Author's response to Anonymous Referee #1

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.

This paper was extremely interesting to read and presents a new record from eastern South Africa spanning from the LGM to the present. I have a few minor comments:

Line 40 - 52: Instead of starting by discussing the uncertainties about regional climate, first motivate the study with why we should care about the region. What challenges to water availability might future climate change pose, and how can paleoclimate help us address those uncertainties?

- We will add an additional paragraph explaining why the region is important to the beginning of the introduction.

Line 61-65: There needs to be a citation for the evidence the claim that the Indian Ocean Walker circulation weakened in response to glacial forcings - see DiNezio and Tierney, 2014; DiNezio et al., 2018. In any case, your site may be too far south to be directly influenced by Walker circulation changes in the Indian Ocean.

- We will add the reference DiNezio et al. (2018) to this sentence.
- According to the conceptual models developed by Tyson (1986) and then by Cockcroft et al. (1987), a weakened Walker Circulation, having its ascending limb further east, reduces tropical-temperate interactions and results in a northward shift of westerly storm tracks and then dry conditions within the SRZ. Thus, we think that changes in the Walker circulation could influence the climate of the region.

I find the interpretation of the leaf wax data convincing - the d13C shifts are quite small and are unlikely to majorly influence the dD signature as a result of major shifts in apparent fractionation. The changes in dD are consistent with an amount effect and/or changes in evapotranspiration. It is interesting, however, that at least in Figure 4 it looks like the most modern dD value seems to look similar to LGM values. Can the authors comment on this?

- Yes, it is interesting that modern day δD_{wax} values are similar to those within the glacial period. Reduced precipitation over the last *c.* 2 ka is also evident in the precipitation stack (Chevalier and Chase, 2015), although values do not quite reach glacial values. Nevertheless, the aridity stack (Chevalier and Chase, 2016) indicates extremely arid conditions during the last few thousand years. Although this period is not specifically commented on within this paper, they stress the importance of temperature in controlling aridity. We speculate that the conditions between LGM and modern were not similar in terms of precipitation amount, but that temperature may have played a role in controlling the aridity, resulting in similar δD_{wax} values. Mean annual temperatures for the last *c.* 2 kyr cal BP were *c.* 2 degs C higher compared with those reconstructed for the LGM (see Chevalier and Chase, 2015). Thus, higher mean annual temperatures over the last *c.* 2 kyr cal BP would have increased ET, resulting in less 'effective precipitation' and drier conditions, despite possibly increased rainfall amount. Whereas during the LGM, lower temperatures would have reduced ET leading to apparently more humid conditions, despite actually reduced rainfall amount. This will now be discussed within the manuscript.

Figure 1 would be improved if it showed instead regional currents and winds for winter vs. summer and seasonal rainfall totals in two panels.

- We will change figure 1 as requested, it will improve and further aid the discussion and interpretation section. It will now show the different climatological features in summer and winter across the region as well as the rainfall zones.

I would like a more detailed discussion of all the regional time series included in Figure 4 in the discussion, especially on the stacks of regional aridity and precipitation.

- We will add more detailed discussion regarding the regional stacks and also regarding the pollen data (as requested from reviewer #2) into the discussion section (6.2).

Line 425 - 436: It actually appears to me that there is a reasonably good correspondence between SST and the dD record at your site during the deglaciation and in the early Holocene - there is just a lack of correspondence after 5 ka and during the LGM itself. It is therefore possible that SSTs played a key role in the response during the deglaciation itself.

- We agree and will better discuss SSTs as a driver of change within section 6.3. In addition, as requested by reviewer #2 we will use the SST record of Sonzogni et al. (1998), which actually shows a better correspondence with our data. Now we can see that lower SSTs during the last *c*. 5 ka may have actually played a role in driving aridity at Mfabeni.

Something that might be useful to consider as well - how might the westerlies, as a result of the wind-evaporation-SST feedback and/or changes in Ekman transport influence SSTs and local ocean dynamics? It is possible that there is a link between the two.

- Firstly, as a request of reviewer #2, we will remove mostly all reference to the westerlies providing more wind and resulting in increased ET. As the other reviewer pointed out, this hypothesis was problematic and we do not have a good independent proxy for wind strength in this region.
- We do not think that this discussion is within the scope of this paper. Nevertheless, high SSTs within the South Atlantic Ocean have previously been related to a poleward shift of the westerlies (see Walker, 1989) and moist conditions over the continent (Tyson, 1986).

As a broader point, calling the northeasterly flow into the region the ITCZ is probably overstated -the term means something very specific - a zonal band of rainfall most accurately applied to ocean regions (i.e. the marine ITCZ). It would be more accurate to characterize it as northeasterly flow that brings tropical moisture.

- We agree and will change all mention of the ITCZ to ‘tropical rainbelt’ (or tropical easterlies) (e.g. Nicholson, 2008).

References cited in this rebuttal:

Chevalier, M., Chase, B.M.: Southeast African records reveal a coherent shift from high- to low-latitude forcing mechanisms along the east African margin across last glacial–interglacial transition, Quaternary Sci.Rev., 125, 117-130, 2015.

Chevalier, M., Chase, B.M.: Determining the drivers of long-term aridity variability: a southern African case study. J. Quaternary Sci., 31, 143-151, 2016.

Cockcroft, M.J., Wilkinson, M.J., Tyson, P.D.: The application of a present-day climatic model to the late quaternary in southern Africa. Clim. Change, 10, 161-181, 1987.

DiNezio, P. N., Tierney, J. E., Otto-Bliesner, B. L., Timmermann, A., Bhattacharya, T., Rosenbloom, N., and Brady, E.: Glacial changes in tropical climate amplified by the Indian Ocean, Science advances, 4, 2018.

Nicholson, S. E.: The intensity, location and structure of the tropical rainbelt over west Africa as factors in interannual variability, International Journal of Climatology, 28, 1775-1785, 2008.

Sonzogni, C., Bard, E., and Rostek, F.: Tropical sea-surface temperatures during the Last Glacial Period: A view based on alkenones in Indian Ocean sediments, Quaternary Science Reviews, 17, 1185-1201, 1998.

Tyson, P. D.: Climatic Change and Variability in Southern Africa, Oxford University Press, Cape Town, 1986.

Walker, N. D.: Sea surface temperature-rainfall relationships and associated ocean-atmosphere coupling mechanisms in the southern African region., PhD, University of Cape Town, 1989.

Author's response to Anonymous Referee #2

We would like to thank this reviewer for their thorough review and useful comments. We reply to each comment individually below. The reviewer comments are in black and our responses are in blue.

This paper presents interesting new data from the Mfabeni wetland in SE Africa, a site that has received considerable attention in recent years. The data presented here is, in my mind, critical for the fuller understanding of how the other proxy records relate to past climate change.

- We appreciate the acknowledgement from the reviewer about the relevance and importance of our work, and thank him/her for their comments, which we address below. We will address all comments and will make substantial amendments to the manuscript as such.

While I find the data to be extremely interesting, the text itself becomes slightly contorted at times and I think there are several aspects that could be reconsidered, or at least clarified.

The first aspect is a more thorough integration of the pollen data from the site. Much of the text relates to vegetation change, but the pollen data is only included in an extremely rough summary form in the final figure. Significantly more effort should go into including the pollen data in comparative diagrams with the original data presented here. This may require some work (creating comparable chronologies, etc.) but it is necessary.

- We will include pollen % data of key taxa from the original palynological dataset (Finch and Hill, 2008) to our paper (in Fig. 6). In order to compare the datasets we have produced a new age model for the old palynological data (from core; MF1). We will add this new age model to the supplementary material of our paper (S2). Throughout the paper, we will discuss our results in light of the palynological results.

The authors also appear to base their palaeoclimatic interpretations on the supposition that the amplitude of dD change (~53‰ cannot be explained by changes in source water dD, and must be (even primarily be) the result of changes in evapotranspiration (only explanation for observed changes in Fig 4c). This contrasts with most authors' interpretations, which acknowledge the potential role of ET on leaf wax dD, but focus more on precipitation amount/intensity (and the general observation that dDwax and mean annual precipitation are strongly correlated in the tropics (e.g. Sachse et al., 2012 and references therein)).

- The peatland today is extremely sensitive to changes to changes in both precipitation and evaporation, with the modern water balance dominated by the interplay between evapotranspiration (ET; 1035 mm) and precipitation (1053 mm). Consequently, at Mfabeni, any changes in both ET and precipitation have great potential to influence the δD_{wax} values. However, it is difficult to disentangle whether changes in δD_{wax} are due to changes in ET or precipitation amount, and indeed the isotopic variability within the Mfabeni record could be a result of either. Nevertheless, both high (low) ET and low (high) precipitation amount lead to δD_{wax} enrichment (depletion) and thus imply generally drier (wetter) climatic conditions. We will amend the label of Fig. 4c to 'drier and wetter' and will clarify the δD_{wax} interpretation throughout the manuscript.

In the African tropics, dDwax records from lake and marine sediments exhibit ranges of ~35‰ 55‰ usually with the lowest values occurring during the last glacial period/Last Glacial Maximum. Considering changes in temperature alone, cooler conditions at this time would have lowered ET, rather than raising it.

- We agree with the reviewer that lower temperatures during the last glacial period would result in lower ET. For example using the equation of Kosa (2009), a 2 degree annual temperature increase from the last glacial period (*c*. 24°C) to the present day (average temperatures *c*. 26°C) would equate to a change in mean actual evapotranspiration of 0.72 mm/day. Nevertheless, numerous additional factors control evapotranspiration, not only temperature, such as the amount of incoming solar radiation, humidity and wind speed. We will amend the 'Interpretation of the proxy signals' section to make clearer the drivers of Mfabeni δD_{wax}.

At Mfabeni, the authors seem to conclude that it is wind strength that drives what is inferred to be increased glacial-age ET, thus essentially interpreting their dDwax record as a wind strength proxy.

- We agree with the reviewer and will change this interpretation throughout the manuscript. Previously, in places, too much focus was on ET and how this could be influenced by wind. δD_{wax} values at Mfabeni are driven by changes in **both** ET and precipitation amount. We will now discuss more thoroughly the climatic mechanisms driving ET and precipitation amount, especially focussing on the latitudinal position of the westerlies, the South African high-pressure cell and the South Indian Ocean Convergence Zone (SICZ).

Considering the remarkable similarities between their dDwax record and pollen-based precipitation reconstructions from the region this seems to become unnecessarily contorted through the discussion and conclusions. Acknowledging that ET can certainly have an impact on dDwax values, is it not still more parsimonious to interpret the Mfabeni record as primarily reflecting rainfall amount/intensity? Or is the suggestion that the other dDwax records from tropical Africa be revisited, and mechanisms for increased glacial-age ET at each site be found?

- With the δD_{wax} record similar to both the pollen-based precipitation and aridity stacks of Chevalier and Chase, 2015 & 2016, and as it is impossible to disentangle the effects of precipitation and ET amount on Mfabeni δD_{wax} values, we suggest that the pollen-based precipitation stacks may also include an element of ET variability. We do not suggest that other δD_{wax} records from tropical Africa be revisited, but as the system at Mfabeni is extremely sensitive to changes in ET we should not exclude the role that ET plays in the water balance at this site.

The authors also focus on the southern westerlies as being the/one of the primary drivers of the changes observed in their record. This strikes me as an odd perspective, as the westerlies are not implicated as being a significant moisture-bearing system in the region. Rather, the authors focus on shifts in the westerlies as somehow (not well- described) inhibiting precipitation and probably increasing wind strength (no reliable data provided (grain size data from the sediment core don't satisfy this requirement) at the site. To follow their interpretive logic further, they strongly associate the position of the westerlies with Antarctic sea ice extent, and thus that changes in Mfabeni hydroclimate are primarily driven by changes in sea ice. Of course the myriad elements of the Earth's system are inter-related, but this focus on the mid- to high latitudes without more detailed description of tropical dynamics and those systems that are responsible for precipitation at the site seems not to be the clearest, most straight-forward way of describing the changes observed.

- We aim to amend our discussion and interpretation as follows. We will include more details on how the position of the southern westerlies (and the South African high-pressure cell and the SICZ) may control precipitation amount/ET at Mfabeni. For example, rainfall along the eastern coast of South Africa is today strongly correlated to the latitudinal position of the subtropical high-pressure system (e.g. Dyson and van Heerden, 2002). Following the modelling results of Vigaud et al. (2009), we now attribute a northward shift of the South African high-pressure cell to reduced moisture in eastern South Africa. When the cell is shifted southward, during the summer, the tropical easterlies are able to penetrate further inland, resulting in more continental moisture availability. Conversely, when the cell is shifted northward, during the winter, monsoonal circulation south of 25°S is weakened, creating a deficit in moisture advection from the ocean to the continent. The more northerly location of the high pressure cell (which we propose occurs during the LGM) reduces the ability for the tropical easterlies to penetrate inland, limiting the advection of moisture over the continent, resulting in arid conditions at Mfabeni. We will also discuss the role of the SICZ and the rain-bearing cloud band associated with tropical temperate trough (TTT) development. A northward migration of the westerlies is associated with a weaker South Indian Anticyclone, which results in a north-eastward shift of the SICZ, and thus lower precipitation at Mfabeni. We will also amend the regional settings section, including more information about how the westerlies, the subtropical high-pressure cell and the SICZ control climate across South Africa.

TITLE: This seems a grand title for a paper describing a single site. It is not sufficiently synthetic to make this claim.

- We agree and will change the title to 'Late Quaternary climate variability at Mfabeni Peatland, eastern South Africa'.

ABSTRACT: Lines 23-24: Saying that the conditions ARE a consequence of low SSTs and westerlies/sea-ice is a pretty strong statement. It suggests that the authors are SURE this is the case, but considering the quality of the sea-ice and westerlies proxies this does not seem like a realistic claim. I suggest dialling back and being more circumspect.

- We agree and will amend this sentence.

Lines 32-34: I imagine this will be explained more fully in the text, but this statement here is unsubstantiated in the sense that a mechanism for the influence of the westerlies at a site so far north. There should be at least some further clue given here as to what the thought process is.

- It is difficult to go into much detail in the abstract but we will add a little more detail to include the climatic process behind how changes in the latitudinal position of the westerlies (and the high-pressure cell and SICZ) may result in dry conditions at Mfabeni.

INTRODUCTION: Line 40: "Last glacial" what? Period? Maximum? In any case, 21 ka is neither. MIS 2 ended at 11.7 ka, and the Last Glacial Maximum ended at 19 ka (Clark et al., 2009). This should probably be stated differently. Also, (pedantically) "BP" should be removed here as it refers specifically to radiocarbon chronologies and indicates before AD 1950. "ka" alone is more appropriate.

- We agree and will change the sentence to 'Changes in vegetation, precipitation and temperature from the beginning of the Last Glacial Maximum (c. 26.5 ka; Clark et al., 2009) to present-day, in eastern South Africa are poorly constrained.'

Line 45: Define LGM here, as first instance in body of text. The authors should also define what it is. I suggest the 19-26.5 ka Clark et al., 2009 definition (as the authors use in Line 268).

- We will first define the LGM here. The duration of the LGM will be defined slightly later, following the definition of Clark et al. (2009).

Line 48: The authors may want to consider/add Chevalier et al., 2017, as this paper deals directly with this question.

- We will add the reference Chevalier et al., 2017.

Line 61: Here the authors may want to consider Otto-Bliesner et al., 2014 and Chevalier and Chase, 2015, as they discuss the relative importance of these and other drivers specifically, providing some detail on precisely how they operate in SE Africa.

- We will add reference to the greenhouse gas climate driving mechanism from Otto-Bliesner et al., 2014. We will also add the Chevalier and Chase, 2015 reference to the parts of the paragraph here describing insolation and SST forcing.

Line 62: The ITCZ is really only clearly defined over the oceans. Have a look at the works of Nicholson et al., but in any case Mfabeni is well to the south, and easterly flow better describes the moisture-bearing vector.

- We agree with the reviewer and will change the term ITCZ to 'tropical rainbelt'.

Lines 63-67: Certainly the component parts of the Earth's climate system are linked, but I wonder from these statements and the abstract how the relationships are described in this paper. Here, for instance, it is said that that the winter rainfall zone gets wetter because the ascending limb of the Walker Circulation shifts eastward. It may well have shifted eastward, and it doing so may have either allowed or been the result of changes in circulation systems that brought more moisture to the winter rainfall zone, but the position of the Walker Circulation over the Indian Ocean does not directly impact the winter rainfall zone.

- We agree with the reviewer and will remove reference to the winter rainfall zone here. We will change the sentence to 'This possibly resulted in an eastward displacement of the coastal cloud band and thus a drier summer rainfall zone (SRZ; Tyson, 1999).' We will also changed the reference to Tyson (1999).

Said differently, these dynamics are most clearly expressed when the primary moisture-bearing system is included in the discussion. Here, the authors account in part for the SRZ, but not at all for the WRZ as the authors do not include the westerlies and associated storm track. I suggest the authors take a few lines to describe the regional climate systems, and define things like the WRZ and SRZ more clearly so that the reader can have a clearer spatial understanding of how it all fits together. I see the authors do this later, but by then it is a bit late. Rework these sections to create a logical development of information.

- We will amend the regional setting section thoroughly giving a better overview of climate systems. We will start off broad-scale, focussing on the whole of SA, and then in the following paragraphs focus on Mfabeni and eastern South Africa. This will create a better flow and assist later in the interpretation. Furthermore, we will change figure 1 (as requested by reviewer #1) to include regional climate dynamics and differences between the summer and winter seasons.

Line 65: "OVER the Indian Ocean"

- We will correct to 'over' the Indian Ocean.

Line 71: These references are only Holocene. Considering that this paper has greater scope surely some of the seminal works like van Zinderen Bakker 1976, Cowling et al., 1999, etc. should be mentioned? The papers mentioned hardly define the concept.

- We will remove the two previous references and replace them with van Zinderen Bakker 1976 and Stuut et al., 2004.

Line 73-74: Chase et al., 2017 isn't a very apt reference for this statement. Chevalier and Chase., 2015, perhaps? This paper addresses this region/topic more specifically, and comes to the stated conclusions, at least to some extent.

- We agree, and will remove the reference Chase et al., 2017 and replace it with the reference Chevalier and Chase., 2015.

Line 75: "last Glacial" what? Period? Maximum? Glacial shouldn't be capitalised here, unless it is for the LGM.

- We will remove the capital and add 'period'.

Lines 75-78: Glacial shouldn't be capitalise here.

- Again we will remove the capital.

Line 80: Again, and here and elsewhere, the ITCZ isn't the correct term here. African or tropical rainbelt is preferable (again, see the papers of Nicholson et al., esp. 2009).

- We agree and will replace ITCZ in all cases throughout the paper with 'tropical rainbelt'.

Line 84: Chevalier and Chase should be removed as that is neither a marine nor speleothem study. Holmgren et al., 2003 is a better speleothem reference, as it extends to the 25 ka.

- We will remove the Chevalier and Chase and Holmgren et al., 1996 and add instead Holmgren et al., 2003 as a reference here.

REGIONAL SETTING: Line 93: CAN be divided. That is just one way of looking at it, and it is not without its shortcomings.

- We agree and will change this to 'can be divided'.

Line 97: tropical-temperate troughs are only one of the composite synoptic systems to consider. Chase et al., 2017 lumped the ensemble as "TTIs", tropical-temperate interactions, but look at Tyson, 1986 and Tyson and Preston-Whyte, 2000 for more specific details.

- Firstly, we will discuss tropical temperate troughs (TTT) more thoroughly throughout the paper, beginning in the regional settings section and then later in light of how changes in the SICZ result in changes in the location of TTT formation. We appreciate that TTTs are lumped together as TTIs in Chase et al. (2017) but as the majority of literature (both old and new) refer to TTTs we would like to keep our terminology consistent with those papers. We will also add a paragraph later in section 6.3 where we will discuss TTTs and millennial-scale variability over the G-IG transition.

It might also be worth mentioning that suggestions have been made that at least two subdivisions of the SRZ have been made, with, according to Chevalier and Chase, 2015, the northern and central/eastern summer rainfall zones operating in substantially different fashions.

- We agree and will include this within the 'regional setting' section.

Line 100: "temporal frontal systems"? Temperate rather? And Tyson, 1986 should perhaps be referenced here?

- We will change temporal to temperate, and reference Tyson, 1986 and Tyson and Preston-Whyte, 2000.

Line 101: "Sandwiched" doesn't suggest the transitional nature of the YRZ very well. For me it suggests solidity for an extremely ephemeral region. Maybe rephrase?

- We agree and will re-phrase this sentence.

Line 104: Probably don't need a reference for the Namib Desert.

- We agree, and will remove this reference.

Line 110: Here and elsewhere in the text, "ka BP" does not indicate if the ages were calibrated. If this is based on a radiocarbon chronology ("BP"), it should be "cal kyr BP" (ka is for an age, kyr is for a span of time). If the chronology includes OSL ages (that do not require calibration) "ka" is appropriate for mixed or non-radiocarbon chronologies.

- We will indicate where ages are calibrated with 'cal'.
- We will use ka for ages, and kyr for durations.

Line 121: The paper referenced focusses on wind, but strong winds do not bring precipitation, as could be inferred from this sentence. Reword or reconsider in terms of the circulation dynamics that the passage of cold fronts induce?

- We will change this sentence and remove the wind focus and the Kruger reference. The sentence will now read 'Occasional rainfall during the winter months at Mfabeni is associated with the passage of cold fronts, which develop in the western Atlantic and move across southern Africa (Grab and Simpson, 2000).'

Lines 105-144: These paragraphs mix around elements of topography/geology and climate. Perhaps disentangle? With topopgraphy and geology coming first (as the backdrop) and then the second paragraph looking at climate?

- We agree and will disentangle and re-structure this section to improve the flow. The section will now follow on from the regional climate settings to Mfabeni local climate. The next paragraph will focus on the geological and morphological settings and how this effects groundwater. Finally the section will end with a paragraph on the sensitivity of the modern water balance at Mfabeni and the suitability of Mfabeni for palaeoenvironmental reconstruction.

Lines 177-179: Maybe specify where C3 vs C4 grasses generally grow?

- This will now be explained in detail in the section '6.1 interpretation of the proxy signals'.

DISCUSSION: Line 307: C3 grasses are found in the WRZ (and YRZ: : :) AND at higher elevations. As mentioned previously, some clue as to the climatic mechanisms that drive C3 vs C4 grass distributions would be very helpful.

- We will add an extra paragraph here explaining the environmental conditions favouring the expansion of C_3/C_4 distribution.

Line 314: "higher" d13C values would be clearer.

- We will change this to 'higher'.

Line 315: How would colder conditions lead to an expansion of C4 grassland? The authors are walking a fine line here, presumably citing frost-intolerance of arboreal and shrub taxa, but not to the point of enabling significant C3 grass expansion? Please clarify.

- We agree and will remove colder conditions from the list of conditions enabling C_4 vegetation expansion. In fact, higher temperatures result in C_4 vegetation expansion. This paragraph will now be more comprehensive, and will include additional factors.

Line 315-316: Why just less tropical/summer rain? If compensated for by an increase in winter rain (not likely, I'd suggest) conditions would actually favour arboreal taxa like *Podocarpus*. Maybe just say less precipitation (as in the figure).

- We agree and will remove reference to tropical rain provided by the easterlies here. It will now state 'less precipitation'.

Line 321: I'd be careful about that temperature reconstruction for the site. It appears rather insensitive to my eye, especially as other reconstruction of both continental and sea-surface temperatures are more like 4-6 degC. Perhaps include a more conservative estimate? The point remains the same, but the basis is sounder, perhaps.

- We agree and will change the LGM–Holocene temperature change estimate to 6°C (from Gasse et al., 2008). This will give a potential LGM to Holocene δD_{precip} *enrichment* of 24‰. Like the reviewer states, the point here in the paper remains the same, our δD_{wax} data displays a LGM–Holocene *depletion*, and thus temperature did not exert a dominant control on Mfabeni δD_{wax}.

Lines 331-332: Please describe what those physiological differences have on the dD signal.

- We will rewrite this section to explain how different vegetation types have different δD_{wax} signals.

Lines 334-335: It isn't entirely clear how the paragraph leads to the statement that the influence of ET is dominant over precipitation amount. Please clarify.

- Agreed. We will remove this statement, as it was unclear and unnecessary.

Lines 336-241: Again here, how do the authors determine that ET is significantly more important that amount effect? I agree it can have significant influence, but the authors have not described amount effect at all, but the authors do say ' amount/heightened ET' in several places.

- We will re-write this section to include more information about the amount effect.

Lines 342-353: OK. The authors get there now. Can the authors please reorganise these paragraphs to first outline the mechanism and role of each factor considered, and then bring them together?

- We will re-write and restructure the whole δD_{wax} interpretation section. We will first discount temperature and vegetation effects, and then go on to discuss the importance of the amount effect and evaporation in controlling δD_{wax} at Mfabeni.

There is an issue here though. The inference (and statement in Figure 4) is that there is stronger ET during the last glacial period and it is lower during the Holocene. Considering that the former was significantly cooler, what is driving higher ET? It seems to be the suggestion that 53‰ of dD variability can only be explained if ET is included (and perhaps even dominant). The cited Gat et al. study may suggest this, but other work like Wu et al. (2015), the examples in the cited Gat et al. paper and the Harris et al. study from Cape Town (2010) show very large changes in inter- and intra-annual dDprecip (60‰). Can the authors expand their discussion here to include consideration of this more clearly? An aspect that hasn't been considered is moisture source. Considering the types of synoptic systems that have been suggested to dominate at the timescales considered here, how might these changes have influenced the dD record?

- The δD_{wax} interpretation section will be re-written and re-structured. Our results do not invoke that ET is necessary to explain the δD_{wax}, rather, that we cannot disentangle whether ET or precipitation amount is responsible for the changes. Ultimately, both ET and precipitation amount are highly related and the effect is the same: high δD_{wax} values = drier, low values = wetter.
- In section 6.3, 'climate driving mechanisms', we will explain the mechanisms further. We will attribute the arid climate and the associated expansion of drought tolerant C_4 plants and a low water table at Mfabeni during the LGM, to both lower SSTs and to a northward displacement of the southern hemisphere westerlies, the South African high-pressure cell and the SICZ, shifting the hydroclimate to a more evaporative regime, where ET exceeds precipitation. We understand that cooler temperatures during the LGM would limit ET, but with the latitudinal position of these systems directly controlling the amount of moisture advection towards the southern African subtropics (e.g. Vigaud et al., 2009), we believe that this would easily shift the sensitive balance between ET and precipitation at Mfabeni to a more ET regime.
- The source of precipitation is always from the Indian Ocean, whether the precipitation comes from frontal systems during the winter months or from monsoonal precipitation during the summer (see Gimeno et al., 2010). Thus, as the dD is mainly controlled by ET and precipitation amount, it is not possible to differentiate between winter or summer rain-bearing systems, only the balance between precipitation and ET.

Line 360: Not then "drier conditions" as such necessarily, but potentially just a lower water table (to explain specifically Paq)?

- We will change this sentence to 'Drier conditions during the LGM correspond with low Paq values that indicate a higher relative contribution of terrestrial-over-aquatic n-alkanes, likely a consequence of a lower water table (Fig. 4f).'

Line 362: How do the authors infer "summer" precipitation specifically from the dD record?

- With the majority of precipitation at Mfabeni (76%) falling within the summer months, sourced from the tropical easterlies (and no evidence for any change in water source during the last *c.* 30 ka, we infer changes in precipitation to represent changes in summer precipitation.

Line 363-366: How would changes in the water table change the dD record?

- A lower water table and drier soil conditions would likely serve to increase the D enrichment, having a similar effect as increased ET/reduced precipitation. We will now include this in the manuscript.

Line 379-381: How do the authors explain an increase in temperature resulting in more C3 grasses? This really must be explained.

- We will re-write this section to explain how the climatic changes from the LGM to the Holocene could have resulted in an expansion of more C_3 grasses. We do not think that temperature drove the changes in vegetation at Mfabeni.

The pollen data from the site is included in descriptive form in Figure 6, but please add real percentage data to a figure for comparison with the d13C and dD data. A summary of C3 arboreal and shrub taxa might be one idea. (keep in mind the differences in the chronologies, and make sure to plot using comparable chronologies (perhaps using lithology as a basis for correction).

- We will add the pollen % data of key taxa to Fig. 6 and replace the descriptive pollen summary. We will discuss this pollen data in light of our new isotope data.
-

Lines 382-383: 'plateau indicating continued expansion' is awkward wording. Please rephrase.

- We will rephrase.

Line 389-392: Studies such as Chevalier and Chase., 2015, Schefuß et al., 2011 and other have indicated that direct insolation was unlikely a dominant control on precipitation until the Holocene, with Northern Hemisphere influence dominating. Thus, this is something of a logical fallacy, considering the data available. Insolation did not apparently drive precipitation variability at this time in this region.

- We agree with the reviewer here and will add a little more information to the climate driving mechanisms section, that the results here provide support to previous studies e.g. Schefuß et al., 2011; Chevalier and Chase, 2015) that insolation controls on precipitation variability was only significant since the Holocene.

Regarding ET, the authors are saying that wind strength drove the inferred variability? Are the authors thus saying that the dD record is predominantly a proxy for wind strength (if ET is dominant, and ET is driven by wind)? If that is the case it should 1) be stated more clearly, and 2) be substantiated with some independent records of wind strength variability. The grain size records from the site is not convincing here, as it show a general increase in grain size from 23-16 cal ka BP, when the authors interpret a reduction in ET (lower dD, Figure 4), and otherwise shows little similarity.

- The δD_{wax} record at Mfabeni is predominantly a proxy for moisture availability, whether it be changes in precipitation or ET amount. We will make our interpretation of the proxy signal section clearer (section 6.1).
- The key question is what is driving the changes in precipitation and ET amount? We agree that the grain size records from the site are not overly convincing. We will amend the climate driving mechanisms section and remove focus from wind strength as a driver for changes in ET amount. Although the focus of our discussion will still involve the latitudinal position of the westerlies in driving environmental change at Mfabeni, we will provide more discussion regarding the processes behind this mechanism and how this influences Mfabeni climate.

Line 421: The SST record the authors have chosen may indicate little change, but the Sonzogni et al. (1998) record from the Mozambique channel seems to be compared quite convincingly with continental temperature reconstructions in Chevalier and Chase, 2015, both showing temperature declines in the mid-Holocene. I understand what the authors are saying, in that there is not a consistent, linear relationship between insolation, SSTs and Mfabeni/regional hydroclimates. What would have been the cause of the late Holocene increase in ET that the authors infer? Their Figure 5e is interpreted as indicating LOWER ET during the late Holocene, after the pulse in higher ET from 2-3 ka. Also, this pulse, seems much more consistent in a multi-millennial context as occurring in time with a period of particularly low dD values, which the authors have said indicate lower ET. It may be that the authors' ET focus is becoming problematic for their interpretations.

- Firstly, we will completely remove the final sentence 'The long-term drying trend is unlikely to be caused by decreased summer precipitation because local summer insolation and Mozambique Channel SSTs are high (Fig. 5a & Fig. 4j). Instead, the general drying trend is more likely a result of heightened ET during the late Holocene.' This will keep the section (climate and environmental conditions) as solely for describing the climate and environmental conditions, with no suggestion of driving mechanisms. That said, our evidence points to the period between *c*. 5–0 cal kyr BP as arid.
- We agree with the reviewer and will replace the SST record of Wang et al. (2013), with the older record of Sonzogni et al. (1998) in our Fig. 4j. We will also remove the record of Wang (GIK16160-3) from Fig. 1 and replace it with MD79257 (Sonzogni et al., 1998).
- Nevertheless, even with our data plotted up against the Sonzogni et al. (1998) record, as suggested by the reviewer, we still do not see a clear relationship between SSTs and climatic change at Mfabeni. For example SSTs during the last glacial period remain low until c. 16 ka, whereas a switch from arid to wet conditions at Mfabeni (and within the precipitation stack) appears to occur well before this, at *c*. 19 ka. Thus we stick by our original statement that 'Mfabeni vegetation and hydrology reconstructions over the last 32 cal kyr BP do not show a clear relationship with changes in southwest Indian Ocean SSTs'. Thus we suggest an additional driver to SSTs (i.e. the migration of the westerlies, high-pressure cell and the SICZ).
- We agree with the reviewer that previously too much focus was on ET, when the δD_{wax} record here is predominantly a proxy for moisture availability. We will remove this from the discussion here and now will give multiple possible reasons that the climate at Mfabeni was arid during the last 5 ka. These are: i) equatorward migration of the westerlies, the high-pressure cell and the SICZ (and the mechanisms of these will be explained in the 'climate driving mechanisms' section), ii) a decrease in Mozambique Channel SSTs, iii) anthropogenic influences and/or, iv) ENSO activity.
- Finally, we much appreciate the reviewer highlighting the Sonzogni et al. (1998) SST record. The slight decrease in SST over the last 5 ka is particularly interesting and potentially a cause for the aridity evidenced at Mfabeni. We will now add this observation to the manuscript.

Lines 425-436: I think the authors may have missed some points in the papers they have cited. And the authors' expectation may be to find a single mechanism that explains the whole of the record the authors present.

- In this paragraph we point out that SST is not the **<u>sole</u>** mechanism driving vegetation and hydrological changes at Mfabeni. We stand by this statement. We acknowledge that SSTs may be important during the LGM and the last *c*. 5 ka. Our data supports previous hypothesis that various mechanisms are important, with insolation only becoming important during the Holocene when the westerlies are located poleward.

If the authors compare SST records such as Sonzogni et al., 1998, the authors will see period of similarity, such as during the LGM and MIS 3, from HS1 to the early Holocene and to a lesser/less visible extent the late Holocene.

- As discussed above we will now compare the Mfabeni record with the SST record of Sonzogni et al. (1998). Indeed there certainly is similarity, but the SST record does not explain the entirety of the Mfabeni record.

The significant differences that are evident occur during HS1 and the mid-Holocene. These were highlighted by Chevalier and Chase (2015), and cited as an important distinction between the northern SRZ, where a simpler relationship appears to exist between SSTs (glacial period), orbital forcing (Holocene) and precipitation. Chevalier and Chase concluded that these mechanisms did not so simply drive central/eastern SRZ region. Instead, they find that in this portion of the SRZ, which includes Mfabeni, climates "may have been significantly modulated by the position and influence of the westerly storm track". This idea has subsequently been developed significantly in Chase et al., 2017, where the combined influences of tropical and temperate systems, and the significance of the development of composite synoptic systems has been described in detail. Thus, it comes as some surprise the potential role of the westerlies is raised as a novel suggestion in lines 435-436.

- We will amend the discussion to incorporate the previous hypothesis of Chevalier and Chase (2015) and Chase et al. (2017). Our data support the previous hypothesis that the differences observed between SSTs and the records comprising the 'central and eastern precipitation stack' was due to the modulation of precipitation by the position of the westerlies.
- We will also add a paragraph of discussion about the interaction between tropical and temperate systems resulting in the formation of 'tropical temperate interactions'. In the Chase et al. (2017) paper these interactions are used to explain millennial-scale climatic variability especially during the glacial-interglacial transition.

Lines 437-443: And then the descriptive logic becomes rather twisted about, from my point of view. The southern westerlies are a 'driver of changes in hydroclimate', but this is done by shifting northward and NOT bringing moisture to Mfabeni. How are the westerlies a driver in this case? What is the mechanism? What is the link between the westerlies and the systems that are perceived as bringing moisture to the region at these times? How do the westerlies induce "a more evaporative regime"?

- We will re-write the climate driving mechanisms and discussion section. Here we will explain how the westerlies drive climate in this region. Our hypothesis differs from that of Chase et al. (2017). We invoke that the position of the westerlies is important (and possibly SSTs), no matter the state of local summer insolation. Although millennial scale variability may be explained by TTTs, they are not necessary to explain the climatic variability within the Mfabeni record. Our hypothesis is quite simple and we will better explain. When Antarctic sea ice is expanded this causes an equatorward shift of the westerlies (i.e. during the LGM and last *c*. 5 ka). The expanded sea ice also causes a northward displacement of the high-pressure system (e.g. Vigaud et al., 2009), which is responsible for limiting rainfall across much of the interior of south Africa during the austral winter months. When the high-pressure system is shifted northwards, then monsoonal circulation south of 25°S is weakened, because the monsoon cannot penetrate the continent. This results in aridity at Mfabeni (lower precipitation and/or higher ET). Furthermore, a more north-easterly displaced SICZ, as a result of the equatorward shifted westerlies (e.g. Cook, 2000), may also play a role in driving aridity at Mfabeni.

Line 445: Cockcroft et al is not a climate model simulation. It is a theoretical model.

- Noted, we will change this sentence.

Lines 447-455: Most people don't think, based on the evidence available, that the WRZ (>66% winter rain) expand so far. Strong frontal systems impact the region today, so what are we really talking about? They moved north, but not far enough to bring increased rain (but enough to bring wind), and the easterlies affected the region for a shorter period each year, extending the dry season. Based on the available evidence this seems like quite a story, and one not firmly based in evidence. The authors should be aware too that more sand in the Mfabeni sequence is not necessarily just a function of wind strength. Shifting sea levels and changes in sediment supply, precipitation and vegetation could also have changed the nature of aeolian sediment fluxes. For the wind story to be solid the authors should seek another record for support.

- We agree with the reviewer. This section will now be re-written and reference to wind as a mechanism for explaining the aridity removed.

Lines 456-465: Where is the imagined moisture source for the frontal systems that influence this region? It isn't the SE Atlantic. And really, a northerly shift of the westerlies would (based on observance of the modern annual cycle) probably be manifested through the development of composite systems that primarily draw moisture from off of SE Africa, albeit with a slightly more southerly component. I am having a hard time understanding this logic, so please include a clearer map of moisture sources and transport vectors

- We agree and will re-write this section. The source of precipitation in eastern South Africa is always from the Indian Ocean/Agulhas region, whether it be from frontal systems during the winter months or from direct monsoonal precipitation during the summer (see Gimeno et al., 2010).
- With more northerly-displaced westerlies (i.e. during the LGM), more frontal systems would likely reach the region. Although today frontal systems do bring moisture during the winter months to the region, an increase in the occurrence of frontal systems (with more northerly displaced westerlies during the LGM) did not result in an increase in moisture availability at Mfabeni. We propose that the more northerly location of the high-pressure cell over the South African interior during the LGM may have limited the amount of moisture advection towards Mfabeni, thus even with increased cyclone occurrence, arid conditions persisted.

Lines 469-470: Are the authors saying that Chevalier and Chase, 2015 suggested that increased precipitation after 19 ka was related to insolation? This is not my reading, and they rather suggest that the region is still dominated by Northern Hemisphere influences at that time.

- Agreed, we will remove this reference here.

Lines 471-474: The link between vegetation and hydroclimate is rarely strictly linear, but looking at the dD and central and eastern SRZ precipitation reconstruction, it seems pretty straightforward. More rain/moisture, more trees.

- We completely agree, the switch in vegetation is surely related to more rainfall. However, please note, the G-IG shift in vegetation we observe is likely not a shift from grasses to trees (with the palynological data suggesting a continuation of grassland). Furthermore, with the vegetation shift abrupt and the dD record displaying a more gradual shift to wetter conditions it is worth noting that a critical threshold in moisture availability must have been crossed to allow the establishment of C_3 type vegetation.

Lines 474-475: Need again to clarify what the authors mean by the westerlies having an influence. This regional dynamic needs to be described in much more detail and clarity. Where does this moisture come from?

- This sentence will now be removed and the influence of the westerlies explained in more detail within the climate driving mechanisms section.

Lines 475-477: The site described in Chase et al., 2017 is in the YRZ, not the WRZ. From the WRZ there is the site at De Rif (Chase et al., 2015). It does not a show a shift to more arid conditions.

- We will remove this reference and this sentence.

Lines 486-487: In fact, Schefuß et al., 2011 and Chevalier and Chase, 2015 have said that insolation is only significant during the Holocene, not the late Quaternary as a whole (the data of Partridge et al., 1997 also show this, albeit more coarsely).

- We agree and will amend this sentence accordingly.

Again here, the westerlies shift south and then what occurs to increase precipitation/moisture? A fuller perspective including all the related systems is required.

- The westerlies and the mechanisms for controlling climate will be better explained within the Climate driving mechanisms section.

Lines 488-492: Neither Seweweeksport or Eilandvlei are in the WRZ. Both are in the YRZ.

- Very good point. This sentence will be amended.

Lines 534-541: As the authors state, the difference in sampling resolution means that little can be conclusively determined from the apparent increase in variability. It would be wiser to step back from commenting on this, I feel.

- Agree, we will remove this section regarding the enhanced climatic variability during the last *c*. 5 kyr BP.

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~~The drivers of late~~Late Quaternary climate variability ~~in~~at Mfabeni peatland, eastern South Africa

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Abstract

The scarcity of continuous, terrestrial, palaeoenvironmental records in eastern South Africa leaves the evolution of late Quaternary climate and its driving mechanisms uncertain. Here we use a ~7-m long core from Mfabeni peatland (KwaZulu-Natal, South Africa) to reconstruct climate variability for the last 32 thousand years (cal ka BP). We infer past vegetation and hydrological variability using stable carbon ($\delta^{13}C_{wax}$) and hydrogen isotopes (δD_{wax}) of plant-wax *n*-alkanes and use P_{aq} to reconstruct water table changes. Our results indicate that late Quaternary climate in eastern South Africa did not respond directly to orbital forcing nor to changes in sea surface temperatures (SSTs) in the western Indian Ocean. ~~The~~We attribute the arid conditions evidenced at Mfabeni during the Last Glacial Maximum (LGM) ~~are a consequence of both~~to low SSTs and an equatorward displacement of: i) the southern hemisphere westerlies, ii) the subtropical high-pressure cell and iii) the South Indian Ocean Convergence Zone (SIOCZ), due to increased Antarctic sea ice extent. The northerly location of the high-pressure cell and the SIOCZ inhibited moisture advection inland and pushed the rain-bearing cloud band north of Mfabeni, respectively. The increased humidity at Mfabeni between 19–14 ~~ka~~cal kyr BP likely resulted from decreased Antarctic sea ice, which led to a southward retreat of the westerlies, the high-pressure cell and ~~increased the influence of~~ the ~~moisture-bearing tropical easterlies~~SIOCZ. Between 14–5 ~~ka~~cal kyr BP, when the westerlies, the high-pressure cell and the SIOCZ were in their southernmost position, local insolation became the dominant control, leading to stronger atmospheric convection and an enhanced tropical easterly monsoon. Generally drier conditions persisted during the past *c*. 5 ~~kyrs, but were overlain by high amplitude, millennial-scale environmental variability~~cal ka BP, probably resulting from an equatorward return of the ~~southern hemisphere~~

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westerlies, the high-pressure cell and the SIOCZ. Higher SSTs and heightened ENSO activity~~.~~ may have played a role in enhancing climatic variability during the past *c*. 5 cal ka BP. Our findings ~~stress~~highlight the influence of the ~~southern hemisphere~~latitudinal position of the westerlies, the high-pressure cell and SIOCZ in driving climatological and environmental changes in eastern South Africa.

Key words: Mfabeni; eastern South Africa; *n*-alkanes; hydrogen isotopes; carbon isotopes; southern hemisphere westerlies; tropical easterlies

1. Introduction

~~Last glacial (*c*. 21 ka BP) to present-day changes~~Eastern South Africa is an important region for scientific focus, specifically for furthering our understanding of regional and global climate dynamics. The region is particularly dynamic and sensitive to long-term climate change as it lies within a climatic transition zone, where it is strongly influenced by both temperate (southern westerlies) and tropical (tropical easterlies) climate systems. In eastern South Africa, modelled precipitation reductions and projected regional warming (3–6°C by 2099), threaten the stability of current ecosystems in a region populated by communities already economically vulnerable to the effects of climate change (IPCC, 2013). Past climate and environmental reconstruction and the determination of climate driving mechanisms will provide valuable information for assessing future climate and environmental trends in the region.

Changes in vegetation, precipitation and temperature ~~in eastern South Africa~~from the beginning of the Last Glacial Maximum (LGM; *c*. 26.5 ka; Clark et al., 2009) to present-day, are poorly constrained~~. Whether~~ in eastern South Africa. Whether this region was characterized by aridity or increased humidity during the last glacial period remains unclear. Proxy data show spatial complexity (e.g. Baker et al., 2016; Chase et al., 2017; Chevalier and Chase, 2015 & 2016, Dupont et al., 2011; Schefuß et al., 2011; Scott et al., 2012; Scott, 2016; Schmidt et al., 2014; Simon et al., 2015), and modelled LGM (26.5–19 ka; Clark et al., 2009) precipitation for the region are highly variable and often do not even agree on the sign of precipitation change. For example, the PMIP3 model ensemble mean suggests increased ~~last glacial~~LGM precipitation in the east of South Africa with dry conditions towards the north ~~and south~~ (compared to the present day; Braconnot et al., 2007; Chevalier et al., 2017). Conversely, the NCAR CCSM3 model indicates drier than present conditions in the centre of South Africa and along the eastern coast (Otto-Bliesner et al., 2006). These contrasting simulations for the last glacial period highlight the difficulty in simulating past precipitation in South Africa, with a lack of ~~proper~~a comprehensive understanding ~~of~~regarding the relevant climate processes involved (Stone, 2014).

The mechanisms driving Quaternary climate variability in South Africa are complex and spatially heterogeneous. For example, hydroclimate may be paced by austral summer insolation fluctuations, resulting from changes in the Earth's orbital precession on 23–19 ka timescales. Strong summer insolation (during precession maxima) causes stronger atmospheric convection and an increase in the land/ocean temperature contrast, which results in higher moisture transport by the tropical easterlies and higher precipitation in eastern South Africa (e.g. Simon et al., 2015; Chevalier and Chase, 2015). Climate may also be influenced by high-latitude forcing related to changes in the Earth's orbital obliquity and eccentricity on longer, i.e. glacial-interglacial timescales, which may result in the latitudinal contraction and expansion of the climatic belts (e.g. Dupont, 2011). The model of Nicholson and Flohn (1980) suggests an equatorward displacement of the ~~intertropical convergence zone (ITCZ; Fig. 1~~tropical rainbelt (Nicholson, 2008) during the last glacial period, although proxy data from South Africa provide no conclusive support for this scenario. In addition, during glacial periods, the Walker Circulation may have been weaker with its ascending limb further to the east, ~~within~~over the Indian Ocean~~.~~ (e.g. DiNezio et al., 2018). This possibly resulted in an eastward displacement of the ~~coastal~~ cloud band (SIOCZ) and thus a drier summer rainfall zone (SRZ~~) and a wetter winter rainfall zone (WRZ~~; Tyson, ~~1986~~1999). Furthermore, changes in the latitudinal position of the southern hemisphere westerlies (as a response to fluctuations in Antarctic sea ice extent) have been invoked to influence climate in South Africa (Chase and Meadows, 2007; Chevalier and Chase, 2015; Chase et al., 2017). The western South African region has received most focus regarding the southern hemisphere westerly influence in controlling climate variability (e.g. ~~Zhao~~Stuut et al., ~~2016; Burdanowitz et al., 2018).~~2004; van Zinderen Bakker, 1976). Some studies also suggest possible mechanistic links between SSTs in the Agulhas Current and the Indian Ocean and rainfall variability in South Africa, with high SSTs linked to increasing South African summer precipitation (e.g. Baker et al., 2017; Chevalier and Chase~~ et al., 2017~~, 2015; Dupont, 2011; Dupont et al., 2011; Reason and Mulenga, 1999). Climate forcing experiments also indicate that changes in greenhouse gas concentrations may have driven eastern South African rainfall changes, increasing precipitation between 17–11 kyr (Otto-Bliesner et al., 2014).

The spatially heterogeneous nature of climate variability in South Africa from the last ~~Glacial~~glacial period to the present-day, and the multiple possible climate drivers render the region an important focus for palaeoclimate research. Two important questions remain: i) what was the climate like in eastern South Africa during the last ~~Glacial~~glacial period? and, ii) what were the causes for the climate variability? These questions are difficult to answer with the majority of long, continuous, terrestrial records situated further north, within the range of the modern ~~ITCZ (Fig. 1; *c.* 14°S,~~ tropical rainbelt (e.g. Barker et al., 2007; Tierney et al., 2008), making it hard to assess the long-term climate drivers ~~further~~in the south, in particular in eastern South Africa. In this area, terrestrial sediment archives suitable for palaeoenvironmental reconstruction are scarce, in particular those extending

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~~into~~to the LGM. Marine and speleothem archives have hitherto mostly formed the basis of Quaternary climate research in this region (e.g. ~~Chevalier and Chase, 2015;~~ Dupont et al., 2011; Holmgren et al., ~~1999~~2003). Here we provide stable carbon ($\delta^{13}C$) and hydrogen (δD) isotope records of terrestrial plant-waxes (long-chained *n*-alkanes) from Mfabeni peatland~~.~~, one of the longest continuous terrestrial archives from South Africa. Our vegetation and hydroclimate reconstructions are compared with a previous biomarker-palaeoclimate study from Mfabeni (Baker et al., 2014, 2016 & 2017). We more than double the temporal resolution of the previous plant-wax $\delta^{13}C$ record from Baker et al. (2017), from *c*. 1200 to *c*. 500 years, revealing important and previously undocumented environmental variability.

2. Regional setting

~~South Africa is divided into three main rainfall zones, i) the summer rainfall zone (SRZ), ii) the winter rainfall zone (WRZ), and iii) the overlapping year-round rainfall zone (YRZ; Fig. 1; Chase and Meadows, 2007). The~~The mid-latitude westerlies, in association with the subtropical high (and low) pressure cells and the SIOCZ, play critical roles in determining climatic conditions across the whole South African continent (Fig. 1; Dyson and van Heerden, 2002; Macron et al., 2014; Munday and Washington, 2017). During austral winter, an anti-clockwise rotating subtropical high-pressure cell is located over southern Africa, which drives large-scale subsidence and suppresses rainfall (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), except for in the winter rainfall zone (WRZ), where the westerlies bring rainfall. During summer, the high-pressure cell shifts to the south, and the Angola and Kalahari low pressure cells dominate the continent, enabling monsoonal systems (tropical easterlies) to penetrate southern Africa, bringing rainfall to the summer rainfall zone (SRZ; Fig. 1a; Tyson and Preston-Whyte, 2000; Munday and Washington, 2017). The dominant rain-producing mechanism in the SRZ during the summer are tropical temperature troughs (TTTs), which are embedded within the SIOCZ and form a northwest-southeast orientated cloud band, extending over the continent into the southwest Indian Ocean (Fig. 1a; Todd and Washington 1999; Tyson and Preston-Whyte, 2000). TTTs form from interactions between tropical convection and mid-latitude perturbations, which result in heavy precipitation events (Tyson, 1986; Macron et al., 2014; Chase et al., 2017). A combination of strong easterly flux from the Indian Ocean and low pressure over the continent during the summer results in the development of TTTs (Fig. 1a; Cook, 2000; Macron et al., 2014; Rácz and Smith, 1999; Todd and Washington 1999).

South Africa can be divided into several climate zones: the SRZ lies in the north and east where 66 % of the mean annual precipitation falls between October and March (Fig. 1a; Chase and Meadows, 2007). ~~Within~~ Based on late Quaternary precipitation reconstructions, further subdivisions of the SRZ

~~the climate is dominated~~(northern SRZ, central/eastern SRZ) have been suggested by ~~tropical temperate troughs~~Chevalier and ~~easterly flow, which brings moisture from the Indian Ocean to eastern South Africa.~~Chase (2015). In the extreme south and west of South Africa lies the WRZ~~,~~ (Fig. 1a), where 66 % of the mean annual precipitation falls between April and September (Chase and Meadows, 2007). This rainfall is associated with ~~temporal~~temperate frontal systems related to the southern hemisphere westerlies (Fig. 1b; Mason and Jury, 1997~~). Sandwiched~~; Tyson, 1986; Tyson and Preston-Whyte, 2000). In between the SRZ and WRZ lies the year-round rainfall zone (YRZ~~,~~) which receives precipitation both in summer and winter seasons~~.~~ (Fig. 1a; Chase and Meadows, 2007). This zone comprises much of the southern Cape of South Africa and is highly heterogeneous in terms of precipitation seasonality and amount, spanning some of the wettest (e.g. along the south coast), and driest (e.g. Namib Desert~~; Williamson, 1997~~) regions in South Africa.

Mfabeni peatland is located within the SRZ, on the coastal plain of northern KwaZulu-Natal (28°09′8.1″S; 32°31′9.4″E; 9 m above sea level; Fig. 1; Fig. 2). The dominating subtropical high-pressure cell across the majority of South Africa during the austral winter months leads to mild and dry winter conditions at Mfabeni. Occasional rainfall during the winter months at Mfabeni is associated with the passage of cold fronts, which develop in the western Atlantic and move across southern Africa (Fig. 1b; Grab and Simpson, 2000). These cold fronts trigger rainout of atmospheric moisture, which is sourced from the Indian Ocean and Agulhas region (Gimeno et al., 2010). When the subtropical high-pressure cell has moved south during the austral summer, the tropical easterlies dominate, TTTs form and conditions at Mfabeni are hot and humid. The average annual rainfall amount between 2010 and 2018 at Mfabeni in the winter months (June–August) was measured at 134 mm compared to 426 mm during the summer months (December–February), meaning the majority of rainfall (76 %) falls during the summer months (data from World Weather Online). A northeast–southwest precipitation gradient is present across the peatbog, with 1200 mm $year^{-1}$ of precipitation in the east decreasing to 900 mm $year^{-1}$ westwards towards Lake St. Lucia (Fig. 1; Fig. 2; Taylor et al., 2006). The main source of water to Mfabeni is precipitation, predominantly provided by the tropical easterlies and TTTs, sourced from the Indian Ocean and Agulhas Current region (Fig. 1; Tyson, 1999; Gimeno et al., 2010). Mean summer temperatures (November to March) surpass 21 °C. The wind regime is characterised by moderate northeasterly winds during the summer and more intense southwesterly winds during winter.~~Mfabeni peatland is located within the SRZ, on the coastal plain in~~

Mfabeni~~ northern KwaZulu-Natal (28°09′8.1″S; 32°31′9.4″E; 9 m above sea level; Fig. 1; Fig. 2). It~~ is one of the oldest, continuously growing peatlands in South Africa (Grundling et al., 2013). It lies within a topographical inter-dunal depression between the Indian Ocean to the east and Lake St. Lucia to the west (Fig. 2; Grundling et al., 2013). Towards the ocean, it is bordered by an 80–100 m high vegetated dune barrier, and to the west by the 15–70 m high Embomveni sand dune ridge (Fig. 2).

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Over the last 44 ka~~ BP~~, the mire accumulated *c*. 11 m of peat, deposited on top of a basal clay layer (Grundling et al., 2015). This clay layer was crucial in the formation and development of the mire, limiting water loss during low sea level stands (Grundling et al., 2013). Mfabeni is bound to the north and south by beach ridges isolating it from Lake Bhangazi and Lake St. Lucia, respectively (Fig. 2; Grundling et al., 2013). When lake levels in Lake Bhangazi are high, minor water exchange between Mfabeni and Bhangazi occurs, but there are no fluvial inputs to either system. Surface drainage occurs southwards towards Lake St Lucia (Fig~~ 1; Fig. 2). The modern climate at Mfabeni is subtropical, with hot and humid summers and relatively mild and dry winters. Mean summer temperatures (November to March) surpass 21 °C and the majority of the annual precipitation occurs during the summer months. The main source of water to Mfabeni is precipitation, predominantly provided in the summer by the tropical easterlies (Fig. 1; Tyson, 1999). Occasional rainfall during the winter months is associated with the passage of cold fronts and strong winds from the south (Kruger et al., 2010). The average annual rainfall amount between 2010 and 2018 at Mfabeni in the winter months (June–August) was measured at 134 mm compared to 426 mm during the summer months (December–February), meaning the majority of rainfall (76 %) falls during the summer months (data from World Weather). A northeast–southwest precipitation gradient is present~~. 2). ~~, with 1200 mm year^{-1} of precipitation in the east decreasing to 900 mm year^{-1} westwards towards Lake St. Lucia (Fig. 1; Fig. 2; Taylor et al., 2006). The wind regime is characterised by moderate northeasterly winds during the summer and more intense southwesterly winds during winter.~~

The peatland receives groundwater via the swamp forest and the western dunes. This groundwater, which is important in keeping the mire wet during the dry season, discharges towards the center of the peatland and then flows within a sub-surface layer towards the east (Grundling et al., 2015). In the northern and eastern part of the peatland, the vegetation is sedge and reed fen (comprising of sedges and grasses). In the western and southern parts of Mfabeni is swamp forest (Venter, 2003).

The modern water balance at Mfabeni is dominated by the interplay between evapotranspiration (ET; 1035 mm) and precipitation (1053 mm). Groundwater inflow (14 mm) and stream outflow (9 mm) have a minor contribution to the modern water balance (all measured between May 2008 and April 2009; Grundling et al., 2015). Changes in regional climate have ~~the~~much potential to influence the fine balance between ET and precipitation. For example, ET is suppressed when cloud cover is increased during the summer months and increased during times of higher wind speed (Grundling et al., 2015). ~~Furthermore, ~~ET is higher in the swamp forest than in the sedge and reed fen, therefore a change in vegetation composition ~~at Mfabeni~~also has the potential to impact ET rates. The depositional setting of the Mfabeni peatland provides a unique opportunity to reconstruct past eastern

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South African climate variability at centennial-scale resolution from the Late Pleistocene to the present day.

3. Methodological background

To reconstruct past vegetation and hydroclimate ~~changes~~ we use the distribution, and the carbon and hydrogen isotopic composition, of long chain *n*-alkanes derived from plant-waxes.

3.1 Distributions of plant-waxes

To obtain information on water table variations, we quantify the relative contribution of plant-waxes derived from submerged and floating macrophytes relative to that of emergent and terrestrial plants (P_{aq}). Odd-numbered *n*-alkanes (C_{25}–C_{35}) are derived from the epicuticular wax coating of terrestrial higher plants (Eglinton and Hamilton, 1967). Conversely, aquatic plant-waxes (of submerged macrophyte origin) are dominated by mid-chain *n*-alkanes (typically C_{23} and C_{25}; e.g. Baker et al., 2016; Ficken et al., 2002). Thus we quantify P_{aq} using Equation 1 (Ficken et al., 2000).

$$P_{aq} = (C_{23} + C_{25})/(C_{23} + C_{25} + C_{29} + C_{31}) \qquad \text{Eq. 1}$$

with C_x the amount of each homologue.

To assess *n*-alkane degradation we used the carbon preference index (CPI; Bray and Evans, 1961). The CPI reflects the molecular distribution of odd-to-even *n*-alkanes, within a certain carbon number range (here, *n*-C_{26} to *n*-C_{34}; Equation 2). High CPI values indicate a higher contribution of odd-numbered *n*-alkanes (relative to even), indicating the *n*-alkanes are derived from higher terrestrial plants. Low CPI values indicate either low contribution from terrestrial higher plants or high organic matter degradation (Eglinton and Hamilton, 1967).

$$CPI_{27\text{-}33} = 0.5 * (\Sigma C_{odd27\text{-}33}/\Sigma C_{even26\text{-}32} + \Sigma C_{odd27\text{-}33}/\Sigma C_{even28\text{-}34}) \qquad \text{Eq. 2}$$

with C_x the amount of each homologue.

3.2 Carbon and hydrogen isotopes of terrestrial plant-waxes

To reconstruct vegetation changes, we use the carbon isotopic composition of terrestrial plant-waxes ($\delta^{13}C_{wax}$). On late Quaternary timescales, the primary factor determining the amplitude of fractionation between the $\delta^{13}C$ of atmospheric CO_2 ($\delta^{13}C_{atm}$) and the carbon isotopic composition of the plant ($\delta^{13}C_{plant}$) is the plant carbon fixation pathway (C_3/C_4/CAM; e.g. Diefendorf and Freimuth, 2017). On these timescales, changes in the $\delta^{13}C_{atm}$ are too small to significantly influence $\delta^{13}C_{wax}$ (Tipple et al., 2010). Shrubs and trees use the C_3 photosynthetic pathway and show the largest fractionation. Grasses utilize either the C_3 or the C_4 pathway, with C_4 plants having the smallest net fractionation (Collister et al., 1994). The differences in carbon isotope fractionation during carbon uptake leads to different $\delta^{13}C_{wax}$ ~~isotopic~~ signatures, and allows the determination of past vegetation types: *n*-alkane $\delta^{13}C$ values of C_3 plants are *c*. -36‰ VPDB (Vienna Pee Dee Belemnite) and *c*. -20‰ VPDB for C_4 plants (e.g. Diefendorf and Freimuth, 2017).

The hydrogen isotope composition of plant-waxes (δD_{wax}) reflects the isotopic composition of the water used during lipid biosynthesis (Sachse et al., 2012), rendering it a valuable tool for reconstructing past hydrological conditions (e.g. Collins et al., 2013; Schefuß et al., 2005). δD_{wax} is influenced by three main factors: i) the isotopic composition of precipitation; ii) enrichment of soil and leaf water due to ET; and iii) differences in the apparent isotopic fractionation between source water and plant-waxes due to differences in vegetation type. The importance of each factor varies by study site and with time. The detailed interpretation of the Mfabeni $\delta^{13}C_{wax}$ and δD_{wax} is discussed in section 6.1.

4. Methods: compound specific C and H isotope analyses

Core MF4-12 (6.96 m recovery, 8.77 m penetration) was recovered from the centre of Mfabeni peatland during January 2012 using a vibrocoring device (Fig. 2). The chronology of the core is established by 24 ^{14}C AMS (accelerator mass spectrometry) dates from bulk peat (Fig. 3, S1). The chronology is extended from that published in Humphries et al. (2017) and the age model is made using Bacon 2.2 program (Blaauw and Christen, 2011). Radiocarbon ages were calibrated using the southern hemisphere calibration curve, ShCal13 (Hogg et al., 2016) and the post-bomb southern hemisphere curve, zone 1–2, for the uppermost modern dates (Hua et al., 2016).

Freeze-dried, bulk peat samples were ground and homogenized ~~with~~using a pestle and mortar~~,~~ and lipids were extracted from *c*. 2 g of peat ~~with~~using a DIONEX Accelerated Solvent Extractor (ASE 200) at 100 °C and at 1000 psi for 5 minutes (repeated 3 times) using a dichloromethane (DCM):methanol (MeOH) (9:1, v/v) mixture. ~~Squalane was added prior~~Prior to extraction~~.~~, squalane was added as an internal standard. ~~Elemental~~Copper turnings were used to remove elemental sulfur ~~was removed~~ from the ~~TLEs using copper turnings, and~~total lipid extract (TLE). To remove water, the TLE was ~~removed by passing~~passed over a Na_2SO_4 column~~,~~ (eluting with hexane~~. After~~). Subsequent to saponification~~,~~ (by adding 6 % KOH in MeOH~~,~~) and extraction ~~of the neutral fractions~~ (with hexane~~,~~), the neutral fractions were split into a further three fractions: hydrocarbon, ketone, and polar ~~fractions using.~~, by silica gel column chromatography (~~with a~~ mesh size ~~of~~ 60 µm) and elution with hexane, DCM and DCM:MeOH (1:1), respectively. ~~Subsequent elution~~ By eluting the hydrocarbon fractions with hexane over $AgNO_3$-impregnated silica columns we obtained the saturated hydrocarbon fractions ~~from~~

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~~the hydrocarbon fractions~~. The~~ concentrations of long-chain *n*-alkanes in the~~ saturated hydrocarbon fractions were measured using a Thermo Fischer Scientific Focus gas-chromatograph (GC) with flame-ionization-detection (FID) equipped with a Restek Rxi 5ms column (30 m x 0.25 mm x 0.25 µm~~).~~), in order to determine the concentrations of long-chain *n*-alkanes. The ~~split/splitless inlet temperature was 260 °C, the~~ GC oven temperature was ~~programmed~~set at 60 °C, held for 2 ~~min~~minutes, increased at 20 °C/~~min~~minute to 150 °C and then at 4 °C/~~min~~minute to 320 °C and held for 11 minutes. ~~Concentrations~~The split/splitless inlet temperature was 260 °C. To estimate the sample concentrations needed for isotope analyses, samples were ~~estimated by comparison~~compared with an external standard ~~containing~~that was run every 5 samples, which contained *n*-alkanes (C_{19}–C_{34}) at a concentration of 10 ng/µl~~ that was run every 5 samples. Replicate~~. A quantification uncertainty of <5% was yielded through replicate analyses of the external standard~~ yielded a quantification uncertainty of <5%.~~ .

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The $\delta^{13}C$ values of the long-chain *n*-alkanes were measured using a Thermo Trace GC Ultra equipped with an Agilent DB-5 column (30m x 0.25mm x 0.25µm) coupled to a Finnigan MAT 252 isotope ratio ~~monitoring~~ mass spectrometer (IR-MS) via a combustion interface operated at 1000 °C. The GC temperature was programmed from 120 °C (hold time~~:~~ 3 min), followed by heating at 5 °C/~~min~~minute to 320 °C (hold time~~:~~ 15 ~~min~~minutes). The ~~$\delta^{13}C$ values were calibrated against~~ external CO_2 reference gas was used to calibrate the $\delta^{13}C$ values and they are reported in ‰ VPDB. Samples were analysed in duplicate when *n*-alkane concentrations were adequate for multiple runs. The internal standard (squalane, $\delta^{13}C$= -19.9‰), yielded an accuracy of 0.6‰ and a precision of 0.2‰ (n=37). The external standard mixture was analysed every 6 runs. The long-term precision and accuracy of the external *n*-alkane standard was 0.2 and 0.15‰, respectively. For $\delta^{13}C$ the average precision of the *n*-C_{29} and *n*-C_{31} alkane in replicates was 0.2‰ and 0.1‰ (n=22), respectively.

The δD compositions of long-chain *n*-alkanes were measured using a Thermo Trace GC coupled via a pyrolysis reactor (operated at 1420 °C) to a Thermo Fisher MAT 253 ~~isotope ratio mass spectrometer (GC/~~IR-MS~~)~~. The GC column and temperature program was similar to that used for the $\delta^{13}C$ analysis. The ~~δD values were calibrated against~~ external H_2 reference gas was used to calibrate the δD values and they are reported in ‰ VSMOW. The H^{3+} factor was monitored daily and fluctuated around 5.2 ppm nA^{-1} during analyses. ~~An~~After every sixth measurement, an *n*-alkane standard of 16 externally calibrated alkanes was measured~~ every sixth measurement~~. The long-term precision and accuracy of the external *n*-alkane standard was 2.7 and 2‰, respectively. Samples were analysed in duplicate when *n*-alkane concentrations were adequate for multiple runs. The internal standard (squalane, δD= -180‰; ±2), yielded an accuracy of 0.9‰ and a precision of 1.9‰ (n=36). For δD the average precision in replicates was 1‰ for both *n*-C_{29} and *n*-C_{31} alkanes (n=52).

~~Last~~The last glacial period Mfabani δD_{wax} values were corrected to account for the effect of changes in global ice volume (Collins et al., 2013; Schefuß et al., 2005). For this, the benthic foraminifera-based oxygen isotope curve (Waelbroeck et al., 2002) was interpolated to each sample age and then converted to *δ*D values using the global meteoric water line (Craig, 1961).

5. Results

This study focusses on the last 32 cal ka BP (*c*. 590 cm). The average temporal resolution between the 62 samples analysed for δ^{13}C and *δ*D is *c*. 500 years. From 590 ~~cm (32 ka BP)~~ to 70 cm (*c*. 32–2 ~~ka~~cal kyr BP) the core is very dark brown in colour containing peat with humus, fine detritus and silt. From 70 cm to core top, the sediments are similar in colour to the peat below and contain fibrous peat with humus and herbaceous fine detritus (Humphries et al., 2017). Between 457 and 358 cm (*c*. 23–14 ~~ka~~cal kyr BP; comprising the LGM) mean grain sizes average at 110 µm, with smaller diameters averaging at 50 µm between 298 and core top (*c*. 11 ~~ka~~cal kyr BP–present, Holocene; Fig. 4g). The lithology of core MF4-12 does not exactly match with that observed from core SL6 (Baker et al., 2014; 2016; 2017), although sandy peat is observed during the LGM at both locations. This result is not surprising~~,~~ as multiple cores taken in transects across the bog indicate peat heterogeneity (Grundling et al., 2013).

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Long chain *n*-alkane CPI values are generally around 6 (ranging from 2–13), indicating good *n*-alkane preservation. The two samples with CPI values of 2, potentially containing more degraded *n*-alkanes, are highlighted in red (Fig. 4b & c; Fig. 5b & c; Fig. ~~6~~6f & g). However, the in- or exclusion of these samples does not affect the observed pattern of changes and we thus consider the record to be suitable for palaeoclimate reconstruction. The samples contain *n*-alkanes with carbon chain lengths ranging from C_{17}–C_{35}, with C_{29} and C_{31} generally having the highest abundance. The high abundances of C_{29} and C_{31} enabled reliable isotopic analyses. The relationship between the δD and δ^{13}C of the C_{29} and C_{31} *n*-alkanes is strong, with R^2 values of 0.8 and 0.9, respectively. Consequently, for the $\delta^{13}C_{wax}$ and δD_{wax}, we use the amount-weighted mean of the C_{29} and C_{31} *n*-alkanes.

The $\delta^{13}C_{wax}$ values range from -29‰ to -21‰ (Fig. 4b). The ice volume *δ*D correction decreases the glacial Mfabeni δD_{wax} values by <8 ‰ (Fig. 4c). The ice-corrected δD_{wax} values of the *n*-C_{29} and *n*-C_{31} alkanes range from -181‰ to -128‰ (Fig. 4c). P_{aq} values range from 0.02–0.7, averaging at 0.2 (Fig. 4f).

During the LGM ~~(26.5–19 ka BP; Clark et al., 2009),~~, $\delta^{13}C_{wax}$ and δD_{wax} values are relatively high averaging at -23‰ and *c*. -136‰, respectively (Fig. 4b & c) and P_{aq} values are low (*c*. 0.24; Fig. 4f). At *c*. 19 cal ka BP a 4‰ negative shift in $\delta^{13}C_{wax}$ values occurs (Fig. 4b). This negative shift in $\delta^{13}C_{wax}$ is concurrent with a gradual shift to lower δD_{wax} values (Fig. 4c) and an increase in P_{aq} values (Fig. 4f). Between 14 and 5 ~~ka~~cal kyr BP, $\delta^{13}C_{wax}$ values are relatively stable and average at -28‰ (Fig. 4b). δD_{wax}

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values become gradually lower during this period reaching -173‰ at 7.5 cal ka BP. At 5 cal ka BP, δD_{wax} values shift towards more positive values by 16‰ (Fig. 4c). Relatively high P_{aq} values occur between 14–5 ~~ka~~cal kyr BP (Fig. 4f). After *c*. 5 cal ka BP several high amplitude millennial-scale fluctuations in both $\delta^{13}C_{wax}$ and δD_{wax} values are evident. These fluctuations interrupt a trend where the isotope values of both $\delta^{13}C_{wax}$ and δD_{wax} gradually increase towards present day. A pronounced shift to higher $\delta^{13}C_{wax}$ and δD_{wax} values occurs at 2.8 cal ka BP. From *c*. 900 cal yr BP, $\delta^{13}C_{wax}$ and δD_{wax} values become higher reaching core top values of -21 and -128‰, respectively (Fig. 4b ~~and~~& c). Generally high, but variable and rapidly fluctuating P_{aq} values are evident between *c*. 5–0 ~~ka~~cal kyr BP. P_{aq} values decrease substantially after 1.3 cal ka BP from 0.6 to a core top value of *c*. 0 (Fig. 4f).

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6. Discussion

6.1 Interpretation of the proxy signals

The relatively high CPI_{27-33} values indicate that the long-chain *n*-alkanes within the peat are derived from terrestrial higher plants and are relatively non-degraded. The long-chain *n*-alkanes are likely sourced directly from the local vegetation surrounding the coring location. It is possible that during times of stronger wind strength (i.e. during the LGM; Humphries et al., 2017) increased aeolian transport resulted in a higher biomarker contribution from more distal sources (i.e. the surrounding dune vegetation). Good preservation of *n*-alkanes in Mfabeni peat was also observed in nearby core SL6, but this was based on a CPI calculated using *n*-C_{21-31} (Baker et al., 2016). No relationship exists between the CPI and P_{aq} (R^2 = 0.11), ~~this~~which suggests that CPI variations at the location of core MF4-12 are not related to changes in organic matter preservation due to water table level variations.

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The main source of carbon for terrestrial higher plants (the source of the C_{29} and C_{31} *n*-alkanes) is atmospheric CO_2, whereas aquatics also assimilate dissolved carbon, complicating the interpretation of their carbon isotope signal. We thus focus solely on C_{29} and C_{31} *n*-alkanes that are predominantly derived from terrestrial plants (Eglinton and Hamilton, 1967). The majority of the samples (67 %) have dominant *n*-alkane chain lengths of C_{29} and C_{31}. For the remaining 33 % of the samples, concentrated between 6 and 1.1 ~~ka~~cal kyr BP, the dominant chain length switched to *n*-C_{25}, indicating a higher *n*-alkane input from submerged macrophytes (Ficken et al., 2000). The *n*-C_{25} are unlikely to be sourced from mosses, as mosses are rare in subtropical peatland environments (Baker et al., 2016). Instead, the C_{25} is likely mainly derived from aquatic plants, which produce mid-chain *n*-alkanes as dominant homologues (C_{20}–C_{25}; Ficken et al., 2000). This increase of *n*-alkanes sourced from aquatic plants *c*. 6–1.1 ~~ka~~cal kyr BP is unlikely to have had any impact on the isotopic composition of the long-chain *n*-alkanes (C_{29} and C_{31}) as these are minor components in aquatic plants (e.g. Aichner et al., 2010). Therefore, we interpret the $\delta^{13}C_{wax}$ as changes in the C_3/C_4 ratio of terrestrial higher plants.

~~C_4 grasses are dominant within the SRZ, with C_3 grasses more prevalent in the WRZ at higher altitudes (Vogel et al., 1978).~~Grasses exhibiting the C_4 or C_3 photosynthetic pathway in South Africa are largely geographically separated, with C_4 grasses dominant within the SRZ and C_3 grasses more prevalent in the YRZ, WRZ and at higher altitudes (Vogel et al., 1978). As C_4 grasses require less water to fix CO_2, thus having greater water-use efficiency than C_3 grasses, C_4 photosynthesis is favored in arid regions (e.g. Downes, 1969; Osborne and Sack, 2012). C_4 grasses also have the potential to achieve higher rates of photosynthesis than C_3 particularly at high irradiance and temperature levels (Black et al., 1969; Monteith, 1978), as their more efficient carbon fixation has a higher energy demand (Sage, 2004). Today growing season temperatures are a controlling factor for the distribution of C_4 and C_3 grasses (with C_4 grasses having an advantage over C_3 grasses at higher temperatures; Sage et al., 1999). Consequently C_4 grasses are mainly found in warm and dry environments such as the African savannas (Beerling and Osborne, 2006). Furthermore, under reduced atmospheric (i.e. glacial) CO_2, the higher carbon fixation efficiency of C_4 grasses provides an important advantage over C_3 grasses (Sage, 2004; Pinto et al., 2014). Previous palynological studies indicate that the dominant components of the pollen assemblage at Mfabeni are Poaceae and Cyperaceae (Finch and Hill, 2008). Although Cyperaceae species can be either C_3 or C_4, most Cyperaceae in eastern South Africa (67 %) are of the C_4-type (Stock et al., 2004). The C_4 vegetation at Mfabeni is thus mostly Poaceae or Cyperaceae from the sedge and reed fen. The C_3 vegetation at Mfabeni is comprised of arboreal taxa from the swamp forest (e.g. Myrtaceae and *Ficus*) and locally distributed *Podocarpus* (Finch and Hill, 2008; Venter, 2003). Shifts to ~~heavier~~higher $\delta^{13}C_{wax}$ values (more C_4-type vegetation) at Mfabeni could indicate an expansion of grassland~~, as a result~~ (at the expense of ~~either~~arboreal taxa), or a shift from C_3 to C_4 grasses, resulting from: i) ~~colder conditions, ii) lesser~~less precipitation~~ provided by the tropical easterlies (weaker summer rains),~~, iii) a longer/more intense dry season, iv) heightened ET, v) reduced water table height, ~~or vi~~vi) higher temperatures, vii) reduced atmospheric CO_2, or viii) increased insolation levels (or any combination of the above).

The δD_{wax} reflects the δD_{precip}, ET~~ amount~~ and vegetation type. The δD_{precip} can be influenced by changes in air temperature, with an estimated temperature effect of *c*. 0.5‰ per 1°C for $\delta^{18}O_{precip}$ (Dansgaard, 1964). The maximum estimated temperature change of *c*. ~~2~~6 °C ~~at Mfabeni~~in the SRZ of South Africa from the LGM to Holocene (~~Chevalier and Chase, 2015~~Gasse et al., 2008), would thus correspond to a change in $\delta^{18}O_{precip}$ of ~~1~~3‰. Conversion to changes in δD_{precip} using the global meteoric water line would thus lead to a potential LGM to Holocene δD_{precip} enrichment of ~~8~~24‰ (Craig, 1961). However, the Mfabeni δD_{wax} record shows a depletion in δD_{wax} from the LGM to the Holocene, rather than an enrichment. The observed glacial δD depletion is therefore a conservative estimate. Consequently, changes in temperature from the LGM to the Holocene did not exert a dominant control on Mfabeni δD_{wax}.

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Changes in vegetation type (C_3/C_4) have the potential to reduce or exaggerate shifts in δD_{wax}.

There are differences in the apparent fractionation (the integrated isotopic fractionation between precipitation and plant-wax lipids) between plant types using different photosynthetic pathways. C_3-type shrubs and trees fractionate the least, C_4-type grasses slightly more, while C_3-type grasses show the highest apparent fractionation (Sachse et al., 2012). ~~This relationship occurs due to physiological differences, with grasses monocotyledonous and~~ The difference in δD_{wax} between dicots (C_3, shrubs ~~and~~, trees ~~dicotyledonous. Nevertheless~~and forbs) and monocots (C_4, grasses) is likely the result of leaf architecture and the nature of water movement in the leaf. Monocots display progressive evaporative enrichment along parallel veins along the leaf, which does not occur in dicots. This grass-blade enrichment results in higher δD_{wax} values in C_4 grasses (Helliker and Ehleringer, 2000). However, recent data suggest that the effect of C_3-tree to C_4-grass vegetation type changes on δD_{wax} is likely ~~is~~ relatively small (Collins et al., 2013; Vogts et al., 2016). ~~Therefore, the observed variability in δD_{wax} at Mfabeni is most likely the result of relative changes in the amount of ET versus changes in precipitation.~~

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The δD_{precip} is strongly controlled by the 'amount effect', where there is a negative correlation between monthly precipitation amount and δD_{precip} (Dansgaard, 1964). Close to the equator, passage of the tropical rainbelt can result in precipitation that is extremely depleted in D. Conversely, in arid regions, rainfall tends to be enriched in D because of enhanced evaporation of the raindrops as they fall (Risi et al., 2008). Studies investigating the present-day relationship between precipitation amount and the isotopic variations in rainfall indicate shifts in $\delta^{18}O$ of up to 15‰ (*c.* 120‰ in δD) with the passage of the tropical rainbelt and shifts in $\delta^{18}O$ of 7‰ (*c.* 56‰ in δD) with the passage of convective storms (Gat et al., 2001). ~~During times of heightened ET and/or lower precipitation amount, soil waters become enriched in D (e.g. Sprenger et al., 2017). Furthermore~~During times of heightened ET and/or lower precipitation amount, soil waters become enriched in D (Sprenger et al., 2017). In addition, under conditions of low ambient relative humidity, leaf water becomes enriched in D through increased transpiration (Kahmen et al., ~~2013).~~ 2013). Large values of isotopic enrichment (*c.* 40‰ in $\delta^{18}O$, 180‰ in δD) are associated with the effects of evaporation (e.g. Cappa et al., 2003).

Mfabeni has high rates of ET, which can equal, or even exceed precipitation during dry periods (Grundling et al., 2015). Consequently, both precipitation amount and ET ~~is therefore~~are likely ~~a dominant factor controlling~~to control the ~~enrichment~~isotopic composition of ~~D within~~ soil and leaf waters, and ~~consequently in~~subsequently of the leaf waxes at Mfabeni.

High δD_{wax} values at Mfabeni likely result from decreased summer precipitation amount and/or heightened ET. ~~Studies investigating the present-day relationship between precipitation amount and δD indicate 'extreme' shifts by up to 15‰ with the passage of the ITCZ, 7‰ with the passage of convective storms or around 1.5‰/100 mm of monthly precipitation (Gat et al., 2001). Much larger values of isotopic enrichment (c. 55‰) are associated with the effects of evaporation (Kim~~

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~~and Lee, 2011). The large isotopic variability observed within the Mfabeni record (*c.* 53‰) therefore implies that both changes in precipitation and ET amount are needed to explain the δD_{wax} variability over the past 32 ka BP.~~ The similarity between the δD_{wax} pattern and the regional precipitation/aridity stacks (Fig. 4d & e; Chevalier and Chase, 2015; 2016) ~~support this~~supports the inference ~~and indicate~~that precipitation amount and ET drive Mfabeni δD_{wax}. Furthermore, this similarity indicates that the hydrological fluctuations in the Mfabeni record represent hydrological change at a broader spatial scale (Fig. 4c–e), but also suggest that the pollen-based precipitation stacks may also include an element of ET variability.

6.2 Climatic and environmental conditions at Mfabeni over the last 32 cal ka BP

The $\delta^{13}C_{wax}$, δD_{wax} and P_{aq} data from Mfabeni indicate that the vegetation, hydrology and the water table varied considerably over the last 32 cal ka BP (Fig. 4 ~~& Fig.~~&. 5). The high $\delta^{13}C_{wax}$ values during the LGM indicate that the vegetation was likely dominated by more drought-tolerant C_4 plant types (Fig. 4b). Similar LGM $\delta^{13}C_{wax}$ depletion was observed previously at Mfabeni (Fig. 4h; Baker et al., 2017). Drier conditions during the LGM ~~are also consistent~~correspond with ~~the observed~~ low P_{aq} values that indicate a higher relative contribution of terrestrial-over-aquatic *n*-alkanes, likely a consequence of a lower water table (Fig. 4f). The high δD_{wax} values during the LGM suggest decreased ~~summer~~ precipitation amount and/or higher ET ~~amount~~, which are both consistent with a drier environment (Fig. 4c). We cannot completely rule out the possible impact of increased drainage of the peatbog during the LGM due to low eustatic sea level (Grundling et al., 2013~~), however~~). A lower water table during the LGM would likely serve to further soil water D enrichment. Nevertheless, the fact that the peat continued to grow during the LGM suggests that the sea level effect was minor. ~~Indeed, the~~The organic geochemical proxies agree with palynological data indicating regional grassland dominance (high Poaceae, Cyperaceae and Asteraceae) with low amounts of arboreal taxa (Fig. 6; Finch and Hill, 2008). Regional aridity and~~/or stronger~~ increased wind strength during the LGM at Mfabeni are also indicated by increased mean grain size of the lithogenic sediment fraction ~~at Mfabeni~~ (Fig. 4g~~;~~), and the modal grain size of the distal aeolian component (Humphries et al., 2017). ~~The~~Evidence for reduced precipitation (from the regional precipitation stack; Fig. 4d) and high aridity (from the regional aridity stack; Fig. 4e) during the LGM, provide evidence that the dry conditions at Mfabeni ~~during the LGM~~ appear to be part of a wider eastern South African pattern~~, since they are consistent with regional precipitation and the aridity stacks (Fig.~~ (Fig. 4d & e).

The shift to more negative $\delta^{13}C_{wax}$ values following the LGM, at *c.* 19 cal ka BP, indicating that the vegetation at Mfabeni changed to more C_3-type ~~plants~~ (Fig. 4b), is also evident in Mfabeni core SL6 (Fig. 4h; Baker et al., 2017). This change is thus likely representative of ~~C_3/~~a C_4 ~~changes~~–C_3 change across the peat bog. ~~However, the~~ The palynological record indicates no shift towards arboreal taxa at

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this time but instead a continuation of grasslands (Fig. ~~6~~6a & b; Finch and Hill, 2008). ~~Baker et al. (2017) suggested that this carbon isotope~~ A decrease in Cyperaceae percentages (as most Cyperaceae is of C_4 type), may be responsible for the C_4 to C_3 shift ~~resulted from the~~observed in the $\delta^{13}C_{wax}$ record, but the gradual nature of the Cyperaceae decrease points to an additional driver (Fig. 6c). The shift is more likely the result of a switch from C_4 to C_3 grasses. If temperature was driving the vegetation shifts at Mfabeni, we would expect a shift from C_3 to C_4 grasses from the LGM to the Holocene (with a *c.* 6°C increase in temperature~~ across the glacial-interglacial transition. However, higher growing-season temperatures would favour C_4 over C_3 grasses (Ehleringer, 1997) and we therefore~~). Nevertheless, the LGM to Holocene shift from C_4 to C_3 grasses suggests that temperature did not drive the vegetation change at Mfabeni. We suggest that the ~~carbon isotope decrease represents a~~ shift from C_4 ~~grasses~~ to C_3 grasses ~~(~~may have been caused instead by i) more precipitation, ii) a shorter/less intense dry season, iii) lower ET, and/or ~~from C_4 sedges to C_3 sedges).~~iv) increased water table height. Furthermore, with C_3 vegetation favored under lower insolation conditions, a decrease in local summer insolation from the LGM to Holocene (Fig. 4a) could have played a role in driving the vegetation shifts.

After *c.* 19 cal ka BP, the $\delta^{13}C_{wax}$ values continue to decrease to -29‰ until they ~~plateau~~stabilize at *c.* 14 ~~ka BP, indicating continued expansion of C_3 vegetation.~~cal ka BP. This trend in $\delta^{13}C_{wax}$ values between *c.* 19 and 14 ~~ka~~cal kyr BP, indicating an expansion of C_3 vegetation, corresponds well with ~~a decrease in aeolian dust (Humphries et al., 2017) and~~ the $\delta^{13}C_{wax}$ record from Mfabeni core SL6 (Baker et al., 2017; Fig. 4b & 4h). There are, however, some minor differences between the two $\delta^{13}C_{wax}$ records. We attribute these to small-scale variations in vegetation across the peatbog, the lower sampling resolution of core SL6 and to dating uncertainties in both records. The shift to lower $\delta^{13}C_{wax}$ values at *c.* 19 cal ka BP occurs at the same time as a rise in the water table as documented by an increase in P_{aq} values (Fig. 4f). ~~The gradual shift to lower δD_{wax} values around 19 ka BP occurs during decreasing local summer insolation, suggesting that this moisture shift was unlikely to be a result of increased precipitation, but more likely resulting from lower ET rates due to decreasing wind strength~~An abrupt increase in precipitation amount and a decrease in aridity is evident in the precipitation and aridity stacks at *c.* 19 cal ka BP. All proxy records for precipitation (the regional stacks and the Mfabeni δD_{wax} data; Fig. 4) strongly suggest a switch to wetter conditions after *c.* 19 cal ka BP.

The $\delta^{13}C_{wax}$ values between 14–5 ~~ka~~cal kyr BP reflect a stable period of C_3-type vegetation (Fig. 4b). At the same time, gradually decreasing δD_{wax} values indicate increasing humidity. ~~Pollen data from Mfabeni provide evidence for an expansion of arboreal type vegetation at *c.* 12 ka BP (Fig. 6~~The gradual increase in precipitation is also evident in the precipitation stack, but this trend is interrupted by an abrupt return to aridity at *c.* 14.2 cal ka BP, coinciding with the Antarctic Cold Reversal (Chase et al., 2017). This abrupt arid event is only evident in one sample at Mfabeni and thus higher resolution sampling is needed across this interval. The aridity stack indicates low aridity during this interval, but

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high variability suggests a complex interplay between high ET (from increased temperatures, resulting in less effective precipitation) and generally more precipitation (Fig. 4e). Pollen data from Mfabeni provide evidence for an expansion of arboreal type vegetation at *c.* 12 cal ka BP (Fig. 6a; Finch and Hill, 2008). The pollen data thus suggest the establishment of swamp forest vegetation during the early Holocene, indicative of a moist climate (Fig. ~~6~~6a). Mfabeni aeolian sediment ~~grain size are~~flux is low and stable throughout this period, also suggesting a moist climate ~~and low wind strength (Fig. 4g~~(Humphries et al., 2017). The moist climate likely resulted in vegetated dunes, reducing the amount of material available for aeolian transport. The relatively high P_{aq} values between 14–5 ~~ka~~cal kyr BP indicate a high and stable water table at this time (Fig. 4f). Elevated total organic carbon percentages within Mfabeni core SL6 during the Holocene, also suggest increased water levels (Baker et al., 2017).

Between *c.* 5–0 ~~ka~~cal kyr BP several high-amplitude millennial-scale C_3/C_4 vegetation changes are evident superimposed on an overall shift from predominantly C_3 to more C_4-type vegetation towards the present-day (Fig. 5b). This variability contrasts with the more gradual C_4/C_3 vegetation transition from the ~~Glacial~~glacial period to Holocene. The $\delta^{13}C_{wax}$ values from Mfabeni core SL6 between *c.* 6–1 ~~ka~~cal kyr BP also indicate a period of predominantly C_4-type vegetation, implying arid conditions during this time (Baker et al., 2017; Fig. 4h). A similar pattern of a long-term trend with superimposed short-term variability is visible in the in δD_{wax} record. The general enrichment in D reflects gradual drying, punctuated by millennial-scale pulses of aridity, with the most pronounced arid event at *c.* 2.8 cal ka BP (Fig. 5c). Counterintuitively, the high abundance of *n*-C_{25} alkanes and high but variable P_{aq} values between *c.* 5–0 ~~ka~~cal kyr BP indicate a generally high water table, interrupted by brief periods of a lower water table (Fig. 5d). After 2.3 cal ka BP, both $\delta^{13}C_{wax}$ and δD_{wax} values become higher and P_{aq} values lower (Fig. 5b–d). This suggests increased C_4-type vegetation cover, decreased summer precipitation amount and/or higher ET ~~amount~~ and low water table levels. A slight increase in precipitation followed by gradually decreasing precipitation over the last *c.* 5 ka is evidenced in the precipitation stack (Fig. 4d). This initial increase in precipitation at *c.* 5 cal ka BP corresponds to an abrupt decrease in aridity (Fig. 4d & e). The increased variability observed in our records between 5–0 ~~ka~~cal kyr BP could be an artefact of the high temporal resolution of our record during this interval (~220 vs ~700 years per sample for the remainder of the record). Nevertheless, (Fig. 4d) other data from the region (e.g. Baker et al., 2017, Humphries et al., 2017; 2016, Finch and Hill, 2008, Neumann et al., 2010) also indicate climatic instability and ~~pulses of~~ arid climatic conditions during the last *c.* 5 cal ka BP, suggesting that the observed variability is likely real (Fig. 5e). ~~The long-term drying trend is unlikely to be caused by decreased summer precipitation because local summer insolation and Mozambique Channel SSTs are high (Fig. 5a & Fig. 4j). Instead, the general drying trend is more likely a result of heightened ET during the late Holocene.~~

It is interesting that modern δD_{wax} values and those during the LGM appear similar (Fig. 4c), implying similarly arid conditions during both periods. The southern aridity stack also indicates extremely arid conditions during the last few thousand years and the authors stress the importance of temperature in controlling aridity (Fig. 4e; Chevalier and Chase, 2016). It is possible that modern high mean annual temperatures drove these modern-day δD_{wax} values to appear similar to those from the LGM. High modern day temperatures, increase ET and result in less 'effective precipitation' and arid conditions, even when rainfall is high (Cheavlier and Chase, 2016). During the LGM, lower temperatures would have reduced ET, leading to apparent humid conditions, despite reduced rainfall amount.

6.3 Climate driving mechanisms

Modern observations suggest that ~~warm~~high SSTs within the Mozambique Channel and Agulhas Current induce increased evaporation (e.g. Walker, 1990), resulting in ~~increased onshore airflow and advection of moist air and~~ higher rainfall in the SRZ (Tyson, 1999). Variations in local ~~SST~~SSTs are thus thought to be an important driver of hydroclimate in eastern South Africa. This mechanism may also play a ~~key~~ role on longer ~~time scales~~timescales. Indeed, Chevalier and Chase (2015) ~~mention this hypothesis, invoking~~invoke SSTs as the dominant driver of precipitation variability during the LGM~~, and Baker et al. (2017) argue for the existence of a short arid event at 7.1 ka BP that corresponds to a decrease in Mozambique Channel SSTs. However,~~. Mfabeni vegetation and hydrology reconstructions over the last 32 cal ka BP do not show a ~~close~~clear relationship with changes in southwest Indian Ocean SSTs (Fig. ~~4j, Wang et al., 2013). This suggests~~ 4j, Sonzogni et al., 1998). For example if SSTs drove the climate at Mfabeni then the abrupt shift to more C_3 type vegetation and the gradual shift to a wetter climate at *c*. 19 cal ka BP would be expected to correspond with an increase in SSTs. This is not the case, and SSTs do not increase until *c*. 15.7 ka (Sonzogni et al., 1998; Fig. 4). The lowest temperatures within the Mozambique Channel correspond to Heinrich Event 1 (SSTs *c*. 3°C colder than present day), an event which is not evident as a particularly arid period in the Mfabeni dataset. Mozambique Channel SSTs thus do not fully explain the variability observed in the records comprising the precipitation stack. These differences, as proposed previously by Chevalier and Chase (2015), suggest that SST variability is unlikely to be the sole driver of the changes in hydroclimate ~~at Mfabeni over the 32 ka. Thus, we suggest an additional role, namely the southern hemisphere~~within this part of the SRZ. Chevalier and Chase (2015) proposed that the differences observed between SSTs and the records comprising the precipitation stack is due to the modulation of precipitation by the position of the westerlies.

We attribute the arid climate and the associated expansion of drought tolerant C_4 plants and a low water table at Mfabeni during the LGM, in part, to a northward displacement of the ~~southern~~

~~hemisphere westerly winds~~westerlies, the SIOCZ and the subtropical high-pressure cell, shifting the hydroclimate to a more evaporative regime, where ET exceeds precipitation. In addition, lower SSTs (Fig. 4j) in the Mozambique Channel at this time likely reduced moisture availability. It is possible that the combination of a northward displacement of these three systems (the ~~southern hemisphere~~ westerlies, SIOCZ and subtropical high-pressure cell) and lower SSTs shifted the fine balance between precipitation and ET at Mfabeni towards higher ET rates during the LGM.

~~Numerous palaeoenvironmental~~Palaeoenvironmental studies (e.g. Lamy et al., 2001; Lamy et al., 2010; Stuut and Lamy, 2004~~; Chase et al., 2017) and~~), climate model simulations (e.g. ~~Cockcroft et al., 1987;~~ Rojas et al., 2009; Toggweiler et al., 2006~~), indicate~~) and theoretical models (e.g. Cockcroft et al., 1987) provide evidence for an ~~intensification and~~ equatorward migration and strengthening of the southern hemisphere westerlies in response to the increased extent of Antarctic sea ice during the LGM. ~~Such changes~~Records from the present WRZ such as Elands Bay Cave (Baxter, 1996), Pakhuis Pass (Scott, 1994) and Driehoek Vlei (Meadows and Sugden, 1993) indicate increased winter rainfall, interpreted as a northward shift and strengthening of the westerlies during the LGM (Chase and Meadows, 2007). An equatorward migration of the westerlies may have expanded the limit of the WRZ in South Africa northward, to around 25°S in the west and 30°S in the east (Cockcroft et al., 1987). This would have put Mfabeni (at 28°S) within the range of the southern westerlies. ~~Regions on the east coast, such as Mfabeni, then experienced stronger winter winds, causing heightened ET (Humphries et al., 2017). With more northerly westerlies, the duration of the dry season at Mfabeni may also have been extended diminishing the influence of the easterlies. This shortened the rain season and heightened ET rates. The northward shift in the westerlies during the LGM is also visible in records from the present WRZ (e.g. Chase et al., 2017), which show increased winter rainfall and moist conditions (Fig. 4i).~~ Although during the LGM the westerlies were in a more northerly position, and had the potential to provide rainfall (via the passage of more cold fronts; Nkoana et al., 2015), we do not see any evidence for increased precipitation at Mfabeni. Today mid-latitude cyclones (frontal systems; Fig. 1b) associated with the westerlies trigger rainout of atmospheric moisture, sourced from the Indian Ocean and Agulhas Current, during the winter months (Gimeno et al., 2010). However, the co-occurring subtropical high-pressure cell over the South African interior may have limited the amount of moisture advection towards Mfabeni, thus even with increased cyclone occurrence, arid conditions persisted. Furthermore, with a northerly displaced subtropical high-pressure cell inhibiting monsoonal penetration, the duration of the dry season at Mfabeni may have been extended, shortening the rain season and heightened ET rates.

~~Although during the LGM the southern hemisphere westerlies were in a more northerly position and had the potential to provide rainfall, we do not see any evidence that the source of precipitation changed. Today the moisture at Mfabeni is mainly provided by the tropical easterlies~~

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~~(Kruger et al., 2010; Tyson, 1999), and thus we see enriched D values on the east coast and more depleted values further inland (δD_{precip} data from GNIP (IAEA, 2018) and groundwater δD data from West et al., 2014). Precipitation sourced from the southern hemisphere westerlies would be strongly depleted in D because of the large continental transport distance from the Atlantic Ocean. However, there is no evidence of lower δD_{wax} values during the LGM at Mfabeni, indicating the westerlies did not bring moisture to Mfabeni during the LGM but that increased wind strength led to increased ET and a more arid climate.~~

A northward migration and strengthening of the westerlies is also associated with a northerly displaced and weaker South Indian Anticyclone (Fig. 1; Cohen and Tyson, 1995). A weakening of the western portion of the South Indian Anticyclone results in a northeastward shift of the SIOCZ (and the rain-bearing cloud band associated with TTTs; Cook, 2000). This northeastward shift results in higher precipitation over coastal Africa (around 15°N) and Madagascar and lower than normal precipitation to the south, in eastern South Africa (Cook, 2000). We propose that a northeastward shift of the SIOCZ during the LGM may have also played a key role in driving aridity at Mfabeni.

The latitudinal position of the subtropical high-pressure cell is highly correlated to rainfall variability along the eastern coast of South Africa (Dyson and van Heerden, 2002). Multivariate analysis of zonal moisture fluxes in South Africa indicates that the latitudinal position of the subtropical high-pressure cell directly controls the amount of moisture advection (monsoonal penetration) towards the southern African interior during the summer months (Vigaud et al., 2009). When the cell is shifted southward, during the summer, the tropical easterlies are able to penetrate further inland, resulting in higher continental moisture availability (Vigaud et al., 2009). Conversely, when the cell is shifted northward, during the winter, monsoonal circulation south of 25°S is impeded, creating a deficit in moisture advection from the ocean to the continent (Tyson and Preston-Whyte, 2000; Vigaud et al., 2009). A more northerly location of the subtropical high-pressure cell, during the LGM, would have lengthened the dry season, resulting in aridity at Mfabeni.

We suggest that the shift to more humid conditions at *c*. 19 cal ka BP was related to the retreat of the ~~southern hemisphere~~ westerlies ~~from this area~~, the subtropical high-pressure cell and the SIOCZ, as Antarctic sea ice began to retreat poleward ~~at this time~~ (Fig. 4k), ~~leading to less ET and~~ allowing an increased influence of the moist tropical easterlies. ~~This shift~~ With the subtropical high-pressure cell further south, stronger easterly flux from the Indian Ocean likely enhanced the development of TTTs in the region leading to increased precipitation. This shift at *c*. 19 cal ka BP was unlikely driven by a change in local summer insolation ~~(i.e. Chevalier and Chase, 2015)~~ because insolation was decreasing at this time~~, which would have caused reduced, instead of enhanced, summer precipitation~~. We suggest that the abrupt shift to more C_3 vegetation was a non-linear response to increasing moisture availability in the ~~peatbog~~region (Fig. 4c). Precipitation amount may have reached a critical threshold at *c.* 19 cal ka BP for the establishment of C_3 type vegetation, resulting in the observed abrupt vegetation shift (Fig. ~~4b). We propose that at *c.* 19 ka BP, the position of the southern westerlies had a greater climatic influence than the local insolation forcing. Further south, within the present WRZ, the retreat of the westerlies at this time resulted in a shift to more arid conditions (Fig. 4i; Chase et al., 2017). The timing of increased humidity at Mfabeni at *c.* 19 ka BP corresponds well to a reduction in Antarctic sea ice extent, which is thought to be the main driver of the latitudinal position of the westerlies (Fig. 4k; Fischer et al., 2007).~~4b).

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Between 14–5 ~~ka, low~~ kyr BP, a reduced extent of Antarctic sea ice (Fig. 4k & 5g), resulted in a more poleward position of the westerlies~~.~~ and the subtropical high-pressure cell. The diminished effect of the westerlies and the subtropical high-pressure cell in eastern South Africa at this time permitted the tropical systems (easterlies~~, and thus local summer insolation,~~), to dominate the climatic regime at Mfabeni. ~~Indeed, increasing~~With a strengthened (but poleward displaced) South Indian Anticyclone the SIOCZ was likely situated over Mfabeni resulting in increased rainfall. Strong easterly flux would have increased the development of TTTs in the region, resulting in higher humidity at Mfabeni. Increasing humidity at ~~this time~~Mfabeni during the Holocene, corresponds with increasing southern hemisphere summer insolation (Fig. 4a). The importance of insolation for South African climate variability during the late Quaternary has been suggested before (e.g. Partridge et al., 1997; Simon et al., 2015~~;~~). Our results support the hypothesis that insolation control on precipitation variability was only significant during the Holocene (e.g. Schefuß et al., 2011; Chevalier and Chase, 2015). ~~However, we~~We suggest that direct local insolation forcing is only dominant in this region when the westerlies and subtropical high-pressure cell are located far south~~,~~, which allows monsoonal precipitation to penetrate into the continent during the summer months.

To explain the millennial-scale climatic variability over the glacial-interglacial transition within their central and eastern African sites (which also includes Mfabeni), Chevalier and Chase (2015) and Chase et al. (2017) suggest that this region may be influenced by the position and the intensity of the westerlies, and the interactions between the westerlies and the tropical easterlies (resulting in TTT development). We highlight the importance of the location of TTT development (i.e the SIOCZ) and stress the interconnections between TTT development, the latitudinal position of the westerlies and the subtropical high-pressure cell on glacial-interglacial timescales.

After *c.* 5 cal ka BP, palaeoenvironmental records from both the WRZ and YRZ, such as from Verlorenvlei (Fig. 1; Fig. 5f; Carr et al., 2015), Seweweekspoort (Fig. 1; Fig. ~~4h~~4i; Chase et al., 2017), Klaarfontein (Fig. 1; Meadows and Baxter, 2001), Cecilia Cave (Fig. 1; Baxter, 1989) and Eilandvlei (Wündsch et al., 2018~~)~~), document increased moisture ~~supply to the WRZ~~availability, implying a recurring more northerly location of the ~~westerly storm tracks at this time~~westerlies. Chevalier and Chase et al. (2015) propose that increased precipitation in the WRZ during the late Holocene was due to both the warmer interglacial climate and the northward expansion of the westerly storm tracks. Although no indication for an increase in sea ice is evident from EPICA salt concentration data (Fig. 4k), diatom data (*Fragilariopsis curta* and *F. cylindrus*) from PS2090/ODP1094 in the southern South Atlantic document an increase in sea ice during the late Holocene (Fig. 5g), which may have pushed the southern westerlies equatorward. In addition, climate modelling results imply a northward shift of the southern ~~hemisphere~~ westerlies at this time (Hudson and Hewitson, 2001). Consequently, in a comparable way to the LGM, the increased sea ice during the late Holocene (Fig. 5g), may have displaced (and strengthened) the westerlies, the South African high-pressure system and the SIOCZ equatorward, ~~increasing winter wind strength and the length of the dry season~~resulting in higher aridity at Mfabeni~~, leading to~~. A slight decrease in Mozambique Channel SSTs may have also played a ~~decreased influence of~~role in the ~~moisture bearing tropical easterlies (Mejía~~generally arid climate at Mfabeni during the last *c*. 5 cal ka BP (Fig. 4j; Sonzogni et al., ~~2014; Toggweiler et al., 2006; Williams and Bryan, 2006). Furthermore, although the westerlies may have had a more northerly position during this time, simultaneous high local summer insolation and warm SSTs (causing strong convective rainfall during summer; Fig. 5a) may have been the cause of the relatively high water table (Fig. 5d) and transitory peaks in precipitation and C_3-type vegetation expansion (Fig. 5b and c).~~1998). Interestingly, the hydrological variability at Mfabeni (Fig. 5c) during the last *c*. 5 cal ka BP, is not present in the central and eastern South African precipitation stack (Fig. 4d). We attribute this to the highly sensitive balance between ET and precipitation at Mfabeni (Grundling et al., 2015), and the fact that the precipitation stack ~~smoothes~~smooths local hydrological variability.

It is possible that anthropogenic influences also played a role in shaping the environment at Mfabeni ~~during~~, at least, during the late Holocene. However, unequivocal agricultural and exotic pollen indicators are absent from the pollen record and although pollen data indicate that ~~deforestation~~forest decline occurred during the late Holocene, it is unclear whether this was related to human influence or regional ~~climatic~~climate change (Fig. 6; Finch and Hill, 2008). The ~~deforestation~~forest decline could have affected the water table and increased the relative amount of C_4-type vegetation. The appearance of *Morella* and *Acacia* in the late Holocene may indicate the development of open vegetation or secondary forest due to fire disturbance (Fig. 6e; Finch and Hill, 2008). Human activities or climate change may be responsible for changes in fire regime. With no palaeo-charcoal data available for Mfabeni yet, no direct evidence for increased fire activity during the late Holocene exists. In addition, the palaeoenvironmental evidence available suggests that the arid conditions during the late Holocene were regional in nature (Scott, 1999; 2003; Humphries et al., 2016, Neumann et al., 2010). Thus, any human activity was unlikely the primary cause of the late Holocene regional aridity and the large magnitude of environmental changes observed at Mfabeni.

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Today ENSO activity is one of the most important driving mechanisms for inter-annual climatic variability in South Africa ~~(Tudhope et al., 2001).~~. Southern Africa's seasonal rainfall is linked to ENSO, with dry (wet) conditions associated with El Niño (La Niña) events (Archer et al., 2017; Mason and Jury, 1997). ~~According to the model of Tyson (1986), during El Niño events, high Indian Ocean SSTs produce a weaker landward pressure gradient over the southeast African coast, diminishing low-level confluence over the continent. This results in a longitudinal migration of the ascending limb of the Walker circulation eastwards, leading to lower rainfall on the continent (i.e at Mfabeni) but higher rainfall towards the east (i.e Madagascar; Mason and Jury, 1997).~~Interannual variability in the strength and position of the SIOCZ is linked to ENSO variability (Cook, 2000). During La Niña years, the SIOCZ is located over the continent, resulting in wet conditions in eastern South Africa. During El Niño, the SIOCZ shifts northeastward over the Indian Ocean and as a consequence, dry conditions prevail in eastern South Africa (Lindesay, 1988; Cook, 2001; Hart et al., 2018). Furthermore, during El Niño events, a northward shift of the westerlies may occur, which could increase rainfall over western South Africa but lead to aridity in the east (i.e. at Mfabeni; Lindesay, 1988). ~~Although after *c.* 5 ka BP the Mfabeni sampling resolution is higher, we document some evidence for heightened climatic variability (in comparison with the rest of the record) and generally drier conditions at Mfabeni over the last *c.* 5 ka BP. We speculate that this variability may have been the result of amplified ENSO activity (e.g. Humphries et al., 2017).~~ Palaeoenvironmental studies in the Pacific Basin and South America indicate that during the early Holocene El Niño events were smaller and occurred less frequently, with a shift to stronger ENSO activity after *c.* 5 cal ka BP (Fig. 5h, Moy et al., 2002; Huffman, 2010; Rodbell et al., 1999; Sandweiss et al., 1996~~; Tudhope et al., 2001).~~). It is difficult to disentangle the possible potential drivers of climate variability during the last *c.* 5 cal ka BP at Mfabeni. We therefore invoke a possible combination of northerly-displaced westerlies, lower SSTs and the impact of ENSO variability as potential climatic drivers during this time.

~~We therefore invoke a combination of both the northerly-displaced southern hemisphere westerlies and the impact of high ENSO variability as climatic drivers during the last *c.* 5 ka BP. The high-amplitude, millennial-scale vegetation and hydrological instability documented at Mfabeni during the last *c.* 5 ka BP contrasts with the relatively stable conditions during the LGM and early Holocene. This increased environmental variability during the late Holocene could be the result of increased and strongly fluctuating sea ice extent during this period, overlain by strong ENSO activity (Moy et al., 2002).~~

7. Conclusions

Compound specific carbon and hydrogen isotope data and *n*-alkane distributions (P_{aq}) from Mfabeni peatbog are used to reconstruct climatic conditions, over the last 32 cal ka BP~~,~~ in eastern

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South Africa. The LGM at Mfabeni was characterized by a high contribution of C_4 grasses, low precipitation amount/high ET ~~rates~~ and a low water table. During the LGM, increased Antarctic sea ice extent led to an equatorward displacement (and strengthening) of the southern hemisphere ~~westerly winds~~westerlies, the SIOCZ and the subtropical high-pressure cell, which may have extended the length and increased the intensity of the dry season ~~at~~, as well as shifted the location of TTT formation northeast of Mfabeni. Between *c.* 19–5 ~~ka~~cal kyr BP an expansion of C_3-type vegetation occurred, with more rainfall and a higher water table at Mfabeni. At *c.* 19 cal ka BP, Antarctic sea ice decreased, which resulted in a southward retreat of the ~~southern hemisphere~~ westerlies~~.~~, the SIOCZ and the subtropical high-pressure cell. This ~~southward~~ retreat ~~of the westerlies after *c.* 19 ka BP and~~ combined with an increase in local summer insolation, after *c.* 12 cal ka BP, resulted in more precipitation and an increased wet season length at Mfabeni. When the westerlies, the SIOCZ and the subtropical high-pressure cell were in their southernmost position (*c.* 14–5 ~~ka~~cal kyr BP), local insolation became the dominant control on Mfabeni climate, leading to stronger convection and enhanced monsoonal precipitation from the tropical easterlies. The late Holocene (*c.* <5 cal ka BP) was characterized by increased environmental instability and increasingly arid conditions. We attribute these trends to concurring ~~high local summer insolation~~low SSTs, and the recurring influence of the southern westerlies and/or heighted ENSO activity.

The Mfabeni record indicates that climate and environmental variability in eastern South Africa over the last 32 cal ka BP ~~is~~were driven by a combination of i) enhanced/reduced moisture transport by the tropical easterlies, driven by variations in southern hemisphere summer insolation, and ii) latitudinal displacements (and the strengthening/weakening) of the ~~southern hemisphere~~ westerlies~~.~~, the SIOCZ and the subtropical high-pressure cell. With the expansion and retreat of Antarctic sea ice ultimately responsible for the displacement of ~~the westerlies~~these systems, we invoke high-latitude climate forcing as an important driver of climate in eastern South Africa.

Data availability: Supplementary data for the depth-age~~-~~ model **(S1)** is available with this manuscript. A new depth-age model of core MF1 (Finch and Hill, 2008), produced by Bacon, can be found within the supplementary information **(S2)**. Other data is available on PANGAEA.

Author contributions: CM and ES conducted $\delta^{13}C_{wax}$ and δD_{wax} analyses. Interpretation was carried out by CM, JF, TH, FP, MH, MZ and ES.

Competing interests: The authors declare no competing financial interests.

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Figure captions

Figure 1. ~~a) Rainfall seasonality map for southern~~Map of South Africa in austral summer **(a)** and winter **(b)** showing the major oceanic and atmospheric currents~~,~~ and the position of the ~~intertropical convergence zone (ITCZ) and the~~ Congo Air Boundary (CAB). ~~Red/orange = summer~~ H (L) = high (low)-pressure systems. BC = Benguela Current. AC = Agulhas Current. Rainfall zones are shown in **(a)**: WRZ = winter rainfall zone~~ (SRZ). Green~~, YRZ = year-round rainfall zone~~ (YRZ). Blue is winter rainfall zone (WRZ). The white arrows are atmospheric circulation~~, SRZ = summer rainfall zone. SIA = South Indian Anticyclone. SAA = South Atlantic Anticyclone. SIOCZ = South Indian Ocean Convergence Zone. Note, the westerlies move north during austral winter and the ~~blue arrows are oceanic circulation. Map courtesy of B. Chase (Chase et al., 2017). Letters~~high-pressure system dominates over much of the continent, suppressing rainfall in the SRZ. Squares represent the key study sites mentioned in the text (and shown in Fig. 4 and 5): **a)** ~~GIK16160-3 (Wang~~MD79257 (Sonzogni et al., ~~2013~~1998). **b)** Mfabeni, this study~~.~~ (red square). **c)** Lake St Lucia (Humphries et al., 2016). **d)** Seweweekspoort (Chase et al., 2017). **e)** Cecilia Cave (Baxter, 1989). **f)** Klaarfontein (Meadows and Baxter, 2001) and Verlorenvlei~~,~~ (Carr et al., 2015), one location. Figure modified from Gasse et al. 2008.

Figure 2. Mfabeni peatland and its regional geomorphological features, indicating the location of core MF4-12 (red circle, this study) and the location of core SL6 (black circle, Baker et al., 2014; 2016; 2017). Map is courtesy of B. Gijsbertsen, UKZN Cartography Unit.

Figure 3. Depth-age model of core MF4-12 produced using Bacon, based on 24 ^{14}C AMS dates~~.~~ (S1). Blue symbols are AMS dates and grey shading indicates 95% confidence interval on the mean age (red line).

Figure 4. Climate and environmental change at Mfabeni compared with regional records and orbital insolation. **a**) December-January-February (DJF) insolation for 28°S (blue line; Laskar~~,~~ et al., 2011). **b**) Stable carbon isotope composition (weighted mean) of C_{29}–C_{31} *n*-alkanes from Mfabeni, reflecting changes in C_3/C_4 vegetation type. **c**) Hydrogen isotope composition (weighted mean) of C_{29}–C_{31} *n*-alkanes from Mfabeni, reflecting changes in ~~summer~~ precipitation amount and ET~~ amount~~. Red is the δD_{wax} corrected for ice volume changes. Error bars on isotope data reflect analytical uncertainty of duplicate analyses. **d**) Central and eastern South African regional precipitation stack (red line; Chevalier

and Chase, 2015). **e**) Southern African regional aridity stack (Chevalier and Chase, 2016). **f**) P_{aq} at Mfabeni, indicating the amount of aquatic vs. terrestrial *n*-alkanes (high/low water table). **g**) Mean grain size data of the lithogenic sediment fraction from Mfabeni~~, with increased grain size indicating increased wind strength~~ (Humphries et al., 2017). **h)** Mfabeni core SL6 stable carbon isotope composition (weighted mean) of C_{29}–C_{31} *n*-alkanes (Baker et al., 2017). **i)** Combined nitrogen isotope data from Seweweekspoort rock hyrax middens, reflecting changes in humidity (Chase et al., 2017). **j**) $U^{K'}_{37}$ derived SSTs from core ~~GIK16160-3~~MD79257 in the Mozambique Channel (~~Wang~~Sonzogni et al., ~~2013).~~1998). **k**) Sea salt sodium concentrations from the EPICA DML ice core in Antarctica, reflecting changes in sea ice coverage (Fischer et al., 2007). The two Mfabeni samples with CPI values of *c*. 2 are highlighted in red (4b & c). ~~SHW = southern hemisphere westerlies.~~ Blue shading = Mfabeni wet, orange = Mfabeni arid.

Figure 5. Comparison of Mfabeni data with other records of environmental variability over the last 15 ~~ka~~cal kyr BP. **a)** DJF insolation for 28°S (black line; Laskar~~,~~ et al., 2011). **b)** Carbon isotope composition (weighted mean) of C_{29}–C_{31} *n*-alkanes from Mfabeni, reflecting changes in C_3/C_4 vegetation type. **c)** Hydrogen isotope composition (weighted mean) of C_{29}–C_{31} *n*-alkanes from Mfabeni, reflecting changes in summer precipitation amount and ET~~ amount~~. **d)** P_{aq} at Mfabeni, indicating the amount of aquatic vs. terrestrial n-alkanes (high/low water table). Blue dashed lines highlight trends. **e)** Mfabeni calcium/scandium ratio, indicating changes in water table (Humphries et al., 2017). **f)** Bulk carbon isotope data from Verlorenvlei (Carr et al., 2015). **g)** ~~Extent~~An estimation of the extent of Antarctic sea ice~~. Estimation is~~ based on the abundance of *Fragilariopsis curta* and *Fragilariopsis cylindrus* at site PS2090/ODP1094 (SW of Cape Town; Bianchi and Gersonde, 2004). **h)** Red colour intensity time-series from Laguna Pallcacocha. High values are light coloured inorganic clastic laminae, which were deposited during ENSO-driven episodes (Moy et al., 2002). The Mfabeni sample with a CPI value of *c*. 2 is highlighted in red (5b & c).

Figure 6. Summary figure highlighting the main climate phases and driving mechanisms at Mfabeni. ~~From left: hydrogen~~All pollen data is from Finch and Hill (2008). Note, the new age model for pollen % data is in the supplementary material (S2). **a)** *Podocarpus* % data from Mfabeni. **b)** Poaceae % data from Mfabeni. **c)** Cyperaceae % data from Mfabeni. **d)** Asteraceae % data from Mfabeni. **e)** *Morella serrata* % data from Mfabeni. Poaceae and Cyperaceae were excluded from the regional pollen sum so their percentages are based on total pollen frequencies. *Podocarpus*, Asteraceae and *M. serrata* percentages are based on regional frequencies. See Finch and Hill (2008) for more details. **f)** Stable carbon isotopic composition (weighted mean) of C_{29}–C_{31} *n*-alkanes from Mfabeni. **g)** Hydrogen isotope composition (weighted mean) of C_{29}–C_{31} *n*-alkanes from Mfabeni. Red is the δD_{wax} corrected for ice

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volume changes. ~~Stable carbon isotopic composition (weighted mean) of C_{29}–C_{31} *n*-alkanes from Mfabeni.~~ The two Mfabeni samples with CPI values of *c*. 2 are highlighted in red. ~~Summary of the palynological data from Finch and Hill (2008) and possible climate driving mechanisms at Mfabeni during the last 32 ka BP.~~ Blue shading = Mfabeni wet, orange = Mfabeni arid.

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