Leaf wax δD inferring variable medieval hydroclimate and early initiation of Little Ice Age (LIA) dryness in southern Mozambique

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Abstract

A sediment sequence from a coastal, hydrologically isolated lake in southern Mozambique was analysed for leaf wax δD, n-alkane abundance (ACL) and bulk organic geochemistry (δ13C, TOC, %N), providing a record of past rainfall variability and savanna dynamics over the last 1500 years. The δDmax rainfall reconstruction reveals a stable hydroclimate between 500–700 CE, while ACL and δ13C together with previous pollen data suggest savanna vegetation was characterized by a relatively dense woody cover. Highly variable hydroclimate conditions are inferred by δDmax between 800–1350 CE, with repeated centennial scale intervals of extreme dry and wet conditions overlapping the Medieval Climate Anomaly (MCA; 950–1250 CE). Savanna tree cover stayed relatively intact over this phase. After ca 1250 CE, a progressive change towards drier conditions was initiated, leading up to maximum aridity during the AD 1700s, a period associated with the Little Ice Age (LIA; 1500–1850 CE). Tree cover was now replaced by a more grass-dominated savanna. The clear antiphase rainfall patterns between Nhaucaiti and equatorial East African proxy records gives support to the notion that Indian Ocean sea surface temperature (SST) gradients act as modulator of southern African climate on a multi-decadal time scale, possibly forced by long-term El Niño Southern Oscillation (ENSO) variability. We suggest that strong ENSO variability and greater occurrence of La Niña events triggered the generally wet and unstable MCA in southern Africa. From around 1250 CE, a shift towards a predominance of El Niño induced drier conditions in south-east Africa during the LIA. Our study of vegetation and hydroclimate proxies in parallel suggests that savanna tree and shrub cover was relatively resilient to the abrupt shifts in hydroclimate over the MCA, but more sensitive to the long-term progressive drying over the LIA.

1. Introduction

The amount and general quality of proxy-based climate reconstructions from Africa are currently too limited to support a regional climate change assessment (Masson-Delmotte et al., 2013). As a consequence, the last IPCC report encourages generation of additional unambiguous and high-resolution proxy-records, especially from southern Africa where available paleo-data are particularly sparse (Nash et al., 2016). The south-eastern part of the African continent is a key region for identifying past shifts in the climate system due to its location at the boundary between the tropics and subtropics and thereby sensitivity to hydroclimate change associated with latitudinal fluctuations in atmospheric circulation systems (Schneider et al., 2014).

Southern African regional rainfall dynamics are also strongly linked to eastern equatorial hydroclimate in an anti-phase fashion. On decadal to centennial time scales, rainfall variability along the east African coast is modulated by large-scale processes such as sea surface temperature (SST) gradients over the Indian Ocean (Saji et al., 1999; Tierney et al., 2013) and land-ocean temperature contrasts along the African east coast (Zhang et al., 2015). Moreover, southeastern Africa is sensitive to teleconnective forcing from the El Niño Southern Oscillation (ENSO), causing extreme droughts during strong El Niño phases and wetter conditions during La Niña phases (Camberlin et al., 2001; Nicholson, 1986). As ENSO activity is likely to change character due to present global warming, effects on hydroclimate in southeastern Africa can be expected. Statistical comparison of ENSO activity and South African
summer temperatures between 1940–2016 CE reveals a warming effect from El Niño events which is particularly strong during the last ca 40 years (Lakhraj-Govender and Grab, 2018). However, climate models disagree in whether global warming will amplify or mute the ENSO variability (Guilyardi et al., 2009; Collins et al., 2010; An and Choi, 2015). In this context, the study of paleo-climate over the late Holocene (last 2000 years) is of central importance due to its close analogue with present day boundary conditions, allowing assessment of potential future impacts on hydroclimate and ENSO variability from expected global warming. Moreover, the impact of past climatic shifts on ecosystem response may inform on how modern vegetation analogues will respond to future climate change.

Here we present a water-isotope based hydroclimate record covering the last 1500 years from Lake Nhauacati, a hydrologically isolated coastal lake in southern Mozambique. The lake was previously investigated for pollen and diatoms (Ekblom, 2008; Ekblom and Stabell, 2008), providing a coarse-scale paleo-record which is utilized to frame the new data in a paleo-ecological context, and to deduce potential vegetation biases on the isotopic record.

Water isotopes (typically expressed on a delta scale as δ¹⁸O, δ³⁴S or δD) are excellent recorders of past hydrological conditions, as their composition is known to fluctuate in accordance with climate-related factors (Dansgaard, 1964). The use of the hydrogen isotope composition (δD) of sedimentary organic biomarkers has opened up for a wider geographical coverage of water isotope-based records worldwide (Sachse et al., 2012). Long-chain n-alkanes derived from terrestrial leaf waxes are particularly suitable for paleo-studies as they stay well preserved over geological time scales, yet leaving their isotopic signature essentially intact (Peters et al., 2005). Furthermore, n-alkane chain length distributions are related to plant functional types, as algae and aquatic plants generally form short (n-C₁₇₋₂₁) and mid- (n-C₂₃₋₂₅) chain lengths, while terrestrial plants produce longer carbon chains (n-C₂₇₋₃₅) (Ficken et al., 2000; Aichner et al., 2010; Bush and McInerney, 2013). Although the leaf wax n-alkane hydrogen isotope composition (δDₜₐₜ) is affected by a fractionation during biosynthesis (e.g. Ficken and Sessions, 2010), its temporal variability has proven to reflect the isotopic changes in plant source water, e.g. of the lake water or precipitation (δDₚₑᵣᵢᵣᵢₑ) (e.g. Sachse et al., 2004; Aichner et al., 2010; Sachse et al., 2012; Tipple et al., 2013).

Based on the tropical and coastal setting, Lake Nhauacati leaf-wax δD can primarily be interpreted as a proxy for rainfall amount (Rozanski et al., 1993), although minor secondary fractionation effects from evapotranspiration may be expected. Furthermore, we use the n-alkane abundance and chain length data to confirm, highlight and lift new aspects of previous reports on regional savanna dynamics (Ekblom et al., 2014a; Norström et al., 2018a). We constrain the findings by considering pollen and diatom analysis performed on the same sediment sequence (Ekblom, 2008; Ekblom and Stabell, 2008). The new paleo-hydrological and vegetation records are compared with previous regional paleo-assessments, and tentative drivers of past hydro-climate variability and vegetation dynamics over the late Holocene period are discussed.

2. Regional setting and local site characteristics

Lake Nhauacati is located in southeastern Africa, ca. 5 km south of the town Vilanculos, along the coastal plain of southern Mozambique (22° 02'16.01"S, 35°18'43.48"E, Fig. 1). The lake is situated ca. 2 km from the coast, at an altitude of 20 m a.s.l., covering an area of ca. 40 ha, with an east-west diameter of ca. 750 m. and a north-south diameter of ca. 500 m. The lake is one of many smaller lakes in the Vilanculos region. It has no surface inlets or outlets, and lake levels are instead regulated by direct rainfall and recharge of ground water through shallow aquifers (Coetsee and Hartley, 2001), probably modulated by shifts in evaporation.

Climate is classified as semi-arid as annual potential evapotranspiration (1440 mm) exceeds annual precipitation amounts (830 mm) (FAO, 1984). Air temperatures vary between ca. 27 °C during summer and ca. 19 °C during winter (FAO, 1984). Precipitation in the area is strictly seasonal with majority of rains falling between November and March, when the Intertropical Convergence Zone (ITCZ) is in its southernmost position and easterly trade winds bring moisture over the continent from the Indian Ocean (Tyson and Preston-Whyte, 2000). Although tropical cyclones are the major rainfall source in this coastal area, the tropical temperate troughs (TTTs) constitute another important rainfall contributor, manifested as NW-se stretching cloud bands, connecting the tropical and mid-latitude climate systems, producing convective and frontal summer rains in southern Africa (Harrison, 1984; Hart et al., 2010).

The coastal stretch of southern Mozambique consist of ancient sand dunes of Pleistocene and Holocene age, forming an undulating landscape with ridges stretching in a SE-NW direction (Spalviero et al., 2014). Pleistocene dunes predominate in the area surrounding the lake, covering the Tertiary Jofane Formation limestone (Salman and Abdula, 1995). The thickness of the dune deposit is spatially variable, but generally thicker shoreward and less thick towards inland areas. The dunes consist of semi-consolidated red sand in the lower part and un-consolidated gray sand in the upper part, and are classified as internal dunes (DNG, 2006), with parabolic shapes formed under the dominating easterly wind pattern.

Vegetation in southern coastal Mozambique is characterized by coastal mosaics, forming a mixed floral composition, including Miombo savanna and woodlands with low density of small trees and shrubs combined with a grassy understorey (Wild and Fernandes, 1968). The open savanna vegetation in the Chibune area and around Lake Nhauacati is dominated by herbs and grasses such as Ergrotris ciliariis, Panicum maximum, Cyperus spp., Cerchirus incertus and Digitaria spp., together with small Miombo tree components, mainly represented by Julbernarda globiflora and Strychonis spinosa (Ekblom and Stabell, 2008). Large areas of the land surrounding the lake are used for cultivation. The floral composition changes towards the sea, with a higher proportion of littoral shrubs such as Commiphora zanzibarica, Phyllanthus reticulatus Turraea nilotica, Grewia monticola, Deinbollia oblongifolia, Clerodendrum glabrum and Acalypha glabrata, as well as the climber Cocculus hirta (Ekblom, 2004).

3. Methods

3.1. Coring, subsampling and sample preparation

A 230 cm long sediment sequence was sampled with a Russian corer in the deepest part of the lake. The lithology was homogeneous throughout the sequence, dominated by gyttja of dark brown colour, but with increasing input of minerogenic components in the lowermost section, below 215 cm. The core was subsampled every 1 cm and stored in a freezer prior to chemical treatment. Details on coring methodology and lake bathymetry are available in Ekblom and Stabell (2008) and Ekblom (2008).

Ca. 2–6 g of sediment from 93 levels of the Nhauacati core were freeze-dried and homogenized. A small amount from each sample (5–10 mg) was weighed for bulk isotope measurements. The rest of each sample was used for lipid extraction. Total lipid extracts obtained via sonication of the dried sediments were fractionated into fractions containing only saturated hydrocarbons – dominated by the long-chain n-alkanes – using silica-gel based solid phase extraction. This procedure is described in detail elsewhere (Norström et al., 2017).

3.2. Bulk organic geochemistry and carbon isotope measurements

Bulk plant δ¹³C was measured using a Carlo Erba NC2500 analyser, connected via a split interface to a Finnigan MAT Delta V mass spectrometer, at the Stable Isotope Laboratory, Stockholm University.
Reproducibility was calculated to be higher than 0.15‰ for $\delta^{13}$C and $\delta^{15}$N. Total organic carbon and nitrogen content (TOC, %N) were also determined when measuring the isotope ratios. The relative error was < 1% for both these measurements. $^{13}$C content is reported as $\delta^{13}$C against VPDB and $\delta^{15}$N against standard air.

### 3.3. n-Alkane quantification and molecular hydrogen isotope measurements

For the n-alkane quantification, hydrocarbon fractions were analysed on a GCMS-QP2010 Ultra system (Shimadzu), equipped with an AOC-20i auto sampler and a split-splitless injector operated in splitless mode. For separation, a GC column with the following dimensions was used: 30 m × 0.25 mm × 0.25 μm (Zebron ZB-5HT Inferno). The GC oven was programmed for a temperature increase from 60 °C to 180 °C by 20 °C/min, then up to 320 °C by 4 °C/min, followed by a hold time of 20 mins. Quantification was performed using an external calibration made using a solution of C21–C40 n-alkanes (Fluka Analytical), which was run several times daily at different concentrations. The software MS Solution 4.11 (Shimadzu) was used for peak identification and area integration. The n-alkane average chain length (ACL) was calculated as: $\Sigma (n \times C_n) / \Sigma C_n$ where $C_n$ is the concentration (mg/g dry weight) of the odd chain n-alkane with n number of carbon atoms ranging from 21 to 35 for ACLall and ranging from 27 to 35 for ACLterr.

The hydrogen isotopic composition (reported as $\delta^D$ versus VSMOW) of the n-alkanes was determined by gas chromatography-isotope ratio mass spectrometry (GC-IRMS) system (Thermo-Finnigan) equipped with a ConFlo IV Universal interface for continuous flow and GC-isolink for combustion and pyrolysis. A DB-5 column was used, provided by Agilent Technologies (50 m × 0.250 mm × 0.25 μm). Injections were performed with inlet operating in a splitless mode during 1.5 min under pressure of 50 p.s.i. Pressure was then released and the split flow of 100 ml/min was reduced to 40 ml/min after 5 mins (gas saver mode). The GC oven was programmed to increase at a rate of 15 °C/min between 100 and 250 °C, and then by a rate of 10 °C/min until reaching 320 °C. This temperature was kept for 9 mins. Alkane isotopic compositions were determined using H2 reference gas pulses at the beginning and end of each run, which in its turn was calibrated against an n-alkane isotopic standard (A6 standard mixture provided by Arndt Schimmelmann, Indiana University, USA). The A6 standard was measured several times daily, reporting a precision of 1.6‰ for H2. The fossil samples were measured in duplicate or triple, yielding a precision of 3.7‰ for average $\deltaD_{\text{ave}}$ and more specifically: 4.8‰ for $n$-C23, 2.3‰ for $n$-C25, 3.1‰ for $n$-C27, 3.8‰ for $n$-C29 and 4.3‰ for $n$-C31. The $\deltaD$ measurements of $n$-C21, $n$-C23, $n$-C25 and $n$-C31 resulted in unsatisfying reproducibility (> 5‰) and are therefore not reported. Throughout the analysed sequence, a stable inter-sample concentration was aimed for in order to avoid biases by peak intensity variations between runs. However, sometimes limited n-alkane abundance

Fig. 1. Orientation maps. (a) South-eastern Africa with annual precipitation amounts (averages 1960–1990 CE, from WorldClim.2), and (b) topographic map of the Chibuene area, indicating the setting of Lake Nhaucati. Sites mentioned in the text are indicated as follows: 1. Lake Nhaucati (this study; Ekblom, 2008; Ekblom and Stabell, 2008; Ekblom et al., 2014a) 2. Lake Nhauhache (Holmgren et al., 2012); 3. Lake Chilau (Norström et al., 2018a); 4. Limpopo floodplain (Stoe et al., 2015; SiTo, 2017); 5. Pafuri/Mapungubwe trees (Woodborne et al., 2016); 6. Makapangat cave (Holmgren et al., 1999, 2003; Sundqvist et al., 2013); 7. Lydenburg fen (Sjöström et al., 2017); 8. Lake Sibaya (Stager et al., 2013); 9. Braamhoek wetland (Norström et al., 2009, 2014); 10. Lower Limpopo Valley (Ekblom et al., 2012); 11. Matumi trees (Norström et al., 2005, 2008); 12. Lake Malawi (Brown and Johnson, 2005; Johnson and McCave, 2008); 13. Mafadi wetland (Fitchett et al., 2016); 14. Ladybird wetland (Norström et al., 2018b); 15. Lake Etzea (Neumann et al., 2010; Scott et al., 2012) and Lake St. Lucia (Humphries et al., 2016); 16. Mfabeni mire (Baker et al., 2016, 2017; Humphries et al., 2017). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
resulted in higher standard deviations, particularly for the homologues mentioned above. The H3+-factor was controlled on daily basis securing a stable value throughout the sequence.

4. Radiocarbon dating and age-model construction

AMS radiocarbon measurements were performed on 13 bulk samples and one sample containing charcoal macrofossils (Table 1, after Ekblom and Stabell, 2008). Analysis was performed at Poznan Radiocarbon Laboratory, Poland, and Uppsala Ångström Laboratory, Sweden. Similar ages of analysed bulk and charcoal macrofossils at the basal section of the core (228–230 cm) ensure absence of reservoir effects. The previous calibration procedure and age model construction performed by Ekblom (2008) was updated using the most recent southern hemisphere calibration curve SHCal13 (Hogg et al., 2013), followed by a Bayesian approach through the Bacon 2.2 software for age-depth modelling (Fig. 2) (Blaauw and Christen, 2011). The updated age model presented here supports the previous model. Sedimentation rates were calculated based on the modelled calibrated date of each cm (Fig. 2), showing slightly slower sedimentation rates in the lower and upper parts of the sequence (ca. 0.1 cm/yr and 0.2 cm/yr respectively), while maximum sedimentation rates are observed in the middle section (average ca. 0.35 cm/yr), i.e. between 1000 CE and 1240. Due to the particularly slow sedimentation rates after ca 1750 CE, the possibility of a hiatus in sedimentation cannot be excluded. Data covering the last 2 centuries should therefore be interpreted with caution.

5. Results

5.1. Bulk organic geochemistry

Total organic carbon (TOC) shifts from 0 to 10% in the oldest part of the sequence (500–700 CE), towards peak values above 30% at ca. 800–900 CE (Fig. 3). After that TOC declines towards relatively stable values between 15 and 20% from ca 1200 CE until modern times. Nitrogen content (N) follows the TOC throughout the record, with values fluctuating between 0 and 3%. The C/N ratio varies between ca. 15 and 25, with highest values around 700 CE, and 1100–1200 CE. In the bottom section, i.e. prior to 650 CE, C/N results are not considered due to unreliable ratio from low concentrations.

The carbon isotope composition on the bulk organic matter (δ13C) is relatively stable around −25% before 1150 CE, followed by an abrupt increase to ca. −18% around 1100 CE. Minor excursions towards more negative values were evident at ca. 1200 CE and after 1800 CE. Due to low sample resolution from 1750 CE it is not possible to assess whether the latter excursions was a consistent feature for this period, or not.

5.2. Relative n-alkane abundance and average chain length (ACL)

The n-C21 and n-C23-alkanes, which are the ones most strongly linked to aquatic biomes (Ficken et al., 2000; Aichner et al., 2010), show the highest relative abundance before ca. 1150 CE (Fig. 4). Consequently, lowest average chain lengths (ACL) are observed in this period. After 1150 CE the aquatic and semi-aquatic components (n-C21 and n-C23) are less prominent but show an occasional increase again around 1250–1300 CE, paralleled by a decline in ACL. From ca. 1300 CE the assemblage is dominated by n-alkanes associated with terrestrial vegetation (n-C27–35), resulting in a drastic increase in ACL, which is finally stabilizing at relatively high values around 28.5. While some terrestrial homologues are relatively stable in their abundance throughout the sequence (e.g. n-C29), others show more temporal variability (e.g. n-C31 and n-C35). ACLterr (based on n-C27–35) was calculated in order to visualize chain length-related shifts within the terrestrial biome only, without interference from the aquatic and semi-aquatic homologues. ACLterr varies between ca. 29.2 and 30.5.

5.3. Hydrogen isotopic composition (δD)

The δD values vary between −198 and −140‰ for n-C23 (average −170‰), between −131 and −173‰ for n-C25 (average −155‰), between −126 and −168 for n-C27 (average −150‰), between −121 and −200‰ for n-C29 (average −152‰) and between −122 and −210‰ for n-C31 (average −160‰) (Fig. 5). Thus, the average δD-values are relatively uniform across the n-alkane spectrum, while δD variability (ΔδD) is greater for the longest n-alkanes (78% for n-C29, 88% for n-C31) compared to the rest (41% for n-C27 and n-C25, 58% for n-C33). The long-term temporal changes in δD generally reveal co-variability between all homologues, with values being generally lower in the older part of the record, followed by a continuous increase towards the younger part of the sequence. Although high-frequency variability is observed for different homologues, some trends are clearly visible over the entire spectrum of n-alkanes: low values between 800–900 CE, high values between 950–1000 CE, occasionally high values around 1150 CE, and a steady increase from ca 1400 CE until around 1700 CE.

6. Discussion

6.1. ACL and δ13C as vegetation proxies – Interpretative rationale

C3 and C4 plants use different photosynthetic pathways, resulting in higher δ13C values in C4 plants compared to C3 plants (e.g. Ehleringer et al., 1997; Vogel et al., 1978). In southern Mozambique, grasses are
Fig. 2. Lake Nhaucati age-depth model and sedimentation rates. (a) The age-depth model based on Bayesian approach in Bacon software (Blaauw and Christen, 2011). (b) Sedimentation rates calculated on centimeter-basis, based on the calibrated age of each cm-section of the core.

Fig. 3. Lake Nhaucati bulk organic geochemistry. (a) Carbon and nitrogen content (TOC, N) expressed in %, (b) C/N-ratio, (c) carbon isotope composition ($\delta^{13}C$) of the organic matter, expressed as ‰ versus Vienna Pee Dee Belemnite (VPDB).
According to fossil pollen data, sedges were relatively abundant around Lake Nhaucati during this phase compared to later phases. This is accordance with the pollen record that shows high abundance of pollen from the aquatic plant *Nymphaea* over the same period (Ekblom, 2008).

The chain-length response within the terrestrial vegetation i.e. the increase in ACL*ter* over time, combined with the shift towards higher δ13C values from around 1150 CE, suggests a vegetation shift with gradually decreasing arboreal components, probably concurring with more arid conditions. Although the major driving mechanisms of ACL*ter* and δ13C are difficult to disentangle, both proxy responses in this warm and semi-dry environment are probably associated with aridity: either as an indirect response (biochemical) or as a direct response (vegetation type), or a combination of the two. It should be stressed however, that both these parameters may be biased by potential species dynamics within the Cyperaceae family (sedges). The Cyperaceae family includes a wide range of species, with varying potential species dynamics within the Cyperaceae species distribution at the site is relatively homogenous, dominated by *Cyperus* sp. (Ekblom and Stabell, 2008). This homogeneity, together with independent and supporting proxy indications (e.g. the Arboreal/Poaceae pollen ratio, Fig. 6), act in favour to our interpretation of δ13C and ACL as aridity and/or land cover proxies, although biases linked to past sedge species dynamics cannot be completely ruled out.

Although C₄ grasses are generally less productive in terms of n-alkanes compared to C₃ trees, shrubs and herbs (Garci et al., 2014), it is clear that amongst African plants, woody angiosperms contain proportionally more of the shorter long-chain alkanes (n-C25–29) while C₄ Grasses and arboreal vegetation. The ACL may however also be related to other factors, such as temperature (Tipple and Pagani, 2013; Bush and McInerney, 2015) and phylloclony controlled by long-term evolutionary processes (Diefendorf et al., 2015). Plants in warm and evaporative environments also show higher ACL values related to higher drought adaptive capacity (Norström et al., 2017), water stress (Pierantozzi et al., 2013) and C₄ preference (Carr et al., 2010). Similarly, there is a large body of evidence suggesting that fractionation against 13C is generally reduced under moisture stress (Cernusak et al., 2013; Diefendorf and Freimuth, 2017). Thus, relatively high ACL* and bulk δ13C-values in the Nhaucati record may also be interpreted as an indicator for aridity.

While ACL*ter* represents changes in the terrestrial biome, the ACL*all* reflects the whole n-alkane spectrum. Here ACL*all* is mainly a result of the relative abundance of the n-C23 alkane, suggesting a relatively large proportion of (semi-)aquatic plants between 600 CE and 1150, indicating higher water levels and persistently wetter conditions during this phase compared to later phases. This is accordance with the pollen record that shows high abundance of pollen from the aquatic plant *Nymphaea* over the same period (Ekblom, 2008).
Fig. 5. Hydrogen isotope composition (δD) in Nhaucati n-alkanes. From top: δD for n-C_{23}, n-C_{25}, n-C_{27}, n-C_{29} and n-C_{31}, expressed as ‰ against Vienna Standard Mean Ocean Water (VSMOW).
6.2. Synthesis of vegetation proxies

The Lake Nhaucati record contains a clear shift in δ¹³C and ACLterr between the ca. 600–1000 CE period and the 1300–2000 CE period, with a transitional phase of high variability (1000–1300 CE) (Fig. 6). This may indicate a shift from a relatively high input from trees and shrubs prior to 1000 CE (pure C₃ signal, average δ¹³C ca. −25‰), towards a proportionally stronger influx of grasses after ca 1300 CE (C₄ signal, average δ¹³C ca. −18‰). Alternatively or additionally, these trends are secondary responses to a shift towards increased dryness. Although sample resolution is low over the last two centuries, the declining δ¹³C and ACLterr from around 1800 CE suggest increasingly higher arboreal input and/or higher moisture access. A comparison with the ratio between arboreal and grass pollen (AP/Poaceae) (after Ekblom, 2008), show that the geochemical data is generally in line with the pollen record (Fig. 6). The compilation and correlation between grass-tree cover inferred by the various proxies provide a detailed understanding of past vegetation shifts. The strongest arboreal signals can be identified between 500–1225 CE and between 1750–2000 CE, with an exception around 800 CE when the proxies give opposing signals. A stronger grass signal can be observed between ca. 1250–1750 CE, with tentative maximum grass cover around 1275 CE, 1425 and 1550 (Fig. 6). In line with this, a phytolith study from the Inhambane area reports an expansion of drought-adapted grasses (Chloridoideae) from ca 1300 CE (Norström et al., 2018a). Pollen analysis from nearby Lake Nhauhache suggests a general decline in trees/shrubs in favour of grasses from 1000 CE onwards most prominent after 1200 CE. A return to more arboreal cover can be seen in the area between 1700–1900 CE.
6.4. Regional hydroclimate synthesis

6.4.1. 500 CE to 1350 CE, including the Medieval climate anomaly (MCA)

The presented hydroclimate reconstruction from Lake Nhaucati $\delta$Dc29 is generally in accordance with other reconstructions from southeastern Africa (Fig. 7). The drastic shifts in proxies from 800 CE and onwards mark the regional onset of the Medieval Climate Anomaly (MCA, 750–1350 CE) (Nash et al., 2016). Although the Nhaucati record suggest an overall trend of high water availability during the MCA, this period was also characterized by repeated, centennial-long intervals of both extremely high and very low precipitation amounts. Comparison with previous diatom analysis (Ekblom and Stabell, 2008) indicates co-variability with $\delta$Dc29 for the last millennium, where high planktonic abundance generally coincides with low $\delta$Dc29 as a response to wetter conditions (Fig. 7). A diatom-based hydroclimatic record from nearby Lake Nhuahache also indicates strong variability between ca. 800 CE and 1150 (Holmgren et al., 2012). Moreover, the extreme wet spells at 800 CE and 1000 CE in Lake Nhaucati, are consistent with abundant planktonic diatom taxa at Lake Sibaya in coastal northern South Africa, suggesting high lake levels (Stager et al., 2013). The Nhaucati data is also consistent with the sharp shift in tree ring $^{18}$O values caused by higher moisture access in Pafuri/Mapungubwe baobab trees from northern South Africa and Zimbabwe (Woodborne et al., 2016) (Fig. 7). In the lower Limpopo valley, water levels became higher from 800–1400 CE (Ekblom et al., 2012). The dry phase suggested in Lake Nhaucati around 850–950 CE concurs with a severe desiccation event of Lake St. Lucia in northeastern coastal South Africa (Humphries et al., 2016). In the northern interior of South Africa, speleothem $^{18}$O at Makapansgat (Holmgren et al., 1999, 2003) and siliceous microfossil assemblages at Lydenburg fen (Sjöstöm et al., 2017) similarly indicate variable MCA conditions. Climate variability over this phase is also inferred from biogenic and geochemical proxies in Lesotho mires (Fitchett et al., 2016; Norström et al., 2018b). At the coastal Mbaseni mire, South Africa, aeolian deposition increased around 2000 years ago, peaking at 350–1150 CE, parallel to increases in grain size and Ti concentration, reflecting a dry and possibly variable climate (Humphries et al., 2017). The increase in Mbaseni n-alkane chain lengths, together with a decline in leaf wax $^{18}$O values, is also indicative of a shift towards drier conditions around the first millennium AD (Baker et al., 2016, 2017). Pollen and magnetic susceptibility at nearby Braamhoek wetland infer a trend towards wetter conditions from ca 500 CE, but with high proxy variability, particularly during the last millennium (Norström et al., 2009, 2014). When compared to the coastal, easterly sites, the interior sites generally report less pronounced variability. Further, the Nhaucati long-term temporal trends are contemporary with variability observed in records from equatorial Africa west of 35°E (e.g. Lake Edward (Russell and Johnson, 2007), but out of phase with records from more easterly equatorial sites (Tierney et al., 2013) (Fig. 7).

6.4.2. 1350 CE to 1850 CE, including the Little Ice Age (LIA)

The trend towards increased dryness from 1350 CE at Nhaucati marks the onset of the southern hemisphere counterpart of the Little Ice Age (1350–1850 CE, LIA) (Moberg et al., 2005; Nash et al., 2016). During this phase, the Nhaucati $\delta$Dc29 co-varies with subtropical southeastern records and tropical hydroclimate proxy records west of 35°E, while being in anti-phase with equatorial records located closer to the coast (Fig. 7) (i.e. Tierney et al., 2013; Nash et al., 2016). The records co-varying with Nhaucati are primarily represented by Lake Malawi terrigenous input (Brown and Johnson, 2005; Johnson and McCave, 2008), Lake Edward Mg/Ca ratio (Russell and Johnson, 2007) and pollen data from coastal Lake Eteza (Neumann et al., 2010; Scott et al., 2012). Common for these sites are an early onset of LIA dryness, and that maximum aridity was reached between 1600–1800 CE (Fig. 7). At Limpopo flood plain there was a pause in major flooding events at the initiation of LIA (Sitoe et al., 2015; Sitoe, 2017). Considering the decline in accumulation rate at Nhaucati, the shift towards drier...
conditions appears to have started already around 1300 CE, similar to Lake Malawi (Brown and Johnson, 2005; Johnson and McCave, 2008). This stands in contrast to hydroclimate indications from the easternmost equatorial region (see synthesized MCEOF proxy in Fig. 7; Tierney et al., 2013), where the MCA was variable but generally dry, followed by a wetter LIA (Nash et al., 2016; Tierney et al., 2013, and...
records are sparse, speleothem δ13C generally appears later than at Nhaucati. Makapansgat speleothem after 1600 CE (Woodborne et al., 2016). Although temperature proxy records are sparse, speleothem δ18O suggest lowest late Holocene temperatures between 1690 CE and 1740 in northern South Africa (Sundqvist et al., 2013).

6.5. Potential climate drivers

Considering the Nhaucati 8ΔC29 record in relation to these wider regional trends, supports previous assumptions of the MCA-LIA transition being a result of latitudinal shift of the atmospheric circulation systems, towards a more northerly position of the ITCZ (Woodborne et al., 2016) and the southern hemisphere westerlies (SHW) (Weldaeab et al., 2013; Abram et al., 2014). Such a mechanism would result in less ITCZ-related rainfall along the south-eastern African coast, while relatively more rainfall would fall along the eastern equatorial coast, as well as in south-western Africa where more rainfall can be associated with the mid-latitude westerlies. Under a more northerly located convergence zone during the LIA, the TTT-related cloud bands would also stretch further north, affecting particularly the more inland sites along a NW-SE transect. Frontal and convective rains could then have been a more important contributor to precipitation at these sites over LIA, as opposed to cyclonic rains. This could explain the delayed indications of dryness at inland sites compared to coastal at the LIA initiation.

The inverse hydroclimate pattern between Nhaucati and the eastern equatorial African sites is striking, particularly when considering proxy indications from Lake Naivasha, Victoria and Challa over the LIA period (MCEOF proxy, Fig. 7). These reverse conditions resemble the present day rainfall response to ENSO events, creating a dipole between east and south-east African hydroclimates on inter-annual time scales. Present-day rainfall in southern Africa is associated with ENSO via ocean-atmospheric teleconnections, where El Niño (La Niña) phases are generally associated with drier (wetter) conditions in south-eastern Africa, while the opposite rainfall signals will appear in equatorial east Africa (Nicholson, 1986; Camberlin et al., 2001). The inverse ENSO rainfall pattern in eastern and southern Africa are both modulated by SSTs but by different dynamics. In coastal equatorial east Africa rainfall is affected as the thermocline heaves westwards at the onset of an El Niño event, creating warmer conditions and increased convection. On the south-east African coast this effect is overruled by the temperature gradient between the central and SW Indian Ocean. The warmer conditions in the central Indian Ocean as opposed to cooler conditions in SW Indian Ocean instead suppress rainfall over south-eastern Africa during an ENSO event (Reason et al., 2000; Jury et al., 2002).

Similarly, but on multi-decadal to centennial time scales, Tierney et al. (2013) found a strong coupling between equatorial east-African rainfall variability and internal Indian Ocean SSTs, i.e. an Indian Ocean Dipole (IOD) forcing on the equatorial region (Saji et al., 1999). Our results from Nhaucati suggest that a similar long-term mechanism may act on south-eastern Africa, in analogue with the inter-annual El Niño-driven ocean-atmospheric mechanisms (Jury et al., 2002). Applying such rationale, the rainfall in south-east Africa would have been suppressed by the relatively stronger warming of the central Indian Ocean compared to the SW Indian Ocean under generally cooler LIA conditions. In contrast, the MCA warming led to higher moisture supply and stronger summer rainfalls.

Climate reconstructions from the tropical eastern Pacific lend support to a temporal link between Nhaucati hydroclimate and ENSO variability. The Bainbridge Crater lake record (Galapagos) shows high gray scale variability between ca. 450–1250 CE (Fig. 7) (Thompson et al., 2017); the Laguna Pallcacocha record from the Ecuadorian Andes (Fig. 7) shows highly variable red intensity between ca. 600–1350 CE (Moy et al., 2002) and Lake El Junco (Galapagos) contains a higher sand content (Conroy et al., 2008), all inferring high frequency of ENSO events over the MCA. Moreover, peaks in carbonate laminations at the Bainbridge record reveal a dominance of La Niña events over this period (Fig. 7) (Thompson et al., 2017). The greater moisture availability and strong hydro-climatic variability at Nhaucati could be driven by teleconnective impacts from the ENSO characteristics during the MCA. During the LIA on the other hand, the frequency of La Niña declines in favour of more frequent El Niño events, as inferred by an increase in Lake Bainbridge siliclastic laminae after 1250 CE (Thompson et al., 2017). Although the overall frequency of ENSO events declined over the LIA (Thompson et al., 2017), their amplitude was higher than during the previous MCA (Li et al., 2011, and references therein). Tentatively, this forced south-eastern Africa into a long-term El Niño-dominated state, with drier conditions prevailing over the LIA.

The delayed inland dryness compared to coastal sites at the initiation of LIA, could be explained by the TTTs providing the interior with rainfall during a transitional period until the LIA conditions finally stabilized. Similarly, the less variable hydroclimate conditions at inland sites over the MCA could be linked to TTTs acting as a mutating agent, providing more continuous rainfall over the interior, as opposed to the dramatic rainfall shifts associated with the Indian Ocean tropical cyclones along the coast.

6.6. Vegetation response to climate change

Parallel vegetation and hydroclimate proxy data provides the possibility to deduce the sensitivity of the savanna system to climate change, particularly when knowledge on human forest management in the area is available (Ekblom, 2004; Ekblom et al., 2014b). The composite vegetation proxies from Nhaucati (δ13C, ACL, pollen) indicate a relatively stable woody savanna cover during the second half of the first millennium AD, and through the main MCA phase until ca 1000 CE, suggesting strong ecosystem resilience to abrupt shifts between drought and wetness reconstructed during this phase. No direct signs of forest management or man-induced fires were found in the area during this early phase (Ekblom et al., 2014b) suggesting climate was the major factor governing changes between woody and non-woody (grass) vegetation.

The overall trend towards drier conditions indicated by higher Nhaucati 8ΔC29 values, from 800 CE onwards, most likely led up to an accumulated moisture deficit. This, together with increased farming activities, including forest clearing and fires from around 1000 CE (Ekblom et al., 2014a, 2014b), probably triggered the decline in Woody cover. This trend was amplified by the progressive drying trend over the LIA, and between 1250–1750 CE savanna vegetation was characterized by a more sparse woody cover than before, occasionally punctuated by maximum grass contribution at 1425 CE and 1550 CE (Fig. 6). Woody cover then recovers after ca 1750 CE. This suggests that, although being relatively resilient to extreme climate variability during the MCA, savanna woodland was sensitive to the prolonged drought over LIA, but also seems to possess capacities to recover when conditions return to more favorable, as over the last two centuries.

7. Conclusions

We present a 1500 year long rainfall reconstruction, based on leaf wax δD (δDwax) from a sedimentary sequence from Lake Nhaucati, southern Mozambique. Vegetation proxies studied in the same core provide parallel indications of savanna vegetation dynamics. The δDwax reflects a stable hydroclimate between 500–700 CE, followed by a generally wet, but highly variable hydroclimate between 800–1350 CE, with repeated centennial-long intervals of dry and wet conditions over

the MCA period. After ca 1250 CE, a progressive change towards drier conditions led up to maximum aridity during the AD 1700s. The relatively early initiation of LIA dryness (from ca. 1250 CE) stands in contrast to later LIA initiation in the interior of South Africa (from ca. 1500–1600 CE), and the wet LIA conditions reported from the easternmost equatorial region (Tierney et al., 2013; Nash et al., 2016). After ca 1750 CE, low sample resolution hampers data interpretation from the Lake Nhaucati record, but tentatively, δDwater suggests increased wetness up until modern times.

The Nhaucati hydroclimate reconstruction, placed in relation to the regional suite of proxy data, suggests a potential northerly shift of the ITCZ and SHW circulation systems at the MCA/LIA transition. Furthermore, the antiphase between Nhaucati and coastal equatorial hydroclimate over the last millennium highlights the significance of Indian Ocean SST gradients for south-eastern African climate on multi-decadal to centennial time scales, and suggests long-term ENSO variability as a potential driver. The wet and unstable hydroclimate in south-east Africa during the MCA was possibly coupled to high ENSO variability and more frequent La Niña events in the eastern Pacific, which forced southern African climate into a long-term El Niño-dominated state over the LIA period.

The ACL and δ13C data, together with previous pollen data (Ekblom, 2008), suggest savanna vegetation was characterised by a relatively dense woody cover in the early part of the record (MCA), which successively transformed into a more grass-dominated savanna (LIA). When synthesising the vegetation and hydro-climate proxies studied in parallel, it can be concluded that the arbooreal component of the savanna ecotone shows adaptive capacities to extreme climate shifts, but is sensitive to long-term, progressive aridity.

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