12.2	-20.1

967

968 Table 3: Statistical relationships between isotopic and elemental profiles for LSR stages in
 969 Core SL6. (Two tail test, P=probability level, df=degrees of freedom)

	$\delta^{15}\text{N} : \text{C/N}$	$\delta^{13}\text{C} : \text{C/N}$	$\delta^{15}\text{N} : \delta^{13}\text{C}$	%N : %C
LSR Stage 1				
r	0.34	0.61	0.39	0.90
(P=0.01; df=36)		Sig		Sig
LSR Stage 2				
r	-0.78	-0.66	0.52	0.99
(P=0.01; df=15)	Sig	Sig		Sig
LSR Stage 3				
r	-0.69	0.83	-0.36	0.84
(P=0.01; df =20)	Sig	Sig		Sig
LSR Stage 4				
r	0.25	0.84	-0.36	0.98
(P=0.01; df=18)		Sig		Sig
LSR Stage 5				
r	0.06	0.16	-0.18	0.46
(P=0.01; df=97)				Sig

Figure Captions

Figure 1: Mfabeni Peatland, iSimangaliso Wetland Park, St Lucia, northern KwaZulu Natal, South Africa with major habitat groups and dominant water bodies represented.

Figure 2: Core SL6 stratigraphic profile with modelled ages.

Figure 3: Core SL6 age-depth model with uncertainty ranges. AMS ^{14}C dates were calibrated with IntCal09 curve (Reimer et al., 2006) with a southern Hemisphere offset of $c. 40 \pm 20$ ^{14}C years. Postbomb dates calibrated with southern Hemisphere postbomb curve by Hua and Barbetti (2004). All dates were adjusted within the Bayesian framework, using Bacon age depth modelling software (Blaauw and Christen 2011).

Figure 4: Downcore profile of Core SL6 presenting linear sedimentation rate (LSR), mass accumulation rate (MAR), total organic carbon (TOC) and carbon accumulation rate (CAR) with age and depth. Red line represents total core average. Hol = Holocene, YD = Younger Dryas; H1-5 = Heinrich events, LGM = Last Glacial Maximum. Heinrich dates after Hemming (2004).

Figure 5: SL6 down core profile representing stable C and N isotope signals, elemental weight percentages for C and N and atomic C/N ratio. Hol = Holocene, H1-5 = Heinrich events, LGM = Last Glacial Maximum. Heinrich dates after Hemming (2004).

Figure 6: Proportion of carbon input from C4 plants (Equation 1, see methods section 2.5). Hol = Holocene, H1-5 = Heinrich events, LGM = Last Glacial Maximum. Heinrich dates after Hemming (2004).

994 **Figure 1**

995

996

997

Core SL6

Dark brown to black, fine grained, consolidated "fibrous" peat with extensive rootlets 0 – 61 cm	10 cm	-16 cal yr BP
		61 cm	1703 cal yr BP
Semi-consolidated black fine grained peat, rootlets, more advanced stage of decomposing 61 – 110 cm	110 cm	3471 cal yr BP
Semi-consolidated black amorphous peat with minimal rootlets 110 – 340 cm	340 cm	13114 cal yr BP
Semi-consolidated black amorphous peat with increasingly sandy texture 340 – 440 cm	440 cm	22323 cal yr BP
Semi-consolidated black peat with grey sand mottling alternating with amorphous peat 440 – 535 cm	535 cm	28559 cal yr BP
Semi-consolidated dark brown medium grained peat with grey sand mottling alternating with medium grained sand and amorphous peat 535 - 610 cm	610 cm	31898 cal yr BP
Semi-consolidated black amorphous peat 610-810cm	810 cm	47272 cal yr BP

Not to Scale

1000 **Figure 3**

1001

1002

1003

1004 **Figure 4**

1005

1006

1007 **Figure 5**

1008
1009

1010 **Figure 6**

1011

1012