Glacial to interglacial climate variability in the southeastern African subtropics (25- 20°S)

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Abstract

We present a continuous and well-resolved record of climatic variability for the past 100,000 yrs from a marine sediment core taken in Delagoa Bight, off southeastern Africa. In addition to providing a sea surface temperature reconstruction for the past ca. 100,000 yrs, this record also allows a high-resolution continental climatic reconstruction. Climate sensitive organic proxies, like the distribution and isotopic composition of plant-wax lipids as well as elemental indicators for fluvial input and weathering type provide information on climatic changes in the adjacent catchment areas (Incomati, Matola, and Lusutfu rivers). At the transition between glacials and interglacials, shifts in vegetation correlate with changes in sea surface temperature in the Agulhas current. The local hydrology, however, does not follow these orbital-paced shifts. Instead, precipitation patterns follow millennial scale variations with different forcing mechanisms in glacial versus interglacial climatic states. During glacials, southward displacement of the Intertropical Convergence Zone facilitates a transmission of northern hemispheric signals (e.g. Heinrich events) to the southern hemispheric subtropics. Furthermore, the southern hemispheric westerlies become a more direct source of precipitation as they shift northward over the study site, especially during Antarctic cold phases. During interglacials, the observed short-term hydrological variability is also a function of Antarctic climate variability, however, it is driven by the indirect influence of the southern hemispheric westerlies and the associated South African high-pressure cell blocking the South Indian Ocean Convergence Zone related precipitation. As a consequence of the interplay of these effects, small scale climatic zones exist. We propose a conceptual model describing latitudinal shifts of these zones along the southeastern African coast as tropical and temperate climate systems shift over glacial and interglacial cycles. The proposed

model explains some of the apparent contradictions between several paleoclimate records in the region.

Key words: Delagoa Bight; southern hemisphere westerlies; South Indian Ocean Convergence Zone; sea surface temperatures; hydrogen isotopes; carbon isotopes; elemental composition

1. Introduction

Despite the increasing number of southern African paleoclimate studies, large data gaps and unresolved debates remain. Controversies concern both the interpretation of the climate records as well as the contradictory major climate forcings that have been proposed for the region. In southeastern Africa, the main moisture source is the warm Indian Ocean (Tyson and Preston-Whyte, 2000), the mechanisms controlling the intensity and duration of the easterly rainfall over time remain, however, uncertain. Climate variations on glacial-interglacial timescales in southernmost Africa were reported to be directly forced by local (southern hemispheric) insolation (Partridge et al., 1997; Schefuß et al., 2011; Simon et al., 2015; Caley et al., 2018). Strong southern hemispheric summer insolation was hypothesized to cause wet climatic conditions along the east African coast due to a stronger atmospheric convection and an increase in the land/ocean temperature contrast, which results in higher moisture transport by the tropical easterlies. However, recent paleo-reconstructions suggested a synchrony with northern hemisphere climate signals, which are inversely correlated to southern hemispheric insolation (e.g. Truc et al., 2013). As a mechanism of transmitting the northern hemispheric signal to southern Africa, ocean circulation variability (Agulhas current strength; i.e. sea surface temperatures [SST]) has often been proposed (Biastoch et al., 1999; Reason and Rouault, 2005; Dupont et al., 2011; Tierney et al., 2008; Stager et al., 2011; Scott et al., 2012; Truc et al., 2013; Baker et al., 2017; Chase et al., 2017). In terms of vegetation shifts, atmospheric CO_2 variability and temperature have been suggested as major driving mechanisms over glacial-interglacial cycles (Dupont et al., 2019). Nowadays, eastern South Africa is not under the direct influence of the intertropical convergence zone (ITCZ) as its modern maximum southern extension is ca. 13-14°S (Gasse et al., 2008). However, the position of the ITCZ was more southerly during glacial periods (Nicholson and Flohn, 1980; Chiang et al., 2003; Chiang and Bitz, 2005), which may have allowed ITCZ shifts to reach much further south along the east African coast than today (c.f. Johnson et al., 2002; Schefuß et al., 2011; Ziegler et al., 2013; Simon et al., 2015). At the same time, the southern hemispheric westerlies (SHW), which presently influence only the southernmost tip of Africa, are hypothesized to have moved northward during glacial periods of increased south Atlantic sea ice extent (Anderson et al., 2009; Sigman et al., 2010; Miller et al., 2019b). As suggested by Miller et al., (2019b), in such a scenario the temperate systems may have brought winter moisture to the southeast African coast and/or blocked South Indian Ocean Convergence Zone (SIOCZ) related precipitation during the summer months. Regional studies integrating many of the available records have found that; i) several small-scale climatic dipoles exist due to the interaction of various driving mechanisms and that ii) the spatial extent of these climatic regions has varied considerably since the last Glacial (Chevalier et al., 2017, Chase et al., 2018; Miller et al., 2019b). Miller et al., (2019b) compile paleorecords along the southeastern African coast and propose a conceptual model of climatic variability during the Holocene. The authors describe three climatic zones; a *northern SRZ* where the climate is driven by local insolation, and *a central and eastern SRZ* and *southern South African* zone where climate is driven by shifts of the southern hemisphere westerlies, the South African high-pressure cell and the SIOCZ. Equatorward shifts of the southern hemisphere westerlies, the South African high-pressure cell and the SIOCZ result in humid conditions in the *southern South African zone*, whereas they cause arid conditions in the *central and eastern SRZ.* We analyze a marine core located within the *central and eastern SRZ* that offers a continuous high-resolution record of the past ca. 100,000 yrs allowing us to add to the existing conceptual models of southeastern African climate dynamics, and to gain an understanding of glacial climate mechanisms in the region. A combination of organic and inorganic geochemical proxies is used in order to decipher the hydrological processes on land, while foraminiferal shell geochemistry serves as a proxy for ocean circulation variability. With this approach we aim to decipher some of the discrepancies concerning the driving mechanisms of southeast African hydroclimate and vegetation shifts during the last glacial-interglacial cycle.

1.2 Regional setting

The coring site is located in an embayment on the southeastern African continental shelf called the Delagoa Bight (Fig. 1a). The southern directed Agulhas Current flows along the East African margin transporting warm and saline water from the tropical Indian Ocean to the tip of

Southern Africa (Zahn et al., 2012). The current system is structured into a series of large-scale (~200 km diameter) anti-cyclonic eddies occurring about 4 to 5 times per year (Quartly and Srokosz, 2004). As they pass the Delagoa Bight, these eddies, together with the Agulhas Current itself, drive the Delagoa Bight eddy; a topographically constrained cyclonic lee eddy at the coring location (Lutjeharms and Da Silva, 1988; Quartly and Srokosz, 2004). Although the coring site is located just west of the mouth of the major Limpopo river system, Schüürman et al., (2019) show that the inorganic material at our site most likely originates from three minor rivers, Incomati, Matola, and Lusutfu, that flow into the Indian Ocean further to the southwest. This is attributed to the eastward deflection of the Limpopo sediments by the Delagoa Bight eddy. The eddy appears to have been stable and strong enough to effectively constrain the drift of the Limpopo sediments eastwards over the late Pleistocene and Holocene (Schüürman et al., 2019). The three rivers, Incomati (also known as Komati), Matola (also known as Umbeluzi), and Lusutfu (also known as Maputo), have catchment areas of ca. 45 300 km^2, 6 600 km^2, and 22 700 km^2, respectively, comprising the coastal region and the eastern flank of the Drakensberg Mountains. Between the Drakensberg escarpment and the coast lies a N-S oriented low ridge, the Lebombo Mountains (400–800 m a.s.l.). The geological formations of this area are the Archaean Kaapvaal Craton, the Karoo Igneous Province, as well as the Quaternary deposits on the coastal plains (de Wit et al., 1992; Sweeney et al., 1994). Climatically these catchments are in the transition zone between tropical and subtropical climate; at the southern limit of the subtropical ridge between the southern Hadley and the Ferrel cell (Tyson and Preston-Whyte, 2000). The average annual temperature ranges from 16°C in the highlands to 24°C in the lowland area. (Kersberg, 1996). Rain (ca. 1,000 mm annually) falls mostly in summer (ca. 67 % of annual rainfall from November to March) (Xie and Arkin, 1997; Chase and Meadows, 2007). Although the ITCZ currently does not directly affect the region, it does induce latitudinal shifts in the SIOCZ, which can be considered as a southward extension of the ITCZ. When the ITCZ is in its southernmost (summer) position, tropical temperate troughs (TTTs), forming at the SIOCZ bring easterly rainfall from the Indian Ocean (Jury et al., 1993; Reason and Mulenga, 1999) (Fig 1b). During austral summer, a low-pressure cell dominates the Southern African interior, enabling tropical easterlies/TTT to bring rainfall to the region. This rainfall is suppressed during austral winter, when a subtropical high-pressure cell is located over southern Africa, (Fig. 1b). This high-pressure cell creates a blocking effect over the continent, which stops moisture advection inland over the majority of South Africa during winter (Dedekind et al., 2016). The winter rain that does fall (33 % of annual rainfall from April to October) is associated with extratropical cloud bands and thunderstorms linked to frontal systems that develop in the main SHW flow (between 40 °S and 50 °S). As the SHW shift northward during the winter, these frontal systems may become cut off and displaced equatorward as far north as 25°S (c.f. Baray et al., 2003; Mason and Jury, 1997) (Fig 1c). Associated with this climatological and topographic setting we find a vegetation in the Incomati, Matola, and Lusutfu catchment areas that consists mainly of coastal forests and mountain woodlands with savanna elements only in the northernmost parts of the catchment and sedges along the riverbanks and floodplains (see White, (1983) and Dupont et al., (2011) for a more detailed description of the vegetation biomes).

2 Material and methods

2.1 Sediments

Gravity core GeoB20616-1 (958 cm long) was retrieved from 25°35.395´S; 33°20.084´E on 15.02.2016 from a water depth of about 460 m. Shipboard sedimentological analysis showed a lithology of clayey silt with signs of slight bioturbation. The composition was observed as mainly clastic with occurrence of foraminifera and shell fragments (Zabel, 2016).

2.2 Oxygen isotopic composition of planktonic foraminifera

Stable oxygen isotopes values values of planktonic foraminifera (*G. ruber*, white variety, >150 µm) were measured in the interval between 395 and 935 cm at 10 cm resolution for age-modeling (Suppl.1). For each measurement, around eight shells of *G. ruber* were selected and analyzed at the MARUM – Center for Marine Environmental Sciences, University of Bremen, Germany using a ThermoFisher Scientific 253 plus gas isotope ratio mass spectrometer with Kiel IV automated carbonate preparation device. Data were calibrated against an in-house standard (Solnhofen limestone). The results are reported in permil (‰, parts per thousand) versus Vienna Peedee belemnite (VPDB). Standard deviation of in-house standard (Solnhofen limestone) $\delta^{18}O$ over the measurement period was 0.06 ‰.

2.3 Age model

Until the limit of radiocarbon dating the age model used in this study is based on 8 radiocarbon ages of *G. ruber*, one shell fragment and a bulk total organic carbon surface sample (see Table 1). The cleaning procedures as well as the Accelerator Mass Spectrometry (AMS) measurements were carried out in the Poznań Radiocarbon Laboratory, Poland. The modelled ocean average curve (Marine13) (Reimer et al., 2013) and a local marine ΔR of 121±16 ^{14}C yr (Maboya et al., 2017) were applied to calibrate the radiocarbon ages. To perform these calculations the Calib 7.1 software (Stuiver et al., 2019) was used. For flexible Bayesian age-depth modelling of the available ^{14}C dates, the software Bacon (Blaauw and Christen, 2011) (Fig. 2b) was used. The uncertainty of the radiocarbon dates is indicated in Table 1. The uncertainty of the Bacon model is indicated in Fig. 2b (grey lines). However, there is possibly an underestimation of the error in the age model around two periods of slow deposition in the interval from 15 to 6 ka BP and in the interval from 32 to 25 ka BP. The calibrated ^{14}C age of a shell fragment found in this interval (390cm) was used as a ^{14}C-tie-point (see Table 1), additionally 2 $\delta^{18}O$ tie-points were defined and an age model was calculated using the software AnalySeries (Paillard et al., 1996) (Fig. 2a). The age-depth model was extended by planktonic foraminifera $\delta^{18}O$ correlation using major $\delta^{18}O$ shifts in the LR04 stack as a reference (Lisiecki and Raymo, 2005) (Fig. 2a,b). With this low number of tie-points it is difficult to capture heterogeneity in the deposition rate, which must be considered when estimating the error of the age model. For the error estimation of $\delta^{18}O$ tie-points the mean resolution of the GeoB20616-1 $\delta^{18}O$ record and the reference curve around the tie-point depth and age (respectively) was taken into account as well as the absolute age error of the time-scale used for the reference record and a matching error visually estimated when defining tie-points. Figure 2b (grey lines) gives an estimate of the age model error. In this paper, we refer to median age estimations.

2.4 Foraminiferal Mg/Ca

Up to 20 specimens (> 150 µm) of *G. ruber* (white) (> 150 µm) were selected for Mg/Ca analysis (see Suppl.2). Foraminiferal tests were gently crushed prior to standard cleaning procedures for Mg/Ca in foraminifera (Barker et al., 2003). For clay and organic matter removal ultrasonic cleaning was alternated with washes in deionized water and methanol, an oxidizing step with 1 %-H_2O_2 buffered in 0.1M NaOH followed, which was then neutralized by deionized water washes.

A final weak acid leach with 0.001M QD HNO_3 was performed before dissolution in 0.5 mL 0.075 M QD HNO_3 and centrifugation for 10 min (6,000 rpm). The samples were diluted with Seralpur water before analysis with inductively coupled plasma optical emission spectrometry (Agilent Technologies, 700 Series with autosampler ASX-520 CETAC and micro-nebulizer) at MARUM, University of Bremen, Germany. Instrumental precision was monitored after every five samples using analysis of an in-house standard solution with a Mg/Ca of 2.93 mmol mol^{-1} (standard deviation of 0.020 mmol mol^{-1} or 0.67 %). A limestone standard (ECRM752-1, reported Mg/Ca of 3.75 mmol mol^{-1}) was analyzed to allow inter-laboratory comparison (Greaves et al., 2008; Groeneveld and Filipsson, 2013).

2.5 Organic geochemistry

Total lipid extracts (TLEs) were extracted from ca. 9-27 g of the freeze-dried, homogenized samples with a DIONEX Accelerated Solvent Extractor (ASE 200) at 100°C and at 1,000 psi for 5 minutes (repeated 3 times) using a dichloromethane (DCM):methanol (MeOH) (9:1, v/v) mixture. Squalane was added in a known amount to the samples as internal standard before extraction. Elemental sulphur was removed from the TLEs using copper turnings. After saponification by adding 6 % KOH in MeOH and extraction of the neutral fractions with hexane, the neutral fractions were split into hydrocarbon, ketone, and polar fractions using silica gel column chromatography (with a mesh size of 60 µm) and elution with hexane, DCM and DCM:MeOH (1:1), respectively. Subsequently elution of the hydrocarbon fractions with hexane over an $AgNO_3$-impregnated silica column yielded saturated hydrocarbon fractions. The concentrations of long-chain *n*-alkanes in the saturated hydrocarbon fractions were determined using a Thermo Fischer Scientific Focus gas-chromatograph (GC) with flame-ionization-detection (FID) equipped with a Restek Rxi 5ms column (30m x 0.25mm x 0.25µm). Quantities of individual *n*-alkanes were estimated by comparison with an external standard containing *n*-alkanes (C_{19}–C_{34}) at a known concentration. Replicate analyses of the external standard yielded a quantification uncertainty of <5 %. The carbon preference index (CPI) was calculated using the following equation:

$CPI = 0.5 * (\sum C_{odd27\text{-}33} / \sum C_{even26\text{-}32} + \sum C_{odd27\text{-}33} / \sum C_{even28\text{-}34})$ with C_x the amount of each homologue (Bray and Evans 1961).

The δD values of long-chain *n*-alkanes were measured using a Thermo Trace GC equipped with an Agilent DB-5MS (30m length, 0.25 mm ID, 1.00 µm film) coupled via a pyrolysis reactor (operated at 1420°C) to a Thermo Fisher MAT 253 isotope ratio mass spectrometer (GC/IR-MS). The δD values were calibrated against external H_2 reference gas. The H^{3+} factor was monitored daily and varied around 6.23 ± 0.04 ppm nA^{-1}. δD values are reported in permil (‰) versus Vienna Standard Mean Ocean Water (VSMOW). An *n*-alkane standard of 16 externally calibrated alkanes was measured every 6^{th} measurement. Long-term precision and accuracy of the external alkane standard were 3 and <1 ‰, respectively. When *n*-alkane concentrations permitted, samples were run at least in duplicate. Precision and accuracy of the squalane internal standard were 2 and <1 ‰, respectively (n=41). Average precision of the *n*-C_{29} alkane in replicates was 4 ‰. The $\delta^{13}C$ values of the long-chain *n*-alkanes were measured using a Thermo Trace GC Ultra coupled to a Finnigan MAT 252 isotope ratio monitoring mass spectrometer via a combustion interface operated at 1,000°C. The $\delta^{13}C$ values were calibrated against external CO_2 reference gas. $\delta^{13}C$ values are reported in permil (‰) against Vienna Pee Dee Belemnite (VPDB). When concentrations permitted, samples were run at least in duplicate. Precision and accuracy of the squalane internal standard were 0.1 and 0.4 ‰, respectively (n=41). An external standard mixture was analyzed repeatedly every 6 runs and yielded a long-term mean standard deviation of 0.2 ‰ with a mean deviation of 0.1 ‰ from the reference values. Average precision of the *n*-C_{29} alkane in replicates was 0.3 ‰. We focus the discussion on the isotopic signals of the *n* -C_{31} alkane, as this compound is derived from grasses and trees present throughout the study area. Supplement 3 shows, however, that the *n* -C_{29} and *n* -C_{33} alkanes reveal similar trends.

2.6 Inorganic geochemistry

The elemental composition of all onshore and offshore samples was measured using a combination of high resolution (1 cm) semi-quantitative XRF scanning and lower (5 cm) resolution quantitative XRF measurements on discrete samples (see Suppl. 4). XRF core scanning (Avaatech XRF Scanner II at MARUM, University of Bremen) was performed with an excitation potential of 10 kV, a current of 250 mA, and 30 s counting time for Ca, Fe, K and Al. For discrete measurements on 110 dried and ground samples, a PANalytical Epsilon3-XL XRF spectrometer equipped with a rhodium tube, several filters, and a SSD5 detector was used. A calibration based on certified standard materials (e.g. GBW07309, GBW07316, and MAG-1) was used to quantify elemental counts (c.f. Govin et al., 2012).

3 Results and discussion

3.1 Proxy indicators

3.1.1 SST

The magnitude of temperature variability (from ca. 27°C during interglacials to ca. 24°C during glacials) in the GeoB20616-1 Mg/Ca SST record and the timing of changes (postglacial warming at ca. 17 ka BP) correspond to existing regional Mg/Ca SST records (c.f. Fig. 3; Bard et al., 1997; Levi et al., 2007; Wang et al., 2013). They do, however, not correspond to SST calculated from other indicators (i.e. $U^{K'}_{37}$, TEX^{86}) (e.g. Wang et al., 2013; Caley et al., 2011). These indicators show slightly different patterns, which may be attributed to a seasonal bias in the proxies (Wang et al., 2013). Wang et al., (2013) suggest that $U^{K'}_{37}$ SST reflects warm season SST mediated by changes in the Atlantic, whereas the *G. ruber* Mg/Ca SST indicator used in this study records cold season SST mediated by climate changes in the southern hemisphere.

3.1.2 Vegetation signatures

The $\delta^{13}C_{wax}$ record of core GeoB20616-1 shows average values of approximately -24‰ VPDB (c.f. Suppl. 3) and shifts from ca. -25 ‰ to ca. -24 ‰ (at 85 ka BP) and from -24 ‰ to – 25 ‰ (at ca. 10 ka BP). The stable carbon isotopic composition of plant waxes reflects discrimination between ^{12}C and ^{13}C during biosynthesis varying with vegetation type: C_4 plants have higher $\delta^{13}C$ values than C_3 plants (e.g., Collister et al., 1994; Herrmann et al., 2016). The average $\delta^{13}C$ value of the analyzed samples falls into the range between C_3 alkanes (around -35‰) and C_4 alkanes (around -20‰) (Garcin et al., 2014) indicating that the *n*-alkanes were derived from C_3 sources in the catchment such as mountain shrublands and coastal forests, as well as from C_4 sedges which grow along rivers and in the associated swamplands (c.f. Fig. 1a). There is no correlation (R^2=0.15) of $\delta^{13}C_{wax}$ variability and hydrological variability indicated by δD_{wax} (see section 3.1.3 Precipitation indicators for details on this proxy). We therefore suggest that the shifts we see in the $\delta^{13}C_{wax}$ werenot induced by a xeric/mesic adaptation of the same plant community. Instead, we imply that the shifts in the $\delta^{13}C_{wax}$ signal were related to shifts in the vegetation community. Palynological work on a nearby marine sediment core by Dupont et al. (2011) shows that large shifts in vegetation biomes are also observed in the Limpopo catchment which is directly adjacent to the Incomati, Matola and Lusutfu catchments (Fig. 1a). A comparison of the Dupont et al. (2011) palynological data (Fig. 3c) and the $\delta^{13}C_{wax}$ data at our site (Fig. 3a) shows a covariation of major shifts in vegetation and $\delta^{13}C_{wax}$. Although the similarities in the pattern of vegetation shifts detected in the nearby Limpopo river sediment core and at our study site suggest that large scale vegetation shifts took place in the region over glacial – interglacial transitions, this does not necessarily imply the mechanisms behind these trends are the same. Studies of the Limpopo sediment record (Dupont et al. 2011; Caley et al. 2018) reveal a $\delta^{13}C_{wax}$-enriched grassland vegetation for glacial intervals and an increase of woodland vegetation during well-developed interglacial periods, as is the case for MIS 5 and 1 (as opposed to MIS 3), reflected in lighter $\delta^{13}C_{wax}$ values. Caley et al., (2018) attribute the $\delta^{13}C_{wax}$-enrichment in Limpopo river sediments during glacials to an expansion of floodplains and the associated C_4 sedges, as well as discharge from the upper Limpopo catchment which reached well into the grassland interior of southern Africa (almost 1,000 km inland). The headwaters of the Incomati, Matola, and Lusutfu catchment areas, however, are in the Lebombo mountain range located within 200 km of the coast. They do not reach into the interior grassland biomes of South Africa. We therefore propose that in the Incomati, Matola, and Lusutfu catchment areas, the heavier $\delta^{13}C_{wax}$ values for the glacial MIS 4-2 interval reflect retreating forests and an expansion of drought tolerant C_4 plants (grasses) due to growing season aridity, whereas interglacial (MIS 1 and 5) lighter $\delta^{13}C_{wax}$ values reflect the formation of woodlands. Furthermore, sedge-dominated open swamps that fringed rivers during MIS 4-2 may have been replaced by gallery forests during MIS1 and 5 contributing to the glacial to interglacial $\delta^{13}C_{wax}$ depletion.

3.1.3 Precipitation indicators

Hydrogen isotope changes measured in plant waxes are related to the isotope composition of precipitation since hydrogen used for biosynthesis originates directly from the water taken up by the plants (Sessions et al., 1999). In tropical and subtropical areas, the isotopic composition of rainfall (δD_p) mainly reflects the amount of precipitation - with δD_p depletion indicating more rainfall (Dansgaard, 1964). Furthermore, rainfall δD_p signatures may also become deuterium-depleted with altitude (ca. 10–15 ‰ per 1,000 m, Gonfiantini et al., (2001)). The δD values of leaf waxes in the three catchments are probably affected by both the amount as well as the altitude effect. Rainfall at higher altitudes takes place during times of generally increased rainfall, as it is high precipitation events that reach the interior. The altitude effect therefore enhances the δD depletion of the "amount effect". The K/Al ratio of the sediment is a less direct indicator of the precipitation regime: K/Al has been interpreted as an index between illite (K,H_3O) and kaolinite

($Al_2Si_2O_5(OH)_4$) giving an indication of the prevailing weathering regime as illite is a product of physical weathering whereas kaolinite is produced during chemical weathering (Clift et al., 2008; Dickson et al., 2010; Burnett et al., 2011). The Ca/Fe ratio is generally used as a proxy of marine (Ca) versus terrestrial (Fe) input to the core site and thus indicative of changes in terrestrial discharge by the river systems (Hebbeln and Cortés, 2001; Croudace et al., 2006; Rogerson et al., 2006; Rothwell and Rack, 2006; McGregor et al., 2009; Dickson et al., 2010; Nizou et al., 2010). The red/blue ratio of the sediment reflects sediment color nuance and increases with sediment lightness. In Geob20616-1 we interpret the reddish values as a more clastic deposition indicative of arid conditions whereas darker blueish colors may reflect clay and organic rich sediments preferentially deposited during humid phases (see also M123 cruise report Zabel, 2006). In the records of δD_{C31}, red/blue , K/Al and Ca/Fe similar patterns can be observed: They all display relatively high values (up to -144 ‰, 1.4; 12 and 0.25 respectively) in the intervals marked in red/yellow in Fig.4 and lower values (down to -160, 1.1, 1, and 0.2 respectively) in the intervals marked in blue/green in Fig. 4. We associate these variations with (respectively) decreasing (red/yellow) and increasing (blue/green) precipitation over the Incomati, Matola, and Lusutfu catchment areas. We note that the observed correlation, in particular for the inorganic proxies (K/Al and Ca/Fe), is relative rather than absolute in nature. This can be associated with the changing background conditions over glacial and interglacial cycles which may cause shifts in the elemental composition. We also note that of the four proxy indicators (δD_{C31}, red/blue, K/Al and Ca/Fe) only δD_{C31} can be considered as direct indicator of past precipitation change. Red/blue, K/Al and Ca/Fe depend to varying extents on precipitation, erosion and fluvial transport, whereas these factors do not necessarily vary in concert. For instance, erosion is not always directly linked to the amount of precipitation and vegetation density is often an additional and more important factor for erosion rates. Erosion rates can also increase substantially at times of rapid climatic and associated vegetation changes. Because the relationship between precipitation, erosion and riverine transport is not linear we base our precipitation reconstruction (i.e. the definition of the arid and wet intervals described in section 3.2 and colored-coded in Fig. 4) mainly on the δD_{C31} values. We consider the red/blue , K/Al and Ca/Fe values as supportive information; the relative correlation of the four proxies suggests that phases of increased precipitation are, for the most part, associated with an increase in erosion rates, chemical weathering and riverine transport. This underlines the reliability of our paleo-precipitation reconstruction.

3.2 Climatic patterns at different time scales

3.2.1 Orbital time scales

3.2.1.1. Sea surface temperatures and vegetation

Over the past 100,000 yrs the SST and $\delta^{13}C_{C31}$ values show a common trend of high SST and low $\delta^{13}C_{wax}$ values during interglacial MIS 5 and 1 and low SST and high $\delta^{13}C$ values during glacial MIS 4-2 (Fig. 3). Our data reveal an increase in SST of ca. 4°C from glacial to interglacial conditions. This correlation between SST and glacial-interglacial changes cycles is commonly found for this area (Caley et al., 2011; Dupont et al., 2011; Caley et al., 2018). On this glacial-interglacial time scale, variations in local SST are thought to be an important driver of hydroclimate in southeastern Africa (c.f. Dupont et al., 2011). During interglacials, warm SST within the Mozambique Channel and Agulhas Current induce an advection of moist air and higher rainfall in the east South African summer rainfall zone (e.g. Walker, 1990; Reason and Mulenga, 1999; Tyson and Preston-Whyte, 2000). The opposite effect is inferred for glacial periods (Dupont et al., 2011; Chevalier and Chase, 2015). The strong influence of western Indian Ocean surface temperatures on the summer precipitation in northern South Africa and southern Mozambique induces a tight coupling between vegetation dynamics in southeastern Africa and sea surface temperature variations in the Western Indian Ocean. This has been shown for several glacial – interglacial cycles in a palynological study offshore Limpopo River (core MD96-2048; Fig. 1a) by Dupont et al., (2011).

3.2.1.2. Hydrology over glacial-interglacial transitions

δD, XRF, and color data are indicators of catchment precipitation changes: decreases in red/blue, Ca/Fe, K/Al ratios and δD values indicate higher precipitation in the catchment, more fluvial discharge and higher chemical weathering rates (see section 3.1.3). Although there is much variability in the hydrological record of core GeoB20616-1, red/blue, Ca/Fe, K/Al ratios and δD values are surprisingly stable over glacial –interglacial transitions (mean δD value of MIS 1 and 5: -149 ‰ versus mean δD value of MIS 2-4: -150 ‰). It can be assumed that, during glacials, the rainfall from the main rain bearing systems (SIOCZ related tropical temperate troughs) was reduced due to generally lower land- and sea-surface temperatures and a weaker global hydrological cycle. However, a southward shift of the ITCZ during glacials as previously suggested

(Nicholson and Flohn, 1980; Johnson et al., 2002; Chiang et al., 2003; Chiang and Bitz, 2005; Schefuß et al., 2011) would have contributed to increased rainfall in the study area. It is unclear if the region would have been under the direct influence of the ITCZ during glacials or if southward shifts of the ITCZ entailed a southward shift of the SIOCZ and thus increased precipitation via the TTT. Furthermore, SHW related low pressure systems shifting northward to the Incomati, Matola and Lusutfu catchment areas during glacial conditions may have become a major additional precipitation source. The SHW northward shift of ca. 5° latitude is well documented (Chase and Meadows, 2007; Chevalier and Chase, 2015; Chase et al., 2017; Miller et al., 2019a). The possibility of more frequent SHW related low pressure systems bringing moisture to our study area during the LGM has previous been proposed by Scott et al., (2012) in the framework of a regional pollen review paper. It is also suggested by a modelling study showing an LGM scenario of drier summers and wetter winters for the southeastern African coast (Engelbrecht et al., 2019). During glacial periods, a reduced summer (SIOCZ related) rainfall amount and an increase in SHW related frontal systems as an additional winter precipitation source, possibly in combination with precipitation from a more southerly ITCZ, would translate to a relatively stable annual rainfall amount over glacial-interglacial transitions.

3.2.2 Millennial scale hydrological variability

3.2.2.1 During Interglacial MIS 5

During MIS 5 there are several prominent (ca. -10 **‰**) short-term (1-2 ka) decreases in the δD record, which are paralleled with decreases in Ca/Fe, K/Al and red/blue ratios (Fig. 4). We interpret these intervals (approximately 83-80 ka BP and 93-90 ka BP) as wet periods while intervals of high Ca/Fe, K/Al and red/blue ratios and δD values (approximately 97-95 ka BP, 87.5-85 ka BP and 77.5 ka BP) are interpreted as arid intervals (see section 3.1.2. for details on proxy interpretation). During the interglacial MIS 5, millennial scale increases in humidity correlate broadly to periods of warmth in the Antarctic ice core records termed AIM22 and AIM 21 (AIM: Antarctic isotope maxima) (see Fig. 4; EPICA members, 2010). During these Antarctic warm periods, sea ice, the circumpolar circulation and the SHW retracted. This is recorded by Southern Ocean diatom burial rates as well as paleoclimate archives at the southernmost tips of Africa and South America (Lamy et al., 2001; Anderson et al., 2009; Chase et al., 2009; Hahn et al 2016 and references therein; Zhao et al., 2016). It has been hypothesized that southward shifts of the SHW and the South African high-pressure cell, allow the SIOCZ and TTT to shift further south causing an increase in humidity in our study area. Miller et al., (2019b) suggest this mechanism for the region just south of our site (termed *eastern central zone*), which shows Holocene hydroclimatic shifts similar to those recorded in GeoB20616-1. Holocene arid events in this region are attributed to northward shifts of the SHW and the South African high-pressure cell which block the SIOCZ and TTT related moisture. These mechanisms are described in detail by Miller et al., 2019b and our data suggests that they were also active during earlier interglacial periods (e.g. MIS 5) (c.f. schematic model in Fig. 5a). Our current chronology suggests that southward SHW shifts during Antarctic warm periods caused the prominent humid phases during MIS 5 in the Incomati, Matola and Lusutfu catchment areas during the timeframes around 83-80 ka BP (AIM21) and 93-90 ka BP (AIM22). When our best age estimate is applied there is little correspondence between northern or southern insolation maxima and the MIS5 humid phases. In view of the chronological uncertainty in this early part of the record (beyond the ^{14}C dating limit), we cannot exclude that these humid phases are related to precessional variability, in the absence of ice interference, causing the division in MIS5a-e. However, in accordance with the conceptual model by Miller et al., (2019b) for the Holocene, we observe no local insolation control on climate at our study site. We suggest that the major shifts in the large-scale rain-bearing systems may override the local insolation forcing.

3.2.2.2. During MIS 4-2 glacial conditions

During the glacial periods MIS 2 and 4 and the less prominent interglacial MIS 3, the correlation between southeastern African humidity and Antarctic warm periods (AIM events) does not persist. In contrast; the first two prominent humid phases in MIS 4 (around 68-63 ka BP and 56 ka BP) as well as some of the following more short-term humid phases coincide with cold periods in the Antarctic ice core record (Fig.4). The general position of the SHW trajectories is suggested to have been located 5° in latitude further north during glacial periods (c.f. section *3.2.1.2. Hydrology over glacial-interglacial transitions*). The Incomati, Matola and Lusutfu catchment areas would therefore have been in the direct trajectory of the SHW related low pressure systems. Whilst northward shifts of the SHW and the South African high pressure cell during an interglacial cause aridity by blocking the SIOCZ and TTT (as suggested by Miller et al.,(2019b) and as described in section 3.2.2.1 for e.g. MIS 5), we suggest that during a glacial, additional northward shifts of the SHW (e.g. during Antarctic cold events) would have led to an increase in precipitation related to particularly strong direct influence of the SHW and the related low pressure cells (c.f. schematic model Fig 5b). Fig. 4 also shows a correlation between some of the humid phases during MIS 2-4 and Greenland cold phases i.e. Heinrich stadials. The timing of the wet phases at 68-63 ka, 56 ka, 44 ka, 37 ka, and 23 ka BP corresponds roughly to the following Heinrich stadials: HS6 (after 60 ka BP, Rasmussen et al., 2014); HS5a (56 ka BP, Chapman and Shackleton, 1999); HS5 (45 ka BP; Hemming 2004) and HS4 & HS2 (37 ka BP and 23 ka BP, Bond and Lotti, 1995). Wet phases in eastern Africa have previously been associated with Heinrich events (Caley et al., 2018; Dupont et al., 2011; Schefuß et al., 2011). It is well documented that during glacial conditions the large ice masses of the northern hemisphere displace the thermal equator southward (Nicholson and Flohn, 1980; Johnson et al., 2002; Chiang et al., 2003; Chiang and Bitz, 2005; Schefuß et al., 2011). It is therefore hypothesized that the ITCZ reached latitudes further south than its modern maximal extent causing the MIS 2-4 rainfall peaks. There is no notable "blocking" effect of the South African high-pressure cell during glacials (schematic model Fig. 5b). The transitions from cold "stadial" to warm "interstadial" conditions and back during MIS 2-4 are extremely rapid and short term. The sampling resolution and age – control of our record (especially prior to ca. 50 ka BP – the limit of ^{14}C dating) is not always sufficient for capturing these variations (e.g. HS4). The association of humid phases with a northward shifting SHW and/or southward shifting ITCZ is therefore not always clear and a combination of both may also be possible.

3.2.1.3 From the LGM to the Holocene

Relative to the prolonged arid phase during the late MIS 3/early MIS 2 (37-25 ka BP; c.f. Fig. 4), we observe a trend towards more humid conditions during the LGM (25 – 18 ka BP) marked by a decrease in Ca/Fe, K/Al, red/blue ratios and δD values. This is most likely due to the more frequent SHW-related low-pressure systems bringing moisture to our study area during the LGM and/or southward shifts of the ITCZ as discussed in section *3.2.1.2. Hydrology over glacial-interglacial transitions* (see also Fig. 5b). Our record shows a wettening trend after the Last Glacial Maximum and during the deglacial (from ca. 15 ka BP). Several paleoenvironmental records show a common humidity increase for this interval (Meadows 1988; Scott 1989; Norström et al., 2009). Chase et al.,(2017) attribute this to the invigoration of tropical systems with post-glacial warming. The wet conditions prevail until the early Holocene (ca. 8 ka BP). Similar observations of a ca. 15-8 ka BP wet phase have been made in the region (e.g. Norström et al., 2009; Neumann et al., 2010). For this early -Mid Holocene period, we infer from the leaf wax $\delta^{13}C$ values a shift from grassland to woodlands as described in section 3.2.1.1. and in Dupont et al. (2011). This may be related to the rainfall intensification as well as to the global temperature and CO_2 increase (c.f. Dupont et al., 2019). The early/Mid Holocene wet phase in our study region (*eastern central SRZ*) is described by Miller et al., 2019b and associated with a southward shift of the SHW and the South African high-pressure cell allowing for the SIOCZ related rain bearing systems (TTT) to shift southward over the region. The late Holocene (the last 5 kyrs) however, was an arid phase at our study cite as suggested by the precipitation indicators δD, Ca/Fe, K/Al and red/blue ratios. Several regional records (e.g. Mfabeni peatlands and the *eastern-central region*) show similar shifts; from a wet deglacial / Early Holocene (18-5 ka BP) to dry conditions thereafter (Chevalier et al., 2015; Miller et al., 2019a). Miller et al. (2019b) compile eastern African climate records and recognize a late Holocene tripole of increased humidity north of 20°S and south of 25°S and a contrasting aridity trend in the region in-between. Our catchment is located at the northernmost extent of this intermediate region; while we record an aridity trend in the Late Holocene, the adjacent Limpopo catchment just to the north received higher rainfall amounts during this time interval (Miller et al., 2019b). A northward shift in SHW with the South African high-pressure cell blocking the SIOCZ and TTT is a suggested mechanism for this late Holocene aridity (Miller et al., 2019b; also described in section 3.2.2.1). Likewise, Mason and Jury (1997) (based on a conceptual model by Tyson (1984)) suggest that northward shifting SHW induce rain-bearing low pressure cells to shift away from the eastern African coast towards Madagascar. During the Late Holocene the modern climatic situation of the study area was established: during the summer months the SHW and the South African high-pressure cell are in their southernmost position allowing the SIOCZ related TTT to bring rainfall to the region (66 % of annual precipitation). During the winter months the SHW and the South African high-pressure cell shift northward. In this constellation the SIOCZ and TTT influence are blocked by the South African high-pressure cell, however low-pressure cells may become cut from the main SHW flow bringing winter rainfall to the area (33 % of annual precipitation) as described in section 1.2.

Conclusions

Using the organic and inorganic geochemical properties of sediment core GeoB20616-1 from the Delagoa Bight we were able to reconstruct the vegetation changes and rainfall patterns in the Incomati, Matola and Lusutfu catchments as well as SST trends of the Agulhas waters for the past ca. 100,000 yrs offshore southeastern Africa. Our reconstructions underline the existing dipoles or tripoles in southeastern African climate: although the glacial-interglacial variability at our site resembles that observed in the adjacent Limpopo river catchment, the Holocene hydrological trends are exactly inverted in these neighboring catchments. Small-scale climatic zones have been previously described for the region (c.f. Scott et al., 2012; Chevalier and Chase, 2015; Miller et al., 2019b) and each zone has been attributed to a climatic driving mechanism. Our data provide insights into the spatial shifts of these zones as fundamental shifts in the major climate systems occurred over glacial-interglacial cycles. In accordance with Miller et al., (2019b) we identify displacements of the SHW as the main hydro-climate driver during the Holocene in our study area (termed *central and eastern zone*). The main trajectories of the SHW related disturbances remain so far south during the Holocene, that they rarely deliver direct rainfall to the study area. Instead, northward shifts of the SHW and the South African high-pressure cell block the SIOCZ and thus TTT related rainfalls over the region (Fig. 5a). In this manner latitudinal SHW shifts influence the local rainfall indirectly. Our study not only confirms the Miller et al. (2019 b) conceptual model for the Holocene, but also finds the same mechanisms to be active during MIS5. Similar to Miller et al. (2019b) we find an absence of insolation forcing in our study area. We suggest that at these latitudes local insolation as a climatic forcing mechanism is overridden by shifts in the major rain-bearing systems. We conclude that during interglacials regional wet phases are induced by southward shifting westerlies (related to Antarctic warming trends) allowing for the influence of the SIOCZ related TTT. During glacial periods, however, we observe an inverted relationship between Antarctic warm events and regional humidity, and an additional correlation of several humid intervals with extreme northern hemispheric cold events (HS). This suggests that the mechanisms driving the millennial scale hydrological variability during glacials are not the same as during interglacials. We attribute this to the global reorganization of climate systems during the glacial as the large ice masses at both poles induce a southward shift of the thermal equator and the ITCZ as well as a northward shift of the SHW. Our study site is located at the interface of these “compressed” climate systems. As a result, during full glacial conditions, the region may have received precipitation both from SHW related disturbances as well as from SIOCZ related TTT (Fig. 5b). In this “compressed” state the northward shifts of the SHW and the South African high pressure no longer have the net effect of blocking SIOCZ related precipitation; as this is compensated by the increase in winter rains. Overall humidity therefore shows no considerable decrease during MIS 2-4. Nevertheless, a shift in vegetation from woodland to grasslands takes place during glacials; we attribute this to a reduced growing-season (summer) precipitation, probably in combination with low temperatures and atmospheric CO_2. Our study shows that these mechanisms are active in a spatially very restrained area resulting in small-scale variability. These small-scale climatic dipoles or tripoles make the southeastern African coastal area especially sensitive to shifts in the global climatic system.

Acknowledgments

This work was financially supported by Bundesministerium für Bildung und Forschung (BMBF, Bonn, Germany) within the projects "Regional Archives for Integrated Investigation (RAiN)," project number: 03G0840A and "Tracing Human and Climate impacts in South Africa (TRACES)" project number: 03F0798C. The captain, crew, and scientists of the Meteor M123 cruise are acknowledged for facilitating the recovery of the studied material. This study would not have been possible without the MARUM—Center for Marine Environmental Sciences, University of Bremen, Germany and the laboratory help of Dr. Henning Kuhnert, Ralph Kreutz and Silvana Pape. In particular, we thank the GeoB Core Repository at the MARUM and Pangaea (www.pangaea.de) for archiving the sediments and the data used in this paper. Thanks to all RAiN members as well as Stephan Woodborne and the anonymous reviewer of this manuscript for critical comments and helpful advice.

Captions

Fig. 1 A: Modern vegetation of southern Africa and the Incomati, Matola and Lusutfu catchments (after White 1983) and annual SST over the Indian Ocean (Locarnini et al., 2013). Grey arrows represent the main easterly transport of moisture from the warm Indian Ocean. The Mozambique current (MC), Agulhas current (AC), and counter current (cc) forming a coastal eddy are shown in black. Sites mentioned in the discussion are numbered as: 1) Wonderkrater (Truc et al., 2013); 2) Braamhoek (Norström et al., 2009); 3) Mfabeni (Miller et al., 2019a); 4) MD96-2048 (Dupont et al., 2011; Caley et al., 2011, 2018); 5) GeoB20610-1 (Miller et al., 2019b); 6) GIK16160-3 (Wang et al., 2013); 7) MD79-257 (Bard et al., 1997; Sonzogni et al., 1998; Levi et al., 2007); 8) GeoB9307-3 (Schefuß et al., 2011). B: Map of South Africa in austral summer showing the shematic postion of the low-pressure system, the ITCZ (Intertopical Convergence Zone), the SIOCZ (South Indian Ocean convergence zone) and related rain bearing TTT (tropical temperate troughs). C: Map of South Africa in austral winter showing the shematic postion of the high-pressure system, the weaker TTT (tropical temperate troughs) and the frontal systems associated with the northward shifted SHW (southern hemispheric westerlies).

Fig. 2 Reference curves and age–depth model of core GeoB20616-1. A: LR04 benthic foraminifera $\delta^{18}O$ stack (Lisiecki and Raymo, 2005) (black) compared to GeoB20616-1 (red) *G. ruber* foraminifera $\delta^{18}O$ with indicated tie points. B: Age-depth model based on Bacon v. 2.2 (Blaauw and Christen, 2011; green) and $\delta^{18}O$ correlation (blue). Blue circles in panel B represent the positions of calibrated ^{14}C ages whereas blue circles indicate $\delta^{18}O$ tie points. Grey lines indicate uncertainty.

Fig. 3 Climatic patterns at orbital time scales recorded in GeoB20616-1. Panel a) shows down-core $\delta\ ^{13}C$ values of the C_{31} *n*- alkane in ‰ VPDB of GeoB20616-1 as indicators for shifts in vegetation type (C_3 vs. C_4). Panel b) shows SST (sea surface temperatures) recorded by *G. ruber* Mg/Ca (black line) in GeoB20616-1 as well as offshore Limpopo River (core MD96-2048) SST calculated from TEX_{86} (dashed line) and from $U^{K'}_{37}$ (grey line) (Caley et al., 2011). Panel c) shows Limpopo vegetation endmember EM2 from Dupont et al. (2011). The diamonds indicate C^{14} dates (red) and $\delta^{18}O$ tie points (orange).

Fig. 4 Millennial scale hydrological variability recorded in core GeoB20616-1. Organic and inorganic down-core geochemistry (c-f: δD, red/blue, K/Al and Ca/Fe) of GeoB20616-1 as indicators for weathering type, fluvial input and aridity. Intervals identified as wet using these indicators are marked in blue or green, while dry phases are marked in red or yellow. Wet intervals marked in green are associated with southward shifts of the SHW (southern hemispheric westerlies) and the South African high-pressure cell allowing for the SIOCZ (South Indian Ocean convergence zone) and related rain bearing TTT (tropical temperate troughs) to move over the study area during interglacials. In turn, wet intervals marked in blue are associated with northward shifts of the SHW and/or southward shifts of the ITCZ during glacials. Arid phases during interglacials (marked in yellow) are related to northward shifts of the SHW as this induces the moisture-blocking effect of the South African high-pressure cell over the region. During glacials, however, southward shifts of the SHW are often associated with arid phases (marked in red) as the rain-bearing systems related to the SHW move south. Transitional intervals between arid and wet intervals are not colored. XRF scanning data is marked as a line, whereas discrete XRF measurements are represented by points. Panel c represents the δD of the C_{31} *n* alkane in the unit ‰ VSMOW. For comparative purposes local insolation (Laskar, 2011) as well as Arctic and Antarctic ice core d^{18}O records are plotted (NGRIP members, 2004; EPICA members, 2010). The most prominent AIM (Antarctic isotope maxima) and HS (Heinrich Stadial) events are named. The diamonds indicate C^{14} dates (red) and δ^{18}O tie points (orange).

Fig. 5 Conceptual model of precipitation shifts during glacial vs. interglacial (present conditions) intervals. The blue shaded boxes indicate the locations of the major regional rain-bearing systems: i) the TTT (tropical temperate troughs) moisture shifting with the SIOCZ (south Indian Ocean convergence zone) and bearing summer rain (therefore marked as SR) ii) the low-pressure systems related to the SHW (southern hemispheric westerlies), bringing mainly winter rain (therefore marked as WR). The orange shaded box marks South African high-pressure cell (HPC) shifting with the SHW. The HPC blocks SIOCZ and TTT related moisture and therefore causes aridity. The arrows mark the millennial scale variability of the position of these systems over the study area which is marked by a star. Please note that the millennial scale variations that the region experiences differ in the interglacial state (box A) and the glacial state (box B) since the organization of the major climatic systems (marked in red) is different ("decompressed" vs "compressed"). The conceptualization for interglacial states presented in box A is based on a schematic model by (Miller et al., 2019b). In this "decompressed" state latitudinal shifts of the SHW indirectly control precipitation at our study site via the moisture blocking effect of the South African HPC: southward shifts of the SHW and HPC allow the SIOCZ related TTT to bring SR to our site, whereas northward movements block this SR moisture (Miller et al., 2019b). During the "compressed" glacial state (box B) the SHW related WR reaches much further north, directly influencing the study site. The SR, in turn is shifted southward and an HPC blocking effect is not noted at our site.

Table. 1 AMS radiocarbon analyses of material from core GeoB20616-1. The modelled ocean average curve (Marine13) (Reimer et al., 2013) was used for calibration and a local ΔR of 121±16 ^{14}C yr (Maboya et al., 2017) was applied. The ages were calibrated with Calib 7.1 software (Stuiver et al., 2019)

Supplement 1 GeoB20616-1 Oxygen and carbon isotopic composition of planktonic foraminifera (*G.ruber*).

Supplement 2 GeoB20616-1 downcore sea surface temperatures (SST) calculated following Lea et al., 2003 using Mg/Ca analysed on the planktonic foraminifer *G. ruber* (in mmol/mol).

Supplement 3 GeoB20616-1 organic geochemical down-core data. *n*-alkane isotopic composition and distribution descriptive parameters averaged. The elevated CPI values ranging from 3.8 to 14 indicate that the *n*-alkanes within the terrestrial and marine samples were likely derived from non-degraded, terrestrial, higher plant material (Eglinton & Hamilton, 1967). We focus the discussion on the isotopic signals of the *n*-C_{31} alkane but note that the *n*-C_{29} and *n*-C_{33} alkanes reveal similar trends.

Supplement 4 GeoB20616-1 inorganic geochemical down-core data from discrete XRF measurements.

Formatiert: Abstand Nach: 0 Pt., Zeilennummern unterdrücken

hat formatiert: Muster: Transparent (Hintergrund 1)

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Sample and data availability

Samples and data are respectively archived at the GeoB Core Repository and Pangaea (www.pangaea.de) both located at MARUM, University of Bremen.

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First reviewer comments and rebuttal

Summary

In this article, Hahn et al. report on the analysis of a 958 cm sediment core that was taken in the Delagoa Bight off southeastern Africa. The source of the sediment is argued to be from three nearby river catchments that are relatively small, and as a result of this the environmental information derived from the sediments represents a fairly clean signal (incontrast to nearby cores that sample the Limpopo River catchment which is vast and probably includes multiple climate sensitivities). The chronology for the core is generated using 12 radiocarbon dates, of which two are beyond the limit of the method, and δ18O values from benthic foraminifera that are compared to the LR04 benthic stack. The radiocarbon dates from the upper part of the record overlap with the δ18O values from the lower part of the stack so that one of the tie δ18O tie points has an apriori age assignment. The results of the age model demonstrate a relatively constant deposition over the last 100 000 years. Ca/Mg ratios on foraminifera are used to reconstruct past sea surface temperature (SST), and this record demonstrates that SST was almost 4◦C warmer during interglacials (MIS 5 & 1) that it was during glacials (MIS4-2). The main substance of the article is the presentation of multiple geochemical tracers of terrestrial climate and the associated vegetation responses in the catchment. These include δ13C C31 (terrestrial plants community indicator), red/blue ratios (organic vs. classic indicators), K/Al (chemical vs. physical weathering), Ca/Fe (terrestrial vs. marine source indicator) and δDC31 values (rainfall amount indicators). The authors identify a coherent pattern in which all of the geochemical tracers vary in concert with one another, and this is coherently argued to reflect hydrological changes in the associated river catchments. The underlying cause for the alternation between mesic and xeric conditions in the catchment are explored through northern hemisphere forcing in the form of Heinrich Stadials, and southern hemisphere forcing in the form of Antartic ice advances. What emerges is that the forcing during glacials and interglacials differ from one another, and this must be reconciled through synoptic scale changes in the drivers of continental rainfall (rather than insolation variability). The model that is proposed centers on the way in which the two main moisture-bearing systems, the inter-tropical convergence zone (ITCZ) and the southern hemisphere westerlies (SHW), interact and influence the development of the South African high system that is dominant influence on modern rainfall. In essence the argument is that during glacial conditions the SHW migrate northwards because of Antarctic expansion, while the thermal equator (ITCZ) migrates southward because of Arctic expansion, and the South African high system occludes. The extinction of the South African high pressure system during glacials prevents the development of the southern Indian Ocean convergence zone (SIOCZ) and the associated temperate tropical troughs (TTT) that dominate modern summer rainfall (but please note the comment on this subject below). As a result, the catchment maintained a relatively constant water balance between glacials and interglacials as the glacial loss of the TTT/SIOCZ was compensated by direct summer rainfall from the ITCZ and/or winter rainfall from the southern hemisphere westerlies (depending on the dominant Arctic vs. Antarctic forcing at the time).

Scientific merit

Notwithstanding the critique that is presented below, this manuscript makes a valuable contribution to climate science in southern Africa. The dynamics of the climate system are relevant to both future projects of climate change, and the interpretation of the rich archaeological heritage of the region. Several archaeological sequences of a similar age are in close enough proximity for the climate model to be relevant. Sibudu cave and Border Cave contain evidence of mesic/xeric cycles, and they are also well dated, so there is potential to refine the glacial/interglacial climate model. Key palaeoclimate records, such as the Pretoria Salt Pan have been dated using insolation arguments, and if the climate model proposed by Hahn et al. is correct, then the basis for the age model of the Pretoria Salt Pan is flawed. The authors have presented their data in supplementary tables, which is going to be very useful for comparing this dataset to others.

We would like to thank this reviewer for their constructive comments and we have replied to each comment individually below. The reviewer comments are in black and our responses in blue.

Review

The critique of the manuscript is in the form of substantive clarifications, minor issues, and typos.

Substantive clarification

One of the most important contributions of this manuscript is the model for synoptic shifts in the region during glacial periods, and in particular its effect on the SIOCZ. Clarification is required of exactly what the SIOCZ is. Comparing figure 1 with figure 5 it would appear as if the ITCZ and the SIOCZ are synonymous, but the text line 330 couples the SIOCZ to TTT, and line 334-336 clearly decouples the SIOCZ and the ITCZ. Certainly in the modern system the SIOCZ and TTT systems are distinct from the ITCZ. Since figure 1 includes the ITCZ, the SHW and the circulation patterns, it should also indicate the modern SIOCZ and TTT systems.

Yes, the relationship between SIOCZ and ITCZ needs to be better explained. We have added the modern SIOCZ and TTT systems to figure 1. Furthermore, we have detailed the relationship between the ITCZ and the SIOCZ in the text (section: “regional setting” lines 113-116:). Following the request by reviewer 2 we have also detailed the regional climate section a bit more also adding fig 1b and c.

The text that describes the contribution of ITCZ summer rainfall in the relevant catchment during glacials (lines 343-347) but in figure 5 the source of summer rainfall is indicated as the SIOCZ. This needs to be reconciled.

Yes; this an error in Figure 5 that has been corrected now. What was initially marked as “SIOCZ” was actually the ITCZ, this has been changed and the SIOCZ position is now indicated as well. Concerning the text on glacial periods: the southward shifts of the ITCZ (due to a more southerly position of the thermal equator) would entail southward shifts of the SIOCZ. It is unclear if this, or a direct ITCZ influence caused humid phases during Heinrich events. This has been added to the section “3.2.1.2. Hydrology over glacial-interglacial transitions” lines 359-363.

In the discussion of SST (line 237-240), comparison is between core top SST with modern SST data from Fallet et al. 2012 in order to defend a seasonal interpretation of the G. ruber Mg/Ca values. This argument is flawed in many ways. First, the uppermost Mg/Ca result from the sediment core is from 40.5cm depth in the core, which is approximately 5 000 years old according to the radiocarbon dates. This cannot be compared with the “modern” data from Fallet et al. (2012) which is approximately 1000 years old. Indeed the age of the youngest Mg/Ca SST value prevents any verification against modern SST values.

Second, the satellite data for SST in the Mozambique Channel presented by Fallet et al. (2012), and also the SST data based on Locarnini et al. (2013) presented in figure 1 show a strong thermal gradient in the Mozambique Channel. Correlating the Mg/Ca 27°C SST temperature for the “top” of this sediment core does not take in to account this southward cooling gradient. The inshore location of this core in the Delagoa Bight also implies stronger coastal influences that is associated with warmer SST (also based on data in Fallet et al. 2012). The seasonality of this SST reconstruction is not central to the development of the climate forcing argument, but it will need to be tempered as a stand-alone interpretation.

This is true; we have removed the comparison with the core top sample. We have left the reference to Wang et al., (2013) saying that the authors “suggest that $U^{K'}_{37}$ SST reflects warm season SST mediated by changes in the Atlantic, whereas the *G. ruber* Mg/Ca SST indicator used in this study records cold season SST mediated by climate changes in the southern hemisphere.”. Although we agree with the reviewer that the seasonality of the SSTs is not central to our argumentation, but we thought we would leave this suggestion as information for the reader.

Minor issues

The age model for the core is very clearly argued, and is sufficiently convincing for the broad-brush stroke assessment of palaeoenvironmental proxies, but close scrutiny of the radiocarbon dates indicates some heterogeneity in deposition rates. Rapid deposition is indicated between 300cm and 200 cm, and also between 150cm and 100cm (although the Bacon model produces a parsimonious smoothing that downplays the date from 102cm). Placing more emphasis on the outlier date leads to the possibility of very slow deposition in the 15 000-6 000 year range, and also in the 32 000 -25 000 year range. The Bacon model needs more input data to verify this level of heterogeneity, and so there is possibly an underestimation of the error in the age model around these periods of slow deposition. Similarly only 2 δ18O tie points are used in the chronology for the oldest 60 000 year part of the record. This clearly cannot capture heterogeneity in the deposition rate, and again the age model error estimates are probably too small.

We agree with the reviewer and have added this to the age model error estimation in section “2.3. age model” See lines 156-171

The suite of proxies that reflect wet and dry conditions in the catchment are reported to change in concert with one another, and this is clear in a relative sense but not in an absolute sense. Scrutiny of figure4, for example, shows clear oscillations in values that are synchronized between proxies, but within proxies these oscillations are really most apparent because of the contrasting peaks and trough values that are immediately older or younger. The absolute values do not hold up to the wet/dry assignations. The K/Al and Ca/Fe ratios in the wet period around 82 000 years ago, for example, have very similar values to the arid values at around 46 000 and 52 000 years ago, and so the absolute values are seemingly not important. Some discussion of the relative nature of these proxies should be presented.

We agree and have added this idea to the end of the "3.1.3 Precipitation indicators" section lines 313-328

The interpretation of the δ13C record invokes a framework presented by Dupont et al. (2011) in which woodlands and forests with grasslands in the interior during interglacials is contrasted with rivers fringed with gallery forests & sedges in glacials. This scenario may account for the observed trends in the record, but it is a very imprecise science. The entire 13Cvariability noted in the 100000year record all falls very in the range of C3 plants, and even the maximum values that are interpreted as an increased C4 plant community still fall in the C3 range. As much as this represents an integrated C3/C4 environmental shift, it could just as well represent a xeric/mesic environment with exactly the same C3 plant communities.

The entire 13C variability noted in the 100000year record (-26 to -23‰) is indeed relatively small and close to the 13C values expected for alkanes from C4 plants (around -20‰, depleted up to -25‰; compare dataset 'all Africa for C31 alkane in Garcin et al., 2014). However, there are only few C4 plants with 13C values as depleted as -25‰ so the variability observed in our record must be caused by variable contributions from C3 plants. Other indications for this:

A) if shifts in d13C were only dependent on shifts in hydrology (xeric to mesic) then there should be a correlation between the d13C and dD variability (dD is a reliable indicator of rainfall amount). Since this is not the case (R^2=0.15), we assume that a further factor (i.e. shift in vegetation biomes) drives d13C values.
B) In order to definitely answer the question on what is driving d13C values (vegetation change vs hydrological change) we would need pollen data for our core. These would give us reliable infor-mation on the actual vegetation changes in the catchment. Unfortunately, palynological analysis will not be done at the site so that we have to refer to closest neighboring palynological dataset. The core studied by Dupont et al. 2011 is in the direct vicinity of our site; and the catchments from which the material is sourced are adjacent. Although the comparison of our record with the Dupont et al. 2011 pollen data is imprecise, as the reviewer correctly remarks, it is the best we can currently do. In section "3.2.1.1. Sea surface temperatures and vegetation" we describe in detail ow our catchment areas differ from that of the Limpopo and the consequences this has on vegetation signals. We find that in order to answer the question concerning the drivers of d13C values (vegetation change vs hydrological change) the comparison with the Dupont et al. 2011 record is helpful in the sense that Dupont et al. 2011 document large changes in the vegetation biomes of the Limpopo catchment over glacial-interglacial transitions. If this was the case in the Limpopo catchment it is likely that similarly shifts in vegetation biomes took place in the adjacent catchments.

We have tried to bring these arguments forth more clearly in the rewritten version of the section lines250-285 Finally we note that we may not be able to pinpoint exactly what caused shifts in the d13C record, but either mechanism would translate to heavier values during more arid conditions and lighter values during wet conditions. We have tried to underline this in the discussion of the proxies (lines 250-285).

The role of sedges in the δ13C record interpretation also needs closer consideration. Stock et al. (2004AustralEcology) suggest that 14% of sedges are C4 in winter rainfall areas and 67% are C4 in summer rainfall areas. Seasonality of rainfall is clearly a controlling factor in the C3/C4 pathways for sedges, but the interpretation of the sediment core δ13C record seems to hint that they are all C4.

This is an interesting point and it is possible that variations in winter and summer rainfall may affect the vegetation and thus our d13C record. However, without downcore palynology with phytolith analysis to distinguish between C4 and C3 grasses, it is not possible to go into such detail. Looking at the data, it also seems unlikely that such an effect took place: if the shifts from winter to summer rain were driving d13C values, our data would suggest that glacial periods have increased summer rains, whereas interglacials had a more winter rainfall regime. This is highly unlikely; we know that modern (interglacial) climate is a summer rainfall regime and that only during glacials has the winter rainfall zone been inferred to shift northward. This interpretation of the d13C signal would therefore be very difficult to consolidate with what we know about the regional climate.

The association between the wet/dry cycles portrayed in the core, and Heinrich Events and the Antarctic Isotope Maxima events is important in resolving the underlying climate forcing. It should be noted that HS4 is the negative excursion in the NGRIP δ18O record around 37 000 years ago (possibly older as it is portrayed in figure 4 – maybe 38 000 -40 000 years ago). It is associated with a dry interval (red shading in figure 4) but the text associates it with a wet period (lines 395-399). Overall the association between wet/dry phases in the core proxies and the AIM and HS data is dependent on the errors in the age model, which was argued to be underestimated, but still comprises several thousand years in the older portion of the core.

Unfortunately, this is true, and we have pointed this out in the manuscript at the end of the section "*3.2.2.2. During MIS 4-2 glacial conditions*" discussion: lines 438-441

It would be useful for those who will undoubtedly make use of this record in their research if the supplementary tables include a model age assignation, and not just the sample depth in the core.

Good idea. This has been done.

Figures and figure captions Figure 1: Please depict the SIOCZ and TTT because it is relevant in the discussion.

Upon suggestion by the 2nd reviewer, We have added a new sub figure to fig1 that now shows in detail the modern climate system including SIOCZ and TTT…

Wonderkrater is depicted in the wrong place (somewhere in Zimbabwe). In reality it is well within the Limpopo catchment.

Yes! Changed…

Figure 2: The caption mentions "LR04" twice in a redundant manner.

Done

Figure 3: This caption needs to be rewritten. It is difficult to decipher what is being referred to because of a random sprinkling of right parentheses and colons.

We have rewritten the caption for better understanding

Figure 4: This caption attributes blue or green shading as wet, "while wet phases are marked in red or yellow". Presumably one of these is dry.

Yes, this has been corrected, the latter is dry.

What is described as blue appears purple – this may be a personal problem, but possibly re-consider the colour that is used.

we have opted for a pure blue now…

The text “related to low pressure cells” is correct but confusing in its detail and should be revised.

Yes, this detail is confusing in the text, we have removed it.

Typos Line 66: winterly should be winter

ok

Line 67, 114-115, 330, 334-336: Define the SIOCZ, is this the same as TTT (in fig 5 it seem synonymous with the southern extent of the ITCZ, but line 330 couples it to TTT, and line 334-336 clearly decouples the SIOCZ and the ITCZ)

Yes, the relationship between SIOCZ, TTT and ITCZ needs to be better explained. The SIOCZ is a southward extension of the ITCZ. The tropical temperate troughs (TTTs) form at the SIOCZ. We have added the modern ITCZ, SIOCZ and TTT systems to new subfigure of figure 1. Furthermore, we have detailed the relationship between the ITCZ and the SIOCZ and TTT in the text (section: “regional setting” lines 113-116).

and also put it on to fig 1 as it comes up repeatedly

ok

Line 76: Re introduces the SIOCZ acronym

This has been removed….

Line 201: permil, but on line 139 per mil. Please be consistent throughout the text

ok

Line 244: Fig. 1a should be Fig. 3a

Yes!

Second Reviewer Comments and Rebutall

Based on a sediment core from Delagoa Bight offshore southeastern Africa, Hahn and co-authors present a new multi-proxy reconstruction of the continental climate for the last 100,000 years. The new record has high potential to improve our understanding how continental wetness has varied in response to latitudinal shifts in the westerlies and South Indian Ocean convergence zone. The data are certainly of very good quality and the new record has great potential, which, however, is not fully exploited in the current version of the manuscript. In my view there are several major shortcomings (see comments below) and major revisions are therefore required before the manuscript can be accepted for publication in CoP. I would like to emphasize that I will focus only on major issues at this stage of the review process:

We would like to thank this reviewer for their constructive comments and we have replied to each comment individually below. The reviewer comments are in black and our responses in blue.

âA˘ c The study site appears to be ´ ideally situated to record displacements of the westerlies and the South Indian convergence Zone. Unfortunately, the authors do not really present a more detailed figure of the present-day atmospheric circulation patterns, which would help the readers to understand the discussion better. Basically, more detailed information on the atmospheric dynamics and according figures are required, such as the one presented by Charlotte Miller an co-authors in a previously published article in Climate of the Past (Figure 1 in Miller, C., et al. (2019). "Late Quaternary climate variability at Mfabeni peatland, eastern South Africa." Climate of the Past 15(3): 1153-1170.

We have added sub figures 1 b and 1c as well as a more detailed description of the regional atmospheric dynamics in the “regional settings “section (lines 114-128).

âA˘ c Although ´ multiple proxies were measured, there is rather little and very rudimentary information on their paleoclimatic significance and potential uncertainties and limitations are not discussed. For instance, the precipitation indicators δD, K/Al, Ca/Fe and red/blue ratios are only very briefly presented in paragraph 3.1.3. All proxies depend to varying extents on precipitation, erosion and fluvial transport, whereas these factors do not necessarily vary in concert. For instance, erosion is not always directly linked to the amount of precipitation and vegetation density is often an additional and more important factor for erosion rates. Erosion rates can also increase substantially at times of rapid climatic and associated vegetation changes. Because the relationship between precipitation and erosion (and riverine transport) is not linear. I would like to see a more critical discussion about the strength and weaknesses of the proxies.

dD is indeed our only “real” precipitation indicator whereas the remaining proxies reflect erosion, fluvial transport and the weathering of the transported material. All of which are indeed liable to have a non-linear relationship with precipitation amount. However, seeing that the four proxies (mostly) correlate in our record, this does not seem to be the case for the most part of our record. We have added these considerations to the paragraph in question (3.1.3). Lines 317-329

âA˘ c Some ´ of the authors have worked for a long time in this region and published multiple articles on past climate variability in this region. It is therefore quite surprising that there are no attempts to incorporate other continental records from South Africa more effectively into this study. Some of the records are mentioned in the text but not displayed in a figure.

We are unsure as to which records the reviewer is referring to. Records that span the time frame in question and at the same time have a resolution that is comparable to that of core GeoB20616 are very rare in the region. The only available records are located much further north and thus out of the influence of the climatic systems we are describing.

âA˘ c The major precipitation indicators are presented in Figure 4, together with ´ ice core records from both poles. The authors try to mark wet periods associated with different atmospheric circulation regimes. However, it remains absolutely enigmatic which scientific criteria were actually used to determine these periods. The width of the color-coded bars seems to be rather arbitrary as, for instance, indicated by the width of the green bar during MIS 5, which do not really match the minima in the δD and K/Al records. The authors must explain in close detail which criteria were used to determine the different climatic phases. Furthermore, what is actually happening during the white intervals?

We have detailed that the definition of the different climatic phases is mainly based on the dD values, this is our most direct precipitation proxy and the red/blue ratios as well as elemental ratios serve mainly as supportive information, underlining the reliability of our paleo-rainfall reconstruction. We definitely needed to clarify this and have added an according section to lines 323-328.

Concerning the white phases; these we consider as transitional periods, as is now marked in the caption of Fig 4.

âA˘ c Figure 5 is a basic conceptual model, but it also highlights ´ the problem of this study as other records were not really used to support this basic model.

We understand that the reviewer would like to see a more thorough comparison of our record with other regional continental records. However, there are few/none records that span the time frame in question and at the same time have a resolution that is comparable to that of core GeoB20616. The only available records are located much further north and thus out of the influence of the climatic systems we are describing.

The authors suggest that the major changes on glacial interglacial time scales are related to latitudinal shifts of atmospheric boundaries and westerlies. Are there no zonal shifts in the moisture transport?

This is an interesting point; however we find no evidence for zonal shifts in the moisture transport. There is a divide (CAB, Congo Air boundary) between Atlantic and Indian Ocean moisture but it is located very close to the Atlantic coast (the Atlantic moisture simply does not make it to the interior due to the Benguela upwelling). Only under conditions without Benguela upwelling (i.e. before the Miocene essentially) it would have been possible that the CAB was located further east and Atlantic moisture would make it to the eastern coast of SA. Under the modern climate (upwelling, atmosphere) system, even under glacial state, it is simply not possible.

Furthermore, I would like to see a third figure showing the conceptual model for the present-day situation.

The present day situation would correspond to the “interglacial state”. We have marked this accordingly in the caption.

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