Climatic variability in Mfabeni peatlands (South Africa) since the late Pleistocene

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Abstract

It has been postulated that a bipolar seesaw interhemispheric mechanism dominated the relationship between the Northern and Southern hemisphere climates since the late Pleistocene. A key test for this proposition would be to undertake palaeoenvironmental studies on terrestrial archives in climatically sensitive regions. Southern Africa's contemporary C_3 and C_4 terrestrial plant distributions display a definitive geographical pattern dictated by different growing season rainfall and temperature zones; however, the region is generally archive poor due to its overall semi-arid climate and high relief topography. The Mfabeni peatland, with a basal age of c. 47 k yrs calibrated before present (kcal yr BP), is one of the oldest continuous coastal peat deposits in Southern Africa. Molecular leaf wax isotopes ($\delta^{13}C_{wax}$) were generated for a 810 cm long core, and combined with previously published bulk geochemical ($\delta^{13}C_{bulk}$, %TOC), palynological, and stratigraphic data, to reconstruct the late Pleistocene and Holocene palaeoenvironments. We interpreted environmental shifts associated with the Heinrich 4, Last Glacial Maximum, deglacial and Holocene periods, which are consistent with adjacent Indian Ocean sea surface temperature records. However, the other shorter climate perturbations during the Heinrich 5, 3, 2, 1, Antarctic cold reversal and Younger Dryas, were muted, most likely due to local hydrological overprinting on the Mfabeni record. A general anti-phase sequence was observed between the Mfabeni record and better established Northern Hemisphere events, underpinning the bipolar seesaw interhemispheric mechanism proposed for global climate forcing since the Late Pleistocene.

Keywords: Holocene; Pleistocene; Paleoclimatology; Southern Africa; Stable isotopes; Continental biomarkers; Peatland

1 Introduction

There has long been a quest amongst the paleoclimatology community to establish if a North-South interhemispheric climate relationship existed during the late Pleistocene and Holocene. Varying degrees of anti-phase interhemispheric coupling have been shown by several studies (Baker et al., 2014, 2016; Bard et al., 1997; Blunier et al., 1998; Chase et al., 2011; Kaplan et al., 2010; Putnam et al., 2010; Schefuß et al., 2011; Schmittner et al., 2003) with numerous mechanisms for climate forcing postulated, prompting a requirement for additional high resolution regional palaeoenvironment studies to corroborate or challenge these hypotheses. One of the more popular theories is the bipolar seesaw (Broecker, 1998; Stocker, 1998; Stocker and Johnsen, 2003) caused by the slow-down or switching off of the oceanic Atlantic Overturning Meridional Circulation (AMOC) that transfers heat from the equatorial and Southern Oceans towards the higher latitudes in the Northern Hemisphere (Chase et al., 2015). Understanding how regional and global climate forcing mechanisms interacted in the recent geological past is critical for modelling global weather patterns in response to anthropogenic influences, not only for food security but its impact on biodiversity. Majority of climate studies have focused on marine sediment cores that are influenced by different sources upwind and/or upstream during the time of their deposition (Railsback et al., 2016) which can complicate the continental signal. Therefore, much emphasis is now being placed on discovering new terrestrial climate archives to unravel the impact of past climate on hydrologic conditions within the core continental precipitation zones. Southern Africa is situated at the junction between tropical, sub-tropical and temperate climate systems, with varying topography/altitudes, and three distinctive rainfall zones, namely winter rainfall (WRZ), all-year rainfall (ARZ) and summer rainfall (SRZ) zones (Fig. 1). However, this unique region suffers from a lack of continuous terrestrial climate archives, mainly due to a semi-arid climate and high relief topography that is not conducive for archive preservation. The Mfabeni peatland, which is located on the eastern shores of South Africa (Fig. 1), returned a basal ^{14}C age of c. 47 kyrs calibrated before present (kcal yr BP), establishing the record as one of the oldest known continuous terrestrial archive of its kind in Southern Africa (Baker et al., 2014; Grundling et al., 2013), that provides a unique opportunity to conduct high-resolution glacial and interglacial palaeoecological investigations of past climate in Southeast Africa.

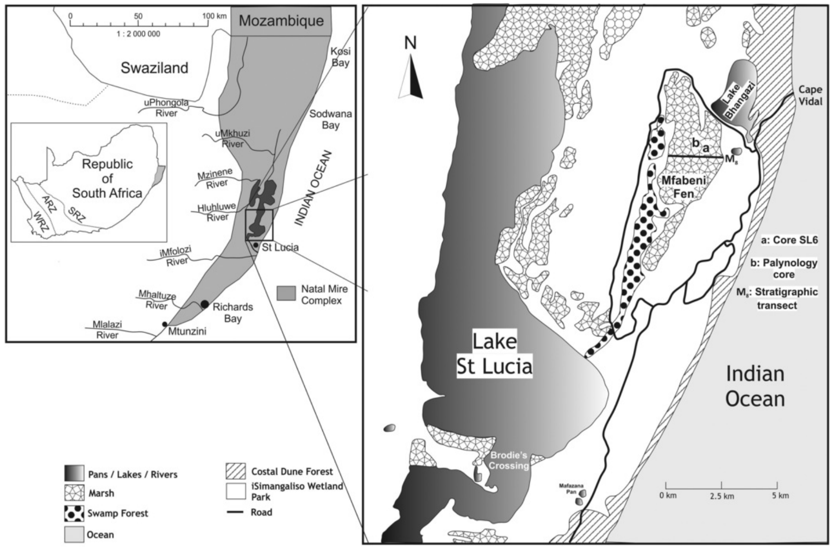

Fig. 1 Location of core SL6 (a) in the Mfabeni peatland, iSimangaliso Wetland Park, northern Kwazulu-Natal, South Africa. Location of palynology core (b; Finch and Hill, 2008) and most proximal and deepest stratigraphic transect ($M_{8;}$ Grundling et al., 2013) included for orientation. WRZ = winter rainfall zone; ARZ = all-year rainfall zone; SRZ = summer rainfall zone. Modified from Baker et al. (2016).

alt-text: Fig. 1

Approximately 85% of all terrestrial plant species use the C_3 photosynthetic pathway to fix carbon (Ehleringer et al., 1997). In contrast, the C_4 pathway is restricted to the Dicotyledonae and Monocotyledonae clades, with the greatest diversity found within the widespread monocots group dominated by the Poaceae (grasses) and Cyperaceae (sedges) families (Ehleringer et al., 1997). Vogel et al. (1978) found that the regional C_4 grasses predominate in the SRZ, whereas C_3 grasses are more prevalent in the WRZ and high altitude regions of the eastern escarpment of South Africa. On the other hand, Stock et al. (2004) showed that temperature variations did not fully account for the C_3 and C_4 Cyperaceae (sedges) family distributions in South Africa, but seem to be more closely related to annual rainfall. They suggest that the C_4 sedges evolved under wetland conditions, but are more abundantly represented in warm and moist (sub) tropical areas (Rommerskirchen et al., 2006), although not at such high proportions as the C_4 Poaceae family (up to 95% in SRZ; Vogel et al., 1978).

The different biochemical pathways employed by C_3 and C_4 plants during photosynthesis causes a variation in degrees of ^{13}C/^{12}C fractionation, resulting in characteristic bulk δ^{13}C values ranging from −22‰ to −30‰ for C_3 and -10‰ to −14‰ for C_4 plants (Vogel et al., 1978). This difference in the biochemical pathways, gives rise to distinct leaf wax δ^{13}C ($\delta^{13}C_{wax}$) signatures that are preserved in peat deposits. The causes of shifts in the proportions of C_3 and C_4 plants over time are still subject to intense scientific debate. Originally it was thought that the Hatch-Slack pathway employed by C_4 plants during photosynthesis was more efficient and therefore favoured under low atmospheric CO_2 (Cerling et al., 1997; Ehleringer et al., 1997). More recent studies have indicated that changes in the balance of C_3/C_4 plants can be affected by local environmental factors such as temperature and aridity (Castañeda et al., 2009; Huang et al., 2001; Khon et al., 2014; Schefuß et al., 2003; Scott, 2002; Xue et al., 2014; Yamamoto et al., 2010). Energy models have previously been employed to predict the range of temperatures and atmospheric ρCO_2 under which C_3 and C_4 pathways may have selective advantages in grasslands (Ehleringer, 1978). Under current CO_2 concentrations, the C_4 pathway was predicted to have a higher photosynthetic efficiency at growing season temperatures in excess of 27 °C, while the C_3 pathway was more efficient at temperatures below 22 °C (Ehleringer et al., 1997). It has also been observed that C_4 tropical grasses display a higher water use efficiency (WUE) rate than their C_3 counterparts (Downes, 1969) which led to a general acceptance that C_4 plants were better adapted to arid environments. However, other studies have found that some C_4 grass communities' productivity increased relative to their competing C_3 equivalents with increased rainfall and in some cases are more abundant than C_3 grasses in wetland habitats (Epstein et al., 1997; Kubien and Sage, 2003; Paruelo et al., 1998; Paruelo and Lauenroth, 1996). Because it has been shown that C_3 and C_4 *Poaceae* (and Cyperaceae) distributions evolved in distinct geographical areas in Southern Africa, determined largely by growing season temperatures, and the interchanges between C_3 and C_4 plants are represented by archived $\delta^{13}C_{wax}$ values (Eglinton and Hamilton, 1967), a link can potentially be made between the changes in relative C_3/C_4 proportions of OM input, substantiated by the palynology and physical peat forming processes archived in the Mfabeni peat deposit, and fluctuations in palaeotemperature and paleohydrology at the time of sedimentation.

The aim of our study is to reconstruct the late Pleistocene and Holocene depositional environment and to explore the climatic controls governing peat formation in the Mfabeni peatland under glacial and interglacial conditions. We show compound specific $\delta^{13}C_{wax}$, bulk $\delta^{13}C$ and total organic carbon (%TOC) records, integrated with information from proximal stratigraphic (Grundling et al., 2013) and palynological (Finch and Hill, 2008) studies undertaken in the peatland, to reconstruct peatland hydrology, sedimentation regimes and proportional inputs of C_3/C_4 plants. We then compare our findings with other regional climate studies to infer the climate controlling these ecological variables, and by presenting the data sets in a Northern Hemisphere climatic framework (Fig. 2), investigate the extent to which the regional Southern African climatic signals correlate with global climatic events since the Late Pleistocene.

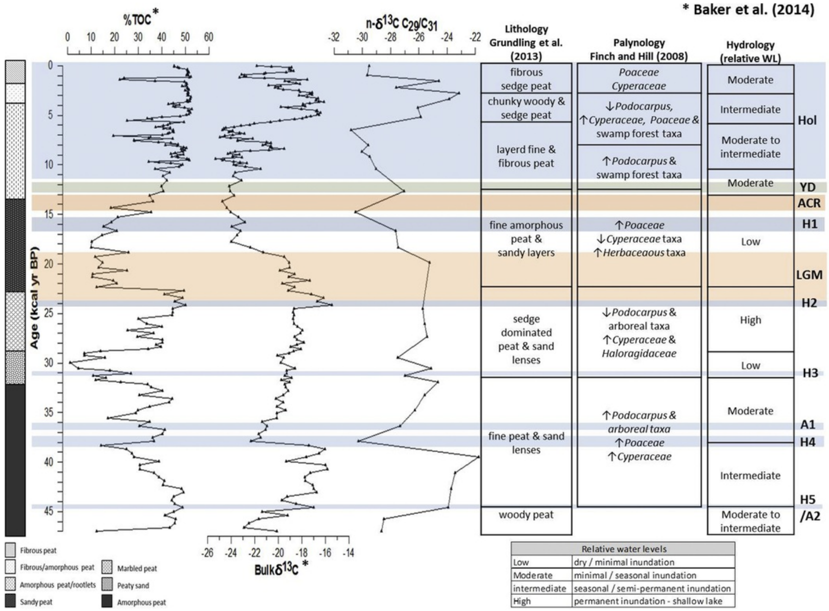

Fig. 2 Core SL6 %TOC (n = 198), bulk carbon (n = 198) and leaf wax (n = 35) isotope profiles in comparison with peatland development chronology by Grundling et al. (2013), palynology record by Finch and Hill (2008) and local hydrology interpretation of the Mfabeni peatland. Core SL6 %TOC and bulk C isotopic data modified from Baker et al. (2014). WL = relative water levels; H1–5 = Heinrich events; A1 and A2 = Antarctic warming events; LGM = Last Glacial Maximum; ACR = Antarctic cold reversal; YD = Younger Dryas; Hol = Holocene.

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2 Methods

2.1 Site description

The UNESCO world Heritage iSimangaliso Wetland Park is situated on the northern coastline of Kwazulu-Natal province, South Africa (Fig. 1). The shallow, but extensive 350 km^2 St Lucia Lake dominates the park and forms part of the largest estuarine wetland system on the African continent (Vrdoljak and Hart, 2007). The Mfabeni fen lies on the eastern shores of Lake St Lucia, within an interdunal basin (Botha and Porat, 2007) measuring c. 10 × 3 km (Clulow et al., 2012), and up to 11 m in depth (Grundling et al., 2013). The fen's hydrology is influenced primarily by the unconfined Maputaland aquifer (Grundling et al., 2013; Taylor et al., 2006a) and local precipitation.

The site falls within a sub-tropical climate, which experiences mainly austral summer rainfall of between 900 and 1200 mm/yr (Taylor et al., 2006b), and is driven by sea surface temperature (SST) anomalies in the adjacent southwest Indian Ocean (Jury et al., 1993; Reason and Mulenga, 1999). However, distinctive cyclical drying events of between 2 and 4 year intervals with a roughly decadal periodicity (Tyson and Preston-Whyte, 2000) have been identified in the contemporary rainfall records from this region (Bate and Taylor, 2008). The Mfabeni fen forms part of the greater Natal Mire Complex (NMC; Fig. 1) that extends from southern Mozambique to Richards Bay, Kwazulu-Natal. It was formed by valley infilling within the KwaMbonanbi formation coastal dune depression (Smuts, 1992), as a result of sustained, but varying groundwater input from the Maputaland aquifer and aggregational blockage of a palaeochannel linking the southern part of the peatland basin with Lake St Lucia (Grundling et al., 2013). The iSimangaliso wetland park vegetation is broadly made up of Maputaland wooded grassland, coastal belt and sub-tropical freshwater wetland and northern coastal forests (Mucina et al., 2006), whereas the fen itself is dominated by herbaceous reed sedges and grasses (Finch and Hill, 2008).

2.2 Sampling

Core SL6 was extracted from the deepest part of the Mfabeni fen (28°9′ 0.76″S; 32°31′ 30.29″E) to a depth of 810 cm in consecutive coring events using a Russian peat corer (5 cm diameter x 50 cm length core chamber). The individual core samples were logged in the field, and later described and sectioned into 1–2 cm intervals in the laboratory, after which the sediments were freeze-dried in preparation for various geochemical analysis.

2.3 Radiocarbon dating

^{14}C dating and age-depth modelling was prepared as per methods listed in Baker et al. (2014, and references therein). Nine evenly spaced samples were chemically pre-treated and measured on a Compact Carbon AMS, and conventional ^{14}C ages were calculated using a correction factor for isotopic fractionation. Ages were then calibrated using the IntCal09 calibration curve with a 40 ± 20 ^{14}C year southern hemisphere offset, while post-bomb ages were calibrated using the southern hemisphere post-bomb curve. Comprehensive core ages were extrapolated within the Bayesian framework, using the age-depth Bacon modelling software.

2.4 Bulk stable C isotope analyses

Bulk stable C and Nitrogen (N) isotope analyses were prepared as Baker et al. (2014 and references therein). The peat samples were analysed on a Thermo Scientific Flash 2000 organic elemental analyser, coupled to a Thermo Scientific Delta V Plus isotope ratio mass spectrometer (detection limit 5 µg), whereas the sand dominated samples were combusted on a Thermo Finnigan Flash EA 1112 series elemental analyser, coupled to a Thermo electron Delta Plus XP isotope ratio mass spectrometer (detection limit 15 µg). The precision for both analytical systems was 0.05 and 0.08‰ for N and C, respectively. N isotopic composition is expressed relative to atmospheric N, whereas C is expressed relative to the Vienna Pee-Dee Belemnite.

2.5 Compound specific *n*-alkane leaf wax isotope analyses

Stable carbon isotope measurements of *n*-alkanes were done using a Delta V Advantage isotope ratio mass spectrometer interfaced with a GC-Isolink Trace Ultra GC Combustion system (Thermo Scientific). Individual *n*-alkanes were separated using a DB-5 capillary column (30 m × 0.32 mm × 0.25 µm). Helium was used as a carrier gas at a flow rate of 1.2 ml/min. The GC oven was programmed from 50 °C (1 min) at 20 °C/min to 150 °C and further at 8 °C/min to 320 °C and held for 10 min isothermally. The saturate hydrocarbon fraction containing *n*-alkanes was injected using programmable temperature vaporization (PTV) injector in splitless mode at 280 °C. Combustion of *n*-alkanes to CO_2 was done at 1030 °C. Carbon isotope compositions of *n*-alkanes are reported based on duplicate analyses of well-resolved peaks. The absolute difference between duplicate *n*-alkane δ^{13}C measurements was 0.5‰ or better.

The weighted average for the δ^{13}C values of *n*-C_{29} and *n*-C_{31} alkanes was determined using the following equation:

$$n{-}\delta^{13}C_{29}/C_{31}=\frac{\left(\delta^{13}C_{29}\times C_{29}\text{ peak area}+\delta^{13}C_{31}\text{ peak area}\right)}{\left(C_{29}\text{ peak area}+C_{31}\text{ peak area}\right)}$$

3 Results

3.1 Leaf wax δ^{13}C signatures ($\delta^{13}C_{wax}$)

The leaf wax compound specific C isotope ($\delta^{13}C_{wax}$) values range between −30.8 and −21.8‰ and display a strong statistical correlation with the bulk δ^{13}C profile (r = 0.87, P = 0.01, df = 35, Fig. 2, Table 1). The $\delta^{13}C_{wax}$ profile exhibits elevated δ^{13}C values from 44.5 to 39.5 kcal yr BP and 4.2 to 2.8 kcal yr BP, with lower values at 38.0, 18.4 to 6.5, 2.7 and 1 kcal yr BP to present. The $\delta^{13}C_{wax}$ values are lower by −8.5 and −3.0‰ with respect to the equivalent bulk δ^{13}C (Table 1).

Table 1 %TOC, bulk and compound specific leaf wax isotopic values for peat samples that underwent biomarker extraction in core SL6. n/a = data not available due to sample contamination or age model restrictions.

alt-text: Table 1

Depth (cm)	Age kcal yr BP	TOC m^{-2}gC m^{-2}	Bulk δ^{13}C	CSIA δ^{13}C^{29}/C^{31}
8	−0.02	n/a	−21.8	−29.5
42	0.98	1163	−22.9	−29.7
57	1.54	540	−19.5	−24.6
75	2.17	1121	−20.3	−27.6
93	2.81	1370	−17.3	−23.2
108	3.39	1491	−16.8	−23.8
135	4.26	1311	−19.2	−26.1
158	5.14	1146	−17.6	−25.9
192	6.46	1227	−24.8	−30.8
209	7.08	371	−24.5	n/a
237	7.99	1194	−21.3	−29.6
258	8.69	1401	−23.7	−30.0
273	9.17	1001	−23.9	−29.5
310	10.40	729	−21.5	−29.1
335	12.64	1005	−24.1	−27.1
355	14.79	871	−24.0	−30.5
375	16.67	657	−23.2	−27.6
390	18.32	295	−22.4	−27.5
405	19.83	469	−19.1	−25.3
424	21.34	393	−19.1	n/a
450	23.06	1131	−17.2	n/a
470	24.50	1118	−18.7	−25.7
490	26.07	955	−18.7	−25.6
510	27.33	850	−18.6	−25.4
539	28.77	434	−18.7	n/a
555	29.46	510	−19.5	−27.5
580	30.58	157	−18.6	−25.2
595	31.24	371	−19.5	−27.0
610	31.90	671	−19.0	−24.6
630	33.21	900	−19.6	−25.6

650	34.76	810	−19.3	−26.3
670	36.33	821	−21.0	−27.3
690	37.92	1063	−22.3	−30.3
710	39.52	922	−17.6	−21.8
730	41.10	1080	−17.3	−23.4
750	42.68	1251	−17.0	−23.7
775	44.61	1602	−17.0	−23.9
790	45.78	1441	−21.7	−28.5
805	46.97	396	−20.1	−28.7

3.2 Bulk $\delta^{13}C$ signatures ($\delta^{13}C_{bulk}$)

As reported in Baker et al. (2014), $\delta^{13}C$ bulk 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. 2). With the exception of the base interval, the $\delta^{13}C$ signal becomes ^{13}C-enriched between c. 46.6 and c. 38.3 kcal yr BP before becoming ^{13}C-depleted at c. 37.9 kcal yr BP. Thereafter, the $\delta^{13}C$ signal increases steadily to the core maximum value at c. 24.2 kcal yr BP, and then declines again to the core minimum value at c. 9.4 kcal yr BP. For the remaining part of the upper core, the $\delta^{13}C$ values are relatively elevated through a succession of fluctuating cycles displaying pronounced ^{13}C-enrichment shifts by 4.5‰ (c. 8.8 to 8.4 kcal yr BP), 8.2‰ (c. 6.5 to 4.7 kcal yr BP), and 4.5‰ (c. 1.2 to 0.15 kcal yr BP).

3.3 Total organic carbon (%TOC)

As reported in Baker et al. (2014), the %TOC values of core SL6, fluctuates between a minimum of 4% (c. 30.6 kcal yr BP) and a maximum of 52% (c. 3.4 kcal yr BP) with an average of 38% (Fig. 2). %TOC increases until c. 44.6 kcal yr BP, after which a steady overall decline occurs reaching the core minimum at c 30.6 kcal yr BP. %TOC values steadily increase again to 41% at c. 23.1 kcal yr BP, followed by a sharp decline and continued below average %TOC values are exhibited until c. 12.6 kcal yr BP. Thereafter, %TOC increases in the lead up to and during the Holocene which shows overall elevated %TOC values with sharp millennial scale declines at c. 7.08 and 1.54 kcal yr BP.

4 Discussion

Due to lipid biomarkers only making up a small percentage of total OM preserved in sediments (Meyers, 1997), multi-proxy bulk geochemical data should be considered in conjunction with compound specific C isotopic values to help delineate physical peat forming processes and organic matter (OM) sources at time of sedimentation. Tropical and sub-tropical peatlands occur in warm and moist environments, conditions that are favourable for OM remineralisation (Kuder et al., 1998). Since peat accumulates when net primary production (NPP) outstrips microbial decomposition (Chimner and Ewel, 2005), the overriding control on peat formation in the tropics is the extent of waterlogging, and to a lesser degree, NPP, which generates anaerobic conditions and results in peat accumulation (Rieley et al., 1996). In this context, the OM-rich Mfabeni fen deposit is ideally suited for palaeoenvironmental study as it is a relatively ancient continuous peat deposit that is a well-preserved record of an autochthonous depositional regime subject only to local precipitation and groundwater recharge (Baker et al., 2014; Grundling et al., 2013).

Grundling et al. (2013) indicated that the Mfabeni fen's geomorphological and consistent groundwater recharge was the overriding factor for the uninterrupted peat accumulation since its inception and this could have buffered the fen against climatic disturbances. Consistent with this, the biomarker study by Baker et al. (2016) showed definitive periods of high water levels that facilitated a dominant submerged macrophyte input in the Mfabeni (c. 44.5–42.6, 29.7, 26.1–23.1, 16.7–7.1 and 2.2 kcal yr BP). This trend highlights the control local hydrology had on the plant type assemblages in the peatland. However, hydrological studies of the eastern shores of the St Lucia estuarine system (Taylor et al., 2006a) have found that regional and temporal variations in rainfall are the driving factors controlling the local hydrological regime. The highest contemporary rainfall (>1200 mm/yr) occurs on the coastal dune barrier that is the main recharge point of the Maputaland aquifer, with approximately 20% of total precipitation arriving at the dune barrier entering the aquifer (Kelbe and Rawlins, 1993).

Bulk $\delta^{13}C$ values encompass many different inorganic and organic inputs of C into sedimentary deposits, which can complicate reconstructions. Long chain *n*-alkane compounds ($>C_{25}$) are predominantly produced in the leaf waxes of higher plants (Eglinton and Hamilton, 1967) and are one of the most diagenetically resistant groups of hydrocarbons found in sediments (Meyers, 1997). Compound specific C isotope

($\delta^{13}C_{wax}$) values of these long chained *n*-alkane compounds are primarily controlled by the C fixation pathways employed during photosynthesis, and to a lesser degree, changes in temperature, moisture and inorganic C source (Chikaraishi and Naraoka, 2003; Hayes et al., 1990). Hence, observed changes in peat $\delta^{13}C_{wax}$ trends can best be explained by the changes in relative abundance between C_3 and C_4 plant input, and thereupon a direct link can be made to the representational proportion of C_3 and C_4 plants that were present at the time of peat formation. A significant and strong statistical correlation exists between the $\delta^{13}C_{bulk}$ and $\delta^{13}C_{wax}$ trends in core SL6 ($r = 0.87$, $P = 0.01$, $df = 35$), which infers that the dominant control on $\delta^{13}C_{bulk}$ values were the relative contributions of terrestrial higher C_3 and C_4 plant types into the peat record. Additionally, a comparison exercise between Mfabeni bulk $\delta^{13}C$ trends and global atmospheric CO_2 concentration records (Ahn and Brook, 2008; Indermuhle et al., 1999) found no association between the records. This suggests that changes in atmospheric ρCO_2 were not a major cause for shifts in the proportion of C_3 and C_4 plants at the site through time.

Based on the strong statistical correlation between the bulk and leaf wax $\delta^{13}C$ signals, (and lack of correlation with global atmospheric CO_2 trends) in core SL6, we assume that the $\delta^{13}C_{bulk}$ values provide a robust record of the proportional input of C_3 and C_4 vegetation into the peatland, as a result of changes in past ambient temperatures and moisture availability. Furthermore, Baker et al. (2014) showed by comparing elemental C and N trends with depth, and the lack of consistent statistical correlations between $\delta^{13}C_{TOC,}$ $\delta^{15}N$ and atomic C/N, that core SL6 exhibits a high degree of preservation, that advocates for minor post-depositional isotopic C fractionation in the Mfabeni peatland.

4.1 Plant assemblage and chronology

To assist in delineating the late Pleistocene and Holocene palaeoenvironmental origins of the significant shifts in leaf wax and bulk C isotopic signature in the Mfabeni fen, we compare our sedimentary records with proximal stratigraphy (Grundling et al., 2013) and palynology studies (Finch and Hill, 2008). The palynology research was conducted on a single core extracted from the deepest part of the fen in close proximity to core SL6, while the stratigraphic investigation was based on several cores drilled along eight east-west transects, spaced between 800 and 1600 m apart from north to south (see Fig. 1 for core locations and most proximal transect). The three different Mfabeni records correlate strongly (Fig. 2) with respect to boundary and transition ages where major deviations in plant assemblages and peat chronology are recorded. By comparing the bulk geochemical and $\delta^{13}C_{wax}$ data generated from core SL6, with the proximal stratigraphic and palynology records, we were able to reconstruct the palaeoenvironmental changes in the Mfabeni peatland since 47 kcal yr BP.

4.1.1 Before the Last Glacial Maximum (LGM)

From c. 47.0 to 44.5 kcal yr BP (Fig. 2), the molecular isotope signal suggests a high proportion of C_3 plant input in conjunction with the basin stratigraphy recoding a riparian/swamp forest dominated environment (Grundling et al., 2013) with increasing %TOC values. These parameters indicate a significant contribution of C_3 swamp forest/arboreal vegetation, implying moderate water levels during this period. At c. 44.5 kcal yr BP, the leaf wax C isotope signal becomes ^{13}C-enriched, suggesting a shift to higher proportions of C_4 plant input, coincidental with a switch to a shallow lake environment (Grundling et al., 2013). The palynology study records dominant local grass and sedge taxa (Finch and Hill, 2008) correlating with elevated core SL6 %TOC signal (Baker et al., 2014). A sustained elevated water level could feasibly have pushed the C_3 forests to the periphery, and facilitated C_4 sedges and grasses to colonise the peatland basin, resulting in higher %TOC and molecular $\delta^{13}C$ values. Similarly, the stable isotope record in the Lobatse II cave in the semi-arid SRZ of Botswana (Fig. 3), displayed a strong C_4 signature between c. 46 and 43 kyr BP, which Holmgren et al. (1995) proposed was as a result of C_4 grass proliferation in response to a drier climate, rather than a reduction of winter precipitation.

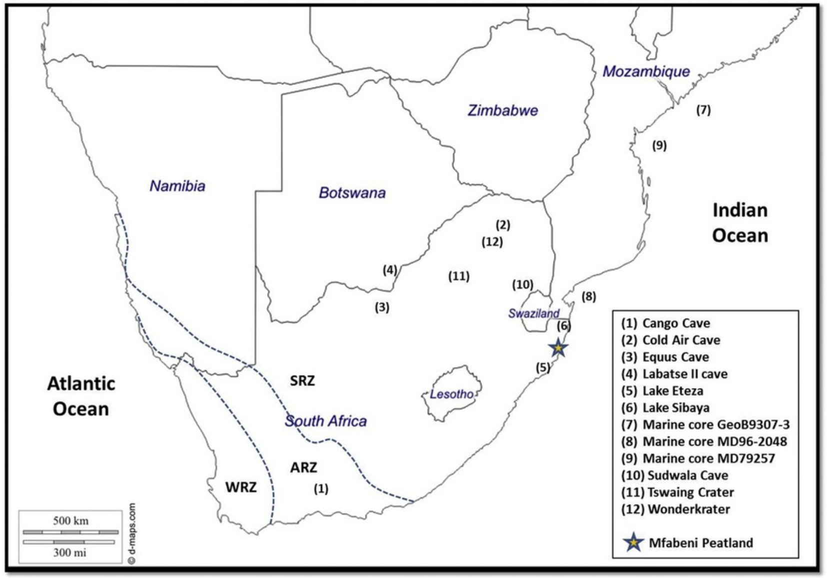

Fig. 3 Regional Site map – 1. Cango Cave (Talma and Vogel, 1992); 2. Cold Air Cave (Holmgren et al., 2003); 3. Equus Cave (Lee-Thorp and Beaumont, 1995); 4. Labatse II Cave (Holmgren et al., 1995); 5. Lake Eteza (Neumann et al., 2010); 6. Lake Sibaya (Walther and Neumann, 2011); 7. Marine core GeoB9307-3 (Scheful3 et al., 2011); 8. Marine core MD96-2048 (Dupont et al., 2011); 9. Marine core MD79257 (Bard et al., 1997); 10. Sudwala Cave (Green et al., 2015); 11. Tswaing Crater (McLean and Scott, 1999; Partridge, 2002); 12. Wonderkrater (Scott et al., 2003; Truc et al., 2013). WRZ = winter rainfall zone; ARZ = all-year rainfall zone; SRZ = summer rainfall zone.

alt-text: Fig. 3

The trend towards higher $\delta^{13}C_{wax}$ values continues until c. 38 kcal yr BP, coeval with slowly decreasing %TOC. Following this, a sharp decrease in $\delta^{13}C_{wax}$ signal occurs more than likely as a result of declining water levels in the peatland and the re-encroachment of the abundant C_3 arboreal forests into the basin (*Podocarpus, Anacardiaceae* and *Celastraceae;* Finch and Hill, 2008). A similar sharp decrease in $\delta^{13}C_{wax}$ values was recorded c. 37 ka in the Sudwala Cave stalagmite, situated in the SRZ transition between the Highveld and Lowveld habitats (Fig. 3) that Green et al. (2015) attributed to the expansion of C_3 forests at the site. Thereafter, the Mfabeni $\delta^{13}C_{wax}$ values steadily increase and stabilise to above average values for the next c. 10 kcal yr BP, with the exception of excursions to below average values at c. 31.2 and 29.5 kcal yr BP (Fig. 2 and Table 1). Both events are accompanied by low %TOC values suggesting a reduced water table with a mixed C_3 arboreal forest and C_4 sedge palynological signal. The fluctuating $\delta^{13}C_{wax}$ signal between c. 31.9 and 28.7 kcal yr BP (accompanied by low, but unstable %TOC signal, sedge dominated peat and a slow increase in sedges at the expense of arboreal forests) could be due to the interplay between encroaching and receding C_3 forests into the sedge dominated peatland in response to fluctuating water levels during this period. After c. 28 kcal yr BP, the %TOC values steadily increase, while the $\delta^{13}C_{wax}$ signal stabilizes at above average values implying a permanently established C_4 sedge fen type environment accompanied by the first appearance of aquatic plant taxa in the palynology record (Finch and Hill, 2008). These parameters, along with a peak in the $\delta^{13}C_{bulk}$ values, advocate sufficiently elevated water table levels in the peatland to support both submerged (Baker et al., 2016) and emergent plant types to the exclusion of arboreal forests up until c. 23 kcal yr BP.

4.1.2 The Last Glacial Maximum and deglacial

A sharp decline in %TOC, decrease in $\delta^{13}C_{wax}$ values and an abrupt change to a grass dominated environment recorded in both the stratigraphic and palynology records occurs after c. 23 kcal yr BP. These trends continue to c. 14.8 kcal yr BP, with the exception of a gradual increase in %TOC values after c. 18.3 kcal yr BP. Both Finch and Hill (2008) and Grundling et al. (2013) interpreted the strong grassland stratigraphic and palynology signal to signify dry conditions in the peatland that facilitated the expansion of grasses to the exclusion of sedges and swamp forests in and around the Mfabeni basin. The sharp excursion in core SL6 $\delta^{13}C_{wax}$ signal to more negative values suggests a large increase in C_3 plant input into the sediments during this same period, inferring a change to more temperate C_3 grasses during the LGM and late glaciation. Consistent with this, Dupont et al. (2011), Rommerskirchen et al. (2006), and Scott (2002) concluded that there is very little evidence for increased C_4 grasses during the LGM in sub-tropical Southern Africa, and that C_3 grasses more than likely dominated the region during this period. This is in direct contrast to records studied in the tropical regions of Africa where there is strong evidence for C_4 plant expansion during the LGM (Castañeda et al., 2009; Sinninghe Damsté et al., 2011). Furthermore, McLean and Scott (1999) undertook a phytolith investigation in the Tswaing Crater, situated on the SRZ escarpment of South Africa (Fig. 3), and observed that *Festucoid* C_3 grasses were dominant during the LGM. An increase of 3.4‰ to higher leaf wax $\delta^{13}C$ values occurs at c. 12.6 kcal yr BP, coeval with increased %TOC and the continued high frequency of grassland taxa. These parameters indicate a switch to predominant C_4 grasses and sedge populations in the peatland, more than likely in response to increasing water levels (and arguably higher temperatures) during the transition from glacial to interglacial conditions.

4.1.3 Holocene

At the start of the Holocene (c. 11.5 kcal yr BP), the $\delta^{13}C_{wax}$ signal once again becomes more negative, continuing on the declining trend until the beginning of the mid-Holocene (c. 6.5 kcal yr BP). This change corresponds with rapid increases in arboreal and a peak in swamp forest taxa (Finch and Hill, 2008), and overall increased %TOC values. Lee-Thorp and Beaumont (1995) similarly observed deviations to higher C_3 plant signatures in tooth enamel samples of grazers found in Equus Cave (Fig. 3), located in the sub-tropical SRZ, during the early Holocene (c. 12–9.5 kyr BP). These parameters infer a switch from overall low water levels of the glacial and deglacial periods to increased ground water inundation and direct precipitation in the Mfabeni; conditions conducive for the intrusion of C_3 swamp forests into the peatland, resulting in lower $\delta^{13}C_{wax}$ signal. Included during this period is an intense, but short excursion to low %TOC values (c. 7.1 kcal yr BP), signifying a dry period (Baker et al., 2014) before the recovery of %TOC values (and water levels) at c. 6.5 kcal yr BP, after which the %TOC signal remains elevated till c. 1.5 kcal yr BP. The ensuing elevated water levels are accompanied by a switch to higher $\delta^{13}C_{wax}$ values and a marked increase in sedges and grasses at the expense of arboreal forests. Scott et al. (2003) also observed higher peat $\delta^{13}C_{bulk}$ values between c. 5 and 3 kyr BP from the Wonderkrater thermal spring located in the sub-tropical SRZ of the Limpopo province (Fig. 3). The Mfabeni stratigraphic study recorded chunky and fibrous peat, remnants of swamp forest OM input during this period according to Grundling et al. (2013). However, peat texture and the palynology record in core SL6, indicates open sedge dominated fen, with reduced input from arboreal and swamp forests. This discrepancy could arguably be linked to dating uncertainties and different age models employed by the different researches. This period of higher $\delta^{13}C_{wax}$ and %TOC values corresponds to a decline in arboreal pollen and a marked increase in sedge and grass taxa, suggesting the increased basin water levels pushed back C_3 forests and allowed C_4 sedges to flourish in the peatland. After c. 2.9 kcal yr BP till present, all three parameters start an overall declining, but fluctuating trend. This represents a general drying trend interspersed with elevated water levels that resulted in the periodic incursion, and subsequent retreat, of peripheral swamp forests into the sedge and grassland dominated peat, as is prevalent in the plant distribution found in and around the peatland today.

4.2 Palaeoclimate reconstruction

The strong and significant correlation between core SL6 $\delta^{13}C_{wax}$ and $\delta^{13}C_{bulk}$ signals surmises that the dominant control on peat $\delta^{13}C$ values are the relative contribution of terrestrial C_3 and C_4 plant types into the Mfabeni sediments. The high resolution C isotope record of terrestrial plant vegetation provides us with an opportunity to correlate the palaeoclimatic record in core SL6 with other climate proxies. To facilitate this investigation, we compared the %TOC, $\delta^{13}C_{wax}$ and $\delta^{13}C_{bulk}$ data with the established northern and southern Hemisphere climatic events, along with interpretations of bulk geochemical (Baker et al., 2014), stratigraphy (Grundling et al., 2013) and palynology (Finch and Hill, 2008) investigations in the Mfabeni deposit (Fig. 2).

The geochemical data suggests several climatic perturbations, either at the onset, during or termination of the H4, LGM, deglaciation and Holocene periods (Fig. 2). The H5/A2 events (c. 44.5 kcal y BP) is marked by the transition from moderate to intermediate water levels and an overall gradual switch from C_3 riparian/swamp forest to C_4 sedge and grass dominated vegetation, coeval with a peak in core %TOC values. This suggests an increase in local precipitation and a switch to seasonal/semi-permanently inundated hydrology in the basin that was conducive for the propagation of water loving sedges to the exclusion of arboreal forests. SST analysed in a proximal Mozambique Channel marine core (MD79257; 20°24′ S; 26°20′ E; Bard et al., 1997, Fig. 3) exhibited a sharp increase during the same A2 warming event that could cause increased continental precipitation over eastern and southern Africa (Dupont et al., 2011; Goddard and Graham, 1999).

At the onset of the H4 event, both bulk and molecular C isotope values decrease by −4.9‰ and −8.5‰, respectively, coeval with a switch back to moderate water levels and high frequencies of C_3 arboreal forest taxa. These parameters suggest a reduction in local precipitation and temperatures, allowing for the intrusion of arboreal forests into the peat basin, supported by a reduction of bacterial activity based on biomarker proxies (*n*-alkanoic acid carbon preference index; CPI_{FA} and *n*-alkanoic acid saturated vs unsaturated concentrations; $sat/unsat_{FA}$) in the Mfabeni (Baker et al., 2016) during the same period. At c. 28.0 kcal yr BP, a sharp recovery in %TOC values occurs, concordant with gradually increasing bulk and molecular $\delta^{13}C$ values and a peak in %TOC at c. 22.7 and c. 24.2 kcal yr BP (H2), respectively. This period is interpreted to represent a shallow lake environment with predominant C_4 sedge and aquatic plant assemblages (biomarker aquatic plant proxy; P_{aq}; Baker et al., 2016). This relatively large increase in local precipitation is corroborated by the Mozambique Channel marine core (MD79257) SST spike (Bard et al., 1997), an increase in ambient temperatures recorded in the ARZ Cango Cave speleothem (Talma and Vogel, 1992, Fig. 3) and increased lake levels of the SRZ Tswaing impact crater lake at around c. 28 kcal yr BP (Partridge, 2002, Fig. 3).

At the peak of LGM, the Mfabeni record exhibits a dramatic shift to dry and cooler conditions (Fig. 2, c. 22.7 kcal yr BP) inferred by the abrupt decline in %TOC values, change to sandy peat and therefore low water levels, and a transition to more dominant C_3 grasslands, which persisted until c. 15.0 kcal yr BP. This interpretation is further supported by several biomarker proxies investigated in the Mfabeni peatland (Baker et al., 2016), where a definite switch from aquatic to terrestrial plant input (P_{aq} and leaf wax proxy; P_{wax}) and a decrease in temperature sensitive microbial decomposition rates (CPI_{FA} and $sat/unsat_{FA}$) occurred after 23.1 kcal yr BP. Consistent with this, additional

Southern African regional climate archives likewise recorded a dry and cool climate during the LGM. Most notably, the Mozambique marine core MD79257, which reported the lowest SST value at c. 20 kcal yr BP (Bard et al., 1997), and a palynology study on marine core MD96-2048 situated ~120 km south of the Limpopo River mouth (Dupont et al., 2011, Fig. 3), suggests lower regional temperatures and moderately less rainfall. Additionally, a SRZ Cold Air Cave stalagmite in the Makapansgat Valley (Fig. 3) exhibited lower temperatures between c. 23 and 21 kyr (Holmgren et al., 2003), and the proximal Wonderkrater fossil pollen sequence (Fig. 3) recorded both 6 ± 2 °C cooler LGM temperatures and c. 50% less precipitation than the Holocene (Truc et al., 2013) during the LGM. These cooler and drier glacial conditions could arguably have been the cause for a shift to C_3 grass dominance in the Mfabeni, which tend to have an advantage over C_4 grasses at lower temperatures (Kotze and O' Conner, 2000; Sage et al., 1999), and appear to have dominated the sub-tropical and WRZ of Southern Africa during the LGM (Dupont et al., 2011; Rommerskirchen et al., 2006; Scott, 2002). These conditions persist with %TOC fluctuating at below average core levels, the $\delta^{13}C_{bulk}$ signal becoming increasingly more negative and dry grassland palynology signature continuing until c. 15 kcal yr BP. Thereafter, even though the %TOC profile steadily increases after the onset of the Antarctic cold reversal (ACR) event, the $\delta^{13}C_{bulk}$ signal remains relatively low until the onset of the early Holocene.

In core SL6, the early Holocene (c. 10.5–7.1 kcal yr BP) is characterised by overall elevated %TOC values compared to the preceding deglacial period (Baker et al., 2014), albeit a brief millennial scale excursion to low %TOC level at c. 7.1 kcal yr BP, and fluctuating but predominant C_3 arboreal and swamp forest molecular and bulk $\delta^{13}C$ signatures. The Mfabeni biomarker investigation (Baker et al., 2016) showed that the early Holocene (c. 10.5–7.1 kcal yr BP) was a period of relatively cool (low microbial activity; CPI_{FA}) and humid (high P_{aq}) conditions, conducive for proliferation of submerged macrophytes in and arboreal swamp forests (Finch and Hill, 2008) around the basin. This interpretation is supported by the Wonderkrater palynology sequence (Truc et al., 2013) and stable isotope speleothem record in the Cold Air Cave (Holmgren et al., 2003).

The Holocene maximum was terminated in the Mfabeni by a short dry event at c. 7.1 kcal yr BP (low %TOC; Baker et al., 2014), corroborated by the proximal Lake Eteza palynology record (Neumann et al., 2010, Fig. 3) and drop in Mozambique channel SST (Bard et al., 1997). After c. 6.5 kcal yr BP, both the $\delta^{13}C_{wax}$ and $\delta^{13}C_{bulk}$ values abruptly increased and remain elevated until c. 2.2 kcal yr BP. These isotopic trends are accompanied by increasing elevated %TOC values, a switch to intermediate water levels and sedge dominated peat (Fig. 2). During this period, the palynological investigation (Finch and Hill, 2008) recorded a marked increase in sedge and grasslands populations, inferring increased ambient temperatures and moisture content supported by the Mfabeni biomarker proxies (Baker et al., 2016), δD and SST records in the adjacent Indian Ocean GeoB937-3 marine core (Schefuß et al., 2011, Fig. 3). Thereafter, an overall decreasing, albeit fluctuating, trend in all three parameters, correspond to a return to moderate water levels in the peatland and oscillating quantities of C_3 peripheral arboreal pollen into the C_4 graminoid dominated peatland in response to fluctuations in the water table. We interpret these trends to signify an overall drying trend with cyclical fluctuations in precipitation, corroborated by similar general drying conditions recorded in proximal Lake Sibaya and Kosi Bay after c. 2 kcal yr BP (Walther and Neumann, 2011, Fig. 3). Notably, the Mfabeni Holocene peat archive agrees with findings of other studies showing that the Holocene was not as stable as previously perceived both regionally (Baker et al., 2014; Bard et al., 1997; Schefuß et al., 2011) and globally (Mayewski et al., 2004).

Our data show that the high-resolution bulk C isotopic trends in core SL6, which has a strong and significant correlation with the molecular leaf wax $\delta^{13}C$ record, reveals perturbations during several of the established climatic events since the late Pleistocene, namely the H4, LGM, deglaciation and Holocene periods. However, some of the shorter, and possibly less intense, climatic events do not exhibit significant shifts in stable C isotope values. Since climatic signals of the Heinrich events were recorded in the northern Atlantic (Hemming, 2004), the lack of a corresponding signal in the Mfabeni archive could be due to the distance between these sites. Nonetheless, we did observe a signal for one of the relatively more extensive Heinrich events (H4), which suggests an alternative reason. We surmise that local hydrological overprinting blurred the climatic $\delta^{13}C_{bulk}$ signal for the more muted events in the Mfabeni peatland, by either attenuating the perturbations or resulting in changes in plant assemblages that did not significantly affect the $^{13}C/^{12}C$ fractionation. Nonetheless, the bulk geochemical (Baker et al., 2014), biomarker (Baker et al., 2016) and leaf wax ^{13}C isotopes signals investigated in the Mfabeni peatland suggest a general antiphase response when compared to the Northern Hemisphere climatic events since the Late Pleistocene, lending support to the North-South interhemispheric seesaw hypothesis.

5 Conclusions

n-Alkane $\delta^{13}C$ record was analysed in the Mfabeni peat core and, along with published bulk geochemical data ($\delta^{13}C_{bulk}$, %TOC) from the same core, compared to palynological data and stratigraphic investigations. These data were then used to reconstruct the late Pleistocene and Holocene environment variability. The $\delta^{13}C_{wax}$ data established large variability in the proportional contributions of C_3 and C_4 terrestrial plants throughout the depositional history of the Mfabeni peatland and correlated strongly with the palynology and stratigraphic records and other regional bulk C isotopic investigations. Due to the strong and significant correlation observed between $\delta^{13}C_{bulk}$ and $\delta^{13}C_{wax}$ data, we conclude that the $\delta^{13}C_{bulk}$ signal represents the proportional contributions of C_3 and C_4 terrestrial plant assemblages, and allows us to link the high-resolution $\delta^{13}C_{bulk}$ with the $\delta^{13}C_{wax}$ data. We found evidence for palaeoenvironmental shifts corresponding to some of the global climatic events since the late Pleistocene (H4, LGM, deglacial period and Holocene). This is consistent with other regional climate records, most notably, the adjacent Indian Ocean SST records. However, due to the rather muted response of $\delta^{13}C_{bulk}$ record in core SL6 with regard to the other shorter climatic events (H5/A2, A1, H3, H2, H1, ACR, and YD), we conclude that the local hydrology of the fen overprinted the signal for climatic events that were less pronounced and/or ephemeral events.

The Mfabeni record clearly exhibits a general antiphase trend with climatic events in the northern Hemisphere. Future work in this area would benefit from higher resolution leaf wax $\delta^{13}C$ analysis to definitively resolve the blurred palaeoenvironmental signals and provide a better understanding of palaeoecological variability in Southern Africa. The Mfabeni peatland has proven to be an important archive of environment and climate variability during the late Pleistocene and Holocene, and a valuable contributor to our understanding of the mechanisms that drove environmental fluctuations in southern Africa during the last glacial and interglacial periods.

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Highlights

- Mfabeni peatland rare c. 47 kcal yr BP old continuous palaeoenvironment archive.
- Bulk C isotopes represents proportional inputs of C3 and C4 terrestrial plants.
- Several climate events since Late Pleistocene chronicled in the peatland archive.
- Shorter climate events signal absent due to local hydrological overprinting.
- General anti-phase trend with Northern Hemisphere climate events.

Queries and Answers

Query: Please note that author’s telephone/fax numbers are not published in Journal articles due to the fact that articles are available online and in print for many years, whereas telephone/fax numbers are changeable and therefore not reliable in the long term.
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