

Manuscript Details

Manuscript number	JQSR_2018_100
Title	Late Quaternary palaeoclimates and human-environment dynamics of the Maloti-Drakensberg region, southern Africa
Article type	Invited review article

Abstract

The Maloti-Drakensberg Mountains are the highest in southern Africa and lie at a crucial interface between the sub-continent's drier, colder, more seasonal interior and its perennially productive sub-tropical coastal belt. The combination of location, high elevation and topography make them ideal for exploring human responses to late Quaternary climatic change. This paper reviews and synthesizes palaeoclimatic and palaeoenvironmental data from the Maloti-Drakensberg region over the past 50,000 years, the approximate limits of the radiocarbon time-scale. It then draws on a database of 325 calibrated radiocarbon dates to examine human occupational trends across the region as a whole and its component sub-regions in order to discuss human-environment dynamics over this time-span and patterning between particular phases of climatic change and the timing, mode, and motives of the region's exploitation by people. An agenda for future palaeoenvironmental and archaeological research is also mapped out.

Keywords	Quaternary; Pleistocene; Holocene; paleoclimatology; southern Africa; geomorphology, glacial; stable isotopes; Maloti-Drakensberg; archaeology; human-environment dynamics
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8 February 2018

Dr. Claude Hillaire-Marcel
Université du Québec à Montréal
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Dear Professor Hillaire-Marcel,

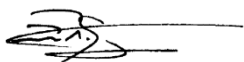
Thank-you for expressing interest in our review paper manuscript, **Late Quaternary palaeoclimates and human-environment dynamics of the Maloti-Drakensberg region, southern Africa**. I am writing this cover letter to accompany our manuscript submission.

As I mentioned in my email of February 2nd 2018, our manuscript reviews 50,000 years of palaeoclimatic, palaeoenvironmental and hunter-gatherer occupational data from the Maloti-Drakensberg region of southern Africa, the continent's highest altitude zone south of Mount Kilimanjaro. Ours represent's the first comprehensive synthesize the Late Quaternary climatic/environmental and human history of this unique southern African bioregion, and the first collation and calibration of all available radiocarbon dates to explore deep-time patterning in the region's hunter-gatherer occupation.

After an introduction and background sections on sources of evidence and the region's modern climate and ecology, the paper summarizes all known proxy data the Maloti-Drakensberg's palaeoclimatic/environmental history before drawing on a database of over 300 calibrated archaeological radiocarbon dates (explored with summed probability distributions and histograms) to investigate associated trends of human occupation and highland cultural adaptations as they relate to climatic picture that we present.

The manuscript is just over 13,000 words (13,284) and includes seven figures, captions for which can be found at the end of the manuscript document (after the references).

Yours sincerely,



Dr. Brian A. Stewart



Late Quaternary palaeoclimates and human-environment dynamics of the Maloti-Drakensberg region, southern Africa

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ABSTRACT

The Maloti-Drakensberg Mountains are the highest in southern Africa and lie at a crucial interface between the sub-continent's drier, colder, more seasonal interior and its perennially productive sub-tropical coastal belt. The combination of location, high elevation and topography make them ideal for exploring human responses to late Quaternary climatic change. This paper reviews and synthesizes palaeoclimatic and palaeoenvironmental data from the Maloti-Drakensberg region over the past 50,000 years, the approximate limits of the radiocarbon time-scale. It then draws on a database of 325 calibrated radiocarbon dates to examine human occupational trends across the region as a whole and its component sub-regions in order to discuss human-environment dynamics over this time-span and patterning between particular phases of climatic change and the timing, mode, and motives of the region's exploitation by people. An agenda for future palaeoenvironmental and archaeological research is also mapped out.

KEYWORDS

Quaternary; Pleistocene; Holocene; paleoclimatology; southern Africa; geomorphology, glacial; stable isotopes; Maloti-Drakensberg; archaeology; human-environment dynamics

HIGHLIGHTS

- Elevation and location make Maloti-Drakensberg highly sensitive to climate change
- Proxy data document greater moisture availability during glacial and stadial maxima
- Diverse climatic conditions saw pulsed human occupation of different sub-regions
- Mountain core provided refuge for human populations during drier/unstable episodes
- Flexible responses to resource stress included intensified fishing in cooler periods

1. Introduction

The Maloti-Drakensberg Mountains are southern Africa's highest, their tallest peak, Thabana Ntlenyana, reaching 3482 metres above sea level (m a.s.l.). One of Africa's topographically most diverse regions, they are among the world's oldest centres of plant endemism (Kingdon, 1989). Several fish, amphibian, reptile, and bird taxa also occur nowhere else (World Wildlife Fund, 2017). Additionally, the region preserves a rich archaeological record that includes one of Africa's best-understood concentrations of rock art (Lewis-Williams, 2003). This combination of cultural and natural heritage was instrumental in establishing the Maloti/Drakensberg Trans-Frontier Conservation Area either side of the mountains' highest peaks and in UNESCO designating part of the region a World Heritage Site (Derwent et al., 2001).

From a palaeoclimatic and archaeological standpoint, however, the real significance of the Maloti-Drakensberg and the lower-lying areas adjoining them is twofold. First, they lie at a crucial interface between the perennially productive sub-tropical coastal forelands of South Africa's KwaZulu-Natal and Eastern Cape provinces and its more seasonal, colder, drier interior. Second, their elevation and topographic diversity reinforced their susceptibility to late Quaternary climatic shifts. They therefore offer an excellent opportunity for exploring human responses to changing environmental conditions along such dimensions as technology, subsistence, demography, settlement and mobility decisions, and the structures used to maintain social connectivity over distance. Current archaeological data allow us to investigate these questions across the last 80,000 years (Stewart et al., 2012, 2016).

A wide range of palaeoenvironmental proxies exists with which to do this, many recovered from archaeological excavations, others from contexts such as peat bogs where human input can be excluded. These archives have not previously been coherently summarized with a view to assessing their potential for understanding late Quaternary climatic change or the latter's implications for human behaviour. The most recent overview of relevance (Fitchett et al., 2016a) focuses solely on palaeoclimatic archives from within the kingdom of Lesotho, which sits in the middle of the wider Maloti-Drakensberg region. However, Lesotho's present boundaries were only defined in the late 1800s (Eldredge, 1993) and are irrelevant in a prehistoric context except as they shape the location and format of ongoing research. To understand the cultural dynamics of past human populations and the effects on them of climatic and ecological change we need to operate at a spatial scale that transcends modern borders.

To do this we first introduce the Maloti-Drakensberg region as a whole, discussing its current climate and ecology and the main controls on these. Next, we critically consider the sources available for understanding its climatic history, emphasizing the importance and challenges of integrating information from as many proxies as possible. The main part of our paper then reviews current knowledge of Maloti-Drakensberg palaeoclimates and palaeoenvironments over the last ~50 kyr. Drawing on a database of radiocarbon dates from archaeological sites for the region that we are developing, we examine human occupational trends across the Maloti-Drakensberg and its sub-regions. Finally, we discuss human-environment dynamics over this time-span, searching for patterning between particular phases of climatic change (relative stability and instability, cooling and warming, greater or lesser precipitation), and the

101 timing, mode, and motives of the region's exploitation by people. In our conclusion
102 we identify the most pressing needs for future palaeoenvironmental and
103 archaeological research in this, the highest part of southern Africa.

105 **2. Defining the Maloti-Drakensberg: climate and ecology**

107 The Maloti-Drakensberg Mountains lie at the heart of the region that takes their name
108 (Fig. 1). Collectively, they form a deeply dissected plateau, anchored in the east by
109 the uKhahlamba-Drakensberg Escarpment, the watershed between the Indian and
110 Atlantic Oceans. This is one of three sub-parallel ranges that collectively fill out a
111 roughly quadrangular 55,000-km² massif and consist of Drakensberg Group flood
112 basalts capping several Karoo Supergroup sedimentary strata. Intense fluvial erosion
113 of these geological strata has produced an intricate network of drainages feeding into
114 southernmost Africa's largest river, the Orange (Gariep), known in Lesotho as the
115 Senqu.

117 The Senqu and its immediate tributaries flow west of the Escarpment. Numerous
118 other rivers (the Thukela, Bushman's, Mooi, Mngeni, Mkhomazi, Mzimkhulu,
119 Mzimvubu etc.) drain east into the Indian Ocean, travelling through the 'Little Berg',
120 a prominent sandstone terrace ~1800-1700 m a.s.l. that runs parallel to, but at some
121 distance from, the top of the Escarpment. West of the Senqu, the Central and the
122 Front Ranges of the Maloti intervene before one reaches Lesotho's lowlands. Two
123 important tributaries of the Senqu, the Senqunyane and the Makhaleng, run southward
124 between them. The Caledon River, which forms Lesotho's western border with South
125 Africa, draws almost all its water from precipitation falling on the Front Range, only

merging with the Orange after both rivers enter South Africa. To the west its valley merges into the highveld grasslands that occupy most of South Africa's Free State province. Finally, to Lesotho's south the Drakensberg Mountains extend into South Africa's Eastern Cape Province. Often termed Nomansland because of its politically contested position in the mid-1800s, this area reaches as far as Elliot and Maclear and includes the Witteberg Range on Lesotho's southwestern border. To facilitate our discussion we divide the Maloti-Drakensberg region into six sub-regions: (1) the northern KwaZulu-Natal uKhahlamba-Drakensberg; (2) the southern KwaZulu-Natal uKhahlamba-Drakensberg; (3) Nomansland; (4) highland Lesotho; (5) lowland Lesotho; and (6) the eastern Free State. A selection of landscape views of each region is offered in Fig. 2.

The Maloti-Drakensberg's climate is continental, with cold, dry winters and warm, humid summers. It lies in southern Africa's summer rainfall zone (SRZ), receiving over 75% of its rainfall between October and March, mostly as a result of easterly airflow from the Indian Ocean. Precipitation totals vary tremendously with altitude and locality, decreasing from east to west because of the pronounced rain-shadow cast by the uKhahlamba-Drakensberg Escarpment. Thus, while estimates of mean annual precipitation for the Escarpment typically exceed 1500 mm (Schulze, 1979), mean values of 580–700 mm are recorded in the upper Senqu and Caledon Valleys (Bawden and Carroll, 1968). Temperatures vary drastically by altitude, as well as seasonally and diurnally. Mean annual values range from ~15°C in the Caledon Valley to ≤6°C on the highest mountains (Grab, 1994, 1997). Corresponding mean mid-summer maxima and mid-winter minima are 29°C and 4.3°C and 17°C and –6.1°C respectively (Grab and Nash, 2010). Snow can fall at any time, but especially

between May and September, and may persist on south-facing slopes for up to six months. Frost is also widespread, with ground freezing estimated to occur in the highest areas up to 200 days per year (Grab, 1997; Schulze, 2008).

Ecologically, the Maloti-Drakensberg lies within southern Africa's Grassland Biome (Fig. 2). As this implies, grasses dominate the region's vegetation, with trees rare except in protected locations such as valleys. The vegetation composition is significantly differentiated by altitude, although locally aspect is also significant and differences in rainfall produce further contrasts between areas below the Escarpment to the east and those west of the Front Range. The highest elevations (≥ 2900 m a.s.l.) support an Afroalpine short shrubland that includes ericaceous and composite taxa along with shorter, less palatable, C_3 -photosynthesizing *Festuca*- and *Merxmuellera*-dominated grasses (Killick, 1978; Mucinda and Rutherford, 2006). Numerous alpine bogs help regulate the flow of rainwater into the Orange-Senqu river system (van Zinderen Bakker and Werger, 1974).

Lying between ~1900 and 2900 m a.s.l. the rest of highland Lesotho, along with adjacent areas of Nomansland, is covered by dense, subalpine, C_4 -dominated *Themeda-Festuca* grassland with patchy shrublands in which *Passerina montana* is common (Mucinda and Rutherford, 2006). Due to its large altitudinal range it contains several altitude-specific vegetation belts. *Themeda triandra*, a C_4 grass that provides excellent pasture (Jacot Guillarmod, 1971), dominates at lower elevations (up to ~1900–2100 m a.s.l. on southern (cooler) slopes, but reaching ~2700 m a.s.l. on northern (warmer) slopes). Intruding into this along the lower reaches of the Senqu and its principal tributaries, a *Cymbopogon-Themeda-Eragrostis* C_4 grassland with

176 numerous trees and evergreen shrubs extends over the Caledon Valley and into the
177 Free State beyond.
178
179 East of the Escarpment, Acocks' (1975) highland sourveld extends between 2150 and
180 1350 m a.s.l. *Themeda triandra* is one of several common grasses here too, but forbs
181 are particularly noteworthy. *Protea* spp. (otherwise strongly associated with the
182 Fynbos Biome of southwestern South Africa) may be common on mountain slopes,
183 while montane forest dominated by *Podocarpus latifolius* characterizes protected
184 southeast-facing slopes. Areas of southern tall grassveld — an open savanna of
185 *Acacia sieberanna* in a mixed grassland dominated by *T. triandra* and *Hyparrhenia*
186 spp. — occur along valleys and at lower elevations below this, particularly in areas
187 draining northeast toward the Thukela River.
188
189 Unsurprisingly, grazers were common here in precolonial times, especially black
190 wildebeest (*Connochaetes gnou*), red hartebeest (*Alcelaphus buselaphus*), plains
191 zebra (*Equus quagga*), blesbok (*Damaliscus pygargus phillipsi*), mountain reedbuck
192 (*Redunca fulvorufula*), springbok (*Antidorcas marsupialis*), and the now extinct blue
193 antelope (*Hippotragus leucophaeus*). Eland (*Taurotragus oryx*), a mixed feeder, was
194 also common, with two small-medium species — oribi (*Ourebia ourebi*) and
195 klipspringer (*Oreotragus oreotragus*) — some of the few antelope to occur in the
196 Afroalpine belt. Geophytes like *Watsonia* spp. and *Moraea* spp. were among the
197 principal edible plants available, although people also consumed a wide variety of
198 fruits, seeds and grasses (Carter, 1978; Pooley, 2003). Along the Senqu (and probably
199 other major rivers too) fish formed another important resource (e.g. Mitchell et al.,
200 2011).

3. Sources of palaeoclimatic and palaeoenvironmental evidence

Initial studies of Maloti-Drakensberg palaeoenvironments focused overwhelmingly on the highest mountains, particularly on the extent to which this Afroalpine zone sustained glaciers during the Last Glacial Maximum (LGM) (Boelhouwers and Meiklejohn, 2002). Happily, the last two decades, in particular, have seen several new palaeoenvironmental archives explored. Derived from both archaeology- and non-archaeology-bearing sedimentary sequences, they reflect the work of multiple researchers over many years, often operating in logistically challenging conditions (e.g. Fitchett et al. 2016a, 2016b, 2017; Grab and Mills, 2011; Grab et al., 2005; Parker et al., 2011; Plug and Mitchell, 2008; Roberts et al., 2013; Smith et al., 2002; Stewart et al., 2012, 2016). Historical sources provide a more finely grained narrative for the nineteenth century (Grab and Nash, 2010; Nash and Grab, 2010). Individually and collectively these archives still contain many spatiotemporal gaps, making it difficult to develop a fully comprehensive picture for the region as a whole. Other lacunae are more methodological and reflect the innovation of new analytical techniques since some sites were first investigated or a dearth in the availability of qualified (and interested) analysts able to study samples that have already been recovered.

More fundamentally, when attempting to synthesize existing palaeoclimatic/palaeoenvironmental data and relate them to the dynamics of past human populations several points must be remembered. First, proxy evidence from archaeological sequences generally relates to what the past inhabitants of human

occupation sites introduced to them: charcoal is principally a by-product of firewood, phytoliths of food waste, artefacts, and bedding materials, and large mammal bones/teeth of animals hunted for their meat and hides. In all these cases people exercised choice in what they harvested and returned to their living sites (e.g. Shackleton and Prins, 1992). Moreover, such archives will, by definition, only have formed when people lived at those sites, although when humans were absent others (e.g. sedimentary records) may have continued to form or be richer (e.g. microfaunal assemblages, since the ovals that often concentrate them are unlikely to co-reside with people; Avery, 1982) Palaeoenvironmental archives where human action can be excluded as a source of proxy accumulation and relevant deposits build up more continuously, such as wetland cores, thus offer a vital complement to what can be learned from archaeological data points.

Because of this, and because different proxy data respond to climatic change at varying rates, it is essential to use as wide a variety of signals as possible, acknowledging that some may reflect change at more local rather than more regional scales or display different degrees of temporal lag (Meadows, 2014). Even so, the changes registered may be subject to equifinality, complicating attempts at directly inferring changes in temperature or moisture: faunal oxygen isotope ratios, for example, reflect both dietary and climatic factors (Smith et al., 2002:691).

Finally, many sequences continue to be hampered by chronologies (mostly of radiocarbon origin) that are still only coarsely resolved and may be varyingly reported using both calibrated and uncalibrated dates. Here, all the radiocarbon dates we employ have been (re-)calibrated using the most recent versions of OxCal (4.2.4) and

the southern hemisphere calibration curve (Hogg et al., 2013), although in some cases (the Mahwaqa wetland record from KwaZulu-Natal and the Tsoaing sediment sequence from western Lesotho) the dates cited depend on interpolation from published age-depth models (Grab et al., 2005; Neumann et al., 2014). We also present a region-wide summed probability distribution (SPD) using OxCal 4.2.4. We restrict our review to the past ~50 kyr because this represents the upper limit of the radiocarbon technique, but note that human history in the Maloti-Drakensberg has at least double — and in all likelihood more than quadruple — this antiquity (Carter, 1978).

4. Maloti-Drakensberg palaeoclimates and palaeoenvironments: a 50-kyr synthesis

4.1. Marine Isotope Stage 3

Fig. 3 presents a proxy data from all palaeoenvironmental archives in the Maloti-Drakensberg region that stretch beyond 16 ka. Also shown are a selection of major palaeoclimatic records from the broader SRZ. Our earliest detailed evidence comes from Melikane (1800 m a.s.l.), a highland Lesotho rockshelter with a pulsed archaeological sequence stretching back to ~83–80 ka (Stewart et al., 2012). Analyses of phytoliths and soil organic matter (SOM) $\delta^{13}\text{C}$ (mean = -23.06‰ , $n = 21$) suggest that all its later Pleistocene sediments were deposited while C_3 -dominated Afroalpine grassland surrounded the site (Stewart et al., 2016) (Fig. 3). Assuming a temperature drop of -0.6 °C per 100 m increase of altitude (Smith et al., 2002), the depression of Afroalpine vegetation belts now found $\geq 2700\text{ m a.s.l.}$ suggests that average

temperatures were $\geq 5^{\circ}\text{C}$ cooler than present-day, consistent with a pioneering study of $\delta^{13}\text{C}$ values in zebra teeth recovered from earlier excavations by Carter (1978) that implied a 75–84% C_3 -rich diet for these grazers in MIS 3 (Vogel, 1983).

Long considered a period of enduring aridity in southern Africa's interior, early Marine Isotope Stage (MIS) 3 is increasingly recognized as a climatically volatile period that included phases of high precipitation and moisture availability. Melikane registers two occupation pulses, the earlier of which ~ 50 ka coincides with the start of our review. These levels (Layers 22–16) contain its most abundant bulliform phytoliths, which preferentially silicify with elevated moisture, and its highest ratio of woody (ligneous dicotyledon) to grassy (Poaceae) morphotypes (Fig. 3). Both suggest a highland landscape with high moisture availability and more woodland than in any other pre-Holocene occupation pulse. Four SOM $\delta^{13}\text{C}$ samples provide a strongly C_3 (cool) signature (mean = -23.16‰), though a high sedimentary input from woody vegetation likely exaggerates this value (Stewart et al., 2016). Broadly contemporaneous records in the wider SRZ agree well with these data. In a series of caves (e.g. Gladysvale, Lobatse and Wolkberg) across South Africa's Savanna Biome conditions were humid enough to allow carbonate deposition for speleothem growth $\sim 56\text{--}42$ ka (e.g. Holzkämper et al., 2009), while at the Tswaing impact crater detrital input from enhanced precipitation was high $\sim 55\text{--}48$ ka (Kristen et al., 2007) (Fig 3). Closer to the Maloti-Drakensberg, higher temperatures, increased summer rainfall, and greater vegetation density are inferred ~ 48 ka compared to older levels at Sibudu Cave in KwaZulu-Natal's coastal belt (Bruch et al., 2012; Glenny, 2006).

Substantial changes are apparent in Melikane's second MIS 3 occupation pulse $\sim 46\text{--}$

38 ka (Layers 15–6). Drops in bulliform phytolith frequencies and, especially after 42 ka, in the ratio of dicotyledon to Poaceae morphotypes suggest major reductions in woodland under markedly drier conditions. C₃ grasses still heavily dominated the landscape (90–95%) at the start of this pulse, but after ~42 ka slightly enriched SOM $\delta^{13}\text{C}$ values (–23.89‰ to –22.77‰) and minor increases in C₄ grass phytoliths probably reflect a shift towards drier conditions and/or reduced atmospheric CO₂ ($p\text{CO}_2$) (Stewart et al., 2016) (Fig 3). A third possibility — higher temperatures — is less likely given that high altitude fynbos elements (*Rhamnus* sp., *Protea* sp., *Leucosidea sericea*, plus *Erica* sp. after ~41.5 ka) dominate the relevant charcoal assemblages (Stewart et al., 2016). Melikane also shows major changes in site formation processes beginning ~42 ka, with episodes of intense roof collapse alternating with recurrent influxes of colluvial gravels signaling heightened climatic instability (Carter, 1976; Stewart et al., 2012). Charcoals from riparian taxa persist throughout this phase, however, suggesting that the upper Senqu River and its larger tributaries sustained ample water flow (Stewart et al., 2016).

This second MIS 3 occupation pulse at Melikane also accords well with other regional proxy data. An orange gravel horizon directly above an organic-rich sand layer with silt and clay lenses radiocarbon dated to ≥ 43 ka in a colluvial/palaeosol sequence (‘Site 1’) at the head of the Sehonghong River near Thabana Ntlenyana in northeastern Lesotho indicates relatively wet, warm conditions before an extended period of colluviation that began ~44–42 ka (Fig. 3), when colder, drier conditions promoted downslope creep or mass wasting events on steep valley slopes destabilized through de-vegetation (Grab and Mills, 2011). Similar changes are registered across the uKhahlamba-Drakensberg Escarpment in the Masotcheni Formation of KwaZulu-

Natal's Midlands (Fig. 1), which consists of massive colluvial mantles deposited during what have been inferred as phases of heightened aridity punctuated by numerous palaeosols reflecting brief shifts to more humid conditions. Various Masotcheni locales are well dated using both OSL and radiocarbon to ~46–37 ka (e.g. Clarke et al., 2003); some also register the temperature drop at ~42 ka apparent at Melikane (Temme et al., 2008).

An occupational and sedimentary hiatus at Melikane between ~38 and 24 ka precluded the accrual of proxy data there later in MIS 3, but this interval is largely captured at Sehonghong rockshelter, 40 km further north (Loftus et al., 2015; Pargeter et al., 2017). Freshly obtained palaeoenvironmental data include SOM $\delta^{13}\text{C}$ values from the entire sequence and $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ measurements of large ungulate tooth enamel from strata dated to ~35–29 ka excavated in an ongoing project (Stewart et al., 2012, 2016). Consistent with Melikane, Sehonghong's SOM $\delta^{13}\text{C}$ values suggest a local landscape heavily dominated by C_3 vegetation before the onset of Holocene warming (mean = -23.7‰ , $n = 20$) with last glacial temperature depressions of $\geq 5^\circ\text{C}$ relative to today (Loftus et al., 2015) (Fig. 3). The lowest SOM $\delta^{13}\text{C}$ value (-24.7‰) ~30 ka matches precisely with the mean of three others from a broadly coeval stratum at Ntloana Tsoana in lowland western Lesotho, while five from the MIS 3 sequence at the adjacent site of Ha Makotoko range between -23.3‰ and -24.5‰ (Roberts et al., 2013) (Fig. 3). The $\delta^{13}\text{C}$ values of herbivore tooth enamel at Sehonghong dated to ~35–29 ka fall between -11.1‰ and -5.8‰ (mean = $-8.1 \pm 1.4\text{‰}$, $n = 29$). As pure C_3 - and C_4 -feeders have enamel value ranges of -14‰ to -12‰ and 0‰ to $+2\text{‰}$, respectively, the diets of these animals must have been predominantly C_3 in composition. Unlike the SOM $\delta^{13}\text{C}$ samples, however, they also included substantial

contributions of C₄ (up to 40%), suggesting that the zebras and alcelaphines in question migrated seasonally to lower altitudes beyond highland Lesotho. A strong deviation from this pattern (mean = −10.8‰, n = 3) implies an almost exclusively C₃ diet ~33 ka consistent with an intense temperature plunge (Fig. 3). Supporting this, the extremely low enamel δ¹⁸O value (−6.8‰) from one sample suggests that this individual drank meltwater from snow, which is substantially depleted in δ¹⁸O (Loftus et al., 2015).

4.2. The Last Glacial Maximum

Both Melikane and Sehonghong register occupation and provide associated environmental data ~25–24 ka, in the early stages of the LGM. SOM δ¹³C and phytolith records at Melikane suggest grassland cover with a strong C₃ signature most likely derived from sour Afroalpine grasses, along with a C₄ component similar to that of mid-MIS 3 (Stewart et al., 2016) (Fig. 3). Once again, however, the C₄ signal probably reflects drying and/or lower pCO₂ levels during the onset of the LGM rather than warmer temperatures since these levels also possess the lowest dicotyledon/Poaceae phytolith ratio of the entire sequence (Fig. 3). Those trees and shrubs still present were likely tightly restricted to deeper river corridors where there was sufficient shelter and surface water to support them. That riverine shrubs and bushes were still available in some density is supported by the presence in the immediately pre-LGM fauna at Sehonghong of common duiker (*Sylvicapra grimmia*) and steenbok (*Raphicerus campestris*), species that browse and/or require cover (Plug and Mitchell, 2008). Drying and coldness is further suggested by the charcoal assemblages from the ~24 ka levels at Melikane. Several mesic streamside taxa

previously present throughout MIS 4 and 3 disappear, leaving a much more limited range of frost-resistant species (*Erica drakensbergensis*, *Protea* sp., *Leucosidea sericea*, and *Olea europaea*). Further west, at Rose Cottage Cave in the Caledon Valley early MIS 2 is also marked by the presence of *Protea* sp., along with *Leucosidea sericea* and other heathland species (Wadley et al., 1992).

Comprising anthropogenic materials mixed with colluvial sediments and host bedrock attrition materials derived from roof fall debris, the sediments from the ~24 ka levels at Melikane also indicate colder and drier conditions, with material derived from landscape erosion as well as freeze-thaw processes (Stewart et al., 2012). Comparable conditions may be indicated at the Thabana Ntlenyana Site 1 sedimentary exposure high on the uKhahlamba-Drakensberg Escarpment, where a thick, organic-poor colluvial layer is dated to between ~27.4 and ~23.1 kya (Grab and Mills, 2011) (Fig. 3).

Conditions throughout the Maloti-Drakensberg deteriorated sharply around ~24 ka, with cold conditions widely recognized across southern Africa at this time (Scott et al., 2012). Geomorphological indications of periglacial conditions along the Escarpment are extensive (Grab and Knight, 2016), accompanied by good evidence of localized niche (cirque) glaciers. This includes debris ridges interpreted as glacial moraines ~3000 m a.s.l. on steep south-facing slopes of the Sekhokong and Tsatsa-La-Mangaung Ranges in easternmost Lesotho; two have dates of ~20.8 ka and ~19.6 ka (Fig. 3), overlapping with the height of the LGM (Mills et al., 2009a, 2009b). The snow accumulation necessary to sustain these glaciers implies that the LGM brought higher precipitation to the Maloti-Drakensberg, and perhaps the wider SRZ as well

(Mills et al., 2012). Snow cover patterns suggest that today the majority (63%) of snowfall events occur in a tightly restricted winter window (June-August), with most (80%) caused by westerly cold fronts and associated cut-off lows (Mulder and Grab, 2009). Together, these data may indicate that during the LGM more of the region's annual precipitation fell in winter than is currently the case (Mills et al., 2012).

Similar periglacial landforms, including blockstreams, protalus ramparts, and solifluction deposits are recorded at several locations further south in Nomansland down to ~1700 m a.s.l., though other claims (including the presence of rock glaciers; Lewis and Hanvey, 1993) cannot be substantiated (Mills et al., 2017). Arguments for temperature depressions of as much as $\geq 10^{\circ}\text{C}$ and snowlines as low as ~2100 m a.s.l. (e.g. at Mount Enterprise; Lewis and Illgner, 2001) are likewise improbable and at odds with mass balance modeling and niche glacier reconstructions for highland Lesotho (Mills et al., 2012). All the observed periglacial features could, in fact, have formed under mean temperature depressions of 6°C , an estimate consistent with other SRZ proxy records (Holmgren et al., 2003; Truc et al., 2013). At the archaeological site of Strathalan B (1800 m a.s.l), pollen spectra after ~24 ka resemble those of Afroalpine environments now found ≥ 2900 m a.s.l. (Opperman and Heydenrych, 1990).

4.3. Marine Isotope Stage 2 after the LGM

Fig. 4 presents Maloti-Drakensberg proxy archives post-dating the LGM, again accompanied by selected major archives from the broader SRZ. The post-LGM period was climatically complex. Heinrich Stadial 1 (HS 1, ~18–15 ka), a North

Atlantic ice–rafting meltwater event that brought pronounced cooling and variable aridity to the Northern Hemisphere, had diverse affects across Africa. Humidity signals in southern Africa are variable, with dry conditions registered north of the Vaal River, but more fluctuating conditions further south (Scott et al., 2012). In the Maloti-Drakensberg peat formation at the high-elevation Braamhoek wetland site ~16 ka implies relatively moist conditions, while pollen spectra there and at Mfabeni, a coastal peatland in KwaZulu-Natal, identify a major summer rainfall spike ~17–15 ka (Chevalier and Chase, 2015). In our region’s southwest, the Aliwal North pollen sequence likewise indicates high moisture availability ~16 ka (Coetzee, 1967; Scott et al., 2012) (Fig. 4). Pdf–based pollen analyses of sites across the SRZ (Chevalier and Chase, 2015; Truc et al., 2013) suggest that post–LGM southeastern Africa experienced sustained temperature depressions (hovering just above LGM minima) through most of HS1 (Fig.4). Locally relevant sequences include Braamhoek and Mahwaqa, another high-elevation wetland on an outlier mountain in KwaZulu-Natal’s Midlands. Basal pollen assemblages at both sites contain high frequencies of fynbos elements, including Ericaceae and Restionaceae, indicating cool and probably fairly humid conditions that may also imply greater winter rainfall (Finné et al., 2010; Neumann et al., 2014; Norström et al., 2014) (Fig. 4). Sharply reduced HS1 temperatures and higher (winter?) rainfall are also attested by glacial moraines in eastern Lesotho’s Leqooa and Tsatsa-La-Mangaung Ranges ~17.2, ~16.5, and ~15.5 ka (Mills et al., 2009b) (Fig. 4).

Two archaeological sites complement these observations. Mean $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of -2.5‰ and 0.6‰ from grazer teeth in the lower part of Layer DB at Rose Cottage Cave ($n = 7$) signal cool temperatures (grazer diets with ~70% C_4 plants) and high

moisture availability ~16 ka, becoming warmer and drier (diets of 90–100% C₄ plants) roughly a millennium later (Smith et al., 2002) (Fig. 4). Supporting this, DB's charcoals contain frost-tolerant taxa typical of high altitude or fynbos vegetation regimes today: *Cliffortia* spp., *Protea* spp., *Leucosidea sericea*, *Erica* spp., and *Rhamnus prinoides*. However, their overall diversity, and factor analysis results confirming high moisture availability (Fig. 4), caution against overstating the ecological impoverishment of the pre-Holocene landscape (Esterhuysen, 1996). Faunal data from Sehonghong in highland Lesotho concur: the site's taxonomically diverse terminal Pleistocene assemblage again includes smaller, browsing and canopy-loving antelope (Plug and Mitchell, 2008), while a SOM $\delta^{13}\text{C}$ value of –19.8‰ from Layer RBL-CLBRF (~15.7–15.2 ka; Pargeter et al., 2017) suggests enhanced presence of C₄ plants, and thus markedly warmer conditions, than at the LGM (Loftus et al., 2015) (Fig. 4).

That warming did not last. Less than half a millennium later the Antarctic Cold Reversal (ACR; 14.6–13 ka), another meltwater-triggered event, precipitated widespread and sustained cooling of the Southern Hemisphere, particularly south of 40°S (Pedro et al., 2016). Although Chase et al. (2011) argue that only latitudes south of 41°S were affected — something supported by recently synthesized SRZ pollen data suggesting that the warming trend initiated late in HS1 continued uninterrupted until ~13 ka (Scott et al., 2012) — evidence is mounting that in the Maloti-Drakensberg the ACR's timeframe was indeed characterized by cooling. In western Lesotho, a heavily depleted $\delta^{13}\text{C}$ value of –10.4‰ from a zebra tooth at Ntloana Tšoana in a context (032 of Layer BLOS) dated ~14.2–13.6 ka indicates a diet almost wholly composed of C₃ plants and, by extension, annual temperatures 6°C

lower than present (Smith et al., 2002) (Fig. 4). Assuming a plant-to-enamel apatite $\delta^{13}\text{C}$ increase of $\sim 13\text{‰}$ (Lee-Thorp et al., 1989), this is highly consistent with $\delta^{13}\text{C}$ analysis of sediments from the same portion of the Ntloana Tšoana sequence (NT21–25, mean of -23.5‰ , Roberts et al., 2013) (Fig. 4).

Similar conditions seem likely at Sehonghong where Layer RF ($\sim 14.9\text{--}13.7$ ka; Pargeter et al., 2017) returned a SOM $\delta^{13}\text{C}$ value of -21.8‰ (Loftus et al., 2015), while closer to the Escarpment's summit two of seven Sekhokong Range glacial moraines produced ages of ~ 14.7 and ~ 13.8 ka (Mills et al., 2009b) (Fig. 4). As with other such phases in the Maloti–Drakensberg, this implies that precipitation — at least in winter — was sufficient to sustain localized glacier formation. Confirming this, pollen, phytoliths, diatoms, and charcoals at Braamhoek all signal wetter conditions (probably linked to northward shift of the westerlies; Fitchett et al., 2017b) from ~ 14 ka, with a peak at ~ 13.6 or ~ 13.2 ka depending on the proxy, while persistent cold is indicated by fynbos pollen (Finné et al., 2010; Norström et al., 2014) (Fig. 4). A comparable picture emerges from pollen and a high silt/clay component in the basal stratum (Unit 9) of the 13 m-deep sedimentary sequence in the Tsoaing Valley, southwestern Lesotho, estimated to date to ~ 14.2 ka (Grab et al., 2005) (Fig. 4). The pollen sequence at nearby Aliwal North registers a moisture peak ~ 14.6 ka, as does that at Craigrossie in the northernmost Maloti–Drakensberg (Scott et al., 2012). Drier conditions follow at both sites ~ 14 ka before a major resurgence of moisture at Craigrossie until ~ 13 ka (Fig. 4). The consistency with which proxy archives register cool temperatures and high humidity between ~ 14.8 and ~ 13 ka may, we suspect, relate to the Maloti-Drakensberg's relatively high altitude since the ACR is already

strongly implicated in major episodes of glacier advance in other Southern Hemisphere mountain systems (e.g. Jomelli et al., 2017; Stansell et al., 2015).

4.4. The Pleistocene-Holocene transition and early Holocene

The period 13–11 ka is, in contrast, severely under-represented in the Maloti-Drakensberg's proxy archives. It corresponds to widespread evidence elsewhere in southern Africa's SRZ for dryness and variable cold, perhaps in response to another abrupt Northern Hemisphere forcing event, the Younger Dryas (YD, ~13–11.5 ka). Reanalysis of the Wonderkrater pollen sequence (Truc et al., 2013), pollen signals from other southern Savanna Biome sites (Scott et al., 2012), and the cessation of speleothem formation at Cold Air Cave (Holmgren et al., 2003) suggest annual temperature depressions to LGM levels and sharply reduced summer rainfall at this time. Increased drought-tolerant pollen elements, depleted $\delta^{13}\text{C}$ N-alkane values, and low magnetic susceptibility values suggesting reduced sedimentation at Braamhoek ~12.6–11.3 ka affirm this on the Maloti-Drakensberg's northern edge, with the persistence of fynbos pollen suggesting that conditions remained cold into the terminal Pleistocene (Norström et al., 2014) (Fig. 4). A similar if less pronounced YD drying trend is evident in the Aliwal North pollen sequence (Scott et al., 2012) (Fig. 4). Between the two in Nomansland, charcoals dominated by the arid-adapted karroid taxon *Euryops* sp. from Layer 4 at Ravenscraig, dated to the very end of the YD (~11.8–11.2 ka), also fit this picture (Tusenius 1989); an abundance of the moisture-loving Afromontane evergreen *Leucosidea sericea* in the underlying unit (Layer 5) does not contradict this since the associated date of ~12.4–11.3 ka (Opperman

1987:147) is probably erroneous given the disappearance of the associated Robberg lithic technocomplex from virtually all other southern African sites before 13 ka.

The end of the YD stadial conventionally marks the interface between the Pleistocene and Holocene. Across the Maloti–Drakensberg, as elsewhere in southern Africa, that transition saw major ecological changes on multiple timescales. On the coarsest analytical level, between ~11.5 and ~8.5 ka the region went from a cool, heavily grass-dominated open landscape to a warmer, more fragmented environment with a larger woody vegetation component. Thus far, however, only in the Caledon Valley do multiple proxies provide a finer-grained image. At the very start of the Holocene (~11.5–11.1 ka), this area experienced mild to considerable warming relative to earlier phases of the Last Glacial-Interglacial Transition. At Ha Makotoko in western Lesotho, SOM $\delta^{13}\text{C}$ values are now enriched by a greater input of C_4 plants relative to those dating to the ACR at nearby Ntloana Tšoana (Roberts et al., 2013; HM14 cf. NT21–25); a mean $\delta^{13}\text{C}$ value of -2.3‰ ($n = 3$) from grazer teeth at the same site implies a ~60% C_4 grassland with temperatures only slightly cooler than present (Smith et al., 2002) (Fig. 4).

However, this brief warming trend was not sustained. The same two proxies show a sudden cold reversal just after 11 ka, followed by high-magnitude oscillations between warmer and cooler temperatures that cannot be resolved closely in absolute time because of rapid sedimentation (at Ntloana Tšoana) combined with the effects of the early Holocene radiocarbon plateau (Mitchell and Arthur, 2012; Roberts et al., 2013). A sharply cold episode is, however, clearly indicated ~10.5 ka when faunal $\delta^{13}\text{C}$ signals at Ha Makotoko imply the prevalence of an almost wholly C_3 grassland

(Smith et al., 2002) (Fig. 3). The *Cliffortia/Protea*-dominated charcoal assemblage from Layer LB at Rose Cottage Cave (~11.1–10.6 ka) likewise suggests an overall prevalence of lower temperatures at this time. Relative to the YD, increasing frequencies of *Leucosidea sericea* in western Lesotho's charcoal record indicate that early Holocene conditions were nevertheless generally wetter (Esterhuysen and Mitchell, 1997). So, too, does the presence of *Xymalos monospora* (lemonwood) at Ntloana Tšoana ~10.4 ka (Deborah Costens, personal communication, 2011) since this rainfall-demanding species only occurs today east of the Escarpment (Palgrave 1983:175).

Data are more plentiful after 10 ka. In western Lesotho faunal and SOM $\delta^{13}\text{C}$ records emphasize a warming trend with conditions perhaps only slightly cooler than today (Roberts et al., 2013) and grazer diets consisting of ~80% C_4 vegetation (Smith et al., 2002) (Fig. 3). At a third site, Tloutle, increased humidity is signaled between ~9.9 and ~9.5 ka (Layer GS) by *Podocarpus* charcoals, calcareous sinter (spring) deposits, and increased numbers of vleirats (*Otomys irroratus*), rodents that prefer densely wooded and streamside habitats (Esterhuysen and Mitchell, 1997; Plug, 1993) (Fig. 4). Wetter conditions are also evident in the sediments at Rose Cottage Cave (Butzer, 1984), where an increasingly diverse range of large mammals provides broader evidence of increased environmental productivity (Plug and Engela, 1992). Specifically, the presence of kudu (*Tragelaphus strepsiceros*) and of both wildebeest taxa (*Connochaetes gnou*; *C. taurinus*) ~9.5–8.8 ka points to the expansion of sweeter C_4 grassland and a greater degree of canopy cover than later in the Holocene (Wadley, 2000a).

Other records from the western Maloti-Drakensberg also imply a generally ameliorating early Holocene climate. A decline in fynbos taxa and higher frequencies of grass species in the Braamhoek pollen sequence indicate increased temperatures and a shift toward a more strongly summer rainfall focus through the millennium 10.2–9.2 ka (Norström et al., 2014). Within this overall trend, however, tooth enamel $\delta^{13}\text{C}$ (Smith et al., 2002) and factor analysis of charcoals at Rose Cottage Cave (Esterhuysen et al., 1999; Esterhuysen and Smith, 2003) document a cooler episode ~9.5–9.3 ka (Layer Cm) that coincides closely with a short-lived drier pulse (Unit 6) in the Tsoaing phytolith and sedimentary record (Grab et al., 2005) (Fig. 4). Drier conditions ~9.5–8.8 ka in the Rose Cottage Cave charcoal assemblage (Esterhuysen, 1996) are mirrored ~9.2–8.6 ka by changes in the xylem structure of *Olea* and *Rhus* charcoals at Bonawe, Nomansland (Tusenius, 1989).

Where present, other palaeoenvironmental archives are too coarsely resolved to offer useful comparisons with those from the Caledon Valley. However, a general early Holocene increase in small antelope and ground game after 11 ka may reflect increasing humidity and warmer temperatures around Sehonghong, just as at Rose Cottage Cave (Plug and Mitchell, 2008). That peat formation only started after 9 ka at a site near Mokhotlong, northeast Lesotho, may nevertheless imply that conditions were not warm enough there before this (van Zinderen Bakker and Werger, 1974).

Charcoal assemblages in the Caledon Valley also usefully remind us how varied conditions may have been. Consistently high frequencies of *Protea* spp. throughout the early Holocene levels at Ha Makotoko and Ntloana Tšoana suggest, for example, that temperatures in this part of Lesotho remained lower than those near

Rose Cottage Cave just 40 km to the west. Conversely, 15 km to the southeast but closer to the Maloti Front Range, charcoals from the ~9.9–9.5 ka GS Layer at Tloutle include not only *Podocarpus*, as previously noted, but also relatively high frequencies of *Olea africana*, *Leucosidea sericea*, and *Myrica pilulifera*, taxa that imply local conditions sufficiently mesic to support patches of closed forest. Spatial variation between broadly contemporaneous charcoal assemblages is to be expected given micro-environmental variation in the availability of woody species and the impact of human fuel selection. Nevertheless, contrasts like these hint that climate change operated on finer spatiotemporal scales than is often appreciated. Subsequent multiproxy isotopic studies at the same sites — $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ analyses of grazer teeth at Rose Cottage, Tloutle, Ntloana Tšoana, and Ha Makotoko (Smith, 1997; Smith et al., 2002), plus higher resolution (≤ 2 cm) SOM $\delta^{13}\text{C}$ analyses of the latter two sequences (Roberts et al., 2013) — bear this out (Figs. 3 and 4).

4.5. The middle Holocene

Although most strongly evident in the North Atlantic, where Greenland ice cores date it to *c.* 8175–8025 BP, the 8.2 ka event was experienced planet-wide (Kobashi et al., 2007). Probably because of its higher elevation, its effects are most strongly evident in southern Africa in the Maloti-Drakensberg (Fitchett et al., 2017). Mean $\delta^{13}\text{C}$ values of -2.6‰ at Rose Cottage Cave (Layer Pt-lower) and -6.4‰ at Tloutle (Layer CSL-LR, basal units of Layer CSL-UP) imply a considerable shift toward C_3 grassland ~8.5–8.2 ka and ~8.2–7.8 ka respectively; both sites also show relatively low mean $\delta^{18}\text{O}$ values (2.3‰ and 0.29‰ respectively), suggesting cooler growing season temperatures (Smith et al., 2002). High frequencies of ice/snow-tolerant fragilarioid

diatoms from the lowest unit (MP5, 8.1–7.6 ka) of the Mafadi core, southern Africa’s highest wetland sample (3390 m a.s.l.), likewise indicate harshly cold, but (locally?) wet conditions, with temperatures too low to support terrestrial plants (Fitchett et al., 2017a) (Fig. 4).

Other proxies suggest that the colder episode encompassing the 8.2 ka event was significantly drier: Tloutle’s CSL-LR charcoal assemblage, for example, lacks *Podocarpus* sp. but includes the only appearance of *Euphorbia* in the site’s sequence; *Buddleia* and *Passerina montana* also increase in frequency (Esterhuysen and Mitchell, 1997). Oxygen isotope analysis of ostrich eggshell from the same layer confirms that CSL-LR was deposited during a much drier episode, while the accompanying fauna is less diverse than immediately above and below; bush-loving (common duiker) and water-dependent (common reedbuck, *Redunca arundinum*; mountain reedbuck, *R. fulvorufula*) species are notable absentees (Mitchell, 2000). Drier conditions are also evident after 8.5 ka in the Mahwaqa pollen record, at Braamhoek (Fig. 4), and in records from the central Free State (Neumann et al., 2014).

Across the region, temperatures subsequently remained relatively high until the start of the late Holocene Neoglacial ~3.5 ka, although moisture availability fluctuated. Tloutle’s charcoal record documents a return to more mesic conditions ~7.9–6.7 ka, with *Podocarpus* again present (Esterhuysen and Mitchell, 1997), while at Rose Cottage faunal $\delta^{13}\text{C}$ values from the upper component of Layer Pt (~6.9–6.6 ka), like those from the bulk of CSL-UP at Tloutle, are positive (means of 2.7‰ and 2.5‰ respectively; $n = 6$ and 6) (Smith et al., 2002) (Fig. 4). Although the single value from

649 Tloutle Layer CCL (-2.7% , $\sim 7.3\text{--}6.7$ ka) suggests that conditions did not remain
650 static, grassland near Rose Cottage was probably 100% C_4 in composition by $\sim 6.9\text{--}$
651 6.6 ka (Smith, 1997). Taking the period $\sim 8.5\text{--}6.6$ ka as a whole, Rose Cottage Cave's
652 charcoals imply a relatively moist, warm climate associated with more wooded
653 vegetation than present, including thermophilous species like *Grewia monticola* and
654 *Acacia karoo* not found in the area today (Esterhuysen, 1996; Wadley et al., 1992).
655 Microfaunal data agree that vegetation was denser than now (Avery, 1997), while
656 high tooth enamel $\delta^{18}\text{O}$ values from both Rose Cottage and Tloutle suggest increased
657 moisture availability, as well as warmer growing season temperatures (Smith, 1997;
658 Smith et al., 2002) (Fig. 4).

659
660 Vervet monkeys (*Chlorocebus pygerythrus*) took advantage of these changes to
661 expand into the Caledon Valley, while records of roan antelope (*Hippotragus* cf.
662 *equinus*) at Rose Cottage ($\sim 8.5\text{--}6.6$ ka; Plug and Engela, 1992) and the western
663 Lesotho site of Fateng Tsa Pholo (<7.7 ka; Shaw Badenhorst, personal
664 communication, 2009) suggest good quality grass and/or expanding savanna with a
665 climate warmer than today. Habitat changes associated with a warming climate
666 following the 8.2 ka event may also be implicated in a major faunal turnover event
667 registered at Sehonghong $\sim 8\text{--}6.5$ ka. Whereas several medium and large grazing
668 ungulates (springbok, bluebuck, and blesbok) that had been present throughout the
669 later Pleistocene disappear, others increase in frequency, among them the common
670 reedbuck, which is restricted to this part of the site's sequence; its heavy water
671 dependence signals a wetter mid-Holocene landscape (Plug and Mitchell, 2008).
672 Comparable changes (the loss of zebra, blesbok, *Hippotragus* sp., and an extinct
673 caprine (Brink, 1999) and a shift toward an ungulate fauna dominated by browsers or

674 mixed-feeders rather than grazers) are also evident above the Escarpment in
675 Nomansland, though less so below it where higher quality pasture was available
676 (Opperman, 1987).

677

678 Other archives also suggest a generally warmer, if not always moister, mid-Holocene
679 climate. Although the Braamhoek pollen core is too coarsely resolved to be helpful,
680 Unit 4 at Tsoaing records wetter conditions soon after the 8.2 ka event (~7.9–7.7 ka),
681 followed by a drier phase ~6.7–6.1 ka (Grab et al., 2005) (Fig. 4). A shift to low
682 (<0.5) Asteraceae/Poaceae ratios (cf. Fitchett and Bamford, 2017) at Mafadi marks
683 the presence of a summer rainfall regime not long after 7 ka, with warming continuing
684 into the Holocene Altithermal as suggested by reduced fragilarioid and increased
685 epiphytic diatom frequencies ~6.6–5.7 ka. Declining lake levels leading to the
686 formation of a locally marshy environment suggest that progressively drier conditions
687 accompanied this (Fitchett et al., 2017a). Extreme, sudden drying of Afroalpine
688 wetlands is also indicated further south at Sekhokong ~7.4 ka, perhaps during a period
689 of cooler temperatures, followed by progressively wetter conditions until ~6.7 ka and
690 then another particularly cold episode marked by a peak in *Fragilaria* spp. in the local
691 diatom record (Fitchett et al., 2016b). Fluctuations in temperature and moisture are
692 also indicated by the Mahwaqa core (Neumann et al., 2014), where the near
693 disappearance of Ericaceae and *Passerina* pollen ~6.8–6.4 ka implies warmer
694 conditions (Fig. 4). Increased Asteraceae counts suggest dryness before a short-lived
695 peak in temperature and moisture ~6.2 ka both there and at Cathedral Peak closer to
696 the Escarpment (Lodder, 2012). The drier conditions implied by high frequencies of
697 *Euryops* charcoals at Colwinton ~7.3–7.0 ka (Tusenius, 1989) are the only relevant
698 contribution from the southern Maloti-Drakensberg at this time.

699

700 The period between 6.0 ka and the onset of the Neoglacial is poorly known, though
701 around 6 ka itself general circulation models predict drier conditions than present
702 across the SRZ (Chevalier et al., 2017). Multiple proxy records from the eastern
703 Karoo, Free State, and KwaZulu-Natal confirm this. Reduced precipitation may have
704 resulted from shifts in the mean latitudinal position of the austral westerlies due to
705 changes in winter sea-ice extent around Antarctica, given that a significant amount of
706 the rain falling over southern Africa's interior results from systems formed when that
707 westerly storm track and tropical easterlies interact (Chevalier and Chase, 2015).

708

709 Archaeology contributes little to understanding this, with several sites lacking
710 evidence of occupation. Leliehoek, a small rockshelter near Rose Cottage Cave, is the
711 principal exception. Its charcoal and faunal records (a proliferation of *Cliffortia*
712 *linearifolia* scrub, increased numbers of vleirats, and the presence of common duiker)
713 suggest that moist conditions prevailed ~5.9–4.9 ka (Esterhuysen et al., 1994). This
714 fits well with the pulse of silt/clay accretion and highest rate of Holocene organic
715 accumulation in Unit 2 at Tsoaing ~5.3–4.9 ka (Fig. 4), which in turn conforms to a
716 regional trend evident further west and south in the central Free State and eastern
717 Karoo (Grab et al., 2005; Scott, 1993; Scott and Nyakale, 2002). Conversely, the
718 Sekhokong record in highland Lesotho shows the driest phase of its entire sequence
719 down to 3.6 ka (Fitchett et al., 2016b), while drier conditions are also evident after 5.6
720 ka at Cathedral Peak (Lodder, 2012) and Mahwaqa (Neumann et al., 2014) (Fig. 4).
721 The lack of conformity between these records and others for the period ~5–3 ka that
722 do not suggest greater aridity in the SRZ may be due to local factors specific to their
723 high altitude location (Chevalier and Chase, 2015).

4.6. The Neoglacial and after

The later Holocene was punctuated by several cold reversals of variable duration and intensity. The longest was the Neoglacial, a period of widespread cooling and humidity registered across much of southern Africa, including the Maloti-Drakensberg, between ~3.5 and 2.0 ka (Nash and Meadows, 2012). The Caledon Valley provides little proxy evidence for most of the later Holocene, but does signal relatively warm conditions either side of the Neoglacial in the form of high frequencies of *Olea africana* charcoals and low (or absent) frequencies of *Leucosidea sericea* and *Erica* spp. at Twyfelpoort and Mauermanshoek ~3.9–3.0 ka (Backwell et al., 1996; Wadley, 2001). Faunal $\delta^{13}\text{C}$ values imply a 100% C_4 grassland in the environs of Rose Cottage Cave ~2.3–2.1 ka (Smith et al., 2002; Wadley, 2000b) (Fig. 4).

A more robust Neoglacial record comes from highland Lesotho. Taken as a whole, the period ~3.4–1.2 ka at Sekhokong saw continuous fluctuations in pollen and diatom assemblages, but was persistently cooler than before. Wet phases are indicated ~3.26–3.19, ~3.05, and ~2.69–1.47 ka, with at least one drier episode before the last of these. The third, and most prolonged, wet event may have included more regular snowfalls (Fitchett et al., 2016b). A return to more humid conditions is also signaled by increases in sedge, *Aponogeton*, and grass pollen compared to that of the Asteraceae at Mahwaqa after 3.5 ka (Neumann et al., 2014). The Mafadi profile, on the other hand, shows little variation between 5.6 and 1.1 ka, although greater cold may have decreased frequencies of Cyperaceae and Asteraceae, something supported by

749 increased numbers of ice/snow-tolerant *Fragilaria* and Apiaceae diatoms (Fitchett et
750 al., 2017a) (Fig. 4).

751

752 Finer detail comes from the archaeological site of Likoaeng near Sehonghong (1725
753 m a.s.l). Here, slackwater sediments deposited by the Senqu River preserve multiple
754 human occupation residues ~3.7–1.7 ka with a further brief episode ~1.2 ka.
755 Phytoliths and SOM $\delta^{13}\text{C}$ values provide most of the evidence, amplified by charcoal
756 and faunal data (Mitchell et al., 2011; Parker et al., 2011). Before 3.0 ka the area
757 supported grassland in which C_4 and C_3 taxa were initially present in roughly equal
758 proportions. Charcoals hint that conditions were cooler (*Erica* spp.) and drier (*Acacia*
759 spp.) than present. Climate cooled markedly after 3.0 ka, with $\delta^{13}\text{C}$ SOM values
760 indicating a shift to C_3 pooid grassland at a scale best understood as implying a
761 significant (≤ 400 m) downslope shift in vegetation equivalent to a temperature
762 decrease of $\sim 2.5^\circ\text{C}$ (Fig. 4). Phytolith and charcoal assemblages document the
763 increased presence of cold-tolerant taxa like *Erica* spp., *Euryops* spp., *Protea* spp.,
764 and *Leucosidea sericea*. Increased numbers of smallmouth yellowfish (*Labeobarbus*
765 *aeneus*) among the fish people caught at the site may have been favored by colder
766 river temperatures, intensified perhaps by greater snowfall (and melt) (Mitchell et al.,
767 2011), while high frequencies of pooid phytoliths suggest that cooler temperatures
768 were accompanied by substantially increased moisture availability (Parker et al.,
769 2011) (Fig. 4). Renewed peat formation ~3.3–2.3 ka at Sani Top near the
770 Escarpment's summit (Marker, 1994), palaeosol and overbank deposit formation at
771 Kilchurn (~3.2–2.3 ka), and gully erosion at Tiffindell (~2.8–2.7 ka) (both in
772 Nomansland; Lewis, 2005) reinforce this. Below the Escarpment, analysis of charcoal
773 xylem vessel morphology at Bonawe agrees that conditions were less dry in the

southeastern Maloti-Drakensberg ~3.3–2.8 ka than previously, but, consistent with the broader regional picture, drier again ~2.3–2.0 ka (Tusenius, 1989). Charcoals from Mhlwazini Cave and Collingham Shelter in the northern half of KwaZulu-Natal’s uKhahlamba-Drakensberg likewise identify decreased rainfall after 2.3 ka, with precipitation 40–80% higher than now before this (February, 1994).

Following the Neoglacial, the Maloti-Drakensberg experienced some of the warmest temperatures of the Holocene. At Likoaeng, a drier, warmer climate is evidenced around 2.1 ka by higher frequencies of C₄ phytoliths and SOM $\delta^{13}\text{C}$ values (–19.75‰ to –16.54‰) consistent with a mixed C₃/C₄ grassland (Fig. 4). *Acacia* charcoals support this and cool-specific plant taxa are absent (Mitchell et al., 2011; Parker et al., 2011). Multiple records from across the SRZ concur that a widespread arid episode occurred around 2000 years ago (Scott et al., 2012; Stager et al., 2013). Conditions became even warmer and drier between 1.6 and 1.0 ka (Parker et al., 2011). SOM $\delta^{13}\text{C}$ values are now at their highest in the Likoaeng sequence, indicating an upper Senqu Valley heavily dominated by C₄ taxa, with frequencies of arid-adapted chloridoid C₄ phytoliths also peaking and the lowest ever phytolith counts from C₃ pooid grasses (Fig. 4). Below the Escarpment, analysis of charcoal vessel diameter at Collingham Shelter agrees that conditions were relatively dry ~1.3–1.0 ka (February, 1994).

These drier, warmer conditions likely represent the local impact of another of the later Holocene’s widely recognized climatic episodes, the Medieval Climatic Anomaly (MCA, ~1.4–0.65 ka). Both it and the ensuing Little Ice Age (LIA, ~0.65–0.15 ka) register in several other regional archives, including peat deposits at Tlaeeng Pass

north of Sehonghong (Hanvey and Marker, 1992), the Craigrossie pollen sequence in the easternmost Free State (Scott, 1989), and the SOM $\delta^{13}\text{C}$ values obtained from open-air sampling locations near Sehonghong itself (Julia Lee-Thorp, personal communication, 2016). In lowland Lesotho a single $\delta^{13}\text{C}$ value (-9.6‰) from Tloutle (for which the associated $\delta^{18}\text{O}$ value is also very low, -0.83‰) documents a significantly colder episode that may equate to the start of the LIA (Smith, 1997) given an associated radiocarbon date of 715 ± 65 BP (OxA-4069; Mitchell, 1993). However, $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values from grazers at Rose Cottage Cave, along with micromammal and charcoal data, suggest that climate was still warmer than present $\sim 0.7\text{--}0.5$ ka with an overwhelmingly C_4 grassland (Avery, 1997; Smith et al., 2002; Wadley, 1997) (Fig. 4). Subsequently, however, reduced charcoal diversity and the presence of cold-tolerant *Protea* spp. (Wadley, 1997), along with lower *Olea* frequencies at Mauermanshoek and Twyfelpoort (plus higher counts of *Cliffortia*, *Erica*, *Rhamnus*, and *Leucosidea* at the latter; Backwell et al., 1996; Wadley, 2001), do signal cooler conditions $\sim 0.5\text{--}0.2$ ka.

Other observations are more difficult to correlate with either the MCA or the LIA because of low temporal resolution and because conditions did not remain static within either episode (Hannaford and Nash, 2016; Stager et al., 2013:118). For example, identifications of a hippotragine antelope and vervet monkey at Lithakong, between the Front and Central Ranges of the Maloti may imply warmer conditions (with more trees along river valleys) at some point during the last 1000 years, but cannot be dated more precisely than this (Brink, 2012). Highland cores document cooler and/or wetter conditions in the early part of this period, followed by drier ones thereafter, but chronologies are again too poor to identify either episode with part or

all of the MCA or LIA (Neumann et al., 2014; Fitchett et al., 2016b, 2017a). Similarly, while the cooler/moister conditions implied by charcoal data from Ravenscraig ~0.5–0.3 ka (Tusenius, 1989) and Mhlwazini ~0.6–0.2 ka (February, 1994) can be linked with the LIA, the wetter conditions indicated at Colwinton ~2.0–1.7 and ~0.9–0.7 ka (Tusenius, 1989) are more difficult to assess. Neither these episodes, nor evidence of pronival rampart formation on the slopes of Thabana Ntlenyana some time after 1.7 ka, which also implies cool, moist conditions (Grab and Mills, 2011), are yet easily reconcilable with other regional proxies.

5. Regional occupation trends in archaeological radiocarbon data

Interest in how changing palaeoclimatic conditions affected people living in the Maloti-Drakensberg has been a longstanding research theme, encouraging archaeologists to collect and analyze palaeoenvironmental proxies from their excavations and facilitating the development of ideas about seasonal mobility over the landscape and its social consequences that remain relevant today (e.g. Carter 1970, 1976, 1978). One strategy for investigating these issues uses radiocarbon dates to establish when people were present in a particular locality, something already explored preliminarily by Wadley (1995) and Mitchell (2009). At the time of writing, a total of 344 radiocarbon dates are available from the Maloti-Drakensberg region. For this paper, we have opted to omit 19 dates from this dataset whose 2σ calibrated range is out of stratigraphic sequence or that returned ages of “modern” or “infinite”. While still small compared to the sample sizes recommended by Williams (2012), we use the remaining 325 dates to identify patterns in human occupation that may, provisionally, relate to the climatic and environmental shifts discussed above. We

acknowledge that variation in the materials sampled and the pretreatment protocols applied to remove contaminants affect the accuracy and precision of individual dates, that techniques have evolved over the past half-century, and that not all sites have been dated with equal thoroughness.

To examine region-wide patterning, Fig. 5 presents a summed probability distribution (SPD) of our complete quality-controlled dataset of 325 dates set against a more conservative histogram of dates binned into 1000-year intervals. The major climatic phases discussed above are also shown. As can be seen, low radiocarbon date frequencies characterize the Late Pleistocene portion of the last 50 kyr, with clear spikes at the onset of the LGM (~25–24 ka) and during the ACR (~14 ka). Each is followed by a trough during which few or no dates have been generated for the region; these correspond to the height of the LGM/early HS1 (~23–17 ka) and the YD (~13–11.5 ka), respectively (Fig. 5). As discussed below, we strongly suspect that the intense cold registered in myriad proxy datasets for these stadial largely discouraged hunter-gatherer settlement of the Maloti-Drakensberg region as a whole. Whether the correspondence between earlier Heinrich Stadials (e.g. HS2–5) and several earlier troughs in the SPD (Fig. 5) hints at similar dynamics recurring throughout the Late Pleistocene must remain speculative until more dates are obtained for these time-depths.

Both the frequencies of dates and their temporal variability increase in the Early Holocene, when regional temperatures and moisture availability rose steeply, with histogram and SPD peaks in the millennia 11–10 ka and 10–9 ka, respectively. Frequencies then decline into the Middle Holocene, with the histogram reducing

874 gradually but the SPD exhibiting sharp fluctuations. Peaks in the latter are evident
875 ~8.5–8.3, ~8.1–7.5, ~7.3–6.5 and ~6–4.9 ka, phases during which the data
876 summarized above suggest humidity was high, with intervening troughs ~9.5–8.6,
877 ~8.3–8.1, ~7.5–7.3, ~6.5–6 and ~4.9–4 ka, which correspond to drier periods (Fig. 5).
878 The downward Middle Holocene trends of both plots bottom out ~5–4 ka, coincident
879 with the late Altithermal when most Maloti-Drakensberg records signal heightened
880 aridity (except in the eastern Free State – see below). The Late Holocene sees
881 radiocarbon date frequencies rise dramatically. Both records indicate the increase
882 began ~4 ka then accelerated sharply in the 2nd millennium BC, shortly after the onset
883 of the cool, wet Neoglacial. Relative to their previous high in the millennium 11–10
884 ka, histogram values more than double for the 2nd and 1st millennia BC, and peak at
885 more than triple for the 2nd millennium AD. The SPD provides sharper resolution,
886 with Late Holocene peaks ~3–2.7, ~2.3–1.5, ~1.3–0.9 and 0.8–0.3 ka (the latter is the
887 record’s highest) separated by short-lived but substantial declines (Fig. 5).

888

889 Because we are also interested in whether the radiocarbon data can inform variability
890 in human settlement preferences for different parts of the Maloti-Drakensberg, Fig. 6
891 breaks the data down by sub-region. Due to the much smaller sample sizes, only
892 millennium-binned histograms are used rather than SPDs. Stark differences in the
893 data are apparent between Maloti-Drakensberg’s sub-regions. Perhaps most striking is
894 the longevity, continuity and intensity of human occupation in highland Lesotho
895 relative to other sub-regions. We stress that this is not an artifact of research bias
896 towards this sub-region, since others have received comparable archaeological
897 attention, particularly the northern KwaZulu-Natal Drakensberg and lowland Lesotho.
898 The Maloti-Drakensberg’s earliest radiocarbon dates come from the highland site of

Melikane, where a series mid-MIS 3-aged dates were obtained ranging ~48–36 ka (constrained by rigorous acid-base-wet oxidation stepped combustion pretreatment and Bayesian modeling to ~43–38 ka (Stewart et al. 2012:50)). The only other sub-region whose radiocarbon data registers human occupation in this timeframe is lowland Lesotho, where several dates concentrate in a single millennium ~44–43 ka. After ~36 ka other sub-regions come online, first the eastern Free State (Rose Cottage Cave) followed two millennia later (~34 ka) by Nomansland (Strathalan B) (Fig. 6). Meanwhile, occupation continued apace in the highland Lesotho, where Sehonghong has a series of tightly constrained late MIS 3 levels dated ~35–29 ka (Loftus et al. 2015; Pargeter et al. 2017).

After a strong highland Lesotho pulse centered on ~24 ka, this and all other sub-regions witnessed very little human activity across the LGM. However, dates with 2σ calibrated brackets of ~19.3–18.6 ka may imply fleeting visits to Sehonghong and Rose Cottage Cave. A further pair of dates ~22.2–20.8 ka may signal earlier, but equally ephemeral, occupations at these sites (Pargeter et al., 2017). Sustained re-occupation of the Maloti-Drakensberg nevertheless only began after 16.5 ka (Rose Cottage Cave) and 15.7 ka (Sehonghong). Although the situation at Rose Cottage requires clarification since Layer DB is currently dated to ~16.5–15.6 *and* ~15.4–14.3 ka, at Sehonghong two distinct occupation pulses are evident. The first (~16–15 ka) coincides with markedly warmer temperatures relative to pre-LGM conditions, while the second (~14.8–13.7 ka) accords with the earlier part of the ACR and is also evident in lowland Lesotho, though both Nomansland and KwaZulu-Natal's uKhahlamba-Drakensberg were unoccupied throughout this period.

The return of globally colder conditions during the YD posed further challenges to Maloti-Drakensberg hunter-gatherers. Except for a stratigraphically thin context at Sehonghong (Layer BARF), poorly constrained by a single conventional date to ~13.4–12.6 ka, neither highland Lesotho nor the eastern Free State has yet delivered evidence of occupation at this time (Fig. 6). Isolated, conventional dates to the contrary from Nomansland are likewise not compelling without confirmation using state-of-the-art pre-treatment protocols and AMS; a determination calibrated to ~12.4–11.3 ka for a Robberg assemblage at Ravenscraig (Opperman, 1987) is, we suspect, erroneously young given the tightness with which the Robberg is now dated in Lesotho (cf. Mitchell and Arthur, 2014:227). There, the Phase 7 occupation at Ha Makotoko nevertheless clearly documents human presence in the Metolong area just before the YD ended (Mitchell and Arthur, 2014).

An upturn in occupation is then quickly evident in other sequences, with both Rose Cottage (eastern Free State) and Sehonghong (highland Lesotho) documenting multiple visits ~11.2–10.2 ka and ~11.5–10.5 ka, respectively. Only in lowland Lesotho, however, is significant occupation evident in this time frame *and* the following millennium 10.5–9.5 ka (Fig. 6). Thereafter, several more sites return dates, with occupation signaled from both sides of the Caledon River ~9.6–8.2 ka, as well as in Nomansland (Bonawe, Strathalan A, and Te Vrede, but probably also undated assemblages from Ravenscraig Layers 3 and 4 and Colwinton Layer 6; Opperman 1987, 1996). Brief(?) occupations ~8.5–8.3 ka at Bellevue and Good Hope represent the first definite sign of human presence along the uKhahlamba-Drakensberg Escarpment since before the LGM (Cable et al., 1980; Carter, 1978).

Major changes occurred either side of the 8.2 ka event. In the Metolong area of lowland Lesotho extensive sequences of early Holocene occupation were overlain by thick aeolian and fluvial sediments soon after ~9.5–9.1 ka. Dates from Rose Cottage Cave's Pt Layer also imply an occupation hiatus after ~8.5–8.2 ka, and KwaZulu-Natal's southern uKhahlamba-Drakensberg appears to have been largely abandoned for several millennia. Occupation during the 8.2 ka event itself is most evident in lowland Lesotho, particularly at Tloutle, but Sehonghong in Lesotho's highlands was also intensively used then and slightly later, followed by further activity at Rose Cottage Cave (eastern Free State) ~7.7–7.6 ka (Fig. 6). The generally warmer, moister conditions that followed the 8.2 ka event saw further occupation in these three areas and probably also in Nomansland, although several relevant contexts there remain undated (Opperman, 1987).

Perhaps significantly, human presence has yet to be registered in the Caledon Valley during the drier climatic phase registered in the Tsoaing sediment profile ~6.7–6.1 ka, although people were present at Rose Cottage Cave just before this and at Tloutle both before and after. Elsewhere, only Sehonghong in highland Lesotho shows activity at this time (Layer GWA, ~6.9–6.5 ka), which the Mahwaqa core suggests was a warmer episode (Neumann et al., 2014). Thereafter, generally drier mid-Holocene conditions ~6.0–3.5 ka (cf. Chevalier et al., 2017) are associated with only minimal evidence of people outside the eastern Free State (at Liphofung on the eastern edge of the Caledon Valley and in the basal occupation at Diamond 1, the oldest signal yet known from the northern KwaZulu-Natal uKhahlamba-Drakensberg). Across the region, the mid-Holocene Altithermal may thus have been generally antithetical to human presence (Fig. 6), though the period clearly

experienced considerably more climatic variability than is sometimes appreciated, particularly with regard to moisture availability.

In sharp contrast, the considerably cooler and wetter Neoglacial is strongly represented in archaeological records from most sub-regions (Fig. 6). Lowland Lesotho is the one exception, but the presence of people in the ecologically similar eastern Free State suggests that this is an artifact of site taphonomy and/or archaeological survey. Of relevance here a river-cut section along the Phuthiatsana River documents deposition of an astonishing 2.6 m of sediment between 4.8 and 3.4 ka, implying substantial erosion and sedimentation on a scale sufficient to remove/cover up open-air sites and affect access to some rockshelters (Charles Arthur, personal communication, 2016). Archaeologists' preference for excavating at the very largest rockshelters at the expense of smaller ones creates a further bias (Mitchell, 1994).

Heavy occupation is also evident immediately after the Neoglacial ~2.0–1.6 ka and the Medieval Climatic Anomaly (~1.4–0.65 ka) appears to have had little effect, notwithstanding the drier conditions signaled at Collingham and in the Senqu Valley. Occupation is evident in KwaZulu-Natal's southern uKhahlamba-Drakensberg, Nomansland, highland Lesotho, lowland Lesotho (but by rock paintings, not excavated assemblages; Bonneau et al., 2017), and the eastern Free State (Fig. 6). The northern uKhahlamba-Drakensberg appears, however, to have been completely abandoned; Mazel (2009) suggests that this was because local hunter-gatherers moved downslope toward farming communities in the central Thukela Basin in order to access metal, grain, and other desirables. The area's subsequent reoccupation may

reflect disruption — and expansion into higher-lying grasslands — of those same agropastoralists as the Little Ice Age began ~0.65 ka (Mazel, 2009; Whitelaw, 2015). Hunter-gatherers relocated into the foothills of the Escarpment, and then into the Escarpment itself, presumably to remove themselves from encroaching farmers.

Discerning chronological detail during the last several hundred years using radiocarbon dates is handicapped by the shortness of the time-scale involved and the well-known wiggles of the calibration curve over recent centuries. At a broad level, however, the return to generally colder, wetter conditions is associated with increased evidence of occupation in all sub-regions bar, perhaps, KwaZulu-Natal's southern uKhahlamba-Drakensberg, where neither of the two excavated sites (Bellevue and Good Hope) preserves well-defined occupations.

6. Human-environment dynamics in the Maloti-Drakensberg

Drawing together the palaeoclimatic and occupational data reviewed above, we now examine the ecological contexts of human exploitation of the Maloti-Drakensberg and the cultural responses to those contexts manifested in the archaeological record in order to explore macro-regional trends in human-environment dynamics. We begin by asking which climatic conditions coincided with a definite human presence in the Maloti-Drakensberg region? While our radiocarbon dataset is not yet large enough to make statistically robust inferences regarding palaeodemographic trends and our archaeological data remain skewed towards rockshelters, we feel confident that clear patterns are emerging as to the climatic conditions in which human settlement was consistent enough to produce spikes in radiocarbon date frequencies.

1024

1025 Elsewhere we have hypothesized that phases of intensified human presence in the
1026 Maloti-Drakensberg should correlate with regional proxy evidence indicative of
1027 warming and/or aridity (Mitchell, 1990; Stewart et al., 2012). The logical basis for
1028 this is the twin observation that this upland region is southern Africa's coldest zone,
1029 with present-day Effective Temperatures (ET) falling below 14°C almost everywhere
1030 (Stewart and Mitchell, in press), and that it possesses some of the subcontinent's most
1031 predictable and abundant freshwater and associated resources due to a combination of
1032 high rainfall and low evapotranspiration. It follows that the Maloti-Drakensberg might
1033 only have become attractive or viable for human settlement when temperatures in the
1034 wider region were relatively high and/or freshwater resources, and the plants and
1035 animals dependent on them, limited. The Maloti-Drakensberg's 'core' — Lesotho's
1036 highlands (and, to a lesser extent, lowlands) — is especially important in this regard
1037 due to the consistently productive and water-rich upper Orange-Senqu catchment
1038 (Stewart et al., 2016) (Fig. 7). In the data reviewed above we can identify several
1039 episodes of intensified highland occupation that plausibly correlate with such phases.

1040

1041 Specifically, the intensive mid-MIS 3 pulse at Melikane, dated to ~43–38 ka, overlaps
1042 neatly with well-dated evidence for widespread aridity-linked colluviation in the
1043 adjacent KwaZulu-Natal Midlands. Humans were again living at Melikane when
1044 phytolith and charcoal evidence suggest a massive reduction of woodland during
1045 the early LGM at ~24 ka, an occupation pulse also registered at nearby Sehonghong.
1046 Moving further forward in time, the 8.2 event, which multiple proxy archives suggest
1047 brought sudden and severe drying to the wider region, is well-represented in terms of
1048 human occupation in both highland (Sehonghong) and lowland (Tloutle) Lesotho.

1049 Slightly later, the arid and warm phase registered at Tsoaing and Mahwaqa ~7–6 ka
1050 again sees Sehonghong intensively inhabited. Strikingly, moreover, each of these
1051 pulses in Lesotho correlates with occupational hiatuses at various lower altitude sites,
1052 including several in other sub-regions of the Maloti-Drakensberg itself, raising the
1053 prospect that the mountain core served a refugium-like function for people at such
1054 times.

1055

1056 Yet the data also show that the climatic conditions corresponding to what tentatively
1057 seem to be the most intensive occupational pulses were more diverse than those we
1058 have previously predicted. At times, for example, regional climatic instability itself
1059 may have played a more central role. We have pointed out that Melikane's mid-MIS 3
1060 (~43–38 ka) pulse is coeval with the deposition of thick colluvial mantles in
1061 KwaZulu-Natal. However, the latter are interspersed with multiple palaeosols
1062 suggesting reversions to more humid conditions that allowed greater vegetation cover
1063 and pedogenesis (Clarke et al., 2003). This is echoed at Melikane itself, where
1064 climatic volatility is strongly suggested by diverse geogenic inputs, including
1065 colluvial gravels and rockfalls, into these deposits and signs of intense chemical
1066 weathering (Stewart et al., 2012). Similarly, a wide range of proxy data from sites in
1067 the Caledon Valley suggest pronounced climatic flux during the early Holocene
1068 (~11.5–9.5 ka), with large and rapid swings of temperature and available moisture
1069 (Esterhuysen and Smith, 2003; Roberts et al., 2013; Smith et al., 2002). Nevertheless,
1070 this period saw dense occupations at numerous sites along drainages of the Maloti
1071 Front Range, attesting to population stability in this area when the climate was
1072 anything but. The highlands also witnessed human presence in the first half of this
1073 interval, with Sehonghong occupied between ~11.5 and ~10.5 ka (Mitchell, 1996a). In

fact, these volatile early Holocene millennia yield the highest frequencies of radiocarbon dates of any time before the Neoglacial (Fig. 5).

Conceivably, it was the pace and magnitude of climatic and associated ecological changes rather than any specific such context that underwrote regional settlement decisions during the early Holocene and mid-MIS 3. What might have attracted foragers into the Maloti-Drakensberg's mountain core at such times? Though the region's pronounced seasonality and ruggedness make it logistically taxing, its wealth of resources is considerable. Broadly, these include greater resource diversity per unit area of terrain and reliable supplies of firewood, plant foods and medicines, opportunities for summer hunting windfalls, abundant rockshelters, high quality toolstone, rich riverine faunas and, of course, freshwater (Fig. 7). The latter — supplied by orographic rainfall and snowmelt and regulated by high altitude bog systems — may have been especially vital when recurrent shifts in precipitation and/or evapotranspiration resulted in unpredictable aquifer conditions further downslope (Stewart et al., 2016). The proxy data reviewed above suggests remarkable consistency in water availability as evidenced by the persistence of moisture-loving vegetation throughout virtually the entire period under investigation. This may be partly responsible for the consistency with which highland Lesotho registers human occupation relative to other sub-regions (Fig. 6).

More surprisingly, several of the most pronounced radiocarbon peaks are associated with regional conditions that an increasingly robust body of proxy data suggests were cool to cold. Indeed, three of the four dry pulses with human occupation just listed — mid-MIS 3 (after ~42 ka), the early LGM (~24 ka) and the 8.2 ka event — were also

characterized by negative temperature excursions. To this we can add the ACR and Neoglacial, both of which were cool (though humid) and associated with radiocarbon peaks (Fig. 5). This seems counterintuitive; why would foragers have enhanced their use of higher altitudes at times when lower ambient temperatures would have increased metabolic requirements, exposure to cold stress, the extent and duration of snow cover, and the downslope distribution of nutrient-poor Afroalpine vegetation belts? The answer, we believe, is that this pattern does not relate to an intensified upland presence, but rather to changes in how pre-existing populations organized themselves in relation to the highland landscape and its resources. Specifically, declines in terrestrial productivity that would have accompanied expansions of Afroalpine grass and fynbos elements likely limited the availability of plant food staples, narrowed windows of reliable hunting, and raised overall terrestrial resource search costs. Groups already present in the area would have responded, we suggest, by assuming even more valley-centered settlement-subsistence foci (Stewart and Mitchell, in press). *Inter alia*, this entailed intensified use of large rockshelters, which likely produced the cold-phase radiocarbon spikes that we have detected, and greater procurement of fish, certain taxa of which, as mentioned above, flourish in colder waters (Mitchell et al., 2011). The latter represents a conspicuous dietary shift to which we return below.

Are there specific climatic phases when we can definitively say that the Maloti-Drakensberg region was abandoned? Until more dates are available, we cannot be certain that gaps in our sub-regional radiocarbon chronologies stem from phases of low population density in, or abandonment of, a given area. In the past, such assumptions for gaps across the Pleistocene-Holocene transition and the Neoglacial

were later demonstrated to be false, with the discovery and dating of new sites showing that both phases are in fact well represented (Mitchell, 1996a; Mitchell et al., 2011). The lesson here is that extreme caution must be exercised to avoid over-extrapolating demographic patterning from radiocarbon ages until more dates become available. Given the time-depth across which humans have exhibited a high degree of adaptive resilience and behavioral plasticity (Wadley 2015), we suspect that permanent human occupation of certain parts of the Maloti-Drakensberg (e.g. highland Lesotho) was the norm for much of the later Pleistocene. Two chronological gaps do, however, seem likely to eventually become demonstrable population lulls, namely the apexes of the LGM and the YD (Fig. 5). Above we reviewed a wealth of proxy data that suggest that low terrestrial evapotranspiration during rainfall-poor phases like the LGM and YD offset losses in available moisture. Nevertheless, severe temperature depressions themselves, though rare, probably pushed resource availabilities (e.g. plant foods, game, fuel), socioeconomic scheduling (e.g. subsistence organization, social network maintenance) and/or technological capacities (e.g. clothing, shelter, exchange items) beyond adaptive thresholds, resulting in genuine regional demographic collapses.

Dispersal is, of course, only one potential response available to hunter-gatherers faced with resource (or other) stress. More often, Maloti-Drakensberg foragers were probably able to adjust their diets, technologies, land-use and social practices on a local basis. Here we ask which of those responses can most clearly help us elucidate human-environment dynamics in the Maloti-Drakensberg? Perhaps the most striking evidence for subsistence reorientation is the recurrent increase in fish bone frequencies in highland Lesotho sites during cold phases (the early LGM- and ACR-

aged levels of Sehonghong, and the Neoglacial levels at Likoaeng) (Stewart and Mitchell, in press). As mentioned above, this expansion of dietary breadth was likely a response to the increased costs of finding game as reduced temperatures and enhanced winter rain simultaneously lowered Afroalpine vegetation belts, increased snowfall and enhanced riverine productivity. Though less pronounced, other instances of dietary intensification coincide with some of the abrupt environmental changes reviewed here, including increased grindstone frequencies suggestive of heavy grass seed processing in the late Pleistocene levels of Ntloana Tšoana (lowland Lesotho), a spike in freshwater mollusc shells associated with the 8.2 event at Tloutle (lowland Lesotho), and an abundance of frog remains in a MCA-aged layer at Colwinton (Nomansland) (Mitchell, 2000; Opperman, 1987; Plug, 1993).

While in the Maloti-Drakensberg, as elsewhere in southern Africa, major transitions between lithic techno-complexes match climate change events only weakly and sometimes not at all, smaller-scale technological adjustments have often been found to tightly track such shifts. One consistent phenomenon is an uptick in bipolar core reduction during phases of reduced terrestrial productivity caused by aridity and/or cooling, including mid-MIS 3 (<42 ka) at Melikane, the early LGM both there and at Sehonghong, the ACR at Sehonghong, the ~9.5–9.3 ka dry/cool interval at Ha Makotoko and Tloutle, and the 8.2 event at Tloutle (Mitchell 2000; Pargeter et al., 2017; Stewart et al., 2012). The independence of the enhanced use of bipolar techniques from any change in the dominance of fine-grained cherts strongly suggests that these shifts reflect raw material economizing behaviors, perhaps as efforts to alleviate scheduling conflicts and time-stress (Mitchell 2000). Indeed, technological reorganizations hint at broader changes in forager land-use and settlement patterns.

We have already suggested that valley-focused settlement arrangements during the early LGM, ACR, and Neoglacial likely encouraged more intensive rockshelter occupation that resulted in our cold-phase radiocarbon peaks. Supporting evidence comes from consistently high artifact densities in the relevant levels of Sehonghong and Likoaeng (Pargeter et al., 2017; Stewart and Mitchell, in press), as well as in the 8.2 event deposits at Tloutle (Mitchell 2000), suggesting settlement regimes in which site visits were more frequent, lasted longer or involved larger groups, perhaps as resource patches became more fragmented in space and time.

It is a near certainty that the variables underwriting at least some of these subsistence-settlement and technological shifts were not strictly resource-related. Many of the phases of climatic downturn and resource stress reviewed here see tantalizing hints of enhanced social networking on various levels, ranging from long-distance importation of marine shell and ostrich eggshell beads (Mitchell, 1996b, 2000) to tentative evidence for group aggregations involving more formalized uses of space (Mitchell, 2000; Ouzman and Wadley, 1997) and, in at least one instance, scheduling to take advantage of seasonal resource bonanzas (Mitchell et al., 2011). Though speculative, if their purpose was to strengthen intra- and inter-group cohesion and co-operation during uncertain times, such social strategies may have entailed an enhanced ritual component. Circumstantial evidence for this may come from increased frequencies in some sites of mineral pigments (e.g. Mitchell, 1996a), but more certainly by phases of intensified production of rock art only now beginning to be recognized through direct dating of the paintings themselves (Bonneau et al., 2017).

7. Conclusions and future prospects

1199

1200 In this paper we have reviewed the current state of knowledge regarding
1201 palaeoclimatic and palaeoenvironmental changes in southern Africa's highest region
1202 — the Maloti-Drakensberg — over the past 50 kyr. We employed a quality-controlled
1203 dataset of 325 archaeological radiocarbon dates to preliminarily explore patterning in
1204 human occupation across the region. Integrating these climatic and chronological
1205 data, we attempted to identify the climatic contexts in which a human presence in the
1206 region was strongest (weakest), to understand what attracted (dissuaded) foragers
1207 from inhabiting the region at such times, and to explore some of the clearest examples
1208 of human-environment dynamics in this unique mountain zone. The region's
1209 altitudinal reach renders it particularly sensitive to climate changes, which, together
1210 with associated shifts in local environments, were considerable and complex. High
1211 amplitude and sometimes rapid shifts are evident not only within phases known for
1212 their volatility, such as MIS 3, but also at numerous other occasions during the later
1213 Pleistocene and Holocene.

1214

1215 Despite this variability, diverse proxy indicators suggest that parts of the region
1216 enjoyed a substantial degree of habitat resilience. Resources and the human
1217 populations dependent on them appear to have been particularly stable in the
1218 mountain 'core' of Lesotho, where the Orange-Senqu fluvial network and other
1219 similarly well-protected and resource-rich deep valley systems (Fig. 7) offered
1220 foragers greater predictability than many other inland regions. Our radiocarbon data
1221 tentatively suggest two instances of intensified human presence in Lesotho coinciding
1222 with regional instability — mid MIS 3 and the early Holocene — complicating and
1223 enriching our previous expectations regarding the climatic contexts of upland

exploitation. Importantly, the region's proxy data also reinforce a picture emerging across the SRZ of cold maxima having been characterized by greater moisture availability than previously postulated on account of reduced evapotranspiration, a compensation that was surely more pronounced in the Maloti-Drakensberg than anywhere else on the subcontinent. Cold-phase precipitation, moreover, may have also been more evenly spread throughout the year than now, with proportionally more falling in winter as snow. Both observations have significant implications for Afromontane foragers, who, our radiocarbon dataset indicates, were surprisingly active at such times perhaps because the environment was less unproductive than previously thought with riverine resources becoming even more abundant as their terrestrial counterparts dwindled. The recurrent intensification of riverine fishing during the early LGM, ACR, and Neoglacial (Stewart and Mitchell, in press) represents one of numerous flexible responses documented archaeologically for Maloti-Drakensberg foragers faced with some of the more acute environmental shifts reviewed here.

While considerable progress has been made towards reconstructing palaeoclimates and human-environment dynamics in the Maloti-Drakensberg, much remains to be done. Topping the list is the need to fill remaining gaps in space and time. The vast bulk of the region's palaeoenvironmental data, for example, derive from archaeological rockshelters whose sequences are punctuated by major occupational and sedimentary hiatuses, with more continuous, non-archaeological archives limited to the past ~16 ka (Norström et al., 2014). More effort must be directed towards locating and sampling deeper-time archives, such as alluvial and colluvial-palaeosol sequences. Similarly, greater spatiotemporal coverage is needed for the region's

archaeology. Particularly pressing is new research in the eastern Free State and Lesotho's Maloti ranges, where entire valley systems and adjacent landscapes remain unexplored, as well as further re-excavation of known rockshelter sequences of great antiquity (e.g. Stewart et al., 2012). Heeding lessons learned from Likoaeng, more such stratified open sites must be located and investigated in order to capture periods, activities, or living arrangements not represented in rockshelters (Mitchell et al., 2011). Methodological advances are necessary for building interpretative bridges between different kinds of archaeological records, such as undated open-air lithic scatters and rockshelter sequences, and between the latter and rock art panels (Bonneau et al., 2017). New approaches to formerly and freshly generated archaeological data are also required if we are to test models of seasonal mobility (Cable 1984; Carter 1970, 1978; Opperman 1987) within the longer-term, climatically diverse occupational pulses discussed here. Finally, developing tools for detecting the presence, antiquity, and ecological impact of landscape modification practices employed by Afromontane foragers, and their feedback on cultural systems, are essential if we wish to illuminate the full spectrum of human-environment dynamics in southern Africa's highest region.

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1722

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1725 **Figure captions**

1726

1727 Fig. 1. Map of the Maloti-Drakensberg region with locales/sites mentioned in the text.
1728 Circles: archaeological sites; squares: non-archaeological sedimentary sequences;
1729 crosses: speleothem records; triangles: glacial/peri-glacial landforms. AN: Aliwal
1730 North; BH: Braamhoek; BV: Belleview; BW: Bonawe; CP: Cathedral Peak; CAC:
1731 Cold Air Cave; CH: Collingham; CW: Colwinton; CR: Craigrossie; D1: Diamond 1;
1732 FTP: Fateng Tsa Pholo; GH: Good Hope; HM: Ha Makotoko; KC: Kilchurn; LIK:
1733 Likoaeng; LIP: Liphofung; LIT: Lithakong; LH: Leliehoek; LQ: Leqooa; MF:
1734 Mafadi; MH: Mahwaqa; MT: Masotcheni Formation; MMS: Mauermanshoek; MEL:

1735 Melikane; MB: Mfabeni; MW: Mhlwazini; MP: Mokhotlong Peats; ME: Mount
 1736 Enterprise; NT: Ntloana Tsoana; RC: Ravenscraig; RCC: Rose Cottage Cave; ST:
 1737 Sani Top; SEH: Sehonghong; SK: Sekhokong; SC: Sibudu Cave; STA: Strathalan A;
 1738 STB: Strathalan B; TV: Te Vrede; TN: Thabana Ntlenyana ‘Site 1’; TD: Tiffindell;
 1739 TP: Tlaeeng Pass; TL: Tloutle; TLM: Tsatsa-La-Mangaung; TS: Tsoaing; TC:
 1740 Tswaing Crater; TP: Twyfelpoort; WC: Wolkberg; WK: Wonderkrater.

1741

1742 Fig. 2. Landscape views of the Maloti-Drakensberg’s major sub-regions. Lowland
 1743 Lesotho: courtesy and copyright Charles Arthur; Eastern Free State: courtesy and
 1744 copyright Creative Commons user JMK under license type CC BY-SA 3.0; other
 1745 photos taken by BAS.

1746

1747 Fig. 3. Maloti-Drakensberg palaeoclimatic/environmental proxy data for archives
 1748 exceeding ~16 ka. Also shown are selected archives from the broader summer rainfall
 1749 zone (SRZ). a: Melikane soil organic matter (SOM) $\delta^{13}\text{C}$ (Stewart et al., 2016); b:
 1750 Sehonghong SOM $\delta^{13}\text{C}$ (Loftus et al., 2015); c: Ntloana Tsoana SOM $\delta^{13}\text{C}$; d: Ha
 1751 Makotoko SOM $\delta^{13}\text{C}$ (Roberts et al., 2013); e: ages of various ^{14}C -dated morains on
 1752 uKhahlamba-Drakensberg Escarpment (Mills et al., 2009b); f: Wonderkrater pollen
 1753 reconstructed mean temperature during the coldest quarter (of the year) (TColdQ); g:
 1754 Wonderkrater pollen reconstructed mean temperature during the warmest quarter
 1755 (TWarmQ) (Truc et al., 2013); h: Cold Air Cave speleothems $\delta^{18}\text{O}$ (Holmgren et al.,
 1756 2003); i: SRZ pollen stack reconstructed mean annual temperature (TmeanAnn)
 1757 (relative to present-day) (Chevalier and Chase, 2015); j: Melikane ratio of dicotyledon
 1758 to Poaceae (D/P) phytoliths; k: Melikane counts of bulliform phytoliths (Stewart et
 1759 al., 2016); l: Thabana Ntlenyana ‘Site 1’ ^{14}C -dated colluvial/paleosol sequence (Grab

and Mills, 2011); m: Wonderkrater pollen reconstructed mean precipitation during the wettest quarter (PWetQ) (Truc et al., 2013); n: Tswaing Impact Crater total inorganic content (TIC) (Kristen et al., 2007); o: central/eastern SRZ pollen stack reconstructed mean precipitation during the wettest quarter (relative to present-day); p: northern SRZ pollen stack reconstructed mean precipitation during the wettest quarter (relative to present-day) (Chevalier and Chase, 2015).

Fig. 4. Maloti-Drakensberg palaeoclimatic/environmental proxy data for archives younger than ~16 ka. Also shown are selected archives from the broader summer rainfall zone (SRZ). a: Braamhoek pollen first principle component (PC1) temperature index; b: Braamhoek frequency of fynbos pollen (Norström et al., 2014); c: Mahwaqa PC1 temperature index (Neumann et al., 2014); d: Rose Cottage Cave grazer tooth enamel $\delta^{13}\text{C}$ (Smith et al., 2002); e: combined Rose Cottage Cave and Tloutle charcoal standardized first factor (SSF1) temperature index (Esterhuysen et al., 1999); f: Mafadi diatoms PC1 temperature(/moisture) index; g: Mafadi pollen PC1 temperature(/moisture) index (Fitchett et al., 2017a); h: Likoaeng soil organic matter (SOM) $\delta^{13}\text{C}$ (Parker et al., 2011); i: Wonderkrater pollen reconstructed mean temperature during the coldest quarter (of the year) (TColdQ); j: Wonderkrater pollen reconstructed mean temperature during the warmest quarter (TWarmQ) (Truc et al., 2013); k: Cold Air Cave speleothems $\delta^{18}\text{O}$ (Holmgren et al., 2003); l: SRZ pollen stack reconstructed mean annual temperature (TmeanAnn) (relative to present-day) (Chevalier and Chase, 2015); m: Braamhoek pollen PC2 moisture index; n: Braamhoek $\delta^{13}\text{C}_{35}$ *n*-alkane biomarker; o: Braamhoek $\delta^{13}\text{C}_{33}$ *n*-alkane biomarker; p: Braamhoek sediment magnetic susceptibility (Norström et al., 2014); q: Aliwal North pollen PC1 moisture index; r: Craigrossie pollen PC1 moisture index (Scott et al.,

1785 2012); s: Mahwaqa PC2 moisture index (Neumann et al., 2014); t: Rose Cottage Cave
 1786 grazer tooth enamel $\delta^{18}\text{O}$ (Smith et al., 2002); u: combined Rose Cottage Cave and
 1787 Tloutle charcoal standardized first factor (SSF2) moisture index (Esterhuysen et al.,
 1788 1999); v: moisture summary of Tsoaing proxies (pollen, phytoliths, sedimentological)
 1789 (Grab, 2005); w: Likoaeng phytolith ‘climatic index’ ($\text{Ic}\% = \text{Pooid} / \text{Pooid} +$
 1790 $\text{Chloridoid} + \text{Panicoid}$) (Parker et al., 2011); x: Wonderkrater pollen reconstructed
 1791 mean precipitation during the wettest quarter (PWetQ) (Truc et al., 2013); y: Tswaing
 1792 Impact Crater reconstructed mean annual precipitation (MAP) (Kristen et al., 2007);
 1793 z: central/eastern SRZ pollen stack reconstructed mean precipitation during the
 1794 wettest quarter (relative to present-day); zz: northern SRZ pollen stack reconstructed
 1795 mean precipitation during the wettest quarter (relative to present-day) (Chevalier and
 1796 Chase, 2015).

1797

1798 Fig. 5. Summed probability distribution (red) of the full quality-controlled database of
 1799 325 radiocarbon dates available for the Maloti-Drakensberg against a histogram of
 1800 dates per millennium (gray). Also shown are the major climatic phases discussed I
 1801 this paper.

1802

1803 Fig. 6. Comparison of radiocarbon date frequencies for the Maloti-Drakensberg’s six
 1804 major sub-regions shown as histograms of dates per millennium.

1805

1806 Fig. 7. Four views of the upper Orange-Senqu River fluvial network in the mountain
 1807 core of highland Lesotho. a: looking south along the Orange-Senqu between
 1808 Makunyapane and Mashai; b: looking west down the Sehonghong River towards its
 1809 confluence with the Orange-Senqu; c: looking west down the Melikane River towards

1810 its confluence with the Orange-Senqu; d: Looking south along the Orange-Senqu
1811 between Sehonghong and Tebalo.

1812

1813

1814 **ACKNOWLEDGEMENTS**

1815

1816 Our work in the Maloti-Drakensberg has been made possible because excavation
1817 permits generously granted by the Protection and Preservation Committee (PPC) of
1818 the Lesotho Department of Culture. We thank the PPC, and especially bo-Mme
1819 Moliehi 'Maneo Ntene, Puseletso Moremi and the late Ntsema Khitsane for their
1820 support. We also thank the people of Lesotho, and all colleagues and students with
1821 whom we have been fortunate enough to work in the Maloti-Drakensberg. Finally, we
1822 gratefully acknowledge research funding from the Arts and Humanities Research
1823 Board (UK), the British Academy, the Leverhulme Trust, the McDonald Institute for
1824 Archaeological Research, the National Research Foundation, the Prehistoric Society,
1825 St. Hugh's College (Oxford), the Swan Fund, the Wenner-Gren Foundation, and the
1826 Universities of Cambridge, Cape Town, Lampeter, Michigan, Oxford and the
1827 Witwatersrand.

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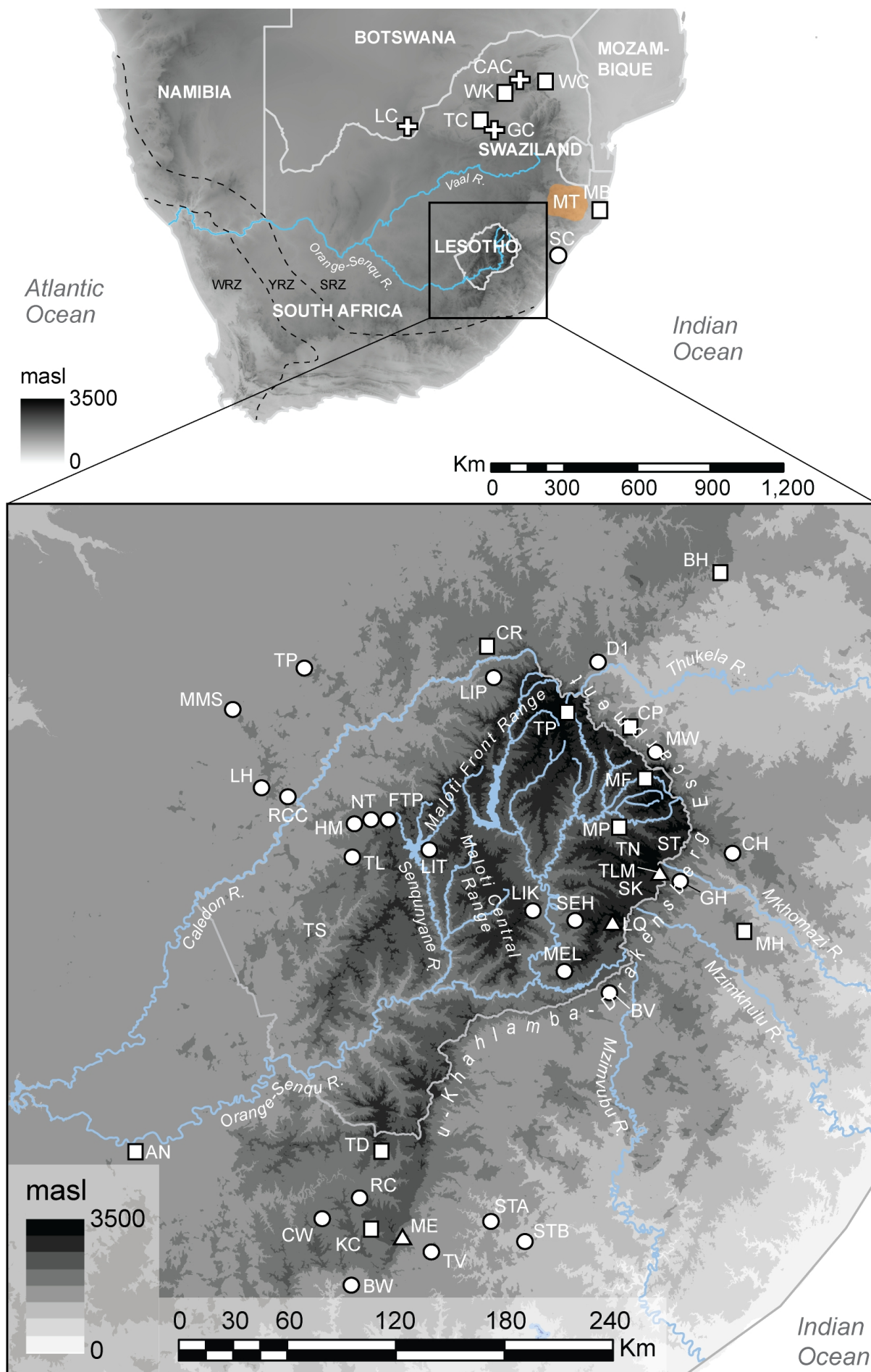
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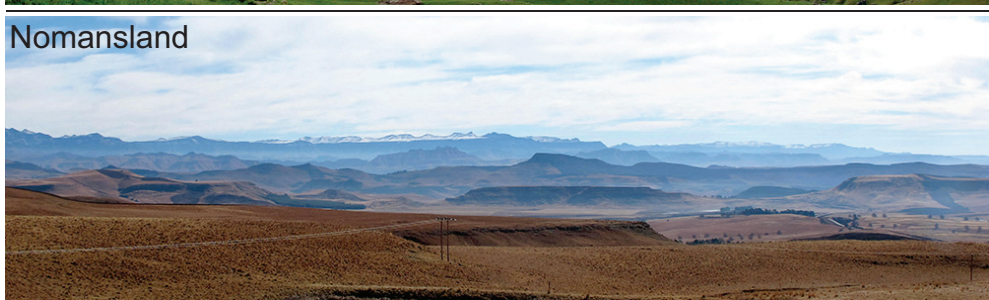
Northern KwaZulu-Natal Drakensberg



Southern KwaZulu-Natal Drakensberg



Nomansland



Lesotho Highlands



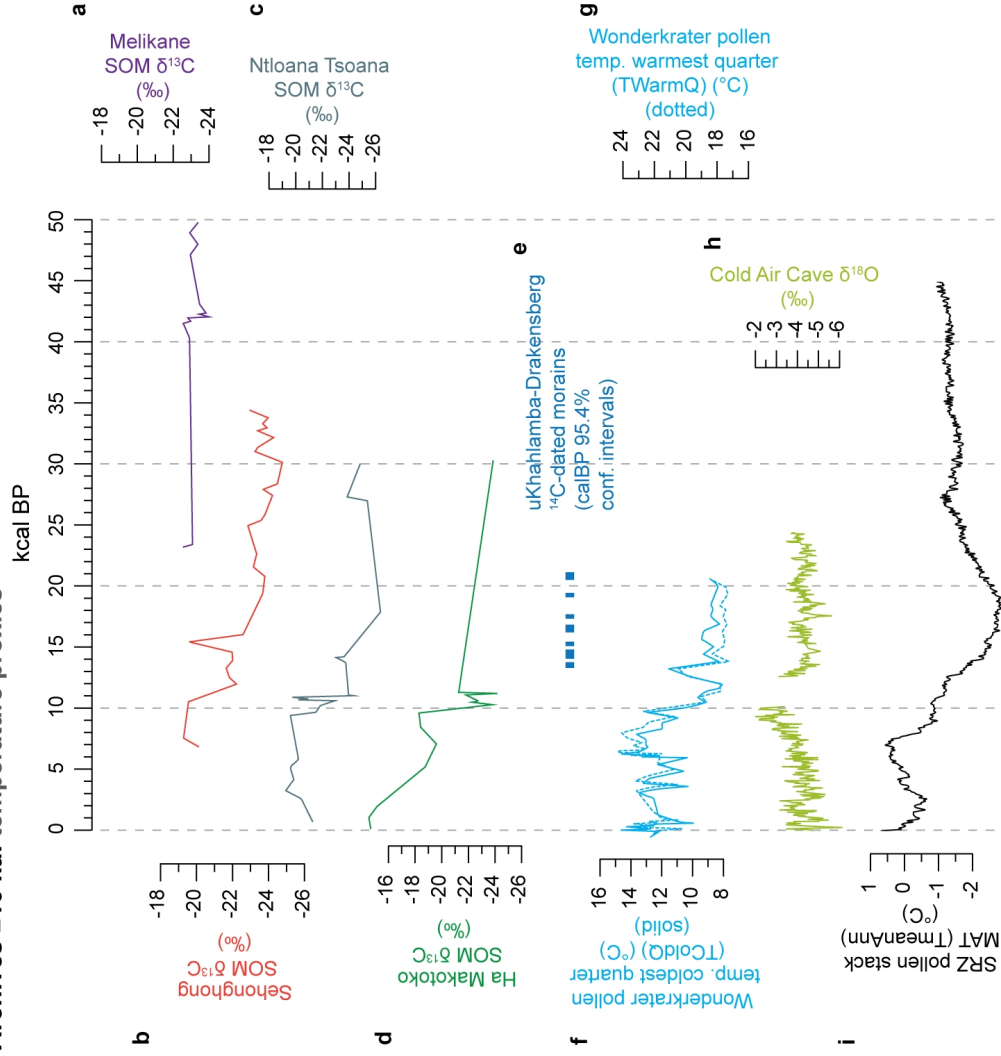
Lesotho Lowlands



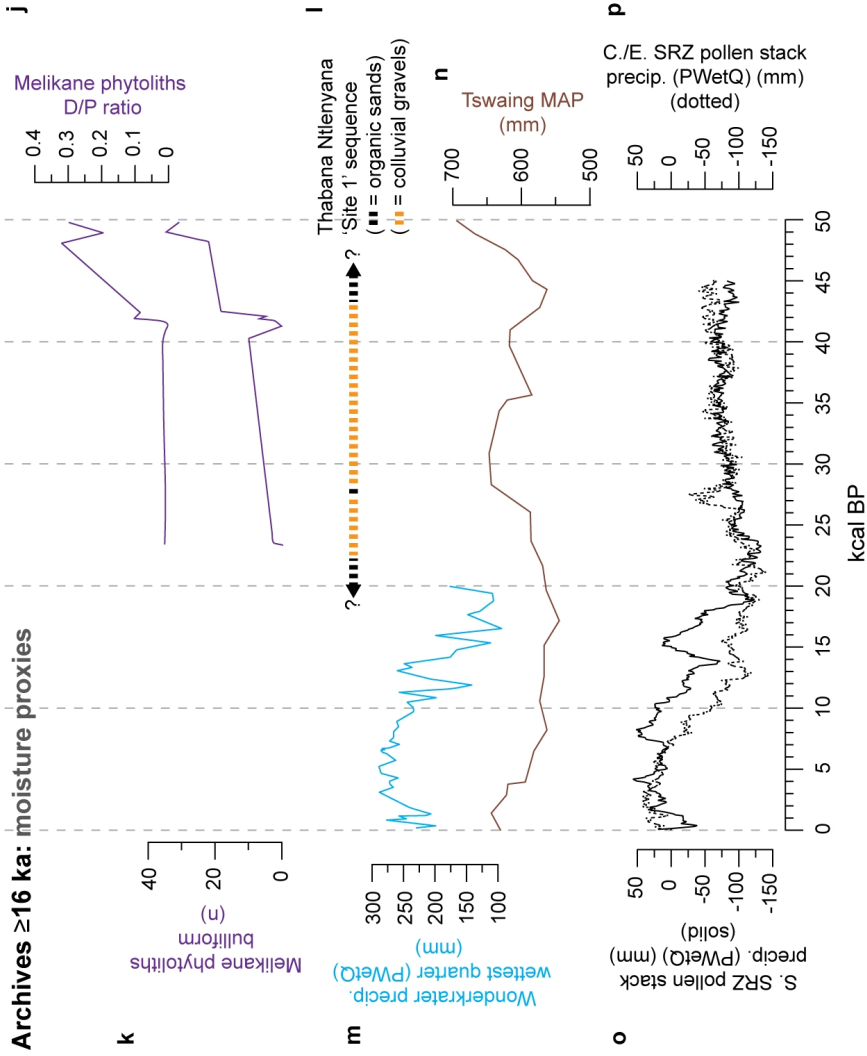
Eastern Free State



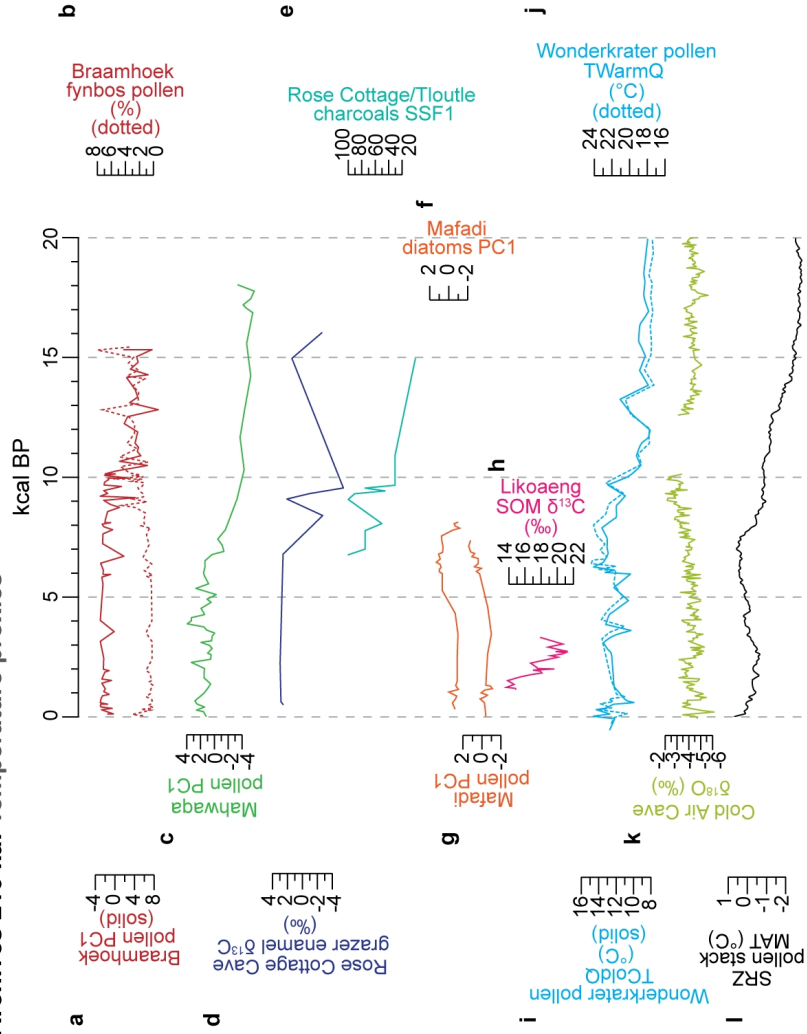
Archives ≥16 ka: temperature proxies



Archives ≥16 ka: moisture proxies



Archives ≤16 ka: Temperature proxies



Archives ≤16 ka: Moisture proxies

