Dryland dunes and other dryland environmental archives as proxies for Late Quaternary stratigraphy and environmental and climate change in southern Africa

Abi Stone

Geography Department, School of Environment, Education and Development, The University of Manchester, Oxford Road, Manchester, M13 9PL

Abstract

The Namib Desert and the Kalahari constitute the drylands of southern Africa, with current relatively humid portions of the latter having experienced periodically drier conditions during the Late Quaternary. This study explores the range of dryland archives and proxies available for the past ~190 ka. These include classic dryland geomorphological proxies, such as sand dunes, as well as water-lain sediments within former lakes and ephemeral fluvial systems, water-lain calcrete and tufa sediments at the interface of surface hydrological and hydrogeological, speleothems and groundwater hydrogeological records, and hyrax middens. Palaeoenvironmental evidence can also be contained within geoarchaeological archives in caves, overhangs and rockshelters. This integration of records is undertaken with the aim of identifying a (or a number of) terrestrial regional chronostratigraphic framework(s) for this time period within southern Africa because this is missing from the Quaternary stratigraphic lexicon. Owing to a lack of long, near-continuous terrestrial sequences in these drylands, the correspondence between nearby terrestrial records are explored as a basis for parasequences to build this chronostratigraphy. Recognising the modern climatological diversity across the subcontinent, four broad subdivisions are used to explore potential sub-regional parasequences, which capture current climatic gradients, including the hyper-arid west coast and the decrease in aridity from the southwest Kalahari toward the north and east. These are the Namib Desert, the northern Kalahari, the southern Kalahari and the eastern fringes of the southern Kalahari. Terrestrial chronostratigraphies must start from premise that climate-driven environmental shifts may have occurred independently to those in other terrestrial locations and may be diachronous compared to the marine oxygen isotope stratigraphy (MIS), which serves as a global-scale master climatostratigraphy relating to global ice volume. The fragmented nature of preserved evidence means that we are still some way from producing unambiguous parasequences. There is however, a rich record to consider, compile and compare, within which seven broad wetter intervals are identified, with breaks between these inferred to be relatively drier. Their onset and cessation do not align with MIS: they occur with greater frequency but not with regular periodicity. Precession-paced insolation forcing has often been invoked as a key control on southern African climate, but this does not explain the pacing of all of these events. Overall the pattern is complex with some corresponding wetter intervals across...
space and others with opposing west-east trends. The evidence for drying over the past 10 ka is pronounced in the west (Namib Desert and the southern Kalahari), but not further east (the northern Kalahari or the eastern fringes of the southern Kalahari). The patterns identified here provide a framework to be scrutinised and to inspire refinements to proposed terrestrial chronostratigraphies for southern Africa. Considering changes across this large geographic area also highlights the complexity in environmental responses across space as we continue to test a range of hypotheses about the nature of climatic forcing in this region.

1. Introduction: southern African drylands and their environmental archives

Dryland and former dryland regions of the Late Quaternary in southern Africa, south of ~15°S encompass the coastal Namib Desert (Goudie and Viles, 2015) and the Kalahari region in the interior of the subcontinent (Thomas and Shaw, 1991) (yellow shading in Figure 1a). Using aridity index (AI) values, calculated by Trabucco and Zomer (2009), nearly all of southern Africa is classified as a dryland using the below sub-humid category (AI < 0.65) and even using below semi-arid (AI < 0.50), all but the eastern and northern fringes of the subcontinent (Figure 1c). Therefore, at the risk of describing all of the environmental archives within southern Africa, the focus here is on the Namib Desert and the Kalahari and its margins, with reference to some neighbouring sites for context (e.g. aeolian deposits along the western margin of South Africa, sites south of the Kalahari, and sites to the east of the Kalahari (dashed box and filled squares in Figure 2). The presence of vegetated and heavily-degraded dune features within northern Namibia, Botswana and in Zimbabwe and Zambia (Figure 1) is testament to a reduction in moisture availability compared to present in these regions in the past, including phases of the Late Quaternary, and earlier.

The Namib Desert spans ~2,000 km from the Carunjamba River in Angola (~14°S) to the Olifants River in south Africa (~32°S). It occupies a relatively narrow band (~100 to 200 km wide) on the coastal plain, bounded to the east by the Great Escarpment. Desert conditions likely established by the mid-Miocene, owing to the establishment of the Benguela upwelling system (Siesser, 1978; Tankard and Rogers, 1978; Ward et al., 1983; Partridge, 1993). The Namib Desert has four main sub-regions (Goudie and Viles, 2015): (i) the southern ‘transitional’ Namib, including the wind-eroded terrain of the Sperrgebeit, and chains of rapidly migrating barchan dunes; (ii) the ~34,000 km² Namib Sand Sea (NSS) between Lüderitz and the Kuiseb River (Lancaster, 1989a; Stone, 2013); (iii) the low gradient surface of the central Namib Plains, containing fluvially-incised terrain, desert pavement, playas and gypsum crusts (Rust and Wienecke, 1974; Eckardt et al., 2001) and inselbergs, including the Brandberg Massif (Goudie and Eckardt, 1999; and (iv) the northern Namib and
Skeleton Coast, containing the Kaokoveld highlands, loess deposits (Brunotte et al., 2009) and a coastal dunefield (Lancaster, 1982).

The Kalahari can be considered as the entire continental area that is covered with the Kalahari Group sediments, stretching from ~29°S at the Orange River in the Northern Cape of South Africa up into the Democratic Republic of Congo, north of ~6°S (Haddon and McCarthy, 2005). This group contains basal gravels and conglomerates, sandstones and unconsolidated sands, and indications of former aridity from dunes (vegetated or degraded), which can be subdivided into five dunefields (the northwest (NWK), northeast (NEK), east (EK), western (WK) and southern (SnK)), as well as pans and aligned drainage (Thomas and Shaw, 1991; Shaw and Goudie, 2002) (Figure 1a, 2).

This special issue of the South African Journal of Geology explores the basis for a preliminary chronostratigraphic subdivision of the latter part of the Quaternary (last ~190 ka) within southern Africa (south of ~15° S (Knight and Fitchett, 2021 this issue)). Hitherto, the absence of a terrestrial-based chronostratigraphy is a substantial oversight, particularly in light of the significance of southern Africa within the picture of human evolution. Therefore, this paper aims to outline and evaluate the nature of palaeoenvironmental/climatic archives and proxies within dryland, and former dryland, regions of southern Africa, and to explore the spatial and temporal patterns within those records over the past ~190 ka. Examples of the most continuous terrestrial palaeoenvironmental sequences available within this region are explored, as are points of correspondence between nearby terrestrial records that might also act as parasequences (correlation of quasi-continuous records in close proximity to build a chronostratigraphy (Gibbard and Hughes, 2021)). This is in order to identify climatically driven environmental responses to form the basis of a regional chronostratigraphic framework for the later part of the Quaternary Period of southern Africa (cf. the approach of Gibbard and West (2000) and Gibbard and Hughes (2021)). It is recognised from the outset that there is a paucity of continuous terrestrial sequences and that finding parasequences may not be straightforward given the nature of the preserved evidence and the complicated connections that geomorphological and sedimentological archives have to past climatic forcing. Four sub-regions, of the Namib Desert, the northern Kalahari, the southern Kalahari and the eastern fringes of the southern Kalahari are used in recognition of the climatological diversity across the subcontinent. In approaching terrestrial stratigraphic division, it is important to first identify major climatic shifts recorded within the landscape and palaeoecology of southern Africa, and to provide a rigorous chronology for these. These shifts can then be compared to those in other terrestrial locations and to the global master-climatostratigraphy of the marine oxygen isotope stratigraphy (MIS) (Lisiecki and Raymo, 2005). We cannot assume that the climatostratigraphy provided by the MIS are reflected in terrestrial and ecosystem processes, and should instead use rigorously dated regional chronostratigraphies to assess when changes are diachronous or asynchronous and then explore the climatic mechanisms driving these spatio-temporal
patterns. Two key challenges for this exercise in southern African dryland regions are that there is a paucity of long, and near-continuous sedimentary sequences, and that within these there is very little preserved palaeoecological evidence, owing to poor preservation conditions for organic carbon and chitin.

2. The nature of environmental archives and proxies in drylands and the chronological approaches applied

2.1 Range of archives and proxies

Archives and proxies within drylands settings include sand dunes, sand ramps former lake shorelines, fluvial sediments, lacustrine and pan sediments, calcrites, tufa carbonates, speleothems and groundwater (Table 1). There are also hyrax midden archives and also hominin middens and rock art within rock shelters, overhangs and caves. Each archive, and the proxies contained within it, ha(s)ve a particular relationship between climatic forcing and the response of the proxy. Proxies occasionally record climatic parameters (temperature, rainfall amounts) relatively directly. For example: (i) biogeochemical proxies within marine sediments can give quantitative estimates of sea surface temperatures (Sikes, 1998), (ii) δ18O in speleothems, whilst subject to the interplay of multiple controls (oceanic source water, atmospheric, soil zone, groundwater and cave system processes), are used to reconstruct temperature quantitatively and rainfall amount effects (Lachniet, 2009), and (iii) botanical-climatological transfer functions, such as for temperature and precipitation from pollen at Wonderkrater in southern African (Truc et al., 2013). More often, the environmental response of sediments or landforms to climatic forcing is qualitative. For example, the deposition of an aeolian or fluvial sediment, and the formation of an inorganic carbonate tell us about decreases or increases in moisture availability as well as changes to sediment availability and wind strength. The development of radiocarbon dating (applicable to organic carbon to identify the timing of death and some inorganic carbonates to identify formation age), U-series dating (applicable to inorganic carbonates to identify formation age), luminescence dating, particularly optically stimulated luminescence (OSL) (applicable to quartz- and feldspar-rich sediment for dating burial age) and Electron Spin Resonance (ESR) dating (applicable to quartz, carbonates, phosphates in fossil teeth and sulphates) have been central to utilising these palaeoclimatic proxies.

2.2 Geoproxies: Dryland dune systems sand ramps, mounds and ridges

Dryland dunes systems have been a key proxy within southern Africa since the late 1960s, which pre-dated the application of chronological methods and relied upon field descriptions and air photographs (e.g. Grove, 1969). The term ‘geoproxy’ for geomorphological landforms such as dryland dunes and ridges/lake shorelines was
popularised by Thomas and Burrough (2012) in the context of African drylands, although this term has perhaps not been widely adopted globally. Linear dunes are the most utilised features, owing to them being the least migratory morphology (Telfer and Hesse, 2013), and pan-fringing lunette dunes are also frequently used (e.g. Telfer and Thomas, 2006). Topographical constrained climbing and falling dunes are rarely utilised, although, the potential of sand ramps as archives have been explored (e.g. Bertram, 2003; Rowell et al., 2017). Hydrological proxies can be found in conjunction with the geoproxies, particularly where aeolian and water-lain sediment interdigitate (e.g. Teller et al., 1990; Stone et al., 2010a in the NSS and Hürkamp et al., 2011; Ramisch et al., 2017 in the southern Kalahari).

The usefulness of geomorphological proxies depends on understanding how their formation relates to climatic forcing. This connection is more straightforward for relict lake shorelines, which are driven by an increase in moisture balance compared to current arid conditions, than it is for dryland dunes. For dryland dunes, the question of exactly what these depositional aeolian geomorphological proxies tell us about environmental and climatic change remains rather complicated, as reflected in the wealth papers that investigate this (e.g. Munyikwa, 2005; Lancaster, 2008; Singhvi and Porat, 2008; Chase, 2009; Roskin et al., 2011; Thomas and Burrough, 2012; Thomas, 2013; Telfer and Hesse, 2013; Leighton et al., 2014; Lancaster, 2016 and references within for discrete regions of the world). There are five key considerations, raised by Singhvi and Porat (2008) and subsequently developed further in individual studies and in reviews of proxies by Thomas (2013) and Stone and Fenn (2020): (1) the conditions that facilitate aeolian deposition and its subsequent preservation, (2) what an age for any particular sample means, (3) record completeness (in terms of preservation and sampling strategy), (4) the need to combine chronological data adequately with stratigraphic and sedimentological data and (5) whether there is mismatch between the rate of an aeolian depositional event and the available temporal resolution of the sedimentary record (relating to the inherent resolution of the sedimentary record, sampling resolution and precision of age estimates). A related problem is how to display chronological data from dune records in order to analyse it most effectively. Telfer and Hesse (2013) develop these themes for linear dunes specifically, and reiterate messages about the importance of accepting multivariate controls on dune accumulation and the potential for response(s) to be regional, or even site-specific. Stone and Fenn (2020) explore to what extent these themes are recognised and examined within the regional reviews for the INQUA Dune Atlas chronological database (Lancaster, 2016).

Sand ramp landforms are topographically-controlled accumulations of hillslope/alluvial and fluvial sediments alternating with aeolian sediments, whose complexity requires detailed understanding of their formation (Bateman et al., 2012). A detailed study of ~75 sand ramps in southern Namibia by Rowell et al. (2017) identifies four classes, based on setting and morphology. From this general conditions for sand ramp formation are identified: (i) a suitable accommodation space against a topographic obstacle, (ii) sufficient sediment
supply, (iii) winds above a threshold for aeolian sand transport and with a directional-coherence to blow
towards the topographic obstacle and (iv) a semi-arid to arid climate with variability at seasonal, or longer
intervals, to drive the variations in the dominance of the different geomorphic processes in operation (Rowell
et al., 2017). Rowell et al. (2017) characterise four sedimentological types within sand ramps, including those
that are dominated by slope-processes, those that are purely aeolian-related, those which represent a mix of
slope and aeolian processes, and calcrite duricrusts units. These same units types are identified within ten
sand ramps, and OSL dating of selected units was undertaken (Rowell et al., 2017).

Lake shorelines have been classified as geoproxies by Burrough and Thomas (2009) and Thomas and Burrough
(2012), on the grounds that they appear as ridges with distinct morphologies, as opposed to being lake-bed
sediments. However, they are perhaps most sensibly grouped as hydrological proxies, noting that they also
require sufficient wind speeds for wave action to assist shoreline building.

2.3 Hydrological proxies

Hydrological proxies include the shorelines of former lakes (e.g. Burrough et al., 2009), and lake basin
sediments (Buch and Rose, 1996; Miller et al., 2010) or smaller pans (e.g. Lancaster, 1978; Telfer et al., 2009;
Schüller et al., 2018; Lukich et al., 2020). There are also sediments for formerly more extensive fluvial systems
in the Namib Desert that flowed east to west (see Stone and Thomas, 2013 for a review), whilst in the Kalahari
studies of such deposits include rivers feeding the Okavango-Makgadikgadi system (e.g. Nash et al., 1997),
studies and also the lower catchment of the Molopo River (Heine, 1990; Hürkamp et al., 2011; Ramisch et al.,
2017). There are dry valleys and misfit ephemeral streams in the southern Kalahari, whose morphology
supports a role for groundwater processes, but dating these is challenging (Shaw and deVries, 1988; Nash et
al., 1994a,b). Studies of the fluvial systems also include a focus on interconnections between fluvial and aeolian
sediments, as is also the case for lacustrine and aeolian sediments at Tsodilo Hills in the northern Kalahari
(Thomas et al., 2003).

Lake shorelines

The altitude of a former lake shoreline gives a clear indication of the extent of the lake at the time of formation,
as well as indicating sufficient wind speeds for wave action that assists shoreline building (Burrough and
Thomas, 2009; Thomas and Burrough, 2012). However, the record is filtered by internal lacustrine mechanisms
that sit between climatic drivers and response, such as currents within lakes (e.g. Cohen, 2003) and the
preservation of subsequent shorelines is influenced by factors including the amount of cementation by
carbonates (which increases resistance to erosion). As with dunes, there can be a mismatch between the rate
of shoreline building event and the available resolution from the record. The lacustrine record of the Okavango-Makgadikgadi rift zone in northern Botswana has a very complex set of hydroclimatic drivers (Burrough et al., 2009; Moore et al., 2012). The inflow through the Okavango River has a catchment that extends to ~ 10°S, which means precipitation outside of the northern Kalahari will contribute to river inflows (Burrough et al., 2009; Moore et al., 2012) and groundwater discharge (McFarlane and Long, 2015) into the Makgadikgadi lake systems, as well as increases in precipitation (and reduced evapotranspirative losses) local to the northern Kalahari. Furthermore, uplift-driven drainage evolution and river diversions through time have altered the importance of inflow from other river inflows including the Proto-Kafue, Kafue-Machili, Cuando, Upper Zambezi and Middle Zambezi, influencing the Late Quaternary drainage evolution of these lakes (Moore et al., 2012). This means it is important to compare any lake-full phases with other northern Kalahari archives and proxies to help determined whether the hydroclimatic shifts were local to the northern Kalahari, rather than solely over Angola.

Lake-bed and pan sediments,

Lake-bed sediments can provide rare opportunities for preservation of palaeoecological proxies, particularly where the current-day climate is toward sub-humid or humid. For example, a sediment core spanning ~16 ka within Lake Ngami contains pollen, spores and microscopic charcoal proxies (Cordova et al., 2017). Sedimentological analysis of lake-bed and pan sediments can provide a rich record of palaeoenvironmental change: (i) variations in grain size relate to depositional process(es) and sediment dynamics (e.g. Hartmann and Flemming, 2007); (ii) variations in organic content and carbonate content provides insights into variations in moisture availability (e.g. Lukich et al., 2020) and (iii) mineralogical analysis gives further details into sediment provenance as well as post-depositional alteration processes (e.g. Vainer and Matmon, 2018; Lukich et al., 2020). This is enhanced by micromorphological analysis of the sediments (e.g. Telfer et al., 2009; Lukich et al., 2020), particularly when a microfacies model (Flügel, 2004) is employed to describe and interpret the sediment formation and alteration processes. Sedimentological analysis has been undertaken at Lake Ngami (site 39) (Cordova et al., 2017), five pans along a north-south gradient in the southern Kalahari (with focus on Omongwa Pan (site 46) and Brandadam East Pan (site 49) (Schüller et al., 2018)), Mamatwan Mine (site 54) (Vainer and Matmon, 2018), and Kathu Pan (site 57) (Lukich et al., 2020). Detailed sedimentological analysis helps untangle some of the complex interactions that pans may have experienced between groundwater and surface water. Whilst the paucity of accessible exposures and the presence of depositional hiatuses may have resulted in a current under-utilisation of pan sediments, their ubiquity, particularly in the southern Kalahari means that they are an archive with great spatial coverage (Lukich et al., 2020). Lacustrine and pan sediments may often be associated with pedogenic calcretes and groundwater calcretes, as can fluvial and alluvial fan sediments (Alonzo-Zarza and Wright, 2009). Owing to the role of infiltrating water in the vadose zone for the
pedogenic calcretes and the water table and capillary fringe for groundwater calcretes, they are considered in section 2.4 on hydrogeological proxies.

Fluvial sediments

As with lacustrine and pan sediments, fluvial sediments can also preserve organic deposits, particularly in the wetland, and former wetland, regions of the northern Kalahari in the Okavango Basin (Goyder et al., 2018) and Ncamasere Valley (site 34) (Nash et al., 1997). The ephemeral rivers of the southern Kalahari contain a number of fluvially eroded scarps, offering stratigraphies of fluvial and fluvial-aeolian deposition, particularly around the confluence of the Nossob and Molopo Rivers (site 53) and further south in the Molopo River (Hürkamp et al., 2011; Ramisch et al., 2017). Along with slope-deposits these dated stratigraphies reveal variations in the intensity of flow and rainfall inputs. There are also a small number of radiocarbon ages for calcretes, freshwater snail shells, ostrich shell and charcoal from dry river valleys (mekgacha sites) in the catchment of the Okavango-Makgadikgadi system (site 35 and site 44), the Okwa River (site 45) in the northern Kalahari and the Kuruman River (site 55) in the southern Kalahari (Shaw et al., 1992).

There are twelve major ephemeral rivers in the Namib Desert (Jacobson and Jacobson, 2013) (Figure 1a, 2), with flow in the Hoarusib reaching the sea almost every year (Srivastava et al., 2005). These rivers contain terraces sequences with a range of older boulder-, or gravel-conglomerates and younger deposits of gravel, sands and silts, which suggest the competence of these fluvial system has declined through time (Korn and Martin, 1995). There has however, been controversy over palaeohydrological and palaeoclimatic interpretation of fluvial silts since the 1970s, relating to climatic forcing-response versus intrinsic geomorphic thresholds and the role of dune-river channel interactions (Lancaster, 2002). For example, the Homeb Silts (site 9) have been interpreted as: (i) dune-dammed lake sediments (Rust and Wienke, 1974), (ii) floodwater deposits during fluvial aggradation (Smith et al., 1993; Ward, 1987) or as (iii) slackwater flood deposits during either intense rainfall events (Heine and Heine, 2002), or during a transition from arid to humid climatic conditions (Srivistava et al., 2006), or (iv) end-point indicators of aridity (Marker and Muller, 1978). More technical approaches to analysis that employ lithofacies approaches offer great promise for reliable palaeoclimatic interpretations (see review by Stone and Thomas (2013) for the Namib fluvial systems). Lithofacies approaches combine sedimentary structures in three-dimensions along with grain size analysis to distinguish between channel and floodplain depositional environments and reconstruct flow energetics (Miall, 1987). This facilitates a nuanced palaeoenvironmental reconstruction with a reliable link to climatic forcing. This should help to overturn the tendency for regional palaeoclimatic syntheses, such as Chase and Meadows (2007), to exclude records from fluvial silts.
2.4 Hydrogeological proxies: pedogenic and groundwater calcretes, tufa, speleothems and groundwater archives

Pedogenic and groundwater calcretes

Calcium carbonate (CaCO$_3$) is often precipitated by displacement or replacive introduction within the vadose zone (in the case of pedogenic calcretes) (Goudie, 1983), or at the water table/capillary fringe (Mann and Horwitz, 1979), which means it is logical to place these within the hydrogeological category, even though they have overlap with lacustrine, pan, alluvial and fluvial archives and processes. They require the presence of water but a moisture deficit that drives the precipitation of the calcrete, and once precipitated, a moisture deficit will help the deposit to survive subsequent leaching (Alzonso-Zarza and Wright, 2009). As with tufa, organisms can have a critical role in helping to fix the pedogenic category of calcrete, whilst groundwater calcretes are abiotic and involve interstitial cementation (displacive or replacive) within shallow aquifer systems driven by strong evapo-transpiration, CO$_2$ degassing and sometimes a common ion effect (Wright and Tucker, 1991). Sediment stratigraphies with multiple calcretes can reveal multiple periods of water availability during high evapotranspiration driven by climatic conditions that lead to a water deficit, which can be radiometrically dated (U-series and sometimes radiocarbon), allowing a record of climatic oscillations through time to be reconstructed. However, sometimes calcrete layers record a number of significant environmental changes in the same layer, because thick deposits require long periods for their formation (Wright, 2007). Strontium ratios are useful indicators for the source of calcium, owing to the strong affinity that strontium shows with calcium, which means that that calcium provenance can be identified, for example from old continental rocks, volcanic rocks, marine carbonates and aeolian dust (and other/wet meteoric fall-out) (Dart et al., 2007). Field morphology (such as the sharpness of contacts in calcrete layers) and micromorphology can also be extremely instructive for revealing the exact processes of calcium carbonate precipitation, as can their lateral variability, which gives insights into micro-topographic differences in formation and preservation of calcretes (see (Alonso-Zarza and Wright, 2009) for more details on the detailed of calcretes as Quaternary archives).

Tufa deposits

Tufa is CaCO$_3$ deposited at the surface when emerging groundwater containing dissolved CaCO$_3$ degasses carbon dioxide, which drives the CaCO$_3$ precipitation (Ford and Pedley, 1996; Pentecost, 2005). There are stromatolitic tufa preserved at the edge of pans in the Kalahari (Lancaster, 1989b). More extensive tufa deposits are found within carbonate-rich bedrock in the upper catchment of the Tsondab River catchment within the semi-arid Naukluft Mountains of the Great Escarpment (Viles et al, 2007; Stone et al, 2010b) (site
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13), the Ga-Mohane Hills (Wilkins et al., 2021) (site 56) and in the Gaap Escarpment at the southeast margin of the Kalahari (site 60) (Butzer et al., 1978). The sequences in the Naukluft Mountains provide evidence for alternations between periods of higher-than-present steady fluvial flow (depositing tufa), and periods of high-magnitude, infrequent flooding that included erosive downcutting followed by the deposition of gravel and boulder bedload, which is later cemented in tufa (Viles et al., 2007; Stone et al., 2010b).

Speleothems

Speleothems are CaCO$_3$ deposits within caves, which will only accumulate when there is sufficient precipitation to infiltrate through soil cover and the limestone bedrock above the cave. This means, like the pedogenic calcrites we can classify them as unsaturated (or vadose) zone proxies. Speleothems are found within CaCO$_3$-rich geology on the fringes of the mega-Kalahari basin and on the central Namib Plains (see Bruan et al., 2019 for an African-wide review). The presence of speleothems in drylands regions is indicative of there being higher levels of moisture availability, and stable isotopic signatures can be used to infer palaeotemperature, rainfall intensity and/or moisture source (Lachniet, 2009).

At Rössing Cave on the central Namib plains (site 8) a NNW-SSE ridge of folded Neoproterozoic granite/marble, contains bands of calc-silicate rock (Basson and Greenway, 2004), which acts as a hydraulic barrier for runoff (Geyh and Heine, 2014). Tinkas Cave (74 km southwest of Rössing Cave, not shown on Figure 2) within a dolomite ridge on the central Namib Plains has also contains speleothems, but no dates within the Quaternary (Heine, 1992). The Orumana Cave (site 12) stalagmite is in the Kaokoveld karst terrain of the Joubert Mountains (Irish et al., 2001), within the upper catchment of the Hoarusib River (Railsback et al., 2016). There are also speleothems in the carbonate succession of the Otavi Group sediments (Miller, 2008), and the Otavi Mountainland is known as the Karstveld (Irish et al., 2001), containing ~50 caves, cenotes (collapsed caves that are flooded) and dolines (Sletten et al., 2013). The Otavi speleothems are found in Dante Cave (site 30) (Sletten et al., 2013), the flooded Dragons-breath (site 27) and Harasib Caves (site 31) (Brook et al., 1998), Aikab and Aigamas cenotes (site 28) and Guinas Meer cenote (site 29) (Brook et al., 1998). Those flooded caves are as the result of a current high groundwater table, so that their speleothem record indicates periods with lower groundwater table (but enough rainfall to facilitate speleothem formation).

In Botswana speleothems have been sampled in the northern Kalahari at Drotsky’s Cave (site 36) in the dolomite marble of the Kwihabe valley (Brook et al., 1990), and at Bone Cave (site 37) in the dolomite marble of the Koanaka (!Ncumste) Hills (Railsback et al., 2018). Sites in eastern fringes of the southern Kalahari include Lobatse II Cave (site 61), within folded dolomite hills (Holmgren et al., 1995). Caves at Gladysvale (site 68), Makapan (site 64) (Hennig et al., 1983), Wolkberg (site 65) (Talma et al., 1974; Holkämper et al., 2009), Cold
Air (site 63) and Echo (site 66) fall outside the Kalahari, but provide a useful reference for the eastern Kalahari sites.

Groundwater as an archive

Palaeoenvironmental/climatic records taken directly from groundwater (below the water table) are rare within the dryland regions of southern Africa, with analyses undertaken for the Stampriet aquifer (site 47) (Stute and Talma, 1998) and the Lethakeng aquifer in eastern Botswana (site 62) (Kulongoski et al., 2004). Low-resolution time-series for groundwater is derived from radiocarbon data (e.g. Selaolo, 1998). The radiocarbon age estimates for groundwater are approximate low-fidelity, particularly where aquifers have not been completely isolated from the atmosphere, therefore not acting as a completely closed system, and if there has been some mixing within groundwater (e.g. Mook, 1980).

2.5 Hyrax middens

Hyrax middens are petrified faecal pellets and hyraceum (a urinary product) that accumulate as the result of communal latrine behaviour of generations of hyrax (Sale, 1960; Louw et al., 1972), usually on rock overhangs, which protects them from rain and erosion. There tends to be no preservation of hyraceum in more humid environments (>800 mm/y) and the middens rapidly decompose into soil, which means dryland regions of southern Africa are ideal locations to utilise these archives (Chase et al., 2012). Radiocarbon dating provides chronological control, and because the hyraceum is not in contact with old carbon there are no reservoir effects (Chase et al., 2012). In addition, there is no evidence that new carbon enters into the hyraceum after deposition, which means that the midden acts as a closed system (Chase et al., 2012), and corrections are made for isotopic fractionation using $\delta^{13}$C. There are four key midden records for the Namib Desert: Austerlitz ((site 5) spanning ~5 ka) (Chase et al., 2010), Spitzköppe ((site 6), one spanning ~11.7 ka, and a composite record back to ~32 ka) (Chase et al., 2009; 2019), Zizou ((site 14) composite of two middens spanning ~36 ka) and Pella ((site 25) a composite of three middens spanning ~50 ka) (Chase et al., 2019). Eight middens sampled from the latter three sites have been combined by Chase et al. (2019) to provide a composite Namib Desert record, and the radiocarbon ages calibrated using SHCal13 (Hogg et al., 2013).

Pollen, charcoal and phytoliths can be recovered from middens, particularly within the pellet regions, although the isotopic signatures from the hyraceum are perhaps the most powerful proxies, with potential to provide a near-continuous time-series (Chase et al., 2012). The hyraceum $\delta^{15}$N signature is a proxy for environmental water availability, as controlled by the signature derived from soil and plants, with higher levels indicating greater aridity (Chase et al., 2012; 2015a,b). Variations in soil $\delta^{15}$N have been shown to transfer through to...
both hyrax body tissue and the hyraceum, with an additional control from dietary protein, and possible additional hyrax metabolic controls also likely mediated by aridity (Carr et al., 2016). The link between soil $\delta^{15}$N and aridity is demonstrated empirically along climatic gradients (e.g. Aranibar et al., 2004; Wang et al., 2014). Nearly all soil systems (with the exception of hyper-arid soils) are open systems with respect to nitrogen, with losses via leaching, net uptake from plants and gaseous losses (Wang et al., 2014). The loss of $^{15}$N-depleted forms of nitrogen in arid conditions is most likely to involve gaseous loss of nitrogen ($\text{NH}_3$, $\text{NO}_x$, $\text{N}_2\text{O}$ and $\text{N}_2$), and short-term increases in plant uptake during pulsed-rainfall event driven soil microbial mineralisation (Wang et al., 2014).

Hyrax $\delta^{13}$C signatures reflect changes to vegetation photosynthetic pathways ($C_4$, $C_3$ and CAM (Smith, 1972)) and plant water-use efficiency (e.g. Pate, 2001), and are influenced by the control long-term variations in atmospheric CO$_2$ have over plant $\delta^{13}$C signatures (e.g. Hedges et al., 2004). Leaf-wax lipid $n$-alkane concentration (and ratios), as well as the $n$-alkane $\delta^{13}$C and $\delta$D are biomarkers that offer proxies for hydrologic conditions (Chase et al., 2012). There is a strong relationship between leaf wax $\delta$D and $\delta$D in precipitation, even with secondary influences from vegetation type and evapotranspiration (e.g. Garcin et al., 2012; Herrmann et al., 2017). Less negative (higher) $\delta$D reflects aridity, based upon the inference that $\delta$D within low-latitude precipitation is controlled by amount effects, and continentality, so that lower rainfall amounts have a higher (more positive) $\delta$D (Dansgaard, 1964).

3. Methodological approach

Dryland archives that contain at least one palaeoenvironmental/palaeoclimatic proxy in a stratigraphic sequence (within sediment, inorganic precipitate or organic deposit) and have absolute, radiometric chronological control (optically stimulated luminescence (OSL), electron spin resonance (ESR), U-series, or radiocarbon (14C) dating) are selected. Records with 14C ages from carbonates that are either close to ~40 ka, or were measured before methodological developments in laboratory preparation and analytical approaches, are excluded. This is because carbonates can be contaminated by old-carbon sources and when they have an open texture they tend to lack closed-systems behaviour, with leaching and re-deposition of carbonate during different time intervals (Holmgren et al., 1994; Shaw et al., 1997; Stone et al., 2010a). Wherever possible (i.e. when un-calibrated 14C dates are reported/published) 14C ages are presented using the SHCal13 calibration (Hogg et al., 2013). Wherever the un-calibrated data is not available in the original sources this is noted in the text for the reader. The SHCal13 is chosen because the continuous age-models for the hyrax midden composite and the Ngami pollen record use this, and avoid having to produce entire new age models for these records, although readers should note the SHCal20 is now available (Hogg et al., 2020). For the purpose of the broad trends described here, small differences in calibration will not change the interpretations made. The utility of
the available records varies as a function of the length of the record, how continuous (or complete) it is and the temporal resolution that is offered (which relates to accumulation rate, sampling resolution and the precision of the chronological control). Significant changes in the records (e.g. switches between aeolian and water-lain sediments, distinct fluvial sediment facies typologies and architecture, and shifts in geochemical properties of carbonates, groundwater and hyrax middens (Table 1)) are identified. The most continuous terrestrial palaeoenvironmental sequences are sought and correlative shifts in nearby quasi-continuous records are examined in an attempt to produce parasequences (cf. Gibbard and Hughes, 2021). This is necessary because of a paucity of long and near-continuous records, particularly within dryland southern Africa.

Southern Africa is a large and climatologically diverse sub-continent, influenced by the Indian Ocean to the east, the south Atlantic to the west and the Southern Ocean to the south, as well as the southward migration of the tropical rainbelt over the northeast, and the Congo Air Boundary over the northwest, of the subcontinent during southern hemisphere summer. Aridity in the Namib and Kalahari relates to high pressure from the descending limb of the Hadley cell. In the Namib aridity is also controlled by the cold Benguela current that suppresses evaporation, whilst the Kalahari experiences a continentality effect that lowers receipt of precipitation. There is a strong northeast-to-southwest gradient in precipitation across the Kalahari (see Figure 1(b)). Under current climatic conditions, tropical-temperate troughs are considered to be the most important summer rainfall producing synoptic system, and involve a cloud band that forms as easterly waves in the tropics link up with westerly waves to the south (Mason and Jury, 1997; Todd and Washington, 1999; Todd et al., 2004). Overall, because the current day climatology is non-uniform, it is sensible to subdivide terrestrial dryland regions within southern Africa when identifying past shifts in climate-environmental response. The choice of how to identify and delimit these regions is not straightforward. Nicholson (1986; 2001) (Nicholson et al., 2012) identified climatically homogenous regions on the basis of modern precipitation for the African continent using linear correlation matrixes, and this gives ~30 rainfall regions for southern Africa, south of 15oS (Figure 1(e)). There are not enough records to explore each of the 30 regions that falls within a dryland/ former dryland area, and a simplified series of regions are used. It is also worth noting that the rainfall zones are only a snapshot of time (modern), covering around ~100 years. These are: (i) the Namib Desert , (ii) the northern Kalahari (extending into northern Namibia, including the Etosha Pan and Otavi Mountainland, and combining the NWK, NEK and EK linear dune sub-regions within Thomas and Burrough (2016)), (iii) the southern Kalahari (combining the WK and SnK linear dune sub-regions in Thomas and Burrough (2016)) and (iv) the eastern fringes of the southern Kalahari , more or less contiguous with the border between Botswana and South Africa (Figure 1(a)). The location of proxy records considered is given in Figure 2 and Table 2.

[INSERT FIGURE 2 and TABLE 2]
4. Results and Discussion

4.1 Spatial and temporal patterns recorded in the arid environmental archives

Region (i) Namib Desert

In this sub-region the speleothem record at Rössing Cave and tufa deposits of the Naukluft Mountains are considered first, before exploring the collated fluvial sedimentary record for west coast Rivers, the sand ramp and sand dune geoproxy record and then the hyrax midden archive. Lastly, insights into environmental change from key archaeological sites are briefly referred to. Figure 3 presents the composite hyrax midden record (Chase et al. 2019) (Figure 3(a)) and then the records are organised from north to south. The long (~191 ka) Rössing Cave speleothem (site 8) is very discontinuous, with only one growth phase (120 to 117 ka) within the ~191 ka timeframe of interest (Geyh and Heine, 2014), and Naukluft Mountain tufa (site 13) indicate a long history (> 350 ka) of hydroclimatic regimes shifts with a phase of deposition between 80.8 ± 2.9 and 72.7 ± 3.3 ka (Stone et al., 2010b). A potential parasequence is provided from the combined record of fluvial lithofacies associations for Khumib (site 2) (Srivastava et al., 2004), Hoarusib (site 3) (Srivastava et al., 2005), Huab (forthcoming, Walsh et al., 2019), Hoanib (site 4) (Eitel et al., 2006), Kuiseb (site 9) (Srivastava et al., 2006), Tsondab (site 12) (Stone et al., 2010a) and Tsauchab (site 16) (Brook et al., 2006) (Figure 3(b)(e)), revealing wetter conditions superimposed on conditions arid and windy enough to maintain dunefields from:

(i) 132 ± 15 to 87 ± 7 ka for 36 m of alternating water-lain muds and aeolian-sands, and between 87 ± 7 and 68 ± 6 ka, at Narabeb (site 12) 60 km northwest of the current ephemeral Tsondab end-point.

(ii) 61.3 ± 12.0 to 24.2 ± 2.4 ka (centred ~61-41 ka and ~34-24 ka) for river-end deposits (low-energy runoff from weak summer monsoon rain) upstream in the Hoanib (site 4), overlapping with 44-40 ka and 29-20 ka Hoarusib facies (site 3) that indicate higher rainfall over upland catchments and the desert, and matched by growth in the Orumana stalagmite (site 1) ~47.5 to ~38.5 ka (Railsback et al., 2016) and low-energy fluvial reworking of sand (steady flow) in the Tsauchab (site 16) at 24.6 ± 2.1 ka.

(iii) 20.8 ± 3.2 to 15.6 ± 2.7 ka, and at 14.4 ± 2.2 ka (cross-bedded gravels grading into coarse sands in the Khumib), with similar timing to Homeb silt deposition (Kuiseb (site 9) ~15 ka (ages out of stratigraphic order 16.3 ± 2.6 to 14.2 ± 1.7 ka), and to bracketed fluvial-silts 16.9 ± 0.9 to 12.3 ± 0.7 ka at Hartmut Pan and 12.8 ± 0.8 to 12.0 ± 0.7 ka at Ancient Tracks (Tsondab (site 12)), and Orumana stalagmite (site 1) growth ~20 to 14.5 ka (Figure 3(b)(c)(e)).

(iv) 11.5 ± 0.5 to 10.5 ± 0.5 ka at Ancient Tracks (Tsondab (site 12)), matching low-energy runoff from 11.7 ± 0.6 to 4.9 ± 0.4 ka in the eastern Hoanib catchment (site 4) (at six locations), an interdune-flood unit 9.1 ± 1.0 ka (Tsauchab (site 16)), flash-flood gravels at 8.1 ± 1.6 ka (Khumib (site 2)), and Homeb Silts in the
Kuiseb River (site 9) from 6.8 ± 1.1 to 4.8 ± 0.7 ka (not in stratigraphic order Srivastava et al., 2006), or from 9.8 ± 0.85 to 6.3 ± 0.16 (Bourke et al., 2003).

Overall, there are some common periods of increased moisture availability, but not all events repeat across all catchments (Figure 3(b)(e)). This may in part be a product of a lack of sampling, or preservation. The divergence between substantially higher rains (Hoarusib (site 3)), or only weak summer rains (Hoanib (site 4)) (61-41 and ~34-20 ka) may reveal different hydrogeomorphic responses to weak climate fluctuations, echoing Gil-Romera et al.’s (2006) interpretation of the heterogeneous proxy data in this region.

Moving now to the geoproxy sedimentary archives of sand ramps and dunes, the sand ramps that fringe the NSS (sites 11, 15, 17, 19, 20, 21) contain mostly aeolian-derived medium sand with sub-angular gravels (from minor slope-process contributions) (Unit 1) (Rowell et al., 2017). Four sand ramps are punctuated by substantial slope process-derived, semi-continuous layers of cobbles and occasional boulders (unit 2), requiring wetter conditions (Rowell et al., 2017). Unit 1 deposition occurred: (ii) ≥224 to ≥175 ka at Samara, and ≥150 to ≥90 ka at both Samara (site 11) and Aus (site 20) at a time of higher moisture availability at Narabeb (Tsondab River (site 12)); (ii) ≥65 ka to ~45 ka at Neuhof 2 (site 15), and 40.4 ± 3.8 ka at Neuhof 1 (site 15), at a similar time to Hoanib (site 4) and Hoarusib (site 3) fluvial deposition and Orumana stalagmite (site 1) growth and (iii) ~30 to ~11 ka at Neuhof 1 (site 15) (with a unit 2 in between 29.6 ± 2.6 and 17.0 ± 1.9 ka) (Figure 3(f)) matching Khumib River (site 2) and Hoarusib River (site 3) sedimentation (Figure 3(b)) as well as Orumana stalagmite (site 1) growth ~20 to 14.5 ka (Figure 3(c)). Unit 4-type aeolian sands are widespread: ≥80 to 12.6 ± 1.8 ka and 12.4 ± 1.0 to 6.7 ± 0.9 ka (3.5 m and 5 m unit at Aus (site 20)), ≥75 to 12.8 ka (4 m unit at Sandkop (site 19)), ≥75 to 1.9 ± 0.1 ka (2.5 m unit at Klein Aus (site 21)), 20.6 ± 2.7 to ≤0.2 ka (2.8 m unit at Samara (site 11)), ≤1.5 ka (1.8 m unit at Jagkop (site 17)) and ≤0.1 ka (1.5 m unit at Sandkop (site 11)) (Rowell et al., 2017) (Figure 3(f)).

It is likely aeolian dune deposition in the NSS occurred throughout the Quaternary, with cosmogenic nuclide (CN) burial dating suggesting a residence time >1 Ma as sand moves from south to north (Vermeeesch et al., 2010). Dune stratigraphies exist for just three complex linear dunes locations, and this very patchy sampling can only confirm phases of aeolian deposition: ≥132.3 ± 15.0, from 111.9 ± 10.0 to 103.6 ± 9.2 ka, and after 68.3 ± 5.7 ka at Narabeb (site 12) (Stone et al., 2010a), 22.5-18 ka and 10-8.5 ka in the south (site 18) (Bubenzer et al., 2007), and dune accumulation and lateral migration eastward in the north (site 10) over the past 6 ka (Bristow et al., 2007).

Further south (~30 to ~32°S), 21 profiles within aeolian cover sands and reticulate dunes on the low-relief plain of the west coast of South Africa (site 24) have been OSL dated (Chase and Thomas, 2006; 2007). Five phases of aeolian activity/deposition are recorded within the composite dataset (rather than a focus on dating...
identified stratigraphic units) at 73-63 ka, 49-43 ka, 33-30 ka, 24-16 ka and 5-4 ka, which contrasts with the rather more continuous record of aeolian activity recorded within the sand ramp archives (Rowell et al., 2017). Chase and Thomas (2006; 2007) suggest the aeolian phases on the west coast of South Africa were driven by changes in wind strength, and influenced by variations in sediment supply. The composite-midden δ\(^{15}\)N record (Chase et al., 2019) (sites 6, 14 and 25) provides powerful insights into hydroclimate over ~50 cal kyr B.P. (age model uses the SHCal13 calibration curve) (Figure 3(a)). There are precession-paced cycles, with higher rainfall negatively correlated to austral summer insolation at 25°S and positively correlated to high-latitude summer insolation in the northern hemisphere (65°N). There is additional variability superimposed on this, with rapid (<200 year) transitions (Chase et al., 2010; 2019). Treating negative excursions in the normalised δ\(^{15}\)N as wetter intervals (Figure 3(a)), these phases are seen:

(i) 50 to 36 cal kyr B.P. (peaking 46-44 and ~40 cal kyr B.P.), showing some correspondence with fluvial records (Hoarusib (site 3) and Hoanib (site 4) (Figure 3(a)(b)).

(ii) ~32 to ~29 cal kyr B.P., overlapping with the start of a phase of fluvial deposition (Khumib (site 2), Hoarusib (site 3)) (Figure 3(a),(b)).

(iii) ~24.5 and ~22 cal kyr B.P., which accord with parts of the fluvial record (Khumib (site 2) and Hoanib (site 4)) (Figure 3(a),(b)).

(iv) ~15 to 14.3 kyr B.P., also seen in the Khumib (site 2), Kuiseb (site 9) and Tsondab (site 12) records (Figure 3(a)(e)) and a slope-process dominated unit at Neuhof-1 sand ramp (site 15) (Figure 3(f)). The start of drying ~10 kyr B.P. is at odds with the fluvial sedimentation between ~11.5 and 6.5 ka, whilst increased aridity ~5 kyr B.P. matches a paucity of fluvial deposition.

Middle Stone Age and Late Stone Age archaeological deposits within the Namib Desert offer only minimal insights into past environmental conditions (Erb Tanks (site 7), Apollo-11 (site 22) and Spitzkloof (site 23). This is because the stratigraphies are predominantly anthropogenic rather than an environmental accumulation (Vogelsang et al., 2010 (site 22); Mccall et al., 2011 (site 7) and Spitzkloof (site 23) has yet to be radiometrically dated (Dewar and Stewart, 2012). However, faunal assemblages at Apollo-11 suggests the persistence of a climate similar to today (arid to semi-arid) with some minor variations. Rockfall deposits at Apollo-11 near the base of the stratigraphy, suggest there may have been a more humid phase sometime prior to the ~70 ka (OSL dated) Still Bay layer (64 cm below in the stratigraphy), whilst the presence of only Chenopodiaceae species in charcoal in the 57.9 ± 2.6 to 42.9 ± 2.7 ka (OSL date) layer suggests slightly greater aridity (Vogelsang et al., 2010). At Spitzkloof the lowermost layer contains similar early MSA artefacts as those found beneath the ~70 ka Still Bay Layer at Apollo-11 (Dewar and Stewart, 2012). The presence of gypsum throughout the Spitzkloof sediments, and two units of a land snail also suggests arid to semi-arid conditions, with occasional episodes slightly more humid (Dewar and Stewart, 2012).
In this sub-section the hydrological proxies of mega-lake Makgadikgadi (sites 39 to 42 combined) and Etosha Pan (site 26) are the major focus, then speleothem records (sites 27, 28, 29, 30, 31, 36, 37) are considered and dune geoproxies (sites grouped as 32, as 38 and as 43). Figure 4 organises records by geographical location from west to east. A candidate for a regional chronostratigraphy for the northern Kalahari comes from OSL-ages for shoreline accumulation from three interconnected lacustrine basins (Ngami, Mababe and Makgadikgadi), which when combined formed mega-lake Makgadikgadi (Burrough et al., 2009). Burrough et al., (2009 p. 1404) argue that independent lake-full phases in Ngami or Mababe might represent “noise(y) background avulsion signal(s)” driven by flow redistribution in the Okavango Delta from tectonics and geomorphic feedback processes, whereas if flow continues (spills over the 930 m asl level) through the Boteti to Makgadikgadi it represents a more substantial regional lake. However, despite the palaeoenvironmental potential of this mega-lake it is important to reiterate that the geography and size of the Okavango catchment, which feeds surface inflow and groundwater resurgence, means that mega-lake phases record an integrated signal of precipitation over a very large area. Therefore, mega-lake phases may record wetter conditions locally (in the northern Kalahari) but may also reflect wetter conditions further north in the catchment in Angola (where the Okavango catchment starts ~10°S) (Burrough et al., 2009; Moore et al., 2012). In addition, uplift-driven river drainage evolution complicates the inputs to this system further (Moore et al., 2012). For this reason, this section uses comparisons between the mega-lake Makgadikgadi record and a range of the other archives and proxies, including Etosha Pan, speleothem records and lacustrine records to the northeast in Malawi, Tanzania and up into the Kenyan Rift Valley to help to assess the spatial scale of hydrological change (local to the northern Kalahari, versus fed from the northerly Okavango catchment).

Age clusters for mega-lake Makgadikgadi phases were determined by Burrough et al. (2009) using cluster analysis, and a Weisburg t-test to establish their statistical significance, and there are eight mega-lake phases recorded: 131 ± 11 ka, 105 ± 4 ka, 92 ± 2 ka, 64 ± 2 ka, 39 ± 2 ka, 27 ± 1 ka, 17 ± 2 ka and 8.5 ± 0.2 ka (Figure 4(g)). The 39 ± 2 ka phase has correspondence from a lakebed sedimentary record from Lake Ngami (site 39) (Huntsman-Mapila et al., 2006) and a basin at Tsodilo Hills (site 33) from 36-32 ka (Robbins et al., 2000; Thomas et al., 2003). The 27 ± 1 ka phase is consistent with dated lake-related sediments 27-22 ka and 19-12 ka (Robbins et al., 2000; Thomas et al., 2003), as does Boteti River (site 41) backflooding at 28 ± 2 ka (Shaw et al., 1997). Fluvial deposits within the now-dry Xaudum Valley (site 35) date to 18.1 to 17.3 cal kyr B.P. and from the Gidikwe Ridge/Okwa Gorge (site 44) date to 18.0 to 17.2 cal kyr B.P and 17.5 to 16.5 cal kyr B.P. (all calibrated for this paper using SHCal13 in OxCal 4.4) (Shaw et al., 1992) correspond to the 17 ± 1 ka mega-lake Makgadikgadi phase. There is later phase of fluvial deposition at Gidikwe Ridge/Okwa Gorge (site 44) between 1.40 and 13.5 cal kyr B.P. (all calibrated for this
paper using SHCal13 in OxCal 4.4) (Shaw et al., 1992). The pollen-derived CREST-based reconstruction of summer zone precipitation at Lake Ngami (site 39) suggests conditions were similar to today, with > 300 mm/y rainfall from 16.6 to 12.5 cal kyr B.P., followed by a phase of lower rainfall (12.5 to 10.0 cal kyr B.P.), reaching a minimum of ~150 mm/y, before rising back to > 300 mm/y and persisting to present day (Cordova et al. (2017) who use the SHCal13 calibration in their age model).

Etosha Pan (site 26) in northern Namibia on the western side of the northern Kalahari is fed by the rivers within Cuvelai Basin (including the Ekuma and Oshigambo Rivers) (Luetkemeier and Liehr, 2018). Evidence combined from lake shorelines, pedogenic calcrites, stromatolites, and some mammal fossils, provide a record of hydroclimatic shifts for northern Namibia and into the Angolan catchments of the Cuvelai Basin (Hipondoka et al., 2014) although these catchments only extend to ~16.5oS, rather than ~10oS for the Okavango-Makgadigkadi system. Five periods of high lake level (21 to 24 m) at Etosha (site 26) are recorded, at ~75-70 ka, ~34-27 ka, ~23-21 ka, ~18-16 ka and ~10 ka, derived from OSL ages of western shoreline (Hipondoka et al., 2014) (Figure 4(a)). Shallow lake conditions (of a few m) occurred in the Holocene at 7.4 ± 0.3 ka and from 5.1 ± 0.2 to 0.27 ± 0.02 ka (Hipondoka et al., 2014), with a 14C age estimate from a semi-aquatic antelope (quagga) pelvis dated to between 5.0 to 4.5 cal kyr B.P. (calibrated for this paper using SHCal13 in OxCal 4.4) (Hipondoka et al., 2006). Faunal remains in the Oshigambo-loess on the west side of the northern shore of Etosha include terrestrial gastropods, ostrich eggshell and termite remains, all indicating sub-humid to semi-arid conditions. Bovid bone and enamel dates to 12.8-12.6, 13.4-13.1 and 14.1-13.8 cal kyr B.P. (Pickford et al., 2009) (calibrated for this paper using SHCal13 in OxCal 4.4). Another reconstruction for Etosha Pan (site 26) is provided by De Cort et al. (2021), who apply a framework to incorporate chronological and proxy-interpretation uncertainties and present six qualitative lake-status classes for the past 25 ka (a higher number indicates a deeper lake or surface area). The class 6 lake status (highest lake level) occur 21.8 to 18.3 cal kyr B.P., 17.7 to 15.7 cal kyr B.P., and 10.3 to 0.1 cal kyr B.P. (De Cort et al., (in press) use SHCal13), corresponding with the ~23-21 ka, ~18-16 ka and ~10 ka lake shoreline phases of Hipondoka et al. (2014) and a further one ~3.5 cal kyr B.P. and three class 5 lake phases which fill some of those gaps, with the other lake class status revealing a more complex lake hydrology than considering shoreline data only (Figure 4(b)). The only overlapping phases of high lake level between Etosha Pan (site 26) and mega-lake Makgadigkadi (sites 39-42) are ~34-27 ka at Etosha (overlapping with mega-lake Makgadigkadi phase 27 ± 1 ka) and ~18-16 ka at Etosha (corresponding with mega-lake Makgadigkadi phase 17 ± 2 ka) (Figure 4). This suggests these phases included increases in precipitation over the northern Kalahari, as well as further north in the large Makgadigkadi catchment.
Deposition of speleothems in now flooded caves and cenotes in the Otavi Mountainland (sites 27 to 31) occurred during eight intervals (indicating a lower groundwater table than today, but high enough moisture availability for speleothem growth): 130 ± 7 ka, 112 ± 5 to 108 ± 7 ka, 61 ± 2 ka, 31 ± 2 ka, 29 ± 1 to 28 ± 1 ka, 15 ± 1 to 14 ± 0.3 ka, 10.6 ± 0.4 ka, and 8.8 ± 0.5 to 7.5 ± 0.3 ka (U-series dated in Brook et al., 1999) (Figure 4(c)). Those at 31 ± 2 ka and 29 ± 1 to 28 ± 1 ka, and at 10.6 ± 0.4 ka correspond with high lake level (21 to 24 m) at Etosha Pan (site 26). Five of the eight phases of Otavi speleothem growth show correspondence/overlap with mega-lake Makgadikgadi: 130 ± 7 ka (corresponding with mega-lake Makgadikgadi phase 131 ± 7 ka), 112 ± 5 to 108 ± 7 ka (overlapping with mega-lake Makgadikgadi phase 105 ± 4 ka), 61 ± 2 ka (overlapping with mega-lake Makgadikgadi phase 64 ± 2 ka), 29 ± 1 to 28 ± 1 ka (overlapping with mega-lake Makgadikgadi phase 27 ± 1 ka), and 8.8 ± 0.5 to 7.5 ± 0.3 ka (overlapping with mega-lake Makgadikgadi phase 8.5 ± 0.2 ka) (Brook et al., 1998) (Figure 4(c)).

Further west, some phases of speleothem growth at Drotsky’s Cave (site 36) and Bone Cave (site 37) in northwest Botswana (>450 km east of Etosha Pan) correspond with Etosha high lake levels - those at 76.8 ± 3.5 to 74.9 ± 2.6 ka, 22.8 ± 4.2 ka, 16 ± 0.9 ka and 5.4 ± 1.4 ka (Brook et al., 1990; Railsback et al., 2018) (Figure 4(f)). The 131 ± 11 ka, 92 ± 2 ka, 39 ± 2 ka and 8.5 ± 0.2 ka mega-lake Makgadikgadi phases correspond with speleothem growth at Drotsky’s Cave (site 36) (U-series dates of 132.9 ± 26.6 ka, 92.9 ± 5.9 ka, 43.9 ± 7.3 ka and 8.2 ± 0.5 ka (Brook et al., 1998)), and the 8.5 ± 0.2 ka phase is also seen at Bone Cave (site 37) (Brook et al., 1998) (Figure 4(f)).

A 4.6 ka (U-series dated) long stalagmite from a non-flooded chamber in Dante Cave (site 30) (Sletten et al., 2013) indicates a gradual transition from wetter ($\delta^{18}$O -12 to -10.5 %) to drier conditions between 4.6 and 3.3 ka ($\delta^{18}$O 10 to -9.5 %), supported by growth-rate and petrological data (Figure 4(d)). This was followed by a pronounced very dry period until 1.8 ka, and since then getting wetter, but with variable wet and dry intervals (100-200 years long) (Sletton et al., 2013). The Ncamasere Valley fluvial deposits (site 34) also record a period of back-flooding between 3.9 and 3.2 cal kyr B.P., as well as phases 0.99 to 0.80 cal kyr B.P. and 0.11 to 0.57 cal kyr B.P. (calibrated for this paper using SHCal13 in OxCal 4.4) (Nash et al., 1997).

To summarise, all eight of the mega-lake Makgadikgadi phases correspond with records of higher moisture availability in one, or other of the records from Etosha Pan (site 26), the Otavi Mountainland cenotes (sites 27 to 31) or Drotsky’s Cave (site 36) and Bone Cave (site 37) (Figure 4). The 131 ± 11 ka phase is seen in speleothems at Drotsky’s cave and further west in the Otavi Mountainland, as is the 8.5 ± 0.2 ka phase. The 27 ± 1 ka and 17 ± 2 ka phases are seen as lake phases in Etosha Pan, and there is overlapping ages with Otavi Mountainland speleothems. Of course, this does
not preclude the scale of hydrological change for these mega-lake phases including higher precipitation in the north of the Okavango catchment.

Lacustrine records outside the northern Kalahari are also instructive in assessing the scale of hydrological change associated with mega-lake Makgadikgadi phases. Regionally wetter conditions during the 17 ± 2 ka phase are recorded at Lake Chilwa (17 ± 1 ka) (Barker et al., 2007) and Lake Malawi (18-17 ka) (Thomas et al., 2009) in Malawi, and further afield at Lake Tritrivakely in Madagascar at 17 cal ky B.P. (Gasse and van Campo, 2001). However, further back in time Lake Malawi and Lake Tanganyika (western border of Tanzania) show low lake levels 135-127 ka and 110-85 ka (Scholz et al., 2007), which lead Burrough et al. (2009) to postulate that positive and negative hydrological extremes may have been occurring during these periods, driven by enhanced variability in climatic conditions (precession-forced insolation changes favouring climatic variability and extremes). Moving further north to the central Kenyan Rift Valley Lakes, all but the 39 ± 2 ka mega-lake Makgadikgadi phase correspond with high levels in the central Kenyan Rift Valley Lakes, particularly the Lake Naivasha chronology (Trauth et al., 2003) (Figure 4(j)).

Dune activity/accumulation across the three dunefields in the northern Kalahari shows dune activity over much of the past 60 ka, with some phases that lack ages in different sub-regions (Figure 4(e)(h)). For example: 60-35 ka, 25 to 18 ka, 14 to 11 ka, 8 to 6 ka, and after 3 ka in the northwest (sites collated as 32); 55 to 50 ka, and after 21 ka in the eastern Kalahari (sites collated as 43) and 60-55 ka, 25-23 ka, 19 to 18 ka, 17 to 15 ka, 12 to 11 ka and after 8 ka in the northwest Kalahari (sites collated as 38). Given these dunes are now heavily vegetated, these terminal dates may represent conditions remaining sufficient wet. In addition, the heavy degradation of dunes in this part of the Kalahari, with evidence for tectonic modification and collapse (McFarlane et al., 2005) require the completeness of the potential (sample-able) chronostratigraphy to be questioned, and the role of post-depositional mixing for OSL dating to be more carefully considered.

**Region (iii) the southern Kalahari**

In this sub-region the long sedimentary record at Mamatwan Mine (site 54) is considered first, alongside the record from the dune geoproxy archive (sites grouped as 48 and sites grouped as 50), and then the semi-continuous stratigraphic sequence preserved within sinkholes at Kathu Pan (site 57), alongside nearby archaeological sites that contain palaeoenvironmental records. Following this, a range of other pan sedimentary records, and pan-fringing lunettes are considered. These include five pans, orientated on a north-south gradient in the southern Kalahari that provide quasi-continuous sedimentary archives, albeit with variable sedimentation rates (0.4 and 26.0 cm/ka), and particularly Omongwa pan (site 46) and Brandamm East pan (site 49; Witpan (site 51) and Koppieskraalpan (site 52). These are considered alongside the Molopo...
River sedimentary records (site 53) where fluvial and aeolian records are interbedded. Finally, the insights from the Stampriet Aquifer record are briefly considered. Figure 5 plots records from the southern Kalahari as well as the eastern fringes of the southern Kalahari (region iv) and records further to the east.

The ~50 m sediment section at Mamatwan Mine (site 54) is a potential candidate for a regional chronostratigraphy. Bateman et al. (2003) described its: basal red clay (Unit I), indicating lacustrine conditions, manganese palaeosol (Unit II, ~2 m), calcareous breccia (Unit III, 10 m), calcrites with a range of morphologies (Units IV to VI totalling ~15 m), mottled sandstone (6.5 m Unit VIII with the lower 5 m bioturbated Unit VII), further calcrites (Unit X, ~2 m), and unconsolidated Kalahari sands (Unit XI ~4 m preserved) (Bateman et al., 2003). Bateman et al.’s (2003) OSL chronology covers the uppermost ~9 m: (1) 527-286 ka for Unit VII (large age range from OSL signal approaching saturation and dose rate uncertainties), (2) 160-108 ka for Unit X, with subsequent calcretisation via groundwater processes, for which there is no age estimate, and (3) 44 ± 5.4 to 6.2 ± 2.2 ka for sand sheet deposition. An updated sedimentological description by Matmon et al. (2015) and Vainer and Matmon (2018) proposes a broad three unit subdivision of: (1) clayey-gravel through to mudstone (53-34 m depth), indicating a fluvial environment replaced by a shallow, brackish, alkaline lake; (2) calcareous, angular gravels, covered by calcrite (34-15 m depth), indicating a high-energy fluvial system giving way to highly evaporative conditions and (3) calcareous sandstone with a palaeosol and topped by unconsolidated sands (15 m depth to surface). The new cosmogenic nuclide chronological control suggests that the lowest unit was deposited around 1.1 to 1.2 Ma (1.18 ± 0.09 Ma and 1.10 ± 0.09 Ma via two different approaches) (Matmon et al., 2015).

The record of linear dune accumulation for 264 OSL ages (217 ages in southern-Kalahari (sites grouped as 48) and 47 ages in the western-Kalahari (sites grouped as 50)) compiled by Thomas and Burrough (2016), puts the OSL ages for the sandstone and unconsolidated sands at Mamatwan into context, showing relatively continuous accumulation over ~190 ka, with an absence of ages (perhaps preservation-related) only during 174-107 ka, 96-87 ka and 38-42 ka (Figure 5(c)). This demonstrates that the apparent episodic dune formation (and aridity) proposed by Stokes et al. (1997) was a spurious product of a very low sampling density within a heterogeneous landscape (e.g. Stone and Thomas, 2008; Stone, 2010; Stone and Larsen, 2011). The oldest ages (>183.3 ± 18.3 and >186.2 ± 15.6 ka) at 6.5-8.0 m depth, east of Stampriet, are near OSL signal saturation (Stone and Thomas, 2008), and Vainer et al. (2018) use a novel, combination of OSL and cosmogenic nuclide burial dating at Mamatwan (site 54) to extend the age range. Modelling reveals that measured cosmogenic nuclide concentrations require up to 22 cycles of sediment and exposure, suggesting dunes accumulated here as early as 4 Ma.
The detailed sedimentological analysis of pan sediments exposed within sinkholes at Kathu Pan (site 57) by Lukich et al. (2019; 2020) may provide an even stronger candidate for a regional chronostratigraphy. Together, three exposures reveal a semi-continuous record spanning nearly 160 ka. A basal aeolian sand unit containing palygorskite is OSL dated from 156 ± 11 ka to 121 ± 6 ka at section KP6, indicating intermediate (semi-arid conditions) (Figure 5e yellow bars on left), whilst the top of a basal sand unit is dated to 119 ± 7 ka at section KP1. At the nearby Ga-Mohana Hill North Rockshelter (site 56), cascade tufa, U-series dated from 110.6 ± 3.0 to 102.1 ± 2.1 ka, indicate abundant water flowing down the hillside, associated with an assemblage of flaked stone Middle Stone Age artefacts (sedimentary associated OSL dated to 105.6 ± 6.7) (Wilkins et al., 2021).

South of the Orange River, the Bundu Farm archaeological site (site 59) contains laminate calcretes related to fluctuations in water table in a pan setting, although there is no chronological control at this site (Hutson, 2018). The wet phase at Go-Mohana Hill is not recorded in the Kathu Pan stratigraphies, where a sand matrix continues until 74.5 ± 5 ka (section KP6), 55 ± 3 ka (section KP9), and 32 ± 3 ka (section KP1, although this OSL sample has higher overdispersion, which may be indicative of post-depositional mixing, so that this age may be too young) (Lukich et al., 2020). Within these sands there are pedogenic units OSL dated (or at least their host sand sediment is dated, noting the host sediment predates the formation of the soil into that sediment) to 95 ± 6 ka (section KP6), 96 ± 5 ka (section KP1) and 84 ± 4 ka (at KP9), which indicate wetter, marshy conditions during two discrete intervals (Lukich et al., 2020) (Figure 5e, blue bars). A massive hardpan unit (at section KP6) and CaCO₃ clasts (at section KP1) are dated to 22.2 ± 1.2 ka and between 32 ± 2 and 10.4 ka, respectively (again noting the high overdispersion in ~32 ka age), and Lukich et al. (2020) interpret this as an increase in aridity, driving the evaporation needed to form the CaCO₃ (red bars in Figure 5e). The upper part of the sections (< 10.4 ± 0.5 ka at KP1, < 5.7 ± 0.3 ka at KP6 and < 8.1 ± 0.5 ka at KP9) remains dominated by deposits of CaCO₃, rather than palygorskite, which suggests the climate continued to be relatively arid (Lukich et al., 2020 (Figure 5(e)). The presence of some organic-rich layers (between 2.7 and 0.81 ka at KP9 and 5.7 and 2.0 ka at KP1) indicates some fluctuations in water availability (perhaps annually, or even seasonally) (Lukich et al., 2020). Fluvial deposits within the Kuruman River (site 55) date to 3.2 to 2.7 cal kyr B.P., 1.8 to 1.5 cal kyr B.P., and 0.6 to 0.5 cal kyr B.P. (all calibrated for this paper using SHCal13 in OxCal 4.4) (Shaw et al., 1992).

In contrast, geochemical proxies at Omongwa Pan (site 46) indicate wetter, and windy conditions, at ~ 42 cal kyr B.P. (caution as un-calibrated data not available and the calibration curve used is not stated by Schüller et al. (2018)), which then shifts to drying, with a peak in dryness 28-27 cal kyr B.P., whilst further south at Branndam East (site 49) there was a later and lower magnitude drier excursion ~19.5 to 18.0 cal kyr B.P., and another ~15.0 to 13.0 cal ky B.P (again caution as noted above) (Schüller et al. (2018) (Figure 5(a))). Further south still, later relatively wet phases are recorded, within OSL-dated pan floor sediments at Witpan (site 51) at 32 ± 4.6 ka (0.8 m depth in the northern sector), as well as at ~20 ka (50-70 cm depth in the southern sector)
(Telfer et al., 2009) (Figure 5(d)). A sedimentological break within the Koppieskraalpan lunette (site 52) from 28.8 ± 1.9 to 0.17 ± 0.02 ka is interpreted by Hürkamp et al. (2011) as evidence for wetter conditions (a flooded pan surface) over a longer period of time. Complex lunette deposition at Witpan (site 51) occurred from 31.9 ± 2.2 to 23.1 ± 1.5 ka, (Telfer and Thomas 2006) (Figure 5(d)), which indicates a shorter period of pan flooding (perhaps with fluctuations in pan water level) at Witpan (requires a dry pan surface) than the Koppieskraalpan lunette (site 52) and pan stratigraphy (Hürkamp et al., 2011). The drying trend at Omongwa pan (site 46) shows a stepped increase ~8 cal ky B.P. and peaking ~5.5 cal ky B.P., with a minor drier excursion ~8 cal ky B.P. also seen at Branddam East (site 49) (Schüller et al., 2018) (Figure 5(a)). Further complex lunette deposition at Witpan (site 51) occurred from, 6.2 ± 0.2 ka, and 2.4 ± 0.1 ka to 0.07 ± 0.01 ka (Telfer and Thomas 2006) (Figure 5(d)). There is a final dry peak at Omongwa (site 46) at ~1 cal ky B.P. (Schüller et al., 2018) (Figure 5(a)).

The wetter-drier shifts indicated at Kathu Pan (site 57), Omongwa Pan (site 46), Branddam East Pan (site 49), Witpan (site 51) and Koppiskraalpan (site 52) are not consistently mirrored in the record of interbedded fluvial sediments and aeolian sediments in the lower Molopo River (site 53). These contain fluvial units of 23.4 to 22.8, 22.3 to 21.7, 21.0 to 20.3 and 16.2 to 15.7 cal kyr B.P. (calibration of Heine’s (1990) ages here using SHCal13 in Oxcal 4.4.) (Figure 5f). The sediments within three sections in the lower Molopo Canyon (site 53) suggest that the Molopo was ephemeral over the past 10 ka (Ramisch et al., 2017). Valley sediment aggradation occurred between ~9 and ~ 6.5 ka (phase 1) with horizontal bedding and absent grading indicating high-magnitude, short-duration events (flash flooding) (Figure 5f). This was followed by alluvial fan aggradation from 6.5 ± 0.6 ka to ~1.5 ka, without any channel deposits, indicating lower-intensity rainfall before a final stage of valley sediment accumulation from ~0.51 ka to 0.15 ka (Figure 5(f)). At the time of the drying trend as Omongwa Pan ~1 ka (site 46), there is some fluvially-reworked dune material in the lower Molopo River (site 53) at 1.08 ± 0.09 ka, which could indicate seasonal flows, covered with young (0.42 ± 0.04 ka) aeolian dune-sand (using OSL) (Hürkamp et al., 2011) (Figure 5(f)).

Recharge to the Stamppriet Aquifer (site 47) occurred from ~42 to 27 kyr B.P. and since ~23 kyr B.P. (calibrated here using the SHCal13 from Stute and Talma (1998) noting the low fidelity dating of groundwater dating)), and a shift in noble gas-derived air temperatures from ~21.2°C prior to ~ 19 cal kyr B.P., compared to the 26.5°C average for the last ~12 cal kyr B.P. (calibrated here using the SHCal13 from Stute and Talma (1998) although noting the low fidelity dating of groundwater dating), (Figure 5(b)). δ¹⁸O signatures of -5.75 to -6.00 ‰ in that later phase, compared to -6.50 to -7.20 for the earlier phase, are interpreted as indicating an Atlantic Ocean moisture source (winter rainfall), as opposed to the current Indian Ocean source (summer rainfall) because the former is subject to a smaller continental effect on isotopic depletion. Wetter conditions at Wonderwerk Cave (site 58) are recorded from ~35.2 to 31.1 cal kyr B.P., ~23.6 to 17.2 cal kyr B.P., (calibrated here using the
SHCal13) (Brook et al., 2010), coinciding with both recharge to the Stampriet aquifer (site 47) (Figure 5b) and fluvial deposition within the Molopo River (site 53 (Figure 5f)).

Region (iv) eastern fringes of the southern Kalahari

In this region, the majority of evidence comes from speleothem records, which includes Lobatse II cave (site 61) in the eastern fringes of the southern Kalahari. Lobatse II is considered first, alongside the caves in the Gauteng Malmani dolomite further to the east (Gladysvale Cave (site 68), Cold Air Cave (site 63), Wolkberg Cave (site 65)), which lie beyond the southeastern margin of the Kalahari Group Sediments (Figure 2). Next, the 90 m sedimentological record from Tswaing Crater (or Pretoria Saltpan) (site 67), also outside the Kalahari margins is considered briefly (further detailed are given elsewhere in this special issue, for example, Knight and Fitchett, (2021 this issue)). The palaeoenvironmental records from the nearby key archaeological sites of Florisbad (site 69) and Erfkoon (site 70) are also considered here. Finally, the insights from the Lethlakeng Aquifer record (site 62) are briefly considered.

The U-series dated Lobatse II speleothem (site 61) has a growth period from ~51-43 ka, a hiatus from ~43 and 27 ka (indicating dry conditions), and growth from ~27 to 21 ka, with δ\(^{18}\)O indicating the first period was warmer and more humid than the second, and that the drying trend continues until growth ceases ~21 ka (Holmgren et al., 1995) (Figure 5(h)). Whilst the Gaap Escarpment tufa (site 60) and their morphostratigraphy are clearly mapped and described (Butzer et al., 1978), the poor fidelity 14C chronology precludes their inclusion here. The caves of Gladysvale (site 68 within the Cradle of Humankind World Heritage Site), Cold Air (site 63), Wolkberg (site 65) and Echo (site 66) all lie further east than the Kalahari Group Sediments (Figure 2). However, a comparison with speleothem growth periods at these sites is useful to determine whether regional climate conditions in the east of South Africa were governing growth. This is particularly useful as local changes in percolating water flowpath direction also influences speleothem growth (Holzkämper et al., 2009). Echo Cave (Brook et al., 1982) (site 66) and Makapans Cave (site 64) are not considered further given limitations of speleothem 14C chronologies (e.g. Holmgren et al., 1994; Hennig et al., 1983). Wolkberg Cave speleothems (site 65) formed from 57.9 ± 0.6 to 46.3 ± 0.3 ka and ~40.1 ± 1.3 ka, and Holzkämper et al. (2009) suggested deposition only occurs during medium-to-high summer insolation (December, 30°S) (Figure 5(i)). At Gladysvale (site 68) there is growth at 56.8 ± 0.35 ka, and at 42.8 ± 6.8 ka (Figure 5(k)) during the hiatus seen at Wolkberg (Pickering et al., 2007) (Figure 5(i)). At Cold Air Cave (site 63), growth is recorded from 24.2 to 12.7 ka and again from 10.2 ka to present (Holmgren et al., 2003) (Figure 5(j)). In the latter phase, δ\(^{18}\)O starts at ~2‰ with a trend to more negative (~5 to ~6‰), interpreted as a shift from persistent rain from mid-altitude...
stratiform clouds (more positive $\delta^{18}O$) to more sporadic rainfall from thunderstorms and short-intensity rainfall events during overall Holocene drying (Holmgren et al., 2003). A similar trend of growth is seen at Gladysvale (site 68) at 16.50 ± 0.15 to 14.39 ± 0.06 ka and then 10.32 ± 2.68 and 7.45 ± 0.35 (Pickering et al., 2007) (Figure 5(k)). The higher resolution studies of speleothems T5 (4.4 ka) and T7 (3 ka) in Cold Air Cave (site 63) reveal fluctuations in $\delta^{18}O$ between -2.0 and -5.5 ‰ (T5 Repinski et al., 1999) and -2.0 and -6.0 ‰ (T7 Stevenson et al., 1999), with some regular periodicities, and record the cooler, and drier, Little Ice Age (1320-1760 AD).

The Tswaing Crater (or Pretoria Saltpan) 90 m sedimentological record (site 67), outside the Kalahari margins is considered in detail elsewhere in this special issue as a site with valuable multiple proxies for climate within the same record (e.g. Knight and Fitchett, 2021 this issue) (Figure 5(l)). As the longest, and most continuous sedimentary record in southern Africa, it has great potential to offer a chronostratigraphy, noting the poor chronological control from 20 to 70 m, where there are no radiometric dates, and instead proxies have been orbitally-tuned. This includes: tuning of sediment texture data to July insolation at 30°S (Partridge et al., 1997) and tuning of total inorganic carbon to January insolation at 30°S (Kristen et al., 2007). This approach to chronological control is remains a severe limitation of this record, leading to circular arguments about the relationship between proxies and climatic forcing. The rainfall proxy indicates > 700 mm/y from 190-180 ka (peak ~790 mm/y ~185 ka), 167-152 ka (peak ~880 mm/y ~161 ka), 145-131 (peak ~760 mm/y ~136 ka), 120-107, 100-86 (peak ~850 mm/y ~114 ka) and 50 ka (Partridge et al., 1997) (Figure 5(l)), with an antiphase relationship to total inorganic carbon % (low % when rainfall is high) (Figure 5(l)) (Kristen et al., 2007). From 200 to 60 ka the timing of these shifts follows precessional-timed pacing, rather than aligning with marine oxygen isotope stages, with higher rainfall seen during peaks in austral summer insolation at 30°S (Partridge et al., 1998; Kristen et al., 2007) (Figure 5(l)). There is a decrease in the amplitude of rainfall change after 80 ka, and the in-phase relationship with precession breaks down from 60 ka onwards. There is a minor peak ~50 cal kyr B.P., to ~ 700 mm/y, during which time there is speleothem growth at Wolkberg Cave (Holzkämper et al., 2009) (Figure 5(h)), and a longer period of growth at Lobatse Cave that starts around this time (~51 ka) (Figure 5(h)) but continues when the Tswaing rainfall proxy decreases. Rainfall is below 650 mm/y from ~67 to 54 cal kyr B.P. and ~48 cal kyr B.P. onwards, and drops below ~590 mm/y around 26 cal kyr B.P through to 4 cal kyr B.P, before rising only slightly (to ~650 mm/y) at the top of the record.

At Florisbad archaeological site (site 69) there is a fossil spring with an associated dune and nearby pan-fringing lunettes. There is a geological and structural control on the topography and accumulation space of this basin, leading to endorheic drainage and controlling the groundwater emergence point (Rabumulu and Holmes, 2012). Above the former spring there are sands (has been named the Florisbad dune) that are likely to represent a migrating dune that has buried the spring (Douglas, 2006), and also contains peat-like deposits (Grün et al., 1996; Rabumulu and Holmes, 2012). An OSL age for sand above a peat deposit within the
stratigraphy is 133 ± 31 ka, with a paired ESR date of 121 ± 31 ka (Grün et al., 1996), which would place the sand incursion toward the end of the increase in rainfall 145-131 ka indicated by the Tswaing Crater (site 67) rainfall proxy. The brown sand, higher up the section fell outside the limits of 14C dating, whilst the uppermost red sand has a calibrated (no detail within Butzer, 1978 as to the calibration) 14C age of 11.7 kyr B.P., which gives an indication of ongoing sand accumulation until that time, although there are not enough dated horizons to get a sense of accumulation rates. At a second Florisian fossil site of Erfkoon (site 70) (Churchill et al., 2000), there is an overbank alluvial-palaeosol sequence associated with the Modder River that contains a three phase record of shifting environmental conditions. The OSL chronology reveals a relatively steady rate of accumulation of ~0.15 mm/y from 44 ka to ~0.83 ka, and fluctuations in mineral magnetics and diffuse reflectance spectroscopy spectra in the sediments reflect variations in precipitation input. During phase 1 (~46-28 ka) there is a phase of seasonal precipitation, not high enough for the formation of pedogenic ferromagnetic minerals, which shifts to higher levels of these minerals from 41-28 ka, indicating wetter (and possibly warmer) conditions (Lyons et al., 2014). This contrasts strongly with the hiatus in speleothem growth from ~43-27 ka at Lobatse II speleothem (site 61) (Figure 5(h)). During phase 2 at Erfkoon/Modder River (~28 to 15.5 ka), there is a decrease in ferromagnetic minerals, indicating climate got cooler and drier, reaching the lowest 18.5 to 15.5 ka, and at this time the presence of CaCO\textsubscript{3} and gypsum formed, also indicating drying (a negative water balance) (Lyons et al., 2014). The Lobatse II speleothem has ceased growing again ~21 ka (Figure 5(h)), which this time fits with a drier episode at Erfkoon/Modder River, whilst speleothem growth continues at Cold Air Cave (site 63) (24.2-12.7 ka) (Figure 5(j) and growth at Gladysvale (site 68) (16.5 to 14.4 ka) (Figure 5(k)). In the final phase (~15.5 to 0.83 ka) another increase in pedogenic ferromagnetic minerals suggests an increase in rainfall, at least seasonally, with a peak at ~0.83 ka where the sedimentary properties suggest it was wetter than today (Lyons et al., 2014). The Cold Air Cave speleothem (site 63) is also being deposits during this interval, although as discussed above the δ\textsuperscript{18}O signature suggests that rainfall is shifting to shorter-intensity rainfall during a drying Holocene.

The Letlhakeng region is currently semi-arid (~400 mm/y precipitation and an estimated ~98.7% loss to evapotranspiration (de Vries et al., 2000)), and proxies in the Letlhakeng aquifer (site 62), ~140 km northwest of Lobatse II Cave (site 61) indicate some broad changes through time. There was cooler infiltrating water in the earlier part of the record (> 40 cal kyr B.P. to ~28 to 32 cal kyr B.P., calibrated here using the SHCal13, noting the low fidelity dating of groundwater dating), compared to the later part (Kulongoski et al., 2004) (Figure 5(h)). It seems there was enough moisture to recharge the Letlhakeng aquifer even when speleothem growth at Lobatse II Cave ceased, which may relate to hydrogeological setting of these two sites. The increase in both excess air (ΔNe) and δ\textsuperscript{18}O through time indicates a shift from drier and cooler in the first interval to wetter (~34 % increase in ΔNe) and warmer (shift from -5.62 to -4.65 % δ\textsuperscript{18}O) after ~26 to 19 cal kyr B.P. (calibrated here using the SHCal13 and data within Kulongoski et al. (2004) again noting the low fidelity dating
of groundwater dating) (Figure 5(g)). At this eastern site (25.03°E) the Indian Ocean appears to have remained the dominant moisture source (summer rainfall) through time (Kulongoski et al., 2004), in contrast to the Stampriet Aquifer (18.45°E) where δ18O composition is best interpreted as indicating an Atlantic winter rainfall receipt after ~19 cal kyr B.P. (calibrated here using the SHCal13 and data within Stute and Talma (1998) again noting the low fidelity dating of groundwater dating) (Figure 5(b)).

4.2 Chronostratigraphic-based phasing relationship

Table 3 sets out preliminary chronostratigraphic units, which in these dryland settings are dominated by proxy evidence that indicates wetter conditions than present, with tentative drier intervals proposed in-between (most often indicated from absence of wetter intervals but others indicated by isotopic signatures). The focus is on the ~190 ka, as per the remit of this special issue, which also reflects the reducing availability of proxy evidence further back in time. A column for each of the four sub-regions is used, and this demonstrates that only some shifts seem to be common to all southern African dryland (and former dryland) regions. Where there is an absence of correspondence this may in part owe to record incompleteness (preservation and sampling) as well as the limits of chronological control for some records. However, given the large size and modern climatological diversity of these dryland regions (Nicholson, 1986; 2001; Nicholson et al., 2012), it is not surprising that different regions might respond differently, with contrasts between the Namib on the west coast and the Kalahari and northeast to southwest climatic gradient (humidity decreasing), as well as an increase in humidity towards the east side of the Kalahari. The climatological controls over these regional contrasts relate to how far south and west the tropical rainbelt and Congo Air Boundary move, the zonal position of the South Atlantic high pressure cell, the continental high pressure cell and the Southern Indian Ocean high pressure cell, and finally any northerly movements (further than the present position at the southern coast of South Africa) of the westerly wave low. The persistent aridity over the Namib is controlled by the cold Benguela Current, with added mediation from the position of the high pressure cells. Overall, it is seen in this syntheses that the start or end of the wetter intervals do not coincide with MIS boundaries.

[INSERT TABLE 3]

There is evidence for a wetter interval ~140 to 120 ka. This is centred ~131 ka in the northern Kalahari with a mega-lake Makgadikgadi phase (sites 39 to 42) (Burrough et al., 2009). The complicated hydroclimatic controls over this archive mean that this record can either reflect conditions in the northern Kalahari or further north in the catchment of the Okavango-Makgadikgadi system (or a combination) (Burrough et al., 2009; Moore et al., 2012). The records of speleothem growth 133 ± 27 ka at Drotsky’s Cave (site 36) in northwest Botswana (Brook et al., 1998) and the Otavi Mountainland 130 ± 7 ka (west of 18°E, (sites 27 to 31)) (Brook et al., 1998;
suggest this interval did involve increased precipitation over the northern Kalahari (note the Etosha Pan chronology does not extend this far back). Together this suggests the tropical rainbelt and Congo Air boundary influenced the northern Kalahari climate at this time. During this interval there is also increase moisture availability in the Namib Desert within the Tsondab River, broadly between ~132 and 87 ka (site 12) (Stone et al., 2010a) and active slope-processes on sand ramps, fairly poorly time-resolved to >150 ka and >90 ka (Rowell et al., 2017). Increased moisture in the Tsondab catchment and the location of sand ramps close to the Great Escarpment is most likely to have involved and easterly source associated with the summer rainfall zone, rather than a deviation from the west coast aridity in the Namib Desert. This could have involved the tropical rainbelt but also moisture from tropical- temperate trough systems as tropical easterly waves and southern western waves connect. In the southern Kalahari there is evidence for semi-arid conditions, rather than a substantial wet interval, from 156 ± 11 to 121 ± 6 ka at Kathu Pan (site 57) (Lukich et al., 2019; 2020), and at this southerly location this is likely due to climatic driving of increased tropical-temperate trough systems.

There is a pause in wetter conditions in the northern Kalahari, with a second wet interval starting up again ~112 to 90 ka, recorded in the northern Kalahari 112 ± 5 to 108 ± 7 ka in the Otavi Mountainland speleothems (sites 27 to 31) (Brook et al., 1998). There are two phases of mega-lake Makgadikgadi (sites 39 to 42) 105 ± 4 and 92 ± 2 ka (Burrough et al., 2009), the first overlapping with the Otavi Mountainland speleothem growth and the second corresponding with Drotsky’s Cave at 93 ± 6 ka (site 36) (Brook et al., 1998), which suggests this involved increased precipitation across parts of the northern Kalahari, perhaps again associated with the tropical rainbelt and Congo Air Boundary. In the Namib Desert the chronological control cannot/does not resolve any such pause. In the southern Kalahari, abundant resurfing water is indicated by cascade tufa deposits between 111 ± 3 and 102 ± 2 ka at Ga-Mohana Hills (site 56) (Wilkins et al., 2021). At Kathu Pan pedogenic units dated to 96 ± 5 and 84 ± 4 ka are indicative of increase moisture availability. The absence of dune ages in the southern Kalahari during these intervals may also support wetter conditions if the absence of dune ages records an absence of accumulation driven by wetter conditions (rather than owing to sampling bias or preservation).

The third wetter interval ~80 to 70 ka is recorded in the Namibian part of the northern Kalahari (at Etosha Pan 75-70 ka (site 26) (Hipondoka et al., 2014) and the Namib Desert (ongoing fluvial units in the Tsondab River (site 12) (Stone et al., 2010a), but not in the mega-lake Makgadikgadi system. A semi-arid phase in the southern Kalahari at Kathu Pan (site 57) is recorded 74 ± 5 ka (Lukich et al., 2020). This spatial pattern suggests an influence of the tropical rainbelt and Congo Air Boundary in the north of the region.

In contrast, the fourth wet interval (~63 to 43 ka) is widespread across three of the four sub-regions: (i) fluvial sediments (centred on ~61 to 41 ka in the Hoanib (site 4) (Eitel et al., 2006), 44-40 ka in the Hoarusib (site 3)
(Srivastava et al., 2005)) and Orumana stalagmite growth starting ~48 ka (site 1) (Railsback et al., 2016) in the Namib Desert, along with a shift to more positive δ¹⁵N in the composite midden record (sites 6, 14 and 25) 46-44 cal kyr B.P. (Chase et al., 2019), (ii) a mega-lake Makgadikgadi (sites 39 to 42) phase 64 ± 2 ka (Burrough et al., 2009), corresponding with Otavi cave speleothem growth ~61 ka (sites 27 to 31) (Brook et al., 1998) in the northern Kalahari, (iii) and speleothem growth ~51-43 ka in Lobatse II (site 61) (Holmgren et al., 1995) in the eastern fringes of the southern Kalahari, as well as at Wolkberg (site 65) and Gladysvale (site 68) caves further east, 58 to 46 ka and 56.8 ± 0.4 to 42.8 ± 6.8 ka respectively (Holzkämper et al., 2009; Pickering et al., 2007).

The north of the Kalahari and the north of the Namib Sand Sea may have been influenced by the tropical rainbelt and Congo Air Boundary. The hydroclimatic shifts recorded in the composite midden (sites 6, 14 and 35) representing a 900 km long north-south transect in the Namib are conceptualised and modelled by Chase et al. (2019), with an explanation for what drives increased aridity, but the corollary mechanism for wetter intervals is perhaps less clear. Chase et al. (2009) suggest increased aridity occurs due to an intensified South Atlantic Anticyclone and related stronger upwelling in the Benguela cold current, and that this is driven by an increased intra-hemispheric temperature gradient with precession pacing which sees austral summer insolation maxima occur at the same time as boreal summer insolation maxima. The 46-44 cal kyr B.P wetter phase is at the opposite insolation forcing, with austral summer insolation minima (boreal summer insolation maxima), which would reduce the temperature gradient and lessen the dominance of the South Atlantic Anticyclone, perhaps allowing easterly rainfall sources to reach the eastern edge of the Namib. However, the lack of a coeval pattern of increased moisture in the southern Kalahari is more challenging to understand. It is not clear why increased rainfall is seen in the eastern fringes of the southern Kalahari and in the eastern Namib, and not in the middle, when the mostly likely source of moisture is from the east, and associated with the tropical temperature trough.

A drier interval ~43 to 27 ka is recorded in the eastern fringes of the southern Kalahari (hiatuses in speleothem growth in Lobatse II (site 61) (Holmgren et al., 1995), whilst sand sheet and sand dune sediments are accumulating in the southern Kalahari (Bateman et al., 2003; Stone and Thomas, 2008; Thomas and Burrough, 2016). The increase in CaCO₃ from 32 ± 2 ka within Kathu Pan sediments (site 57) is also indicative of drying towards a highly negative moisture balance in the southern Kalahari (Lukich et al., 2020). However, the Omongwa Pan (site 46) record suggests wetter conditions ~43 to 30 cal kyr B.P. (Schüller et al., 2018) and recharge to the Stampriet Aquifer (site 47) is occurring (Stute and Talma, 1998). A drier phase is not recorded in the Namib Desert, or the northern Kalahari. Towards the end of this eastern dry phase a fifth wetter interval is starting in the northern Kalahari, and is later recorded in all four sub-regions. In the northern Kalahari, mega-lake Makgadikgadi (sites 39 to 42) 39 ± 2 ka (Burrough et al., 2009), overlaps with the 43.9 ± 7.3 ka growth at Drotsky’s Cave speleothem (site 36) (Brook et al., 1998) and then at Tsodilo Hills from 36-32 ka (site 33) (Thomas et al., 2003), at Etosha Pan (site 26) from 34 to 27 ka (Hipondoka et al., 2014), and speleothem growth
in the Otavi Mountainland 31 ± 2 and from 29 ± 1 to 28 ± 1 ka (sites 27 to 31) (Brook et al., 1998). This is also seen in the Namib Desert fluvial record 34-20 ka (in the Hoanib (site 4) Eitel et al. (2006) and Hoarusib (site 3) Srivastava et al. (2005)), major slope processes on sand ramps ~30 to 17 ka (Rowell et al., 2017), and supported by a negative shift δ15N composition of the composite midden record from ~32 to 29 kyr B.P. (Chase et al., 2019). In the southern Kalahari there is speleothem growth at Wonderwerk Cave (site 58) ~33 ka (Brook et al., 2010), flooding of Witpan surface (site 51) ~32 ka (Telfer et al, 2009). In the eastern fringes of the southern Kalahari Lobatse II speleothem (site 61) growth starts up again ~27 ka (Holmgren et al., 1995) and ~24 ka at Cold Air Cave (site 63) further east (Holmgren et al., 2003). A widespread wet interval would suggest the influence of the tropical rainbelt and Congo Air Boundary in the north and rainfall within the temperature tropical trough influencing the south and reaching the eastern fringes of the Namib.

There is a pause in evidence of wetter conditions, everywhere apart from the eastern fringes of the southern Kalahari, before a widespread sixth wetter interval (~23 to 16 ka), again likely involving the tropical rainbelt in the north and easterly-derived moisture in the tropical temperate trough through the south and into eastern parts of the Namib. This wetter interval is recorded in the Namib Desert within fluvial records ~21 to 16 ka in the Khumib (site 2) (Srivastava et al., 2004) and negative excursions in δ15N composite midden record (sites 6, 14 and 35) ~24.5 and ~22 kyr B.P. (Chase et al., 2019). In the northern Kalahari there are lake phases at both Etosha Pan (site 26) 23-21 and 18-16 ka (Hipondoka et al., 2014), and in mega-lake Makgadikdagi (sites 39 to 42) 17 ± ka (Burrough et al., 2009) In the southern Kalahari there is flooding in the Lower Molopo (site 53) ~23 and ~18 ka (Hürkamp et al., 2011) flooding at Witpan (site 51) ~20 ka (Telfer et al., 2009), and speleothem growth at Wonderwerk Cave 23-17 ka (Brook et al., 2010). In contrast, the 22.2 ± 1.2 ka hardpan unit at Kathu Pan indicates a strong negative moisture balance, so it could be that increased rainfall was highly seasonal. In the the eastern fringes of the southern Kalahari speleothems are still being precipitated at Lobatse II (site 61) through to 21 ka (Holmgren et al., 1995), and further east speleothem growth starts again ~24 ka at Cold Air Cave (site 63) (Holmgren et al., 2003), whilst at Gladysvale (site 68) speleothem growth does not recommence until 16.5 ± 0.2 ka (Pickering et al., 2007). Following this, there is evidence for slightly later wetter conditions in the composite midden (sites 6, 14 and 25) δ15N ~14 and ~14.3 kyr B.P. (Chase et al., 2019), at the same time as wetter conditions in Tsodilo Hills (site 33) (Thomas et al., 2003) in the northern Kalahari, with no corresponding evidence in northern Namibia or the southern Kalahari.

An increase in aridity is seen ~10 ka in the Namib Desert with a start of increasing δ15N in the composite midden record (sites 6, 14 and 25) (Chase et al., 2019), supported by a reduction in fluvial sedimentation in the Khumib (site 2) (Srivastava et al., 2006), Tsondab (site 12) (Stone et al., 2010a) and Tsauchab (site 16) (Brook et al., 2006) in northern Namibia. In contrast, the northern Kalahari has a mega-lake (sites 39 to 42) phase 8 ± 5 ka (Burrough et al., 2009), a lake at ~10 ka at Etosha Pan (site 26) (Hipondoka et al., 2014; De Cort et al. in press)
as well as speleothem growth at Drotsky’s Cave (site 36) and Bone Cave (site 37) (Brook et al., 1998), and in the Otavi Mountainlands (sites 27 to 31) (Brook et al., 1998) and an absence of dune accumulation in the northwest, northeastern and eastern dunefields. In the eastern fringes of the southern Kalahari, speleothem growth has already ceased at Lobatse II (site 61) (Holmgren et al., 1995), although further east there is still deposition at Cold Air Cave (site 63) (Holmgren et al., 2003). There is, however, sufficient precipitation for both Stampriet (site 47) (Stute and Talma, 1998) and the Letlhakeng (site 62) aquifers (Kulongski et al., 2014) to receive some recharge during this time. The aridity intensifies from ~5 ka in the Namib Desert (composite midden record (sites 6, 14 and 25) (Chase et al., 2019)), and the northern Kalahari (e.g. Etosha Pan (Hipondoka et al., 2014), Dante Cave (Sletton et al., 2013), no mega-lake Makgadikgadi (Burrough et al., 2009)), and the isotopic record in Dante Cave (site 30) speleothem indicates a transition to drier conditions after 4.6 ka (Sletton et al., 2013). In the southern Kalahari, Omongwa Pan sediments (site 46) indicate a wetter phase ~5.5 cal kyr B.P. (Schüller et al., 2018), and some fluvial deposits in the Kuruman River (site 55) continue to be deposited during this interval up to ~ 0.5 cal kyr B.P. (Shaw et al., 1992).

The onset and cessation of the wetter phases does not coincide with boundaries in the MIS, nor does the drier interval ~43-40 ka observed in the eastern fringes of the southern Kalahari. The spacing of wetter intervals across the four sub-regions does not demonstrate any regular periodicity. This overview from combined records (pseudo-parasequences) is therefore in contrast with the two key southern African terrestrial records - the Pretoria Salt Pan in the eastern fringes of the southern Kalahari (Partridge et al., 1997; Kristen et al., 2007) and the combined Namib Desert hyrax record in the west (Chase et al., 2019), for which the influence of precessional insolation forcing has been invoked. For the former, noting the circular argument from orbital-tuning of the record, insolation maxima at 30°S are in phase with higher rainfall (as recorded in the sediment-grain size derived rainfall proxy (Partridge et al., 1997) and TIC, TOC and TN (Kristen et al., 2007)) for 200 to ~60 ka, with the climatic mechanism relating to land-sea temperature and the position of the tropical African rainbelt (Tyson, 1999; Schefuß et al., 2011). In contrast, for the latter, the increases in moisture availability over the past ~50 ka are generally negatively correlated to southern hemisphere summer insolation (Chase et al., 2019). Chase et al. (2019) identify how this is contrasts with other archives in the west from marine sediments that do show a positive relationship (e.g. Daniau et al., 2013; Collins et al., 2014). The climatic mechanism for the Namib Desert explored by Chase et al. (2019) and supported by interpretations of climate model simulations (He et al., 2013) is that it is actually the contrast between high-northern-latitude (65°N) summer insolation and low-latitude southern hemisphere (30°S) variations that may be driving the response in the Namib Desert. When NH summer insolation is low, SH summer insolation is high (the ‘out of phase’ precession forcing between hemispheres) and the increase in latitudinal intra-hemispheric temperature gradients drives an intensification in the South Atlantic Anticyclone and increases the strength of the upwelling in the cold Benguela current. This leads to drier air, which advects eastward and the South Atlantic Anticyclone
blocks the incursion of other moisture-bearing atmospheric systems. The land-sea temperature gradient during higher SH summer insolation receipt drives higher summer rainfalls in the east of the subcontinent. Therefore, the west and east are out of phase, with the west in phase with the northern hemisphere peaks in summer insolation (Chase et al., 2012; 2019). Table 3 shows support for this west-east contrast during the drying in the west during the past ~10 ka. However, over the past ~190 ka there is neither consistent 23 ka pacing or a consistent west-east contrast in moisture availability.

4.3 Overview of the significance and limitations of arid archives and proxies

The utility of proxies within dryland regions is influenced by the spatial distribution and the temporal length of the record (and the datable portion of it, in terms of dating method upper age-limit), as well as the achievable temporal resolution, and our ability to understand, and isolate, the connection between climatic forcing and proxy response. Temporal resolution is controlled by the accumulation rate within an archive and any mixing of the signals through time, along with the sampling resolution employed within studies and the precision available with each of the applied radiometric dating techniques. The five key considerations introduced for dunes are also widely applicable to most dryland archives and proxies in Table 1. In summary, these are: (i) being clear what condition/comination of conditions facilitate the deposition and preservation of a sedimentary unit or a chemical precipitate (and its composition); (ii) what an age for a sample means, in terms of event, process or response to climate, and with geoproxies the form of the feature too; (iii) record completeness (considering the potential for reworking and erosion of archives); (iv) the importance of combining chronological information with stratigraphic and sedimentological information (in the case of sedimentary-based archives) and (v) the potential mismatch between the rate of a depositional events and the available temporal resolution of the proxy.

Geoproxies

The most spatially extensive evidence in drylands remains geoproxies, particularly sand dunes, although their palaeoclimatic inference may not to be straightforward, as introduced with the five key considerations. It has been recognised for decades that dune accumulation depends as much on windiness, sediment supply and/or availability, as it does on it being ‘dry enough/not too wet’ as to be impeded by vegetation coverage (e.g. Kocurek, 1998; 1999; Singhvi and Porat, 2008). Three decades of sand dune research in the Kalahari has revealed that some dune accumulation is recorded in most areas for much of the past 120 ka (data from individual studies is summarised in Thomas and Burrough, 2016), whilst the dune accumulation record of the Namib Sand Sea remains to be investigated in detail (Stone, 2013). In this study dune accumulation has been presented simply, in the form of ages with errors (Figures 4(e)(h), 5(c)), and these can be stacked to get an
impression of geomorphic activity across space (cf. Stone, 2009; Thomas and Burrough, 2012). However, binwidths exceeding 1 ka are needed further back in time to reflect the available precision (Figure 6). Considering the intensity of geomorphic activity across space within these sub-regions may prove fruitful in the future, if a large enough dataset can be produced and if the bias introduced by rarely sampling to the base of dunes can be addressed. The use of field portable OSL readers (port-OSL) (Sanderson and Murphy, 2010) may prove fruitful in facilitating this spatially extensive analysis of sample relative age. Furthermore, calibrations of port-OSL signals (e.g. Stone et al., 2015; 2019) can facilitate age estimations, albeit at lower precision that full laboratory dating.

{INSERT FIGURE 6}

Empirical findings from field sampling and dating of dunes are complemented by conceptual models (e.g. the sediment state framework of Kocurek and Havholm, 1993; Kocurek, 1998; 1999) and numerical models (e.g. Telfer et al., 2010; Bailey and Thomas, 2014; Mayaud et al., 2017; Buckland et al., 2019). These models help to capture the “chain of necessary conditions” required for sediment production, deposition and preservation (Singhvi and Porat, 2008, p 540), to which we can add the additional filter of sampling strategy (see Figure 12a within Stone and Fenn, 2020). The 1D numerical models of Telfer et al. (2010) and Bailey and Thomas (2014) both suggest the transition from aeolian activity to inactivity (or the end of a windy and drier phase) is more likely to be preserved than other parts of the window of activity. This contrasts with the conceptual sediment state framework, which suggest the end of a more arid period is be erosive, with sediment preservation beginning after a more humid period has started (e.g. Kocurek and Havholm, 1993), which would mean the start of that transition would be truncated in the record. Numerical models that can capture 3D feedbacks between wind-flow dynamics, sediment flux and vegetation cover also have great potential for more nuanced understandings of dunefield dynamics (e.g. the ViSTA (Vegetation and Sediment TrAnsport) model (Mayaud et al., 2017)). Where dunes and fluvial and/or lacustrine records interdigitate, it is more straightforward to unpick hydroclimatic variations than dune accumulation in isolation, because the two sediment types are driven by opposite ends of the moisture availability spectrum. The mismatch between the rate of sedimentary accumulation events and the achievable resolution of sedimentary records is less tractable; for example, events of decadal, centennial, or even millennial-length cannot be resolved 100 ka years ago, because chronological precision here is ± 10 ka.

There is an additional, fundamental challenge regarding form-process linkages for both dunes and sand ramps. In the case of dunes in the Kalahari there is a clear morphological difference between the southern and northern Kalahari: the former are stabilised and covered in vegetation, with some mobility at the crests (e.g. Wiggs et al., 1995), whilst the latter are heavily vegetated, degraded by surface processes and also tectonic
modification (e.g. McFarlane et al., 2005). This means that the sedimentary record within northern dunes may relate to a different set of geomorphic processes (processes additional to aeolian sediment accumulation) that can expose sediment to light (influencing luminescence chronologies). In addition, the erosion and crest lowering means they their stratigraphic record is likely to be less completely preserved than for southern dunes. For sand ramps, any tectonic, or neotectonic, processes could influence the operation of slope processes, as can the crossing of internal geomorphic thresholds relating to slope angles and stability (e.g. Brunsden and Thornes, 1979), which means that the response is not purely climate-driven. The sand ramps fringing the Namib do not contain a consistent record of slope-process across space (Rowell et al., 2017).

In addition, dunes, lake shoreline sediments and sand-rich units in sand ramps and any other sedimentary units are also subject to, post-depositional modification, including bioturbation, which mixes grains with different OSL burial ages. This requires to think carefully about what process is being dated - the original depositional event in the landform, or a post-depositional process. Soil sampling studies for indicator minerals in mining for diamonds in the Kalahari indicate that termites are capable of moving grains up and down through 30 to 40 m depth of the Kalahari formation sediments, with several tonnes of sediment moved during the construction of individual nests (Dira and Daniels, 2018). Single-grain scale OSL analysis can be helpful to reveal the presence of bioturbation (e.g. Bateman et al., 2007). A small number of dunes in the Namibian portion of the southern Kalahari have been OSL dated at single grain, as well as single aliquot (hundreds of grains together) (Stone and Thomas, 2008). Whilst the overdispersion (scatter) in the data of 21 to 63% does indicate presence of heterogeneous grain distributions, these values are no higher than those found in Australian dune settings with very well preserved (intact/ not bioturbated) stratigraphy (e.g. Lox et al., 2007). In addition, overdispersion can also be driven by heterogeneity in the sedimentary dose-rate, rather than post-depositional mixing (see Cunningham et al. (2012) for technical details). The similarity of age estimates derived at single grain and single aliquot scale gives some confidence that post-depositional mixing in that region is not severe at those Namibian sites. Potential post-depositional mixing should however, be studied in more detail in southern Africa. Ground penetrating radar is extremely valuable for revealing sub-surface stratigraphy in regions where creating large exposures (such as in mines) are not practicable (e.g. Bristow et al. (2005) in the Namib Sand Sea). One attempt in the southern Kalahari did not reveal interpretable imagery, perhaps because of the homogeneity of the sand sediment, or water content, through the profile (Matt Telfer, pers. comm).

The example of lake bed ‘barchans’ versus ‘spring mounds’ on the surface of Ntwetwe Pan within the Makgadikgadi system is a compelling example of the challenge of form-process attribution for some geomorphological proxies. These crescentic-shaped sedimentary accumulations have been interpreted as aeolian barchan dunes (e.g. Grove, 1969), or as subaqueous features (Cook, 1980). These contrasting process attributions have strikingly different environmental and climatic implications - the former requires a dry
lakebed for their accumulation and the latter indicates the presence of a water body, and likely groundwater discharge. Burrough et al. (2012) use a combination of morphometric analysis and OSL signal analyses to suggest the features fall within the range of barchans and that the sediment within the features were well exposed to light (bleached) prior to deposition, as would be expected to occur with aeolian transport. However, they conclude that there is insufficient evidence to reject either of process-form hypothesis. In contrast, McFarlane et al.'s (2015) morphological analysis suggests the forms are inconsistent with barchans, particularly the elevated areas (“eyes”) within the crescentic mounds. In addition, the flat-tops of the mounds themselves, the abrupt shifts in tone banding (discrete colour shifts in remotely sensed images) with undisturbed curvature of the bands suggests these features are relatively undisturbed (so would not fit with an origin as heavily-degraded barchans) (McFarlane et al., 2015). Furthermore, McFarlane et al. (2015) observe groundwater discharge from the flanks of the features up to 7 m above the current dry pan floor, suggesting there is upward movement of groundwater associated with these features. The presence of other small, circular mounds on the pan floor, exhibiting the same tone-banding pattern, the spatial distribution of the round and crescentic mounds within the pan, and the presence of evaporites in the sediment all point towards these being subaqueous mound-springs in a discharge-zone of a hydrogeological system with more vigorous groundwater flow in the past (McFarlane et al., 2015). This interpretation is supported by Franchi et al. (2020), who investigate the internal architecture of the mounds and find they are planar-bedded (which in turn has led to the tone-banding in birds-eye view), from either flowing, or stationary water, and that there is no aeolian depositional architecture, such as foresets.

Hydrological proxies

Lake shorelines have the potential to be a less ambiguous proxy, representing lake-full conditions, although the critiques relating to post-depositional mixing, and challenges about (iii) record completeness, (iv) the importance of combining chronological information with stratigraphic and sedimentological information and (v) the potential mismatch between rate of event and available temporal resolution within the shoreline archive all still stand. In addition, as discussed, the large catchment and complex controls over the Okavango-Makgadikgadi system require careful consideration and comparison against other archives to help confirm the increase in precipitation is local to the northern Kalahari. Moore et al. (2012) go further to suggest that the control over Makgadikgadi lake levels is the inherited landscape, involving tectonic drivers of drainage evolution, rather than being driven by more recent Quaternary climatic events.

There is great potential in lake floor sediment analysis with an ongoing analysis of sediment cores drilled from the Makgadikgadi lakebed (Burrough, pers. comm.). The geochemical analysis and micromorphological analysis that has been undertaken for pan sediments (e.g. Schüller et al., 2018; Vainer and Matmon, 2018;
Lukich et al., 2020) is an extremely valuable approach, situating stratigraphies and chronologies together and providing multiple proxies with which to reconstruct environmental and climatic conditions. The ages provided by OSL for pedogenic units within sand in lacustrine or pan sediments require some scrutiny. OSL dates the last exposure of sand to light, and hence OSL is dating the deposition of the sand unit, which is the host sediment into which the soil developed. This means the OSL age predates the formation of the palaeosol.

The analysis of facies typologies and architecture within the fluvial sediment archive is a powerful approach to nuanced palaeoenvironmental reconstructions (e.g. Srivastava et al., 2004; 2005; 2006). This directly addressed challenge (iv), which is the need to combine stratigraphic and sedimentological information when producing a chronology for deposition. However, fluvial archives are subject to issues of resolvable temporal resolution of processes (challenge (v)), preservation (challenge (i)) and record completeness (challenge (iii)), with Jerolmack and Paola (2010) arguing that external signals, such as climate forcing, can be ‘shredded’ by morphological turbulence (geomorphic thresholds) in the fluvial system.

Hydrogeological proxies

The biggest challenges for pedogenic and groundwater calcrites are firstly that they may form during multiple environmental cycles (which results in a mismatch in, or inability to, resolves the environmental process of interest (challenge (v))). Secondly, 14C and U-series dating may be affected by open-systems behaviour, with carbon/uranium leached and added during multiple depositional cycles or post-depositional alteration, and for 14C dating, that the incorporation of old carbon can lead to age overestimations (Geyh and Eitel, 1998). For U-series dating of calcrites, the incorporation of detrital sources of thorium, common in the terrestrial environment, also presents a challenge for dating, as is also the case for tufa deposits, and to a lesser extent speleothems (e.g. see Schwarcz and Latham, 1989). Approaches to correct for detrital contamination do make U-series dating extremely important in providing the most reliable chronologies. Speleothems have been extremely important archives of palaeoclimate in southern Africa (and Africa more widely) (Braun et al., 2019), even if there are frequent hiatuses to growth in the more arid regions, limiting the potential for long continuous geochemical records. Speleothem records are restricted by lithology, requiring a carbonate-based bedrock composition. Given the tufa deposits within the Naukluft Mountains, a systematic exploration of cave speleothem records in these limestone mountains could be extremely fruitful. Irish et al. (2000) summarise explorations to three sites (only one with coordinates), whilst Lichtenecker (1966) describes the 1931 exploration of a cave (Merkerhoele) near Blasskranz.

The utilisation of groundwater requires a great deal of physical infrastructure, which will limit this to aquifers with extraction boreholes, and there are recent critiques of whether meaningful residence times for
groundwater in arid and semi-arid regions can be determined (Cartwright et al., 2020). The uncertainties in determining groundwater residence times hamper the utility of the geochemical and isotopic insights that groundwater contains beyond broad-brush observations at low temporal resolution, such as changes on glacial-interglacial timescales.

**Hyrax middens**

Accumulation rates for hyrax middens vary from ~20 to 2000 years per cm (Chase et al. 2012), which even at the lower end makes them a high-resolution dryland proxy, aided by the precision of 14C dating. The controls over the geochemical signatures have been well established via empirical studies and correlations with soil and vegetation geochemical signatures. Their limitation lies in their spatial distribution (requiring rock shelters as sites of accumulation and preservation) and their upper time-range, which is ~50 cal kyr B.P. (and the majority < 20 cal kyr B.P.), which may be a limit set by maximum midden sizes, and/or preservation. In addition, there is a humidity/moisture threshold for midden preservation of <480 mm/y mean annual precipitation and the wettest month < 90 mm (Chase et al., 2012).

5. Conclusions and future research directions

Attempts to constrain terrestrial chronostratigraphies are valuable, because they focus on the changes in the terrestrial environment as the starting point to link to environmental response to climatic forcing, rather than starting with the global master climatostratigraphy of the MIS. A terrestrial chronostratigraphy has not been formalised for southern Africa, or its sub-regions, with the exception of some named stages for the Eastern Cape by Lewis (2011). Therefore, syntheses that work towards identifying key terrestrial stratigraphies and parasequences combined from temporally overlapping records are valuable. The dryland regions of southern Africa are widespread and climatologically diverse, and for this reason it is sensible to consider sub-regions whilst exploring parasequences from the available overlapping terrestrial records. This study uses five sub-regions (the Namib Desert, the northern Kalahari, the southern Kalahari and the eastern fringes of the southern Kalahari) as a pragmatic balance between capturing heterogeneity across space whilst also ensuring there are data within each sub-regions.

Whilst dunes are an archetypal dryland-zone archive/proxy, the archives and proxies in drylands are widespread and diverse. They have been classified here into surface, and subsurface, and within this, there are aeolian-, hydrological- and hydrogeological-process driven archives. The archives include geoproxies (e.g. dunes, sand ramps), water-lain deposits in lakes and pans, including calcretes, and fluvial deposits (both sediments and tufa carbonate), speleothems, groundwater and hyrax middens (Table 1). The richest record is
revealed when a range of proxies is considered together. The fragmented nature of preserved evidence means that we are still some way from producing unambiguous parasequences, and at this stage there is rich record to consider and compare (as summarised in Table 3). There are seven wetter intervals identified in two or more of the five subregions (140-120, 112-90, 80-70, 64-43, 43-40 to 34 ka, 34-24 ka, 23-16 and ~15 ka, and a drier phase ~43-40 only observed in the east. There is drying in the west over the past ~10 ka, during which time there are still indicators of wetter conditions in the northern Kalahari, and beyond the eastern fringes of the southern Kalahari. Aridity intensifies during the past ~5 ka, and this is more widespread across all four sub-regions.

What is clear is that the wetter and drier intervals recorded from the rich dryland zone record do not follow global MIS glacial and interglacial stages. In addition, there is no consistent link with precession-paced insolation forcing across the five sub-regions of the southern African drylands. This is despite strong evidence for precession-paced climatic forcing over the Namib Desert over the past ~50 ka within hyrax middens (Chase et al., 2019) and the inferred in-phase southern hemisphere summer insolation forcing (30°S) of wetter conditions in the east of the subcontinent from ~200 to ~60 ka (e.g. Partridge et al., 1997; Kristen et al., 2007), noting the circularity of argument for the latter from the orbitally-tuned chronology. In the former, this is thought to involve intensification and weakening in the intrahemispheric temperature gradient relating to out-of-phase precession-forced insolation between the low-latitudes of the southern hemisphere and mid-to-high latitudes of the northern hemisphere (Chase et al. 2019).

It is hoped that this overview of the currently available evidence will serve the basis for future discussions and explorations of the rich palaeoenvironmental and palaeoclimatic archives and proxies within southern African dryland regions. There is a need to search for more long and near-continuous records, to continue to seek records with which to produce parasequences and to improve and refine chronological control wherever possible. In doing so, it is vital to continue to scrutinise the connection between proxy response and climatic forcing, acknowledging any other forcing (tectonic, intrinsic geomorphic thresholds within geomorphological proxies and post-depositional alteration), in order to accurately reconstruct environmental and climatic conditions from the deposition and preservation of the proxy. Relating to this, it is important to consider what an age for a sample means in the context of process or event that has been recorded (including a potential for post-depositional reworking and alteration) and available resolution of a dated record. The latter is a function of precision of the dating method as well as the combination of sample size and the rate of sample deposition (i.e. what a 1 mm or 1 cm of sample relates to in terms of years of accumulation). We must be aware that there can be a mismatch between the rate of a preserved event and the available temporal resolution for that proxy, so that events of seasonal, yearly, decadal-scale, or even greater time-intervals, remain unresolved. Furthermore, hiatuses within proxies may be the product of erosion, rather than indicating a lack of deposition
of that proxy. For sedimentological proxies, it is important to combine stratigraphic and sedimentological information with chronological control for a nuanced understanding of the chronostratigraphies we are producing. There continue to be breakthroughs in the measurement and analysis of proxy signals, with many notable examples in the geochemical signatures within hydrax midden hyraceum (e.g. Chase et al., 2012 and references therein). In addition, for proxies such as sand dunes, there are advancements in the interpretation of the archive by unpicking changes across different parts of the landscape and using conceptual and numerical modelling to refine our understanding and interpretation of the landscape-response to climatic forcing.

Overall, exploring similarities and differences in dryland environmental responses across space, is a fertile ground to explore, and test, a range of hypotheses about the nature of climatic forcing over southern Africa. The patterns identified here highlight the complexities across space, they provide a framework to be scrutinised, and a basis from which to refine a proposed terrestrial chronostratigraphy for southern Africa, which has hitherto been absent, and is important for this region, particularly given the insights into hominin evolution provided here.

Acknowledgements

I am grateful to Jasper Knight and Jennifer Fitchett for the invitation to write this overview, and to Frank Eckardt and Dave Nash for insightful reviews, which have improved this paper I would also like to thank the authors of the papers and datasets considered here, as well for fascinating and insightful conversations with many of those authors over the years.

References


Bubenzer, O., Bödeker, O. and Besler, H., 2007. A transcontinental comparison between the southern Namib Erg (Namibia) and the southern Great Sand Sea (Egypt). Zentralblatt für Geologie und Paläontologie Teil I, Heft, 1/4, 7-23.


Table and Figure Captions

Table 1. A classification of archives and proxies found in dryland regions.

Table 2. Location of proxy sites referred to in this paper.

Table 3. Preliminary chronostratigraphic time periods for the five sub-regions within southern African drylands for the past ~190 ka, as identified in this study. The chronostratigraphy proposed for southern Africa by Knight and Fitchett (2021), which provides local-names for the Marine Oxygen Isotope Stages (MIS), which are as defined by Lisicko and Raymo (2005) is given in the final column for reference.

Figure 1. Southern Africa, showing (a) locations of the mega-Kalahari and its dunefields (NWK is northwest Kalahari, NEK is northeast Kalahari, EK is eastern Kalahari, WK is western Kalahari and SnK is southern Kalahari), and the Namib Desert and its sand sea and dunefield regions, also showing the four sub-regions considered in this paper (Namib Desert, northern Kalahari, southern Kalahari and eastern fringes of the southern Kalahari) (b) mean annual precipitation and extent of the winter rainfall zone (WRF) and summer rainfall zone (SRZ), with the year round rainfall zone in-between (modified from Chase and Meadows, 2007), (c) aridity index, where <0.5 is boundary for semi-arid (modified from Trabucco and Zomer, 2009) (d) biome distributions (modified from Truc et al., 2013), (e) the rainfall regions and numbering system of Nicholson (1986; 2001; 2012) (redrawn from Nicholson, 2001).

Figure 2. Location of sites referred to in the text, site names and references given in Table 2. The open circles represent sites within the sub-regions of interest, whilst the solid squares are from sites beyond the margin regions that offer useful proxies for comparison. Where there are lines grouping multiples sites to one number these represent individual sites dated within dunefields (32, 38, 43, 48 and 50) or individual sites within the lower Molopo catchment (site 53). The ephemeral rivers in the Namib, from north to south are Khumib, Hoarusib, Hoanib, Uniab, Koigab, Huab, Omaruru, Swakop, Kuiseb, Tsondab, Tsauchab.

Figure 3. Proxy records for the Namib Desert for the past 60 ka, arranged broadly N to S for Namib and then the northern Namibian record. (a) the Namib composition midden δ¹⁵N record of Chase et al. (2019), for which positive values indicate drier, and negative values indicate wetter, conditions, (b) the Khumib (site 2), Hoarusib (site 3) and Hoanib (site 4) ephemeral river records, as summarised within Stone and Thomas (2013) but compiled from Srivastava et al. (2004) (Khumib), Srivastava et al., (2005) (Hoarusib) and Eitel et al., 2006 (Hoanib), where solid fill is wetter and diagonal indicates only weak summer rains, (c) Orumana Cave (site 1) speleothem δ¹⁸O, located in upper catchment of the Hoarusib River (Railsback et al., 2016), (d) δ¹⁵N and δD n-c alkene record from the Spitzkoppe midden (site 6) (Chase et al., 2010), (e) the Kuiseb (site 9), Tsondab (site 12) and Tsauchab (site 16) ephemeral river records, as summarised within Stone and Thomas (2013) but compiled from Srivastava et al. (2006) left hand and Bourke et al. (2002) right hand (Kuiseb), Stone et al. (2010a) (Tsandb) and Brook et al. (2006) (Tsauchab), (f) dated sand ramp units (sites 11, 15, 17, 19, 20 and 21) where orange solid are pure aeolian sand, stripped are minor slope processes and solid blue rectangle is major slope deposit (Rowell et al., 2017).

Figure 4. Proxy records for the northern Kalahari for the past 190 ka, arranged broadly west to east, and the central Kenyan Rift Lake level records for comparison, where: (a) Etosha lake (site 26) levels, solid is lake 21 to 24 m deep and striped is low (a few meters) (Hipondoka et al. 2014), and (b) Etosha lake (site 26) classes, where left to right is 6 (highest) to 1 (lowest) (De Cort et al., 2021), (c) speleothem growth phases combined from Aikab and Aigama (site 28) and Guinas Meer (site 29) cenotes (Brook et al., 1999), (d) Dante Cave (site 30) speleothem δ¹⁸O (Sletten et al., 2013), (e) dune accumulation ages for the northwestern (NWK) dunefield Kalahari (site 32) (see individual citations in Thomas and Burrough, 2016), (f) speleothem growth phases at Drotsky’s Cave (site 36) and Bone Cave (site 37) (Brook et al., 1998), (g) mega-lake Makgadikgadi (sites 39 to 42) (Burrough et al., 2009), (h) dune accumulation ages for the northeast (NE) (site 38) and eastern (EK) (site
dunefields in the northern Kalahari (see individual citations in Thomas and Burrough, 2016), and (i) central Kenya Rift lakes outside the region of interest for comparison (combined within Trauth et al., 2003).

Figure 5. Proxy records for the southern Kalahari and eastern fringes of the southern Kalahari, as well as some records outside eastern boundary of Kalahari for the past 190 ka. Southern Kalahari: (a) index from ratio of (Zr+Ti)/(Al+Ca) in pan sediments where lower values indicate wetter (fluvial inputs) and higher values reflect drier (aeolian inputs) conditions for Omongwa (site 46) and Branddam Pan East (site 49) (Schüller et al., 2018) [*note calibration curve for 14C in Schüller et al. (2018) is not known], (b) Stampriet aquifer (site 47) δ18O and noble-gas derived temperature record (Stute and Talma, 1998), (c) dune accumulation ages for western (site 48) and southern Kalahari (site 50) dunefields (ages compiled within Thomas and Burrough, 2016), (d) Witpan (site 51) pan floor (blue) and lunette (red) record (Telfer et al., 2009), (e) palaeoenvironmental summary for Kathu Pan (site 57), where left (yellow) indicates semi-arid conditions, blue indicates wet conditions and red indicates dry conditions (Lukich et al., 2020), (f) fluvial (blue on left) and aeolian (red on right) sediments within the Lower Molopo River (site 53) (Hürkamp et al., 2011). Eastern fringes of the southern Kalahari: (g) Lethakeng aquifer (site 62) δ18O and noble-gas derived temperature record (left plot) and excess air (ΔNe) (right plot) (Kulongoski et al., 2004), (h) Lobatse II Cave (site 61) speleothem δ18O record (Holmgren et al. 1995), (i) Wolkberg Cave (site 65) speleothem δ18O record (Holzkämper et al., 2009), (j) Cold Air Cave (site 63) speleothem δ18O record (Holmgren et al., 2003), (k) growth record of Gladysvale speleothem (site 68) (Pickering et al., 2007), (l) key proxies from the Tswaing Crater (site 67) record, where left to right is: sediment grain-size derived precipitation record (Partridge et al., 1997), total organic carbon (TOC) and total nitrogen (TN) (Kristen et al., 2007) and δDwax and δ15N (Schmidt et al., 2014). Eastern Kalahari: (f) Lethakeng aquifer δ18O and noble-gas derived temperature record and (g) excess air (ΔNe) (Kulongoski et al., 2004), (h) Lobatse II Cave speleothem δ18O record (Holmgren et al. 1995). Records east of eastern Kalahari (i) Wolkberg Cave speleothem δ18O record (Holzkämper et al., 2009), (j) Cold Air Cave speleothem δ18O record (Holmgren et al., 2003), (k) growth record of Gladysvale speleothem (Pickering et al., 2007), and (l-n) key proxies from the Tswaing Crater record, where (l) is sediment grain-size derived precipitation record (Partridge et al., 1997), (m) is total organic carbon (TOC) and total nitrogen (TN) (Kristen et al., 2007) and (n) is δDwax and δ15N (Schmidt et al., 2014).

Figure 6. Dune accumulation ages for the separate dunefields in the Kalahari in time bins of increasing width to reflect reduced absolute precision through time using a binary system (1 for a date in that bin) (adapted from Thomas and Burrough, 2012), where SnK is southern Kalahari, WK is western Kalahari, NWK is north-western Kalahari, NEK is north-eastern Kalahari and EK is eastern Kalahari.
**Table 1.** A classification of archives and proxies found in dryland regions.

<table>
<thead>
<tr>
<th>Zone</th>
<th>Surface</th>
<th>Subsurface</th>
<th>Hydrogeological</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Archive</strong></td>
<td>Rock shelters including hyrax middens and anthropogenic</td>
<td><strong>&quot;Geoproxies&quot;</strong></td>
<td><strong>Fluvial deposits</strong></td>
</tr>
<tr>
<td><strong>Zone</strong></td>
<td><strong>Surface</strong></td>
<td><strong>Hydrological</strong></td>
<td><strong>Subsurface</strong></td>
</tr>
<tr>
<td><strong>Zone</strong></td>
<td><strong>Dunes</strong></td>
<td><strong>Sand ramps</strong></td>
<td><strong>Lake shorelines</strong></td>
</tr>
<tr>
<td><strong>Zone</strong></td>
<td><strong>1965</strong></td>
<td><strong>1966</strong></td>
<td><strong>1967</strong></td>
</tr>
<tr>
<td><strong>Dating technique applied</strong></td>
<td>$^{14}$C</td>
<td>OSL, $^{14}$C, CN</td>
<td>OSL</td>
</tr>
<tr>
<td><strong>Proxies within</strong></td>
<td>$^{13}$C</td>
<td>Presence/accumulation of sediment</td>
<td>Sediments derived from slope or aeolian processes</td>
</tr>
<tr>
<td><strong>Response to climatic forcing</strong></td>
<td><strong>Indicate:</strong></td>
<td><strong>Indicate:</strong></td>
<td><strong>Indicate:</strong></td>
</tr>
<tr>
<td><strong>Vegetation type</strong></td>
<td><strong>Moisture availability.</strong></td>
<td><strong>Windy Dry enough</strong></td>
<td><strong>Interplay of slope, colluvial and aeolian processes.</strong></td>
</tr>
<tr>
<td><strong>Response to climatic forcing</strong></td>
<td><strong>Vegetation type.</strong></td>
<td><strong>Sediment supply/available</strong></td>
<td><strong>Sediment supply/available</strong></td>
</tr>
<tr>
<td>Lat/Long</td>
<td>Name (reference)</td>
<td>Lat/Long</td>
<td>Name (reference)</td>
</tr>
<tr>
<td>---------</td>
<td>----------------</td>
<td>---------</td>
<td>----------------</td>
</tr>
<tr>
<td>18.26°S, 13.89°E</td>
<td>Various sites (Bubenza et al., 2007)</td>
<td>39</td>
<td>Various sites (Lake Ngami (Huntman-Mapilla et al., 2006; Cordova et al., 2007; Burrough et al., 2009))</td>
</tr>
<tr>
<td>18.82°S, 12.50°E</td>
<td>Khumub (Van der Merwe, 2000)</td>
<td>2</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>18.97°S, 12.70°E</td>
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<td>40</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>19.29°S, 14.04°E</td>
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<td>41</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>20.46°S, 14.44°E</td>
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<td>42</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>21.83°S, 15.20°E</td>
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<td>43</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>approx. 22.34°S, 15.08°E</td>
<td>Various sites (Bubenza et al., 2007)</td>
<td>44</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>22.53°S, 14.80°E</td>
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<td>45</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>23.00°S, 13.90°E</td>
<td>Various sites (Bubenza et al., 2007)</td>
<td>46</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>Northern NSS dunes (Bristow et al. 2007)</td>
<td>Various sites (Bubenza et al., 2007)</td>
<td>47</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>24.07°S, 15.97°E</td>
<td>Various sites (Bubenza et al., 2007)</td>
<td>48</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>24.72°S, 16.05°E</td>
<td>Various sites (Bubenza et al., 2007)</td>
<td>49</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>Tsachua, Sossus Vlei (Brook et al., 2006)</td>
<td>Various sites (Bubenza et al., 2007)</td>
<td>50</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>24.98°S, 15.93°E</td>
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<td>51</td>
<td>Various sites (Mapila et al., 2006)</td>
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<tr>
<td>26.45°S, 16.53°E</td>
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<td>52</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>26.68°S, 16.10°E</td>
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<td>53</td>
<td>Various sites (Mapila et al., 2006)</td>
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<td>26.73°S, 16.16°E</td>
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<td>54</td>
<td>Various sites (Mapila et al., 2006)</td>
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<td>27.70°S, 17.13°E</td>
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<td>55</td>
<td>Various sites (Mapila et al., 2006)</td>
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<tr>
<td>28.86°S, 17.07°E</td>
<td>Various sites (Bubenza et al., 2007)</td>
<td>56</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>Western NSS dunes (Bristow et al. 2007)</td>
<td>Various sites (Bubenza et al., 2007)</td>
<td>57</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>29.00°S, 19.14°E</td>
<td>Various sites (Bubenza et al., 2007)</td>
<td>58</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>29.00°S, 19.14°E</td>
<td>Various sites (Bubenza et al., 2007)</td>
<td>59</td>
<td>Various sites (Mapila et al., 2006)</td>
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<tr>
<td>19.01°S, 17.79°E</td>
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<td>60</td>
<td>Various sites (Mapila et al., 2006)</td>
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<td>19.11°S, 17.07°E</td>
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<td>Various sites (Mapila et al., 2006)</td>
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<td>19.23°S, 17.35°E</td>
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<td>19.40°S, 17.88°E</td>
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<td>Various sites (Mapila et al., 2006)</td>
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<td>Various sites (Mapila et al., 2006)</td>
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<td>68</td>
<td>Various sites (Mapila et al., 2006)</td>
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<td>Various sites (Bubenza et al., 2007)</td>
<td>69</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
<tr>
<td>Various sites (Bubenza et al., 2007)</td>
<td>Various sites (Bubenza et al., 2007)</td>
<td>70</td>
<td>Various sites (Mapila et al., 2006)</td>
</tr>
</tbody>
</table>
Table 3. Preliminary chronostratigraphic time periods for the five sub-regions within southern African drylands for the past ~190 ka, as identified in this study. The chronostratigraphy proposed for southern Africa by Knight and Fitchett (2021), which provides local-names for the Marine Oxygen Isotope Stages (MIS), which are as defined by Lisic and Raymo (2005) is given in the final column for reference.

<table>
<thead>
<tr>
<th>Suggested chronostratigraphic unit</th>
<th>Namib (west coast)</th>
<th>N Kgalagadi</th>
<th>S Kgalagadi</th>
<th>SE-E Kgalagadi fringes</th>
<th>Correspondence with suggested scheme in overview paper (names in brackets) and MIS, using LR04 boundaries</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Wet interval</strong> (1) 140 to 120 ka (peak ~132 ka)</td>
<td>Wet mega-lake Makgadikgadi centred on 131 ± 11 ka (Burrough et al., 2009). Speleothem growth in Drotsky’s Cave in 132.9 ± 26.6 ka (Brook et al., 1998). Otavi caves and cenote speleothems growing 130 ± 7 ka indicate lower groundwater table than present, but wet enough for speleothems to deposit (Brook et al., 1999). [East African Rift Valley Lakes also wet, peaking 139-133 ka (Trauth et al., 2003)].</td>
<td>Kathu Pan sediments with palygorskite indicating semi-arid conditions 156 ± 11 to 121 ± 6 ka (Lukich et al., 2019; 2020). Lack of dune ages (either accumulation of preservation) 174 to 107 ka (collated dataset in Thomas and Burrough, 2016), following deposition (&gt;186 and &gt;183 ka) near Stampsriet (Stone and Thomas, 2008). However, deposition of Mamatwan mottled sandstone unit 160 to 108 ka (Bateman et al., 2003).</td>
<td>No evidence dated to this interval.</td>
<td>No evidence dated to this interval.</td>
<td>This falls across end of MIS 6 (191-127 ka) (Agulhasian) and into MIS 5e (peak 123 ka) (Knysnaian).</td>
</tr>
<tr>
<td><strong>Dry interval?</strong></td>
<td>Wetter conditions, to the extent Tsondab River overcame transmission losses 60 km further into Namib Sand Sea between ~132 and 87 ka (Stone et al., 2010a). Sand ramp sediments with minor slope-process ±150 to ±90 ka at Samara and Aus (Rowell et al., 2017). Individual wetter intervals cannot be resolved given chronological precision and available stratigraphic resolution.</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Wet interval</strong> (2) ~112 to 90 ka (possibly two phases ~105 and ~92)</td>
<td>Otavi caves and cenote speleothems growing 112-108 ka indicate lower groundwater table than present, but wet enough for speleothems to deposit (Brook et al., 1999). Wet mega-lake Makgadikgadi centred on 105 ± 4 ka and 92 ± 2 ka (Burrough et al., 2009). Speleothem growth in Drotsky’s 92.9 ± 5.9 ka (Brook et al., 1998). [East African Rift Valley Lakes also wet – Lake Navaisha 113-108 and ~91 ka (Trauth et al., 2003).]</td>
<td>Cascade tufa deposition at Ga-Mohana Hill North Rockshelter 110.6 ± 3.0 to 102.1 ± 2.1 ka indicating abundant water (Wilkins et al., 2021). Pedagogic unit at Kathu Pan at 96 ± 5 to 84 ± 4 ka, indicating wetter conditions (Lukich et al., 2020). Lack of dune ages ~98-87 ka (collated dataset in Thomas and Burrough, 2016).</td>
<td>No evidence dated to this interval.</td>
<td></td>
<td>Broadly MIS 5d to MIS 5c or 5b (MIS5a-d called Blombosian).</td>
</tr>
<tr>
<td><strong>Dry interval?</strong></td>
<td>Wetter conditions, to the extent Tsondab River can overcome transmission losses 60 km further into Namib Sand Sea at some point between ~87 and 68 ka (Stone et al., 2010a).</td>
<td>Etosha Pan lake 75-70 ka (Hipondoka et al., 2014).</td>
<td>Kathu Pan sediments with palygorskite indicating semi-arid conditions 74 ± 5 ka (Lukich et al., 2020). Dunes accumulating during this period (collated dataset in Thomas and Burrough, 2016).</td>
<td></td>
<td>Falls from around MIS 5a peak toward MIS5/4 boundary. (MIS 4 called Wildernessian).</td>
</tr>
<tr>
<td><strong>Wet interval</strong> (3) ~80 to 70</td>
<td>Wet mega-lake Makgadikgadi centred on 64 ± 2 ka (Burrough et al., 2009). Otavi caves and cenote speleothems growing 61 ± 2 ka indicate lower groundwater table than present, but wet enough for speleothems to deposit (Brook et al., 1999).</td>
<td>Kathu Pan sediments with palygorskite indicating semi-arid conditions 55 ± 3 ka (Lukich et al., 2020). Dunes accumulating during this period (collated dataset in Thomas and Burrough, 2016).</td>
<td>No evidence dated to this interval.</td>
<td></td>
<td>Falls from end of MIS 4 to middle of MIS 3 (Thukelian).</td>
</tr>
<tr>
<td><strong>Dry interval?</strong></td>
<td>Higher rainfall over eastern catchments of rivers: Hoanib ~61-41 ka only weak summer monsoon rain (Eitel et al., 2006), Hoarusib ~44-40 ka over whole</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Interval</td>
<td>Events and Observations</td>
<td></td>
<td></td>
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<td>--------------------------------</td>
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<td>Drier interval in east only</td>
<td>43-40 ka to ~34 ka</td>
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<td>No evidence dated to this interval.</td>
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<td>Minor excursions to wetter conditions in δ15N composite midden record 46-44 cal ky B.P. and ~40 cal ky B.P. (Chase et al., 2019).</td>
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<tr>
<td>Wetter interval (5)</td>
<td>34-24 ka</td>
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<td>Wet mega-lake Makgadikgadi centred on 39 ± 2 ka and again 27 ± 1 ka (Burrough et al., 2009). Wet in Tsodilo Hills basin 36-32 ka and 27-22 ka (Thomas et al., 2003). Boteti River back-flooding 28 ± 2 ka (Shaw et al., 1997). Etosha Pan lake 34-27 ka (Hipondoka et al., 2014). Otavi caves/cenote speleothems growing 31 ± 2, and 29 ± 1 to 28 ± 1 indicate lower groundwater table than present, but wet enough for speleothems (Brook et al., 1999). [East African Rift Valley Lakes also wet for latter – Naukuru-Elmenteita Basin 30-28.5 ky B.P. (Trauth et al., 2003)].</td>
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<td>Witpan pan floor flooding 32 ± 5 ka (Telfer et al., 2009). Stampriet Aquifer recharge ~36 to 33 ka (Stute and Talma, 1998). Wetter at Wonderwerk Cave ~ 35.2 to 31.1 cal ky B.P. to 23.6 to 17.2 cal ky B.P. (Brook et al., 2010). But drier at Omongwa Pan ~30-25 (peak 28-27) cal ky B.P. (*) (Schüller et al., 2018). Dunes accumulating during this period (collated dataset in Thomas and Burrough, 2016). But increased aridity at Kathu Pan to form a hard pan, starting at 32 ± 2 ka (Lukich et al., 2012). To the south the Modder River sediments record cooling and drying from 28 ka through to 15.5 ka (Lyons et al., 2014).</td>
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<td>Wetter interval (6)</td>
<td>23-16 ka</td>
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<td>Cross-bedded gravels grading into coarse sands in the Khumib from 20.8 ± 3.2 to 15.6 ± 2.7 ka (Srivastava et al., 2004). Sand ramp with major slope processes continuing in this interval (Rowell et al., 2017). Excursions to wetter conditions in δ15N composite midden record ~24.5 and ~22 cal ky B.P. (Chase et al., 2019). Etosha Pan lake 23-21 ka and 18-16 ka (Hipondoka et al., 2014). Wet mega-lake Makgadikgadi centred on 17 ± 2 ka (Burrough et al., 2009). Fluvial deposits in Xaudum Valley 18.1 to 17.3 cal ky B.P. and Gilkwe Ridge/Okwa Gorge 18.0 to 17.2 cal ky B.P. (Shaw et al., 2992)., Supported by non-drainable Sn African record at Lake Tritrivakely, Madagascar (Gasse and van Campo, 2001) and E and N of dryland regions at Lakes Chilwa (Barker et al., 2007) and Malawi (Thomas et al., 2009). Lower Molopo fluvial units 23.4 to 22.3 through to 16.2 to 15.7 cal ky B.P. (Hürkamp et al., 2011). Witpan pan floor sediments indicating pan flooding ~20 ka (Telfer et al., 2009). Wetter at Wonderwerk Cave 23-17 ka (Brook et al., 2010). But drier at Branddam East Pan ~19.5-18 cal ky B.P. (<em>) (Schüller et al., 2018), whilst Omongwa Pan gets wetter again ~17.5 cal ky B.P. (</em>) (Schüller et al., 2018). But at Kathu Pan, a hard pan dated to 22 ± 1.2 ka indicates an increase in aridity (Lukich et al., 2020). There is recharge to the Stampriet Aquifer from ~23 cal ky B.P. onwards (Stute and Talma, 1998).</td>
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<td>Event Description</td>
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<td>Wetter (7) Centred on ~15 ka</td>
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<td>Wetter excursions composite midden δ^15N ~15 and ~14.3 kyr B.P. (Chase et al., 2019). The Orumana stalagmite is still growing until 14.5 ka (Railsback et al., 2016). Also seen in the Khumib as cross-bedded gravels grading into coarse sands in the Khumib ~14.4 ± 2.2 ka (Srivistava et al., 2004) and Kuiseb as Homeb silts 16.3 ± 2.6 to 14.2 ± 1.7 ka (Srivistava et al., 2006). Slope processes in Neuhof-sand ramp during this interval (spanning ~30 to ~11 ka) (Rowell et al., 2017). Wet in Tsodilo Hills basin 16-12 ka (Thomas et al., 2003). Otavi caves and cenote speleothems growing 15 ± 1 to 14 ± 0.3 ka indicate lower groundwater table than present, but wet enough for speleothems to deposit (Brook et al., 1999). Braddam East Pan wetter ~15-13 cal kyr B.P. (*) (Schüller et al., 2018). Dunes accumulating during this period (collated dataset in Thomas and Burrough, 2016).</td>
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<td>Drying west, south and east, remains wetter in north ~10-5 ka</td>
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<td>Start of drying trend ~10 kyr B.P. in composite midden δ^15N (Chase et al., 2019). Tsondab drying up with last Ancient Tracks units ending ~ 10.5 ka (Stone et al., 2010a). Flash flood units at ~9 ka in Tsauchab (Brook et al., 2006) and ~8.1 ka in the Khumib (Srivastava et al., 2006). Etosha Pan lake ~10 ka, and then shallower flooding/lake ~7.4 ka (Hippodoka et al., 2014). Wet mega-lake Makgadikgadi centred on 8 ± 5 ka (Burrough et al., 2009). Growth of speleothems in Drotsky's and Bone Cave ~8.2 ka (Brook et al., 1998). Otavi caves and cenote speleothems growing 8.8 ± 0.5 to 7.5 ± 0.3 ka indicate lower groundwater table than present, but wet enough for speleothems to deposit (Brook et al., 1999). Absence of dune accumulation after ~12 and ~8 ka in the northwest and eastern dunefields (collated from Thomas and Burrough, 2016). Ephemeral conditions in the lower Molopo River, with some flash-flood sediments ~9 to 6.5 ka, and then only alluvial fan aggradation from 6.5 ka. Omongwa and Branddam East Pans wetter starting ~8 cal kyr B.P. (*) (Schüller et al., 2018). Increasing aridity at Kathu Pan from ~10.4 ka as calcium carbonate replaces palygorskite (Lukich et al., 2020). Dunes accumulating during this period (collated dataset presented in Thomas and Burrough, 2016).</td>
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<td>Further drying in the west 5 ka - present</td>
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<td>Increase in aridity from ~5 kyr B.P. in δ^15N composite midden record (Chase et al., 2019). Paucity of Namib River deposits. A shallower lake than before at Etosha Pan from ~5.1 to 0.3 ka (Hipondoka et al., 2014). Transition to drier conditions between 4.6 and 3.0 ka (δ^18O shift from -11.5 to -6 ‰) in Dante Cave isotopic record (Sletten et al., 2013). Peak in this wet phase at Omongwa ~5.5 cal kyr B.P. (<em>) (Schüller et al., 2018). Fluvial deposits in the Kuruman River 3.2 to 2.7 cal kyr B.P., 1.8 to 1.5 cal kyr B.P. and 0.6 to 0.5 cal kyr B.P. (Shaw et al., 1992). Dunes accumulating during this period (collated dataset in Thomas and Burrough, 2016). Speleothem growth has already ceased at Lobatse II (Holmgren et al., 1995). Peak in this wet phase at Omongwa ~5.5 cal kyr B.P. (</em>) (Schüller et al., 2018). Fluvial deposits in the Kuruman River 3.2 to 2.7 cal kyr B.P., 1.8 to 1.5 cal kyr B.P. and 0.6 to 0.5 cal kyr B.P. (Shaw et al., 1992). Dunes accumulating during this period (collated dataset in Thomas and Burrough, 2016). Speleothem growth has already ceased at Lobatse II (Holmgren et al., 1995). The cooler, drier, little ice age (1320-1760 AD) is recorded in Cold Air Cave, further east than this region (Holmgren et al., 2003).</td>
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*just before MIS2/1 boundary, (MIS 1 remains called Holocene)
Figure 1. Southern Africa, showing (a) locations of the mega-Kalahari and its dunefields (NWK is northwest Kalahari, NEK is northeast Kalahari, EK is eastern Kalahari, WK is western Kalahari and SnK is southern Kalahari), and the Namib Desert and its sand sea and dunefield regions, also showing the four sub-regions considered in this paper (Namib Desert, northern Kalahari, southern Kalahari and eastern fringes of the southern Kalahari) (b) mean annual precipitation and extent of the winter rainfall zone (WRF) and summer rainfall zone (SRZ), with the year round rainfall zone in-between (modified from Chase and Meadows, 2007), (c) aridity index, where <0.5 is boundary for semi-arid (modified from Trabucco and Zomer, 2009)(d) biome distributions (modified from Truc et al., 2013), (e) the rainfall regions and numbering system of Nicholson (1986; 2001), Nicholson et al., 2012) (redrawn from Nicholson, 2001).
Figure 2. Location of sites referred to in the text, site names and references given in Table 2. The open circles represent sites within the sub-regions of interest, whilst the solid squares are from sites beyond the margin regions that offer useful proxies for comparison. Where there are lines grouping multiples sites to one number these represent individual sites dated within dunefields (32, 38, 43, 48 and 50) or individual sites within the lower Molopo catchment (site 53). The ephemeral rivers in the Namib, from north to south are Khumib, Hoarusib, Hoanib, Uniab, Koigab, Huab, Omaruru, Swakop, Kuiseb, Tsondab, Tsauchab.
Figure 3. Proxy records for the Namib Desert for the past 60 ka, arranged broadly N to S for Namib and then the northern Namibian record. (a) the Namib composition midden ¹⁵N record of Chase et al. (2019), for which positive values indicate drier, and negative values indicate wetter, conditions, (b) the Khumib (site 2), Hoarusib (site 3) and Hoanib (site 4) ephemeral river records, as summarised within Stone and Thomas (2013) but compiled from Srivastava et al. (2004) (Khumib), Srivastava et al., (2005) (Hoaruisb) and Eitel et al., (2006) (Hoanib), where solid fill is wetter and diagonal indicates only weak summer rains, (c) Orumana Cave (site 1) speleothem δ¹⁸O, located in upper catchment of the Hoarusib River (Railsback et al., 2016), (d) ¹⁵N and δD n-C alkene record from the Spitzkoppe midden (site6) (Chase et al., 2010), (e) the Kuiseb (site 9), Tsondab (site 12) and Tsauchab (site 16) ephemeral river records, as summarised within Stone and Thomas (2013) but compiled from Srivastava et al. (2006) left hand and Bourke et al. (2002) right hand (Kuiseb), Stone et al. (2010a) (Tsondab) and Brook et al. (2006) (Tsauchab), (f) dated sand ramp units (sites 11, 15, 17, 19, 20 and 21) where orange solid are pure aeolian sand, striped are minor slope processes and solid blue rectangle is major slope deposit (Rowell et al., 2017).
Figure 4. Proxy records for the northern Kalahari for the past 190 ka, arranged broadly west to east, and the central Kenyan Rift Lake level records for comparison, where: (a) Etosha lake (site 26) levels, solid is lake 21 to 24 m deep and striped is low (a few meters) (Hipondoka et al. 2014), and (b) Etosha lake (site 26) classes, where left to right is 6 (highest) to 1 (lowest) (De Cort et al., 2021), (c) speleothem growth phases combined from Aikab and Aigama (site 28) and Guinas Meer (site 29) cenotes (Brook et al., 1999), (d) Dante Cave (site 30) speleothem δ¹⁸O (Sletten et al., 2013), (e) dune accumulation ages for the northwestern (NWK) dunefield Kalahari (site 32) (see individual citations in Thomas and Burrough, 2016), (f) speleothem growth phases at Drotsky’s Cave (site 36) and Bone Cave (site 37) (Brook et al., 1998), (g) mega-lake Makgadikgadi (sites 39 to 42) (Burrough et al., 2009), (h) dune accumulation ages for the northeast (NE) (site 38) and eastern (EK) (site 43) dunefields in the northern Kalahari (see individual citations in Thomas and Burrough, 2016), and (I) central Kenyan Rift lakes outside the region of interest for comparison (combined within Trauth et al., 2003).
Figure 5. Proxy records for the southern Kalahari and eastern fringes of the southern Kalahari, as well as some records outside eastern boundary of Kalahari for the past 190 ka. Southern Kalahari: (a) index from ratio of (Zr+Ti)/(Al+Ca) in pan sediments where lower values indicate wetter (fluvial inputs) and higher values reflect drier (aeolian inputs) conditions for Omongwa (site 46) and Branddam Pan East (site 49) (Schüller et al., 2018) [*note calibration curve for 14C in Schüller et al. (2018) is not known], (b) Stampriet aquifer (site 47) δ18O and noble-gas derived temperature record (Stute and Talma, 1998), (c) dune accumulation ages for western (site 48) and southern Kalahari (site 50) dunefields (ages compiled within Thomas and Burrough, 2016), (d) Witpan (site 51) pan floor (blue) and lunette (red) record (Telfer et al., 2009), (e) palaeoenvironmental summary for Kathu Pan (site 57), where left (yellow) indicates semi-arid conditions, blue indicates wet conditions and red indicates dry conditions (Lukich et al., 2020), (f) fluviatile (blue on left) and aeolian (red on right) sediments within the Lower Molopo River (site 53) (Hürkamp et al., 2011). Eastern fringes of the southern Kalahari: (g) Letlhakeng aquifer (site 62) δ18O and noble-gas derived temperature record (left plot) and excess air (ΔNe) (right plot) (Kulongoski et al., 2004), (h) Lobatse II Cave (site 61) speleothem δ18O record (Holmgren et al. 1995), (i) Wolkberg Cave (site 65) speleothem δ18O record (Holzkämper et al., 2009), (j) Cold Air Cave (site 63) speleothem δ18O record (Holmgren et al., 2003), (k) growth record of Gladysvale speleothem (site 68) (Pickering et al., 2007), (l) key proxies from the Tswaing Crater (site 67) record, left to right is: sediment grain-size derived precipitation record (Partridge et al., 1997), total organic carbon (TOC) and total nitrogen (TN) (Kristen et al., 2007) and δDwax and δ15N (Schmidt et al., 2014).
Figure 6. Dune accumulation ages for the separate dune fields in the Kalahari in time bins of increasing width to reflect reduced absolute precision through time using a binary system (1 for a date in that bin) (adapted from Thomas and Burrough, 2012), where SnK is southern Kalahari, WK is western Kalahari, NWK is north-western Kalahari, NEK is north-eastern Kalahari and EK is eastern Kalahari.