

Sand ramps as late Quaternary palaeoenvironmental archives: Analysis from Southern Africa

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ABSTRACT

Sand ramps are widespread but understudied landforms which have the potential to provide detailed palaeoenvironmental information in dryland regions. This thesis investigates the utility of sand ramps as late Quaternary palaeoenvironmental archives by addressing two research questions: (1) What are the main controls on sand ramp formation in southern Africa? (2) What does the sand ramp record tell us about late Quaternary palaeoenvironments in southern Africa?

The distribution of sand ramps in southern Africa was surveyed using Google Earth™ and 75 features were identified in southern Namibia. Ten of these sand ramps, and an additional feature from South Africa, were studied in the field. Sediments and morphology were mapped and a total of 64 OSL dates, 96 sediment samples, 10 heavy mineral assemblages and OSL sensitivity data from 8 samples were examined.

The distribution of sand ramps suggests formation is dependent on (1) sediment supply, (2) accommodation space, (3) persistent unidirectional wind and (4) a variable semi-arid to arid climate. Chronologies and sediment analyses indicate individual sand ramp accumulation is locally controlled by sediment supply modulated by the availability, and nature, of the accommodation space. Comparison between the Namibian and South African sand ramps suggests considerable regional variation in the factors controlling sediment supply.

The Namibian sand ramps show an affinity to local ephemeral river channels. Periods of dated sand ramp activity in multiple features are interpreted as periods of increased regional fluvial activity. Significant activity occurred at ~21-12 ka (with peaks at 21-18.5 ka and 14.5-12 ka), ~8.5-7.5 ka and ~2 ka. Some activity is also indicated at 85-65 ka and 45-35 ka. These results correspond well to regional records.

Overall, this study demonstrates that sand ramps can provide palaeoclimatic information on both the local and regional scale but only if a number of caveats are taken into consideration.

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ABBREVIATIONS AND UNITS

NH	Northern Hemisphere	OSL	Optically Stimulated Luminescence
SH	Southern Hemisphere	TL	Thermoluminescence
WRZ	Winter Rainfall Zone	D_e	Equivalent Dose (Gy)
SRZ	Summer Rainfall Zone	SAR	Single Aliquot Regenerative
YRZ	Year-round Rainfall Zone	IR	Infrared
ITCZ	Intertropical Convergence Zone	IRSL	Infrared Stimulated Luminescence
CAB	Congo Air Boundary	PH	Preheat
SAA	South Atlantic Anticyclone	DRC	Dose Response Curve
SIA	South Indian Anticyclone	SSE	Single Saturating Exponential
TTT	Tropical Temperate Trough	SEPL	Saturating Exponential Plus Linear
SST	Sea Surface Temperature	DSE	Double Saturating Exponential
MAR	Mean Annual Rainfall	DRT	Dose Recovery Test
MIS	Marine Isotope Stage	PMT	Photomultiplier Tube
LGM	Last Glacial Maximum	PCA	Principle Component Analysis
LGIT	Last Glacial Interglacial Transition	PDF	Probability Density Function
YD	Younger Dryas		
ACR	Antarctic Cold Reversal		

Units

ka	Thousands of years before present	Gy	Gray (1 Gy = 1 J kg ⁻¹)
		nm	Nanometres
		W	Watt 1 W = 1 J s ⁻¹
Cal BP	Calibrated radiocarbon years before AD1950	J	Joule

Part 1: Introductions

Chapter 1: Introduction

1.1 INTRODUCTION

The late Quaternary (~130 ka - present) is a period of considerable global climatic and environmental change (see Bell and Walker, 2005; Lowe and Walker, 2014 for detailed discussion). Due to information derived from a variety of proxy records (e.g. high resolution ice cores, lake cores and speleothem records, long marine cores and detailed organic and geomorphic records) we have a good understanding of the broad climatic changes, and their environmental manifestations, that occurred during this period. However, there is spatial disparity in the density and quality of available proxy records with the majority clustered in the northern hemisphere (Figure 1-1). Records are particularly sparse, and often of low resolution, in dryland regions. Arid conditions and typically seasonal rainfall inhibit the preservation of traditional organic proxies whilst the more abundant geomorphological records (e.g. dunes) are frequently discontinuous and can be difficult to interpret (Chase, 2009; Chase and Meadows, 2007; Thomas and Burrough, 2012). Therefore, our understanding of late Quaternary climatic changes, and their spatial environmental manifestations, in dryland regions is relatively limited, particularly in the southern hemisphere.

Drylands (hyper-arid to dry-subhumid; $\frac{\text{precipitation}}{\text{potential evapotranspiration}} = <0.65$) cover >40% of global land surface, house ~35% of the global population (UNEP, 1992; UN-MEA, 2005) and are extremely sensitive to perturbations in climate, particularly moisture availability. Desertification, dust flux and dune mobility have far reaching implications and present major concerns when considering

anthropomorphic climate change in recent decades and into the future (e.g. Reed and Stringer, 2015; Wang et al., 2012). Therefore, accurately understanding how these regions responded to past climatic changes and over longer time frames can provide valuable insights into possible future conditions.

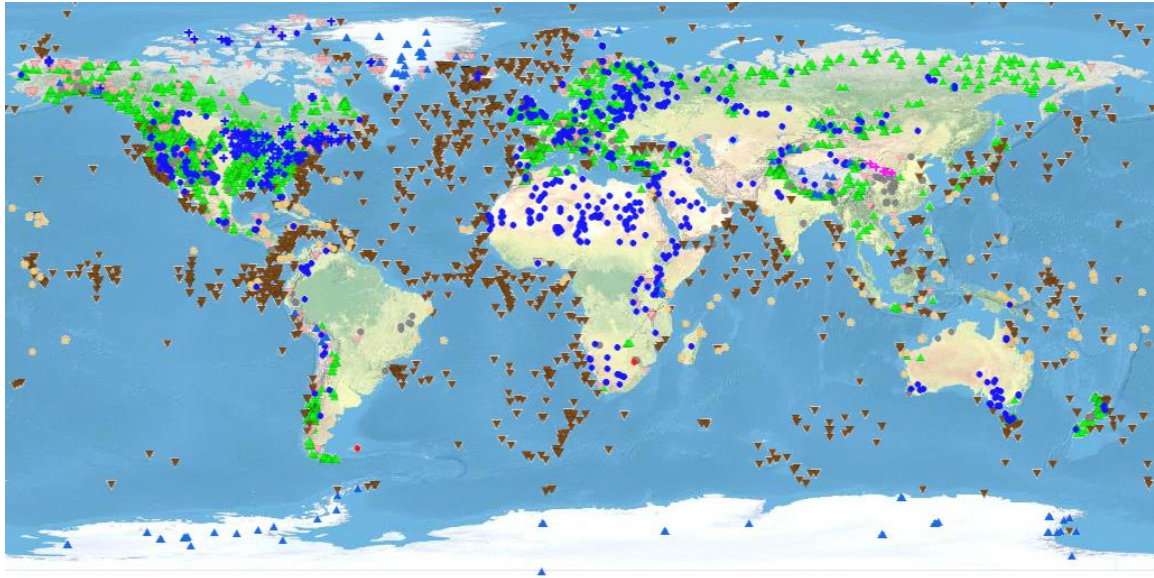


Figure 1-1. Distribution of palaeoclimate data available on NOAA database. Brown triangles represent marine cores, orange circles = coral records, blue triangles = ice cores, blue circles = lake level records, pink crosses = loess, blue crosses = macrofossils, pink triangles = palaeolimnology studies, red circles = pollen records, grey circles = speleothem record, green triangles = tree ring data. Available from:

<http://gis.ncdc.noaa.gov/map/viewer/#app=cdo&cfg=paleo&theme=paleo&layers=0010101000101001>

This dataset is not a complete compilation of all Quaternary records, nor is the resolution or duration of the records apparent from this map. Nevertheless, it provides a representative overview of the distribution of available records

Southern Africa is a significant region for dryland palaeoenvironmental research as it is located at the confluence of both tropical and temperate climate systems. In addition, due to the impact of ice extent on global circulation, southern Africa was influenced by both northern hemisphere (NH) and southern hemisphere (SH) climate forcings over the Late Quaternary (Chase et al., 2015a; Collins et al., 2014; Stone, 2014). However, as with most drylands, our present understanding of environmental change is limited, characterised by spatially and temporally patchy proxy records which present a confusing and sometimes conflicting

picture (Burrough and Thomas, 2013; Chase and Meadows, 2007; Chevalier and Chase, 2015; Gasse et al., 2008; Scott et al., 2012; Stone, 2014).

Advances in Optically Stimulated Luminescence (OSL) dating have permitted significant developments in our understanding of Quaternary environments in southern Africa and other dryland regions. The sediments of depositional landforms (e.g. dune, lake shorelines, ephemeral river channels) can be directly dated using OSL, providing an estimate of the time elapsed since the sediments were deposited and buried from sunlight. These 'geoproxy' reconstructions therefore present a direct and abundant record of landscape response to environmental change (Thomas, 2013; Thomas and Burrough, 2012). However, geoproxy records are typically discontinuous and considerable uncertainty can surround their climatic interpretation (Chase, 2009; Telfer and Hesse, 2013; Thomas, 2013). Therefore, there are still considerable gaps in our understanding of Quaternary dryland behaviour.

Sand ramps, accumulations of aeolian sands interdigitated with fluvial and hill slope deposits anchored against a topographic barrier, have the potential to provide detailed records of Quaternary environmental change in drylands (Kumar et al., 2016; Lancaster and Tchakerian, 1996; Telfer et al., 2012). The topographic barrier stabilises the deposits making long term, continuous deposition possible (Chase, 2009) whilst the changes in depositional mechanism can record prevailing environmental conditions (Lancaster and Tchakerian, 1996) and the sediments themselves can be directly dated using OSL. However, to

date, sand ramp research has been limited, with the majority of research focusing on single features (e.g. Bateman et al., 2012; Telfer et al., 2012; Thomas et al., 1997). Therefore, the controls governing sand ramp formation, accumulation and preservation are not fully understood and the full potential of sand ramps as palaeoenvironmental archives has not been realised. Focusing on southern Africa, and using the following research questions, **the aim of this thesis is to assess the utility of sand ramps as palaeoenvironmental archives.**

1.2 RESEARCH QUESTIONS

This thesis addresses this aim through a programme of field and laboratory analysis that focuses on answering two broad questions:

1. What are the main controls on sand ramp formation in southern Africa?

Understanding the controls operating on sand ramps provides insights into where sand ramps do and do not form, the palaeoenvironmental conditions which lead to accumulation, and the sensitivity and completeness of the environmental record.

A suite of 75 sand ramps from Namibia and a single sand ramp from South Africa are used to assess the conditions needed for sand ramp formation and accumulation in these regions.

2. What does the sand ramp record tell us about late Quaternary palaeoenvironments in southern Africa?

Late Quaternary palaeoenvironmental reconstructions are currently limited by a paucity of records. Using the information gained by answering research question 1, the sand ramp record is used to reconstruct environmental conditions in southern Africa over the past ~100 ka, providing a case study of the palaeoenvironmental utility of the sand ramp record.

1.3 THESIS STRUCTURE

This thesis has been written by papers adhering to the University of Oxford, School of Geography guidelines. The majority of results are included and discussed in four papers which have been submitted for publication in a range of peer-reviewed journals. These are presented in their thesis adapted form in Chapters 4-7 but a certain degree of repetition is unavoidable. The four papers have been written with co-authors and details of their contributions can be found in Appendix A.

The thesis has been split into three parts. Part 1, *Introductions*, details the context of the problem (Chapter 2) and the methods used to tackle the research questions (Chapter 3) are presented. Paper 1 (Chapter 4) which examines the reliability of the OSL dose response curve at high doses (i.e. how accurate the dating of old samples is) is included in this section as it shapes how OSL data are interpreted in subsequent papers.

Part 2, *Investigations and Interpretations*, presents the bulk of the geomorphic and palaeoenvironmental research and the interpretations of the results. This section comprises the remaining three papers. Paper 2 (Chapter 5) directly addresses research question 1 and explores the geomorphic controls operating on sand ramps in southern Namibia. Paper 3 (Chapter 6) straddles both research questions 1 and 2 using a single sand ramp from South Africa to assess the regional variation in controls on sand ramp formation and to explore the challenges of reconstructing palaeoenvironments from sand ramp sediments. Paper 4 (Chapter 7) addresses research question 2 using the suite of Namibian sand ramps to reconstruct regional palaeoenvironmental conditions.

Part 3, *Implications*, draws the research together to provide answers to the research questions and to discuss the reliability of the age estimates underpinning these conclusions (Chapter 8). The conclusions and overall utility of sand ramps as palaeoenvironmental archives are discussed in the final chapter (Chapter 9).

Chapter 2: Context

2.1 INTRODUCTION

The sand ramps studied in this thesis are located in Namibia between 23°-28°S and 15°-19°E and in South Africa at 31°23'35.10"S, 26°41'24.40"E. The Namibian climate is currently hyper-arid to arid whilst the South African site (the Klipkraal Sands) has a semi-arid climate today. The site-specific geological and climatic settings of these locations are discussed in detail in Chapters 5 and 7, and 6 respectively. Southern Africa is an interesting case study for late Quaternary climate change as it sits at the confluence between temperate and tropical climatic systems, allowing changes in their spatial distribution to be examined. In addition, both direct SH insolation forcing and changes in global circulation driven by NH ice extent influence the climate of this region over Quaternary timescales (Chase et al., 2015a; Collins et al., 2014; Stone, 2014). At present, our knowledge on both late Quaternary hydroclimatic change in southern Africa and on the utility of sand ramps as palaeoenvironmental archives is incomplete (Bateman et al., 2012; Chase and Meadows, 2007; Gasse et al., 2008; Stone, 2014). This section examines the current hydroclimate of southern Africa before discussing theories of, and evidence for, Quaternary climate change. The palaeoenvironmental significance of aeolian and hillslope records is examined before focusing on the sand ramp record, displaying the current state of knowledge and identifying areas for further research.

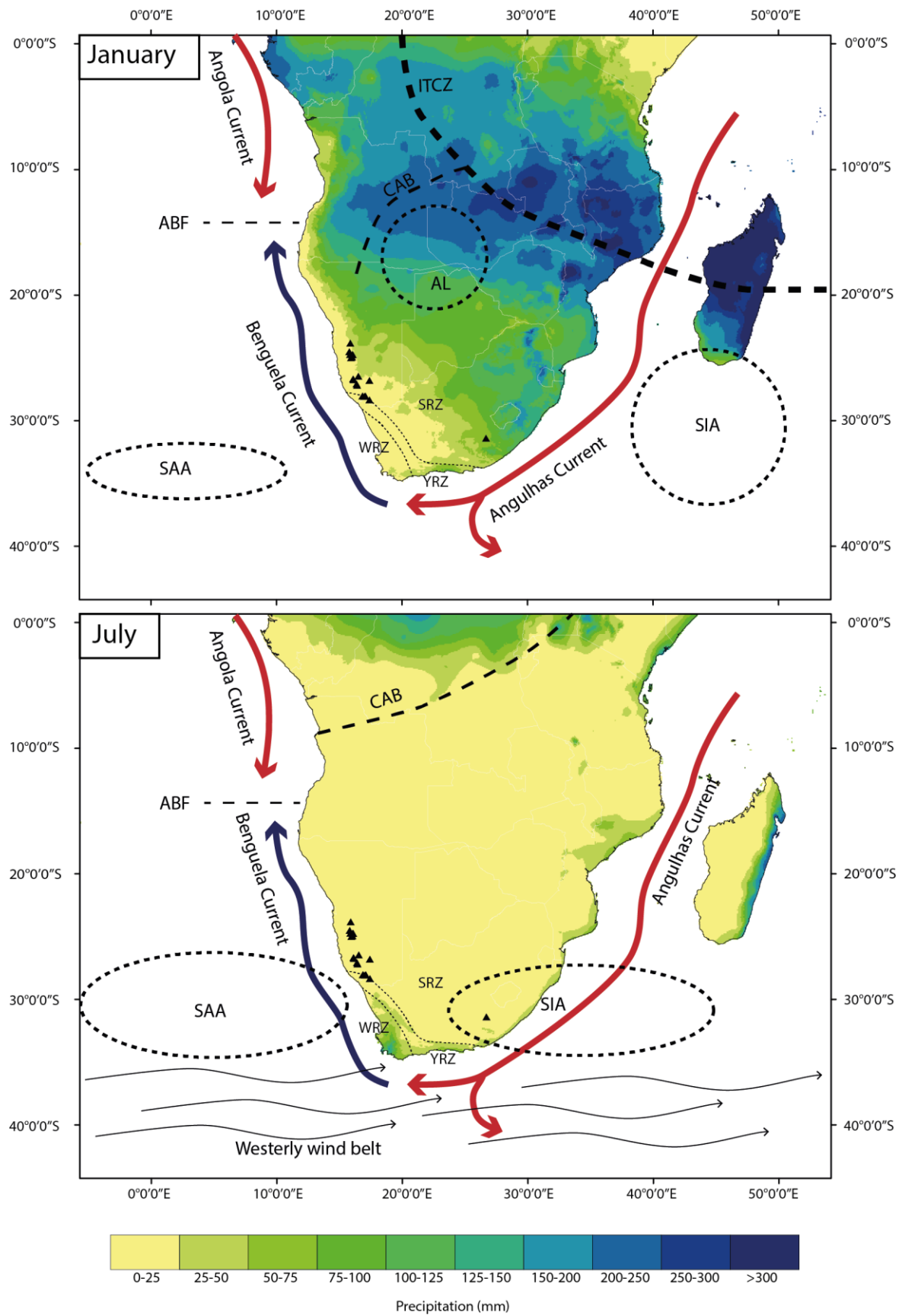


Figure 2-1. Mean precipitation (Hijmans et al., 2005) and position of circulatory systems in January (Austral summer) and June (Austral winter). Black triangles are sand ramps studied in this thesis. See text for explanation of acronyms.

2.2 PRESENT HYDROCLIMATE OF SOUTHERN AFRICA

Precipitation in southern Africa is derived from both tropical and temperate sources and is highly seasonal. Precipitation in the south west (winter rainfall zone; WRZ; >66% of rain falls June-August; Chase and Meadows, 2007) is associated with the equatorial migration and strengthening of the South Atlantic Anticyclone (SAA) and mid-latitude westerlies during the Austral winter (Figure 2-1). Precipitation over the remaining majority of the sub-continent (summer rainfall zone; SRZ; >66% of rain falls December – February; Chase and Meadows, 2007) is convectional and associated with the summer latitudinal migration of the Intertropical Convergence Zone (ITCZ) and zonal migration of the Congo Air Boundary (CAB) (Figure 2-1). The CAB delimits precipitation derived from the Atlantic Ocean via the West African monsoon from precipitation derived from the Indian Ocean from the East African Monsoon (EAM) which flows from the South Indian Anticyclone (SIA). The position of the CAB is determined by the relative strength of the monsoonal systems. (Reason et al., 2006; Tyson, 1986; Tyson and Preston-Whyte, 2000). Tropical-Temperate Troughs (TTTs) (where tropical systems meet mid-latitude low pressure systems) are the source of the majority of moisture in southern Africa (Harrison, 1984; Todd et al., 2004). The year-round rainfall zone (YRZ) (Figure 2-1) receives precipitation from both systems with no notable dry season.

The topography and oceanography of the continent produce a ENE to WSW precipitation gradient (Figure 2-1). In the east the warm Agulhas Current promotes

convective rainfall which is carried into the interior of the sub-continent by the SE trade winds, driven by the EAM and the development of the Angola low pressure system (Figure 2-1). In the west SE trade winds cause the cold Benguela current to upwell between 15-34°S inhibiting convection and moisture transport. Additionally, the Great Escarpment intercepts moisture carried from the Indian Ocean by SE trade winds preventing it from reaching the arid west. The Great Escarpment also restricts the inland penetration of the winter frontal rain derived from the westerlies, further contributing to the aridity of the interior. The Benguela current meets the warm Angola current at ~15-17°S (Angola-Benguela Front; ABF). SE trade winds are deflected clockwise bringing some Atlantic moisture inland at this latitude. (Reason et al., 2006; Tyson, 1986; Tyson and Preston-Whyte, 2000).

2.3 QUATERNARY CLIMATE CHANGE

Reconstructions of late Quaternary palaeotemperatures show good agreement and indicate southern Africa experienced mean cooling of 5-7°C over the Last Glacial compared to mean Holocene conditions (Heaton and Talma, 1986; Holmgren et al., 2003; Kulongoski, 2004; Talma and Vogel, 1992) and that the Holocene Athermal (~8-6ka) was 1-2°C warmer than today (Partridge et al., 1999). However, in this arid environment moisture availability is typically the limiting factor in determining environmental conditions. Reconstructions of hydroclimatic conditions in southern Africa over the late Quaternary are heterogeneous, complex and sometimes contradictory (e.g. Burrough and

Thomas, 2013; Chase and Meadows, 2007; Chevalier and Chase, 2015; Gasse et al., 2008; Scott et al., 2012). In addition, due to its role in determining potential evapotranspiration, temperature also modulates moisture availability, adding increased complexity to hydroclimatic reconstructions (Chevalier and Chase, 2016).

2.4 THEORIES OF QUATERNARY HYDROCLIMATIC CHANGE

There are several theories to explain how the hydroclimate of southern Africa may have behaved during the late Quaternary. For simplicity these can be roughly grouped into the behaviour of the mid-latitude westerlies, the influence of the ITCZ, the role of sea surface temperatures (SSTs), and the location of the CAB. However, these climatic systems, driven by global atmospheric circulation and temperature changes, do not operate independently and concurrent and/or heterogeneous climatic changes may have occurred (Burrough and Thomas, 2013; Singarayer and Burrough, 2015).

2.4.1 BEHAVIOUR OF THE MID-LATITUDE WESTERLIES

Winter rainfall in southern Africa is currently controlled by the latitudinal position of the mid-latitude westerlies. The westerly belt migrates equatorward in winter due to the seasonal expansion of Antarctic sea ice. It has been proposed that glacial expansions in Antarctic sea ice would increase the quantity of, and the area receiving, winter rainfall (Chase and Meadows, 2007; Cockcroft et al., 1987; Van Zinderen Bakker, 1976). Increased ice cover would cause an increased

temperature gradient between the equator and the poles resulting in an equatorial displacement and strengthening of the westerly belt. Meanwhile, the same phenomenon in the northern hemisphere is likely to lead to a southward displacement of the ITCZ, further intensifying circulation in the southern hemisphere (Chiang and Bitz, 2005). Therefore, increases in westerly derived winter rainfall are likely to be associated with increased tradewind strength, increased Benguela upwelling and decreased southern Atlantic SST (Little et al., 1997).

Evidence

Within the present-day WRZ there is evidence for increased moisture during the Last Glacial and Last Glacial-Interglacial Transition (LGIT) (Cowling et al., 1999; Meadows and Baxter, 1999; Parkington et al., 2000; Scott and Woodborne, 2007). Isotopes from bovid teeth at Elands Bay indicate this is a result of winter rainfall (Stowe and Sealy, 2016). In contrast, Chase et al. (2015) suggest arid conditions at De Rif over the Last Glacial Maximum (LGM) to Late Glacial (19.4-16 cal. kBP) and the records of humidity over the Holocene indicate variable, but spatially and temporally heterogeneous changes in moisture (Chase et al., 2013, 2015a, 2015b; Hahn et al., 2016; Meadows et al., 2010; Quick et al., 2011; Scott and Woodborne, 2007; Valsecchi et al., 2013; Zhao et al., 2016). The relationship between winter rainfall and Antarctic sea ice extent has been demonstrated over the past 1400 years (Stager et al., 2011) whilst a causal relationship between winter rainfall, tradewind strength, Benguela upwelling and Angulhas leakage is suggested for the Late and Early Holocene (Weldeab et al., 2013). Offshore fluvial

deposits indicate an antiphase relationship between winter and summer rainfall quantity suggesting a northwards shift in circulatory systems during periods of increased winter rainfall (Hahn et al., 2016; Zhao et al., 2016). However, at present, evidence of the above relationships is only available for the Holocene and the impact of glacial boundary conditions remains unknown.

Palaeoenvironmental evidence for glacial expansion of the westerlies and area receiving winter rainfall is inconclusive. Reviews of the available palaeodata derive different conclusions. Chase and Meadows (2007) suggest an increase and expansion of winter rainfall. Gasse et al., (2008) and Chase and Chevalier (2015) suggest an increase in winter rain without expansion of the WRZ whilst Scott et al., (2012) concludes no changes from present-day patterns.

Marine cores provide continuous records of glacial/interglacial changes but findings may not be directly applicable to terrestrial conditions. Along the west coast there is evidence for both increased (Pichevin et al., 2005) and decreased (Collins et al., 2014; Stuut et al., 2002) aridity during glacial periods. Fynbos-type pollen (currently associated with the WRZ) is found in increased abundance in marine cores during MIS 6,4,3 and 2 and may represent an increase in the extent of the WRZ and the vegetation it supports (Kirst et al., 1999; Schneider et al., 1995; Shi et al., 2000, 2001; Urrego et al., 2015). However, these findings are not presently supported by terrestrial pollen evidence (Lim et al., 2016; Scott et al., 2004) and marine isotopic evidence suggests no expansion in C3 (fynbos) vegetation (Rommerskirchen et al., 2006). Collins et al. (2014) suggest that

increases in moisture observed over glacial periods reflect changes in summer, rather than winter, rainfall.

Within the YRZ evidence for increased winter rain is mixed. The Crevice Cave speleothem indicates increased winter rainfall and C3 vegetation alongside decreased summer rain during interglacial MIS 5, compared to MIS 4 (Bar-Matthews et al., 2010). Pollen from Vankervelsvlei covering the same time period does not provide evidence for changes in winter rainfall but suggests an increase in summer rainfall during MIS 5d (Quick et al., 2016). $\delta^{13}\text{C}$ from Congo Cave suggests low C4 (summer rain associated) plant coverage during the Last Glacial (Talma and Vogel, 1992). Holocene $\delta^{14}\text{N}$ concentrations from Seweweekspoort midden have been correlated to changes in Antarctic sea ice suggesting a link between increased humidity and expanded sea ice (Chase et al., 2013). This is in antiphase with the Katbakkies Pass midden from the WRZ during the same time period. As a result $\delta^{14}\text{N}$ at Katbakkies Pass is thought to reflect changes in summer rainfall (Chase et al., 2015b).

Data from climate models is also conflicting with coupled Atmospheric-ocean models and atmospheric only models showing equatorward (Drost et al., 2007; Kim et al., 2002, 2003; Williams and Bryan, 2006), poleward (Kitoh et al., 2001; Shin et al., 2003; Wyrwoll et al., 2000) or negligible (Otto-bliesner et al., 2006) latitudinal shifts of the westerlies with both strengthening (Otto-bliesner et al., 2006; Shin et al., 2003) and weakening (Kim et al., 2002, 2003) observed.

Additionally, the model that most accurately reflects the palaeoenvironmental data shows a slight poleward shift (Sime et al., 2013).

2.4.2 INFLUENCE OF THE ITCZ

The position of the tropical rainbelt influences summer rainfall over eastern- and central-southern Africa with more southerly positions bringing more rainfall to these regions (Tyson and Preston-Whyte, 2000). The tropical rainbelt is influenced, but not solely determined, by the ITCZ (Nicholson, 2009). For simplicity, and following common convention, this review discusses latitudinal changes in the position of the tropical rainbelt in terms of ITCZ dynamics. However, this is an oversimplification with African Easterly, and Westerly Jets playing an important role in rainfall distribution (Nicholson, 2009; Singarayer and Burrough, 2015).

The continental behaviour of the ITCZ over the Late Quaternary is currently debated. Presently, the ITCZ seasonally tracks the latitude of greatest insolation. Over millennial timescales changes in orbital precession influence the seasonal receipt of insolation between hemispheres on a 23ka cycle. Peak seasonality occurred in the northern hemisphere ~10ka with this region experiencing warm summers and cold winters whilst cool summers and mild winters occurred in the SH. At ~21.5ka the reverse was true and peak seasonality occurred in the southern hemisphere (Berger, 1978). It is therefore suggested that the ITCZ tracked these changes over millennial timescales following peak summer insolation. There is good evidence to support this in the northern hemisphere (e.g.

Pokras and Mix, 1987; Rossignol-Strick, 1985) however, data from the southern hemisphere is less clear. Northern hemisphere cooling may also influence the position of the ITCZ with cooler temperatures producing a southward displacement via changes in ocean and atmospheric circulation (Arbuszewski et al., 2013; Chiang and Bitz, 2005; Mohtadi et al., 2014; Vellinga and Wood, 2002).

Evidence

The 200 ka Tswaing Crater rainfall record suggests a correlation between precessional changes in southern hemisphere insolation and summer rainfall (Partridge et al., 1997). However, this relationship breaks down over the past 40 ka, the only part of the record to be independently dated. The remainder of the record is orbitally tuned, presenting a circular argument for insolation driven changes in precipitation. A second, also orbitally tuned, core shows good agreement with Partridge et al. (1997) prior to ~80 ka when insolation oscillation is strongest (Kristen et al., 2007). After ~80 ka the relationship with insolation breaks down (Kristen et al., 2007; Schmidt et al., 2014). Terrestrially derived Fe/K ratios from the south west Indian Ocean (31°S) indicate a good correlation between moisture availability and SH summer insolation over the past 270 ka. However, akin to Tswaing Crater, the relationship breaks down at ~60 ka (Simon et al., 2015). Periods of speleothem construction, linked to increased summer rainfall, are associated with medium to high summer insolation at Wolberg Cave. Although this relationship is not exact and may be superseded by the impact of Heinrich events (Holzkämper et al., 2009). Terrestrially derived leaf waxes from Namibia and the Kalahari (accumulated in the southeast Atlantic; 23°S) and from the

Zambezi River Basin (18°S) indicate a good correlation between insolation and C4 grasses and moisture throughout the past 140 ka and 17 ka respectively (Collins et al., 2014; Schefuß et al., 2011). However, hyrax middens from Namibia suggest an antiphase relationship between moisture and SH summer insolation over the past 11.7 ka (Chase et al., 2009, 2010). Terrestrially derived organic and sedimentary markers from 8 marine cores between 20°N to 17°S also suggest a symmetrical relationship between precipitation change in the NH and SH, suggesting the SH is responding to NH insolation forcing, possibly indicating expansions/contractions of the African rainbelt rather than migration (Collins et al., 2011). To some extent this may be reflecting the northerly positions of the fluvial catchments which source sediment to these cores or may reflect a low resolution expression of the response to annual interhemispheric temperature gradients under glacial boundary conditions (Singarayer and Burrough, 2015). Increased precipitation over south east Africa during the Younger Dryas and Heinrich event 1 is suggested by Green et al. (2015) and Schefuß et al. (2011). This is attributed to a southward displacement of the ITCZ, forced by NH cooling. Climate models suggest shifts in the tropical rainbelt remained asymmetric between hemispheres throughout the late Quaternary. In addition, global cooling and increases in global ice extent, primarily accumulated in the NH, is shown to cause a southward displacement in the rainbelt (Singarayer and Burrough, 2015 and references therein).

2.4.3 ROLE OF SSTs

Changes in Indian Ocean SST, primarily driven by direct insolation and changes in ocean circulation, currently influence rainfall variability in the SRZ (Jury et al., 1993; Reason and Mulenga, 1999). Therefore, Quaternary SSTs is suggested to exert a first order control on southern African rainfall with periods of increased temperature associated with increased atmospheric moisture and rainfall (Stokes et al., 1997b).

Evidence

Pollen records from the offshore Limpopo River basin suggest a strong correlation between aridity and decreased Agulhas SSTs over the past ~300 ka (Dupont et al., 2011). Inland, pollen and biomarker records from within latitudes 23°S–26°S show good correlation with Indian Ocean SSTs and precipitation (Chase et al., 2015b; Kristen et al., 2007; Truc et al., 2013) although there is some evidence that this relationship breaks down in the mid to late Holocene with direct insolation forcing becoming dominant (Chevalier and Chase, 2015). Summer rainfall in the YRZ can also be linked to Agulhas SSTs (Bar-Matthews et al., 2010; Quick et al., 2016) although the two records suggest antiphase relationships. Precipitation reconstruction from central and eastern South African pollen records (27–33°S, 24–33°E) show some similarity to SSTs but are suggested to be modulated by the position of the mid-latitude westerlies (Chevalier and Chase, 2015). Interpretation of the Makapansgat speleothem record (Holmgren et al., 2003) is contested (Brook et al., 2010; Chevalier and Chase, 2015). The $\delta^{18}\text{O}$ record may either represent moisture source suggesting decreased summer rainfall during the LGM

and the Antarctic Cold Reversal (ACR) (Holmgren et al., 2003) or, if interpreted as the amount effect, suggests changes in moisture corresponding to changes in SSTs (Chevalier and Chase, 2015) and is in good correlation to proximal pollen records (Brook et al., 2010).

Models show a positive correlation between SST and summer rainfall during interglacial periods but the relationship breaks down during glacial periods. This is attributed to large changes in atmospheric circulation caused by eustatically exposed continental shelves in the Indo-Pacific (Singarayer and Burrough, 2015). Evidence for aridity concurrent with the Younger Dryas has been found in some southern African records (Brook et al., 2010; Chase et al., 2013, 2015a; Holmgren et al., 2003; Norström et al., 2009, 2014; Truc et al., 2013) and has been linked to decreased Agulhas SSTs (Truc et al., 2013) but chronological uncertainty limits some of these conclusions (e.g. Norström et al., 2009, 2014; Truc et al., 2013).

2.4.4 LOCATION OF THE CAB

In tropical East Africa the location of the CAB has been shown to play a large role in hydrological balances of the Quaternary. Decreases in the strength of the trade winds increase the region influenced by Atlantic rainfall and therefore increase precipitation in central-eastern Africa (Costa et al., 2014; Singarayer and Burrough, 2015; Tierney et al., 2011a, 2011b). However, below 17°S the impacts of longitudinal changes in CAB are reduced due to Benguela upwelling which inhibits convective rainfall from the Atlantic (Singarayer and Burrough, 2015).

2.4.5 OTHER REGIONAL EVIDENCE

In addition to the above, there is a considerable number of palaeoenvironmental reconstructions from southern Africa which do not fit neatly into the above theories. In some cases, the short or discontinuous nature of the records perturbs full identification of climatic forcings (e.g. few terrestrial organic proxies extend past the Holocene whilst inorganic proxies infrequently preserve a continuous record; see Chase and Meadows 2007 for discussion). In other circumstances the palaeoenvironmental significance of the record can be difficult to interpret (e.g. $\delta^{18}\text{O}$ from speleothems and $\delta^{15}\text{N}$ from hyrax middens can be influenced by multiple factors; Brook et al., 2010; Chase et al., 2012; Green et al., 2015, while the palaeoenvironmental significance of geomorphic records are debated (Section 2.5). For the majority of older studies, poor chronological control precludes meaningful conclusions from being drawn (e.g. ^{14}C dating of open system carbonates see Lancaster (2002); Thomas and Shaw (2002) for discussion and review). Regional syntheses can also be hampered by the variety of proxies available. Different records respond to different factors, which need to be taken into account when comparing records. For example, Chevalier and Chase (2016) stress the difference between moisture availability, which is heavily modulated by temperature, and total rainfall when interpreting the pollen record. As such, pollen records may suggest moist conditions during periods of low temperatures whilst precipitation records (e.g. rivers, lake level) may record no change or arid conditions. Additionally, the spatial distribution of proxy records plays an important role in reconstructions. Southern Africa is a large region with a complex hydrological regime (e.g. Nicholson, 2000; Tyson and Preston-Whyte, 2000).

Therefore, some heterogeneity over glacial timescales is to be expected, especially when the low precision of many dating methods are taken into account. Caution should be taken when extrapolating records over wide areas particularly as apparently contradictory records may simply be reflecting spatial or temporal variation (Nicholson et al., 2012a, 2012b; Thomas and Burrough, 2012).

2.4.6 SUMMARY

Overall, existing palaeoenvironmental records suggest a spatially and temporally complex southern African Quaternary hydroclimate. The observed trends cannot be explained by a single theory but instead suggest a complex interplay of forcing factors modulated by both local and global influences (Chevalier and Chase, 2015; Thomas and Burrough, 2012). This is unsurprising given the current rainfall heterogeneity (e.g. Nicholson, 2000). It is suggested that direct insolation forcing is most significant between 15-20°S (Schefuß et al., 2011; Thomas et al., 2009) whilst SSTs exert a direct control over summer rainfall in eastern South Africa (Chevalier and Chase, 2015; Kristen et al., 2007; Truc et al., 2013) but are moderated by the westerly winds in central and western Southern Africa (Chase and Meadows, 2007; Chevalier and Chase, 2015). However, even this is an oversimplification with relationships dependent on global boundary conditions and subject to change over time and space and as a result of local factors (Chase et al., 2015a; Singarayer and Burrough, 2015)

2.5 THE GEOMORPHIC RECORD

The paucity of biological and geochemical records in southern Africa (and other dryland regions) has necessitated the extensive use of geomorphological analysis to reconstruct past environments ('geoproxies'). These archives, which include aeolian, fluvial, lacustrine and hillslope features, can be independently dated using OSL, are widespread throughout the sub-continent and provide an under-valued, spatially extensive record of landscape change over the Late Quaternary (Thomas and Burrough, 2012). However, by nature, these records are discontinuous and low resolution and can be heavily influenced by a range of local factors, often making them difficult to interpret (Thomas, 2013). Nevertheless, interpretation and understanding are improving (e.g. Bailey and Thomas, 2014; Leighton et al., 2013; Stone and Thomas, 2008; Telfer et al., 2010; Thomas and Burrough, 2012; Thomas and Wiggs, 2008) and with careful consideration these records provide a valuable contribution to our understanding of Late Quaternary environments in southern Africa (Thomas and Burrough, 2012).

2.6 PALAEOENVIRONMENTAL INTERPRETATION OF AEOLIAN OSL DATES

Accumulation histories of geomorphic features are complex with several stages of filtering and aliasing occurring between initial deposition and the interpretation of OSL ages (Bailey and Thomas, 2014 and Figure 2-2). Uncertainties are introduced, and information is lost, at each stage creating the potential for misinterpretation.

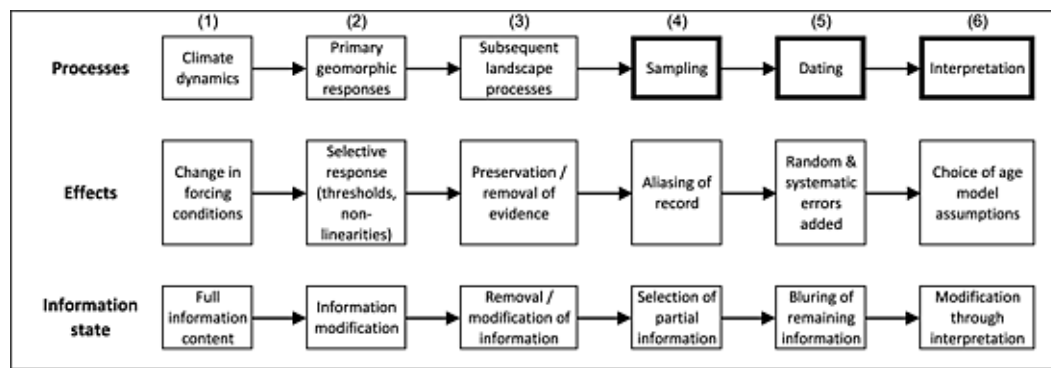


Figure 2-2 synthesis of the 'filtering' that occurs between climatic dynamics and the OSL dated record. At each stage information is lost from the record. From: Bailey and Thomas, 2014

2.6.1 PRIMARY GEOMORPHIC RESPONSE

The sediment state of an aeolian environment is dependent on sediment supply, sediment availability and the transport capacity of the wind (Kocurek and Lancaster, 1999). These are each determined by specified climatic and environmental parameters which interact to influence aeolian geomorphic response.

Sediment supply is simply defined as sediment of a grain-size suitable for aeolian processes. Some sediment is supplied directly through deflation of the strata but the majority of aeolian sediment is supplied from adjacent fluvial, alluvial, lacustrine or coastal sediments. Additionally, internal sediment supply plays an important role in some aeolian systems (i.e. 'cannibalisation' of existing features). Sediment availability is the capability of the surface sediment to be entrained by the wind. This is impacted by grain-size, sorting, surface roughness and cohesion. Therefore, increased moisture decreases sediment availability via increased surface cohesion and via increased vegetation, which increases surface roughness as well as increasing sediment binding. Transport capacity of the wind

is the quantity of sediment able to be entrained if supply is not limited. This is determined by the windspeed and sediment availability (Kocurek and Lancaster, 1999).

Sediment flux into an aeolian system can therefore be limited by supply, availability or transport capacity and sediment influx can be deposited contemporaneously (if transport capacity and availability allow) or, if present conditions are not conducive to aeolian activity, be stored and deposited later. Therefore, the aeolian response may lag behind the geomorphic forcing (Kocurek, 1998).

The aeolian system is non-linear, complex and chaotic (i.e. outputs are not proportional to inputs, the system is characterised by thresholds and is sensitive to minor variations in initial conditions) (Phillips, 2003). Put simply, seemingly identical inputs (climatic/environmental conditions) will not always elicit the same response, as demonstrated numerically by Bailey and Thomas (2014) and empirically by Stone and Thomas (2008). In addition, the system is highly sensitive to current and historical conditions such that the same magnitude of event may have an extreme or negligible effect depending on the conditions that prevailed before (Thomas and Burrough, 2012). Additionally, seemingly small changes may have a large impact if they push the system over a threshold and small intense events may produce a similar geomorphic response to long-term shifts in average conditions.

2.6.2 SUBSEQUENT LANDSCAPE PROCESSES

Subsequent geomorphic processes can alter the preserved record. For free-standing dunes reworking and erosion are common such that the onset (Nanson et al., 1992), end phases (Munyikwa, 2005), or the entirety of aeolian events may be removed. Therefore, there is a preservation bias towards extremely large events and towards young events in the approximate relationship of $\frac{1}{\sqrt{age}}$ (Bailey and Thomas, 2014).

2.6.3 SAMPLE BIAS

By its definition, sampling generates a subset of the information available in the parent population. Due to the non-linearity of aeolian processes the frequency and location of sampling can have a considerable impact on palaeoenvironmental interpretation. Early aeolian studies in southern Africa (e.g. Blümel et al., 1998; Stokes et al., 1997a, 1997b, 1998; Thomas et al., 1997) were based on a limited number of dates, primarily restricted to the top 2 m of sediments in dune bodies. From these, apparently clear patterns of accumulation could be seen. However, as aeolian research in southern Africa has progressed, the production of more dates has reduced the clarity of palaeoenvironmental interpretation with additional ages “filling in” the original gaps in the record and suggesting continuous deposition (Telfer and Thomas, 2007; Thomas and Burrough, 2012, 2016). The location and resolution of sampling has a large impact on the palaeointerpretation based on ages alone. Stone and Thomas (2008) demonstrate this empirically; OSL ages from 3 linear dune cores indicate near continuous deposition at the 0.5

m resolution whilst at 1m resolution apparent patterns start to emerge. These patterns vary significantly depending whether 1 m resolution sampling begins at 0.5 m or 1 m depth, potentially leading to vastly different environmental interpretations. For aeolian studies the regional scale is the most appropriate means to assess environmental patterns as single features are too heavily biased by local factors (Bailey and Thomas, 2014; Stone and Thomas, 2008).

2.6.4 OSL ISSUES

A series of systematic and random uncertainties are added during OSL dating (see Chapter 3). These uncertainties mean that individual paleoenvironmental events can be too short or frequent to be accurately resolved (Bailey and Thomas, 2014; Thomas and Burrough, 2016) with the likelihood of resolving an event decreasing with age (Figure 2-3).

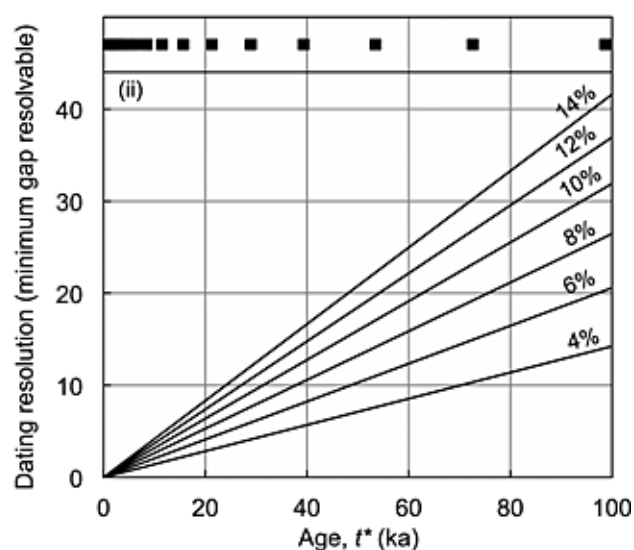


Figure 2-3. Time period needed between two events for them to be accurately resolvable (90% confidence) at different age precisions. E.g. At 40 ka a 5 ka gap is needed for two events to be resolvable at 4% dating precisions, however, if dating precision is only 14% a ~17 ka gap is needed for the gap to be resolved. From Bailey and Thomas (2014).

2.6.5 OVERALL PALAEOENVIRONMENTAL SIGNIFICANCE

Due to the above, interpreting the environmental significance of free-form dunes can be problematic. Internal stratigraphy is often unknown (e.g. Chase and Thomas, 2006; Stone and Thomas, 2008; Telfer and Thomas, 2006) but even when known, the relationship between stratigraphy and chronology is complex (Leighton et al., 2013b). Additionally, the climatic forcing conditions are not straight forward and may be site-dependent (Chase, 2009). Aeolian activity has traditionally been interpreted as a proxy for aridity (e.g. Stokes et al., 1997a, 1997b; Thomas and Shaw, 2002). However, wind strength and sediment supply (which may be associated with increased moisture) are also primary controls (Chase, 2009; Kocurek and Lancaster, 1999), whilst, due to their influence on sediment availability, temperature, vegetation cover, groundwater, seasonality and human influence have all been cited as contributing factors (e.g. Chase, 2009; Goudie et al., 2000; Hesse, 2014; Hesse et al., 2003; Telfer and Hesse, 2013; Wang et al., 2005).

Geoproxy records from within the regions studied in this thesis suggest an interplay of multiple geomorphic processes has occurred during the Quaternary period. Cosmogenic nuclide dating suggest sand has been present, and aeolian processes operating, within the Namib Sand Sea for >1 Ma (Vermeesch et al., 2010, Garzanti et al., 2012). Linear dunes in the north of the Namib Sand Sea suggest ongoing aeolian activity with lateral migration rates of ~0.12 m/yr over the past 2500 years, indicating a complete turnover of sand during the Holocene (Bristow et al., 2005; 2007). Nevertheless, some hiatuses in migration are

recorded in the sedimentary record and are suggested to reflect changes in moisture availability and vegetation cover (Bristow et al., 2007). Water lain deposits from former termini of the Tsondab River indicate increased moisture availability during the late Quaternary relative to the present. Peak flows 65 km and 55 km north west of the present terminus are dated to MIS 5 and ~17-13 ka respectively whilst fluvial deposits indicate that flow reached 6 km west of the present terminus ~13-10 ka (Stone et al., 2010). OSL dating from deposits of the Kumib, Hoarusib, Hoanib, Kuiseb and Tsauchab Rivers present a complex and sometimes contradictory picture of Namibian fluvial activity over the past 40 ka (Srivastava et al., 2004; 2005; 2006; Eitel et al., 2006, Brook et al., 2006: see Stone and Thomas 2013 for review). This may reflect spatial and temporal variations in summer rainfall (Eitel et al., 2005) but may also reflect ambiguity in the interpretation of the fluvial record, as exemplified by the Homeb Silts of the Kuiseb River which have been cited as evidence for both wetter and drier conditions (Stone and Thomas, 2013). In addition, fluvial records from ephemeral dryland rivers can suffer similar preservation bias to the dune record with not all events recorded (Reid, 2009). Dates from former shorelines of Etosha Pan in northern Namibia suggest arid conditions 70-30 ka but that lake levels were >20 m higher than present prior to 10 ka (Hipodonka et al., 2014). However, this is in some disagreement with the stromatolite records which indicate lake levels were 8 m above present at 29 ka before falling and fluctuating from 19 ka onwards (Brook et al., 2011). Shoreline and stromatolite data both point towards fluctuating lacustrine conditions prevailing since 7 ka (Brook et al., 2007; 2011; Hipodonka et al., 2014).

In the region surrounding the Klipkraal Sands, South Africa Late Quaternary geomorphic records are primarily derived from alluvial and colluvial deposits. Near-continuous alluvial records from Modder River suggest fluctuating moisture availability over the past 42 ka possibly linked to changes in the westerly storm belt, or Indian Ocean SSTs (Lyons et al., 2014). Colluvial records suggest that colluviation in the region has occurred for >22 ka (Lyons et al., 2013) and perhaps even for > 135 ka (Clarke et al., 2003) with limited evidence of erosion (Clarke et al., 2003; Lyons et al., 2013; Temme et al., 2008). However, the region is now characterised by badland topography. The cause of this gullying is debated with climatic change associated with the Medieval Climate Anomaly (Lyons et al., 2013, Tooth et al., 2013, McCathy et al., 2011) and human land disturbances (e.g. Kaey-Bright and Boardman, 2006; Boardman et al. 2003; 2014) both suggested as potential causes.

2.7 PALAEOENVIRONMENTAL SIGNIFICANCE OF SLOPE PROCESSES

The main processes operating on slopes can be classed as colluvial or alluvial. In general, colluvial sediments are spatially diffusive deposits, eroded from the hillslope and transported by running water or gravity whereas alluvial deposits are constrained to confined channels associated with water (Fuchs and Lang, 2009). The effectiveness of environmental change on hillslope processes is dependent on the intensity and duration of an event, modulated by the resistivity of the hillslope. The resistivity of the slope to change is dependent on the slope angle, lithology and location/frequency of lines of weakness. As such, prolonged climatic

conditions or very intense shorter term events on high angle slopes or weak lithology illicit the most significant responses. Therefore, hillslopes have been viewed as relatively insensitive recorders of environmental change (Schmidt, 2009). As with the aeolian record the response to forcing events is chaotic and non-linear with events sporadic in both space and time.

Until recently, Quaternary studies on hillslope processes have been limited by dating constraints (Harvey, 2011). Recent advances mean OSL dating has now been successfully applied to multiple hillslope deposits (e.g. Fuchs et al., 2010; Fuchs and Lang, 2009; Kenworthy et al., 2014; Lyons et al., 2014 and references therein). However, the short transport distances and the tendency of material to move as a block, make incomplete signal resetting (partial bleaching) a frequent concern (Fuchs and Lang, 2009). The residual doses can equate to age overestimation up to several thousand years depending on setting (Porat et al., 2001). The multiple phases of transportation and deposition that frequently characterise hillslope deposits increase the likelihood of complete signal resetting but can lead to complex chronologies (Lang and Honscheidt, 1999). Additionally, the poor sorting of sediments means heterogeneity of the beta dose (see section 3.1.3) can be a significant issue (Fuchs and Lang, 2009; Kenworthy et al., 2014).

The palaeoenvironmental significance of alluvial deposits has been shown to be highly site specific and dependent on catchment size, regional climate, lithology and tectonics (Harvey, 2011). Overall, alluvial aggradation has been associated with: a transition to aridity, aridity, a transition from dry to wet conditions, wet

phases and para/periglacial weathering. (see Dorn, 2009 and references therein). Therefore, regional understanding and good chronological control are needed to interpret the hillslope record.

2.8 THE SAND RAMP RECORD

As sand ramps are accumulatory features which are stabilised by a topographic barrier, erosion and reworking are assumed less significant than for the free-standing dune record (Chase, 2009) whilst the varying sedimentary deposits provide additional palaeoenvironmental information. In addition, the aeolian component makes these features suitable for OSL dating. Therefore, if formation controls can be understood, these landforms may provide a valuable geoproxy for southern Africa.

Existing literature on sand ramps is limited, but expanding with 14 published studies and a handful of unpublished PhD, Masters and undergraduate theses referring explicitly to sand ramps (Bateman et al., 2012; Berking and Schütt, 2011; Bertram, 2003; Clemmensen et al., 1997; Ellwein et al., 2015; Kumar et al., 2016; Lancaster and Tchakerian, 1996; Manning, 2015; Pease and Tchakerian, 2003; Rendell and Sheffer, 1996; Telfer et al., 2012, 2014a; Thomas et al., 1997a; Turner and Makhoul, 2002; Tyson, 1999; del Valle et al., 2016)

Section 5.2.1 provides a detailed review of the sand ramp literature. The relatively small body of research, with most studies conducted on single features, means there are still many unanswered questions regarding the sand ramp record. Most

investigations suggest sand ramps are constructed from local sediment sources (Bateman et al., 2012; Bertram, 2003; Clemmensen et al., 1997; Ellwein et al., 2015; Kumar et al., 2016; Telfer et al., 2012, 2014a; del Valle et al., 2016). However, this has only been established at two sites; one in the Mojave, by geochemical fingerprinting (Pease and Tchakerian, 2003) and one by observation in Jordan (Turner and Makhoul, 2002).

The primary outstanding questions regard the climatic, geomorphic and environmental controls on the timing and location of sand ramp accumulation, erosion and preservation. Lancaster and Tchakerian, (1996) suggest that individual sediment units represent discrete periods of average climatic conditions. Stone layers are suggested to represent periods of climatic stability particularly when associated with palaeosols. This is supported by work in the Ladakh Desert, India (Kumar et al., 2016) and observations of episodic deposition at Eivissa, Ibiza (del Valle et al., 2016), Sudan (Berking and Schütt, 2011), South Africa (Telfer et al., 2012) and Black Mesa, USA (Ellwein et al., 2015) although sediment supply is suggested as the controlling factor. In contrast, rapid, opportunistic, deposition is suggested at Solider mountain, Mojave and Ardakan, Iran (Bateman et al., 2012; Thomas et al., 1997a). Bertram (2003) suggest that the accumulation rate of sand ramps varies with size with small sand ramps accumulating in a single phase whilst large sand ramps represent several phases of deposition with palaeosols, calcretes and stone layers representing periods of stability.

2.9 SUMMARY

The above review highlights the complexities of reconstructing southern African palaeoclimate and discusses the gaps in our current understanding. Sand ramps have the potential to provide palaeoenvironmental information which may ‘fill in’ some of these gaps but, presently, the palaeoenvironmental significance of sand ramps has not been tightly constrained. As the majority of previous studies have been conducted on single sand ramps, work is needed to disentangle local and regional controls on accumulation and establish the relationship between accumulation and palaeoclimate. Through a systematic survey of 75 sand ramps and the dating and sediment analysis of 11 features, this thesis aims to better understand the controls on sand ramp formation and to establish what sand ramps can contribute to the southern African climate debate.

Chapter 3: Methods

The choice of methods underpins the results presented in this thesis. This section:

1. Introduces the principles, physics and uncertainties associated with the main applied method; OSL dating.
2. Explains methods used to identify sand ramps and potential field sites, discusses how study sites were selected and explains the sampling strategy.
3. Provides details of the laboratory methods used in the four papers, including justification for the chosen techniques.

3.1 PRINCIPLES OF OSL DATING

The basic physics of luminescence dating has been described and synthesised by Aitken (1985, 1998), Bailey, (2001) and Preusser et al. (2008). Unless otherwise stated this synopsis has been written with reference to these texts.

Luminescence dating provides an estimate of the time elapsed since minerals such as quartz or feldspar were last exposed to heat or light. When sealed from daylight at low or ambient temperatures, charge is trapped at defects within the mineral's crystal lattice, accumulating signal at a rate proportional to background or administered ionising radiation. Exposure to a large source of energy (e.g. heat or light) releases the trapped charge and, given sufficient exposure time, resets the signal to zero (Figure 3-1). Accumulated signal can be measured in the

laboratory and an 'equivalent dose' (D_e) estimated by comparing the natural dose to laboratory doses produced using known quantities of irradiation. Therefore, if the environmental dose (amount of background ionising radiation in the substrate; Dose rate; $\dot{D}$) is known, the time elapsed since signal resetting can be estimated using Equation 3-1.

Equation 3-1:

$$\text{Age(ka)} = \frac{\text{Equivalent dose (Gy)}}{\text{Dose rate (Gy/ka)}}$$

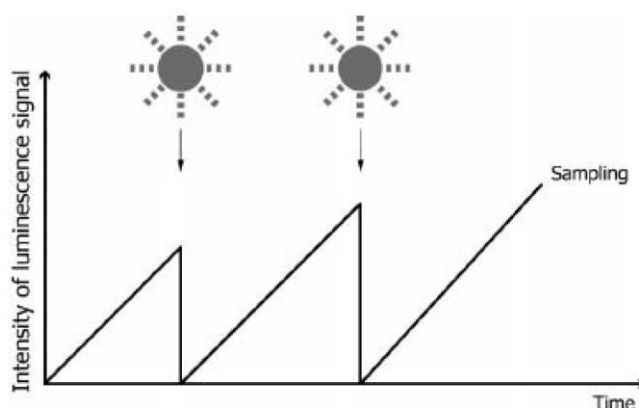


Figure 3-1. Schematic representation of OSL charge accumulation and signal resetting. Charge accumulates when the sample is sealed from daylight at ambient temperatures and is released when the mineral is exposed to sunlight. From Preusser et al. (2008)

3.1.1 PHYSICAL BASIS

The accumulation of charge responsible for producing luminescence is best described using an energy level diagram (Figure 3-2). The crystal lattices of quartz and similar minerals are constructed from a network of positive and negative ions. Defects can occur within these lattices intrinsically or due to impurities or radiation damage and cause an absence or excess of positive or negative ions depending on type (Preusser et al., 2009).

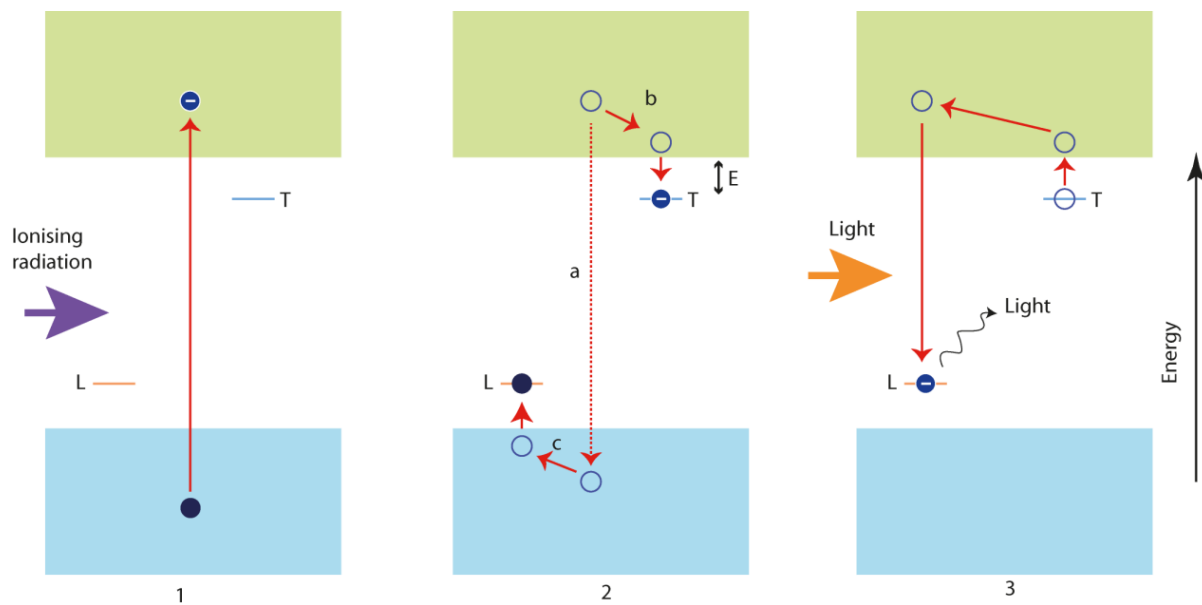
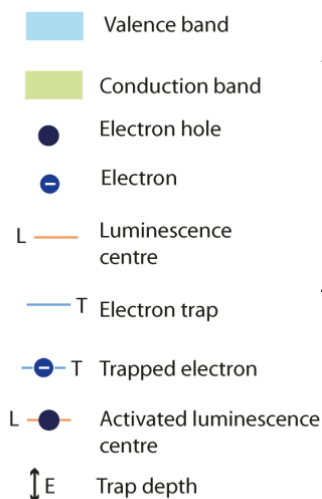


Figure 3-2. Schematic representation of the OSL process at the energy level.



1. Crystal is exposed to ionising radiation. This input of energy causes an electron to be excited into the conduction band. The luminescence centre and the electron trap represent defects in the crystal lattice. The electron trap is attractive to electrons (i.e. has a weak positive charge) whereas luminescence centres need to contain an electron hole to become active (a process which results from exposure to ionising radiation).
2. The after effects of ionisation. a. most electrons will fall back into the valence band. b. Some electrons become trapped in electron traps and are stored. The depth of the trap (E) determines how stable it is. Deeper traps can withstand more natural crystal lattice vibration and thus effectively retain electrons for millions of years. Only traps of 1.6eV or more are stable enough for OSL dating. c. Demonstrates how luminescence centres become active (contain a hole) and attractive to free electrons. Luminescence centres are presumed abundant with numbers increasing with increased ionisation.
3. When the mineral is exposed to light trapped electrons are given enough energy to become excited out of their traps and into the conduction band. They then diffuse around the crystal until a recombination centre is found. If this happens to be a luminescence centre then the energy is released as light. The amount of luminescence given off by a mineral is directly proportional to the number of trapped electrons and thus the amount of ionising radiation the mineral has received. (After Aitkin, 1985, 1998, Preusser et al. 2008).

When stimulated with ionising radiation, electrons detach from their parent ions and become excited to higher energy states (Figure 3-2). Most electrons return to the valence band immediately but a small proportion become trapped below the

conduction band in ‘electron traps’ caused by a negative ion vacancy (Figure 3-2). Once trapped, electrons can only be released due to vibrations within the crystal lattice. The ‘depth’ of the electron trap below the conduction band determines its stability. The average time an electron is expected to spend in a trap can be described by Equation 3-2; where s is the ‘frequency factor’ i.e. the rate of detrapping if activation energy was zero, E_t is the thermal activation energy (i.e. trap depth; eV), k_B is the Boltzmann’s constant (8.614×10^{-5} eV/K) and T is the absolute temperature (Bailey et al., 1997).

Equation 3-2:

$$\tau(s) = s^{-1} \exp\left(\frac{E_t}{k_B T}\right)$$

Only traps with activation energies ~ 1.6 eV or more are suitable for luminescence dating. At this ‘depth’ traps are stable over millions of years, and therefore unlikely to ‘leak’ charge over Quaternary timescales (Preusser et al., 2008).

Once evicted from the trap an electron can either be re-trapped, in the same or different electron trap, or can recombine with an ‘electron hole’/ ‘recombination centre’ (Figure 3-2). The quantity of emitted light is determined by the number of electrons which recombine at ‘luminescence centres’ (recombination centres which emit light). This is directly proportional to the number of trapped electrons which in turn is proportional to the quantity of ionising radiation the crystal has been exposed to since the signal was last reset. However, as the number of electron traps is finite, this relationship is not linear and emitted luminescence tends towards saturation.

The Quartz OSL signal

Both quartz and feldspar can be used for OSL dating but only quartz is used in this thesis. Quartz bleaches more rapidly than feldspar, taking ~10 s to deplete the signal to ~1% compared to ~1 hr for feldspar (Godfrey-Smith et al., 1988) and, unlike feldspar, quartz does not suffer from ‘anomalous fading’ (an unaccounted for loss of charge over laboratory timescales) (Spooner, 1994a, 1994b). However, quartz typically saturates at lower doses and produces a ‘dimmer’ signal (fewer photon counts per given dose) than feldspar. Therefore, the development of measurement techniques which isolate non-fading feldspar signals is rapidly advancing (e.g. Buylaert et al., 2012; Fu and Li, 2013; Li and Li, 2012; Thomsen et al., 2011). Nevertheless, quartz has been the dominant material utilised in luminescence dating for the past 20 years and the physical processes and potential problems are well understood (Duller, 2004; Roberts et al., 2015; Wintle, 2008).

Quartz OSL Components

The decay of quartz OSL with stimulation time cannot be described by a single decaying exponential as would be expected if the signal arose linearly from a single electron trap (Huntley et al., 1985). Using mathematical deconvolution Bailey et al. (1997) identified 3 components which, based on their rate of signal decay, are termed the ‘fast’, ‘medium’ and ‘slow’ components (Bailey et al., 1997). Subsequent research has identified at least 3 separate ‘slow’ (S) components (Jain et al., 2003; Schilles et al., 2001; Singarayer and Bailey, 2003) and an additional ‘ultrafast’ component (Jain et al., 2003, 2008). Each of the separate

components has its own unique thermal and optical properties, indicating the presence of unique traps (Durcan and Duller, 2011; Jain et al., 2003; Singarayer and Bailey, 2003)

The fast component is preferable for dating as it has a long lifetime (850 Ma at 20°C; Wintle and Murray, 1998), is thermally stable, readily bleachable and is relatively quick to decay (Jain et al., 2003; Singarayer and Bailey, 2003). Fast-dominated signals generally produce accurate age estimates (Bailey, 2010) with the ubiquitously applied Single Aliquot Regenerative (SAR) protocol (Murray and Wintle, 2000, 2003, section 3.3.6) developed and tested on fast dominated samples. Fast-only OSL can be directly measured by Infrared stimulated luminescence (IRSL) at high temperatures (Bailey, 2010; Jain et al., 2003; Singarayer and Bailey, 2003) but this requires long measurement times and can only be successfully applied to very bright samples making it impractical for routine use (Bailey, 2010).

The ultrafast component and the S2 (S3 of Jain et al., 2003) are unstable and therefore unsuitable for Quaternary dating (Jain et al., 2003; Singarayer and Bailey, 2003). In most cases this is not a significant problem as they can be removed from the dated signal by stringent preheating (240-280°C) (Jain et al., 2003; Singarayer and Bailey, 2003). Thermal transfer (movement of charge from light insensitive to light sensitive traps during heating) can be a problem for the medium and slow components (Jain et al., 2003) and significant age underestimations have been demonstrated for samples with a dominant, thermally

unstable medium component (Steffen et al., 2009). Additionally, medium and slow components may not sensitise in the same way as the fast component making the SAR protocol unsuitable for non-fast dominated samples (Jain et al., 2003).

The relative contribution from different components can be calculated mathematically using curve deconvolution (e.g. Choi et al., 2006) but individually deconvolving the decay curves of all measured aliquots is time consuming and labour-intensive. Steffen et al. (2009) suggest non-fast dominated signals can be identified by 'lazy' decay curves, non-distinct 110°C thermoluminescence (TL) peaks, NR(t) (natural to regenerated signal over time) plots or $D_e(t)$ plots (D_e as a function of increasing integration limits; Bailey, 2003a, 2003b; Bailey et al., 2003). However, these methods are either non-quantifiable, relatively timing-consuming and/or ambiguous. The application of an early background method can increase the relative contribution from the fast component (Ballarini et al., 2007; Cunningham and Wallinga, 2010) but is associated with a loss of precision in all but the brightest samples.

The Fast-Ratio (Durcan and Duller, 2011, Equation 3-3) provides a simple numerical method to determine whether a sample is fast-dominated. Samples with fast ratios >20 are considered fast-dominated (~90% of the initial signal is likely to originate from the fast component). This method shows good agreement with the results of component fitting but can produce erroneous results for very dim samples.

Equation 3-3:

$$\text{Fast Ratio} = \frac{L1 - L3}{L2 - L3}$$

L1 = signal from the first channel and represents the greatest dominance of the fast component. **L2** = photon count from the channel after 75 mJ.cm⁻² of cumulative stimulation power and represents the portion of the signal where the medium component dominates. **L3** = mean photon counts between 455 and 683 mJ.cm⁻² and represents the slow component. For this calculation a channel represents ~5 mJ.cm⁻² of stimulation.

3.1.2 THE ENVIRONMENTAL DOSE

The measurement of the annual environmental dose (dose rate) is of equal importance to the palaeodose but whilst the precision and accuracy of D_e determination has been dramatically improved over the past 30 years (see Duller, 2004; Roberts et al., 2015; Wintle, 2008 for reviews), relatively little attention has been paid to dose rate calculations and considerable uncertainties still exist (e.g. (Murray et al., 2015)).

Basic Principle

The annual dose is delivered to the sample via the alpha, beta and gamma radiation emitted by the radioactive decay of Thorium (Th), Uranium (U) and Potassium (K) (and to some extent rubidium) in the substrate. A small contribution is also made by cosmic rays. Therefore, calculation of the dose rate can be

summarised by Equation 3-4 where D_α = environmental dose from alpha radiation, D_β = environmental dose from beta radiation, D_γ = environmental dose from gamma radiation, D_{cosmic} = environmental dose from cosmic radiation.

Equation 3-4:
$$D_r = D_\alpha + D_\beta + D_\gamma + D_{cosmic}$$

The calculations involved in dose rate determination are not complex but involve several stages and the choice of different published conversion and attenuation values. Figure 3-3 summarises the process.

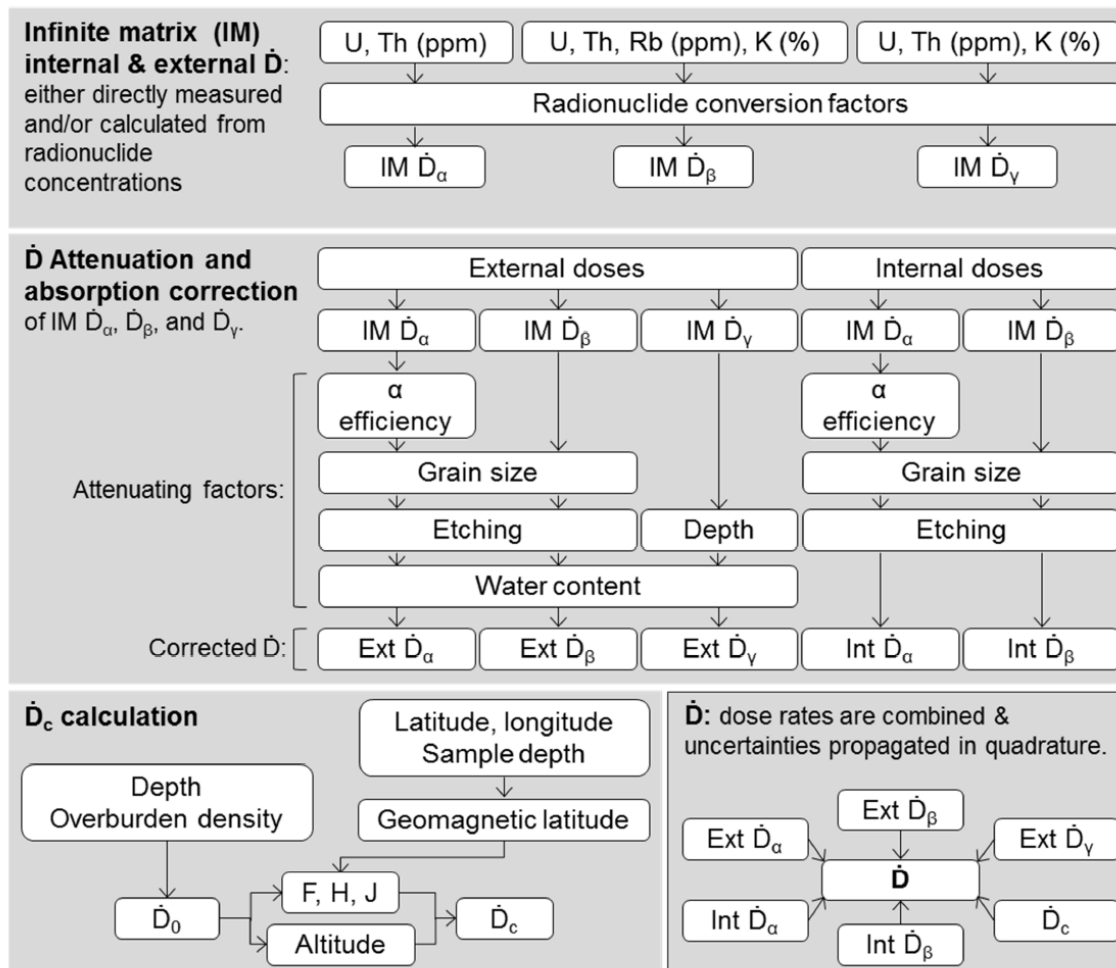


Figure 3-3. Stages involved in calculating the environmental dose. IM = infinite matrix, ext = external, int. = internal From Durcan et al. (2015)

Based on the infinite matrix assumption the alpha, beta and/or gamma contribution can be assessed by measuring emissions directly from the sediment or, assuming constant concentrations over Quaternary timescales, can be calculated from the concentration of radionuclides (^{40}K , ^{232}Th , ^{238}U) in the substrate.

Conversion factors

Conversion factors translate radionuclide concentrations into emitted β and γ dose rates. The values are calculated from complex nuclear data tables which undergo frequent refinement (Adamiec and Aitken, 1998). The most recent values, and the only ones to include errors, have been published by Liritzis et al. (2013). Despite 40 years of updates, conversion factors show less than 8% variation and there is no difference in calculated dose rates based on the historical mean and current values (Liritzis et al., 2013).

Attenuation

α -particles (two neutrons, two protons) and β -particles (fast-moving electron) interact readily with material as they pass through, resulting in energy loss (attenuation of charge). α -particles and β -particles have ranges in quartz of 20-30 μm and $\sim 3\text{mm}$ respectively. Therefore, α - and β -attenuation produce a heterogeneous distribution of ionising radiation within and between grains that needs to be corrected for. As γ -rays (photons) can penetrate $\sim 30\text{ cm}$ of sediment this distribution of radiation is assumed homogenous.

Based on the self-dose principle (how much radiation is lost when emitted from the centre of an infinite quartz grain), β -attenuation factors (e.g. Bell, 1979; Brennan, 2003; Fain et al., 1999; Guérin et al., 2012; Mejdahl, 1979) allow the β -dose received by grain of different sizes to be estimated. However, the self-dose approach has the potential to underestimate the absorbed dose by up to 15%, 30% and 45% for ^{40}K , ^{238}U and ^{232}Th respectively when quartz is the sole absorber. In reality, this underestimation is often compensated for by the self-dose properties of K-feldspars (Guérin et al., 2012). Despite concerns, grain shape has been shown to have a negligible effect on absorbed β -dose at 200 μm (Fain et al., 1999; Nathan, 2010)

Etching, routinely performed when isolating quartz grains for dating, removes the outer 8-10 μm rind of quartz grains. Therefore, alpha contribution is assumed negligible and removed from dose rate calculations in most studies. Additionally, β -attenuation adjustments need to be made as the most highly irradiated portion of the grain is removed. Etched attenuation values have been calculated by Nathan (2010), Brennan (2003), Fain et al. (1999), Bell (1979) and Mejdahl (1979).

Moisture content

Water both dilutes the concentration of radionuclides and absorbs radiation more readily than quartz (Guérin et al., 2012). Therefore, the higher the moisture content, the lower the annual dose absorbed by quartz, irrespective of radionuclide concentrations (Aitken and Xie, 1990). Water content is corrected for

by using Equation 3-5, where D_w is the wet dose, D_d is the dry dose rate, x is the factor used to correct for the difference between energy absorption in water and in quartz, W is saturation water content (weight of water/dry weight) and F is the fraction of the pore space occupied by water. WF can be simplified to the percentage of time the substrate was saturated.

Equation 3-5:

$$D_w = \frac{D_d}{1 + xWF}$$

The x values of the different radionuclides are within <2% of each other. Therefore, mean values, calculated by Zimmerman (1971), of 1.25 for beta and 1.14 for gamma are typically used (Aitken and Xie, 1990; Guérin and Mercier, 2012; Zimmerman, 1971). Recalculations of the beta (Aitken and Xie, 1990; Nathan and Mauz, 2008) and gamma (Aitken and Xie, 1990; Guérin and Mercier, 2012; Nathan and Mauz, 2008) x values have been conducted. Gamma results are similar, apart from Nathan and Mauz, (2008), but beta is reduced by 5% (Aitken and Xie, 1990; Nathan and Mauz, 2008). For carbonate sediments Nathan and Mauz, (2008) suggest the use of updated x factors to account for the reduction in pore space available for water over time.

Given its impact on absorbed dose and the impossibility of accurately reconstructing the concentration of water in the substrate over Quaternary time scales, moisture content provides one of the greatest uncertainties in dose rate (and age) calculation. Water contents are usually based on present moisture values or are an educated guess based on the environment.

Choice of conversion and attenuation factors

The choice of conversion, attenuation and water values can cumulatively cause up to ~3% change in dose rate (Figure 3-4). The choice of attenuation factor has a negligible impact on overall dose rate. Different water attenuation values (x values) create <0.5% dose rate variation. Choice of beta etch attenuation factors can cause 1% variation whilst choice of conversion factors can lead to ~1.6% variation. Within the scope of dose rate uncertainties these variations are small. Nevertheless, they are worth considering when comparing to other studies. The combination of Liritzis et al. (2013) conversion factors, Bell (1979) etch attenuation, Guerin et al. (2012) beta attenuation and Aitkin and Xie (1990) water values used in this study (see section 3.3.13) are in good agreement with the average dose rate produced from the various combination of factors; 5.55 Gy/ka and 5.56 Gy/ka respectively for this example.

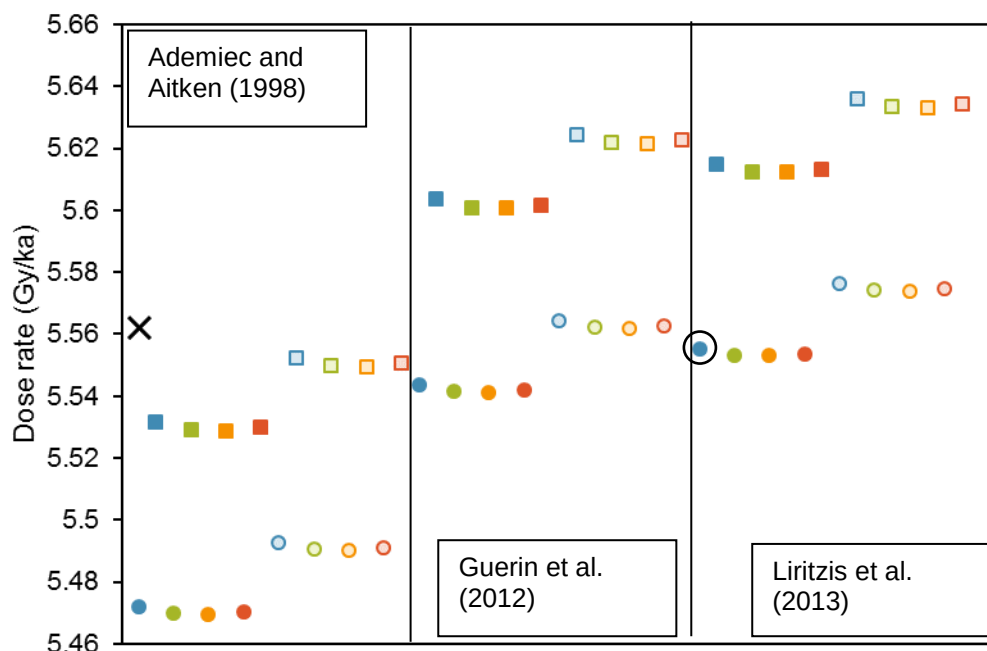


Figure 3-4. Dose rate of an exemplary sample using different calculation values. Conversion factors are indicated by position on the graph. Colours represent attenuation factors: Blue = Guerin et al., (2012), Green = Nathan (2011), Orange = Mejahl (1979), Red = Brennan (2003). Shapes represent etch depth attenuation: Circles = Bell (1979), Squares = Brennan (2003). Shade represents water attenuation factors: Bold = Aitken and Xie (1990), Light = Nathan and Mauz (2008). The black cross is the average of all values. The black circle marks factors used in this study.

The cosmic dose

The cosmic dose is formed of two parts: the hard and soft components. The hard component consists of muons (cosmic particles which interact with the atmosphere), can penetrate up to 6 m of rock and provides the majority of the cosmic dose. The soft component is a secondary shower of electrons which can only penetrate the top 60 cm of the substrate (Prescott and Hutton, 1994). The concentration of cosmic rays increases with altitude and at high latitudes (Prescott and Hutton, 1988). Therefore, sample depth, altitude and latitude are considered during calculation (Prescott and Hutton, 1994). The production of cosmic rays has varied through time due to changes in the strength of Earth's magnetic field and changes in solar output. However, except for high altitude equatorial sites, with low β and γ doses, the impact of this is negligible (Prescott and Hutton, 1994). Additionally, the overburden history will impact the total cosmic dose received by samples but this only has a significant impact on total dose rates at sites with low β and γ contributions. In these circumstances overburden modelling is suggested (e.g. Burrough et al., 2007; Telfer and Thomas, 2007).

3.1.3 POTENTIAL OSL DATING PROBLEMS

Alongside measurement errors which are typically quantified at 2% (Duller, 2007) several factors can cause age inaccuracies or uncertainties and careful consideration is needed to ensure ages are representative.

Disequilibrium

^{238}U and ^{232}Th are the parent atoms in radioactive decay chains consisting of several daughter products. Dose rates calculated from the concentration of ^{238}U and ^{232}Th take into account the radioactivity of all of the daughter products, assuming they are in secular equilibrium (i.e. the production of the daughters is equal to their decay, and their concentration is proportional to the concentration of the parent). However, in natural open systems atoms can be gained or lost during the burial history of a sample, causing disequilibrium and resultant errors in dose rate calculation. Due to the short half-lives of the daughter products Th-series disequilibrium is not considered significant over Quaternary timescales (Olley et al., 1997). However, significant disequilibrium in the U-series is well documented (e.g. Krbetschek et al., 1994; Lahaye et al., 2012; Molodkov, 2007; Olley et al., 1997, 1996; Prescott and Hutton, 1995; Rambaud et al., 2000; Zander et al., 2007).

The most common causes of U-series disequilibrium are the outgassing of radon gas (^{222}Rn) and the solubility and subsequent mobility of ^{234}U (Krbetschek et al., 1994). Therefore, U-series disequilibrium is most common in water-lain sediments (Krbetschek et al., 1994; Olley et al., 1996, 1997; Prescott and Hutton, 1995).

A rough indication of disequilibrium can be obtained by comparing uranium and thorium concentrations derived from in situ gamma spectrometry, which measures the concentrations of late daughters ^{214}Bi and ^{208}Pb , and concentrations derived from inductive coupled plasma mass spectrometry (ICP-MS) which measures the

concentration of the parents ^{238}U and ^{232}Th (Prescott and Hutton, 1995). However, complete identification and characterisation of disequilibrium requires lengthy measurements using a combination of alpha spectrometry and high-resolution, low level gamma spectrometry (Guibert et al., 2009; Krbetschek et al., 1994; Lahaye et al., 2012; Olley et al., 1996; Preusser and Kasper, 2001).

Correcting for disequilibrium is problematic as current radionuclide ratios cannot be used to hindcast geological history. Nevertheless, some attempts have been made (e.g. Guibert et al., 2009; Lahaye et al., 2012; Zander et al., 2007).

Beta-dose Heterogeneity

It is currently only possible to measure the average environmental dose. However, due to the short range (~ 3 mm) of β -particles the dose received by individual grains is strongly dependent on the distance from the emitter. This can cause considerable variation in the beta dose, depending on the concentration and distribution of radionuclides and the grain-size, sorting and composition of the sediment (Chauhan and Singhvi, 2011; Cunningham et al., 2011; Guérin et al., 2015; Mayya et al., 2006).

Due to its high concentration in sand sized feldspar grains (only 7% total feldspar is needed to produce 1% ^{40}K ; Mayya et al., 2006) and high beta emitting capacity, ^{40}K is typically considered the main, and most studied, cause of beta-dose heterogeneity (Chauhan and Singhvi, 2011; Cunningham et al., 2011; Guérin et al., 2015; Mayya et al., 2006). Beta heterogeneity produces a positive skew in D_e

distribution of individual grains (Cunningham et al., 2011; Guérin et al., 2015; Mayya et al., 2006; Nathan et al., 2003). The degree of skew increases with decreased ^{40}K content and increased grain-size as both these factors increase the range of distances from hotspots (areas of high beta radiation) (Guérin et al., 2015; Mayya et al., 2006). 0.1-4% of ^{40}K concentrations have been shown to produce 46-18% relative standard deviation in single grains respectively, therefore contributing 18-25% of total variation for well-bleached samples (Guérin et al., 2015; Kalchgruber et al., 2003; Mayya et al., 2006). Beta heterogeneity is also more pronounced for samples with low dose rates ($<1\text{Gy/ka}$) as the relative variation between doses is greater (Guérin et al., 2015). Carbonate has been demonstrated to cause 'coldspots' leading to a negative skew in dose rates (Guérin et al., 2015; Kalchgruber et al., 2003) whilst ^{232}Th and ^{238}U are also likely to contribute to beta heterogeneity to some degree (Chauhan and Singhvi, 2011).

Beta heterogeneity is less significant for single aliquot studies due to averaging effects. Chauhan and Singhvi, (2011) suggest only 24-35 aliquots are needed to estimate the mean palaeodose to within 5% of the true value when ^{40}K concentration are 1-3%. The higher the ^{40}K concentration, the smaller the number of aliquots needed to overcome the impact of beta heterogeneity.

Partial bleaching

Partial bleaching (incomplete resetting of the OSL signal before burial) can cause significant scatter in D_e values (Duller, 2008; Olley et al., 1999; Wallinga, 2002). Partial bleaching occurs in environments where not all grains are exposed to

sunlight for sufficient time to reset the luminescence signal. As such, it is most commonly observed in fluvial samples (e.g. Olley et al., 2004, 1998; Stokes et al., 2001) but can also be significant in hillslope deposits (e.g. Fuchs and Lang, 2009; Kenworthy et al., 2014; Spencer and Robinson, 2008) or aeolian samples which have been transported at night or during dust storms (Fan, 2013; Olley et al., 2004). In most circumstances, partial bleaching is rare in aeolian samples (Duller, 2008; King et al., 2014). As residual doses are typically small (Fuchs et al., 2005; Jain et al., 2004; Singarayer et al., 2005) the impact of partial bleaching is most significant for young samples (<1 ka) (Cunningham et al., 2011; Jain et al., 2004). Techniques for detecting partial bleaching include analysis of single grain D_e distributions (Olley et al., 1998, 2004) comparison of feldspar and quartz age estimates (Fuchs et al., 2005; Murray et al., 2012), and $D_e(t)$ plots (Bailey, 2003a, 2003b; Bailey et al., 2003a). Statistical techniques (e.g. minimum age model (MAM; Galbraith et al., 1999), Internal-External Consistency Criterion (IEU; Thomsen et al., 2003, 2007) and the lowest 5% model (Olley et al., 1998) can be used to reduce the impact of partial bleaching and produce age estimates in line with independent age control (e.g. Medialdea et al., 2014; Olley et al., 2004).

Post depositional mixing

Post-depositional mixing in the form of bioturbation (disturbance of the sediment by plants and animals) or geomorphic processes (e.g. sediment slumping into gullies) can also cause significant D_e scatter. Bioturbation can cause both upwards and downwards movement of grains creating both positive and negative skew in D_e values (Bateman et al., 2007b). Ants and termites can burrow up to 4

m (Tschinkel, 2003) and are predominantly associated with the upward movement of grains (Kristensen et al., 2015; Rink et al., 2013) although up to 20% downward mixing has been shown (Kristensen et al., 2015). Pedoturbation and plant growth primarily causes downward mixing in the top 30-60 cm (Gliganic et al., 2016; Johnson et al., 2014) and can produce complete overturning over millennial time scales (Gliganic et al., 2016). Bioturbation can be difficult to identify in the field and ages from bioturbated sediments frequently present in stratigraphic order making overdispersion due to bioturbation difficult to diagnose (Bateman et al., 2003a, 2007a, 2007b). However, bleached (zero-dosed) grains are easily identified in single grain (SG) dating and large differences between multigrain single aliquot (SA) and SG ages may also indicate bioturbation (Bateman et al., 2003b, 2007b)

Contamination

Due to variations in the mineral properties, not all quartz grains are suitable for dating (Adamiec, 2000; Duller, 2008, 2012; Thomsen et al., 2005). Significant contribution from unsuitable grains can skew SA ages. This is explored in detail in Chapter 4. Additionally, contribution from components other than the fast component can cause considerable inaccuracies (e.g. Steffen et al., 2009).

3.2 SITE AND SAMPLE SELECTION

3.2.1 SAND RAMP IDENTIFICATION AND SITE SELECTION

Sand ramp identification and site selection was an iterative process. Google Earth™ satellite imagery was used initially to identify potential sand ramps in southern Africa. Firstly, the aerial morphology of known sand ramps (Bateman et al., 2012; Bertram, 2003; Lancaster and Tchakerian, 1996; Telfer et al., 2012; Thomas et al., 1997a) was studied for reference. Bertram (2003) reports abundant sand ramps in south west Namibia. Therefore, identification of potential sand ramps and study sites was first conducted in this area (26°05' - 26°45'S and 16°05' to 16°40' E). Once regional sand ramp geomorphology was better understood, the survey was expanded to cover much of Namibia and South Africa. Regions were spilt into sections of roughly 500 km². These sub-regions were scanned at a resolution of 50-100 km². Potential sand ramps were identified firstly by recognising potential accommodation spaces (frequently isolated inselbergs) and secondly by searching for accumulations of sand against these mountains. All potential sand ramps were marked on Google Earth™ and recorded in a database. Investigations in western and central South Africa also included newspaper advertisements to land owners. However, these were largely fruitless and due to the complex topography and extensive vegetation cover in the region, sand ramps were difficult to identify from aerial imagery alone. Therefore, only one South African sand ramp is studied in this thesis.

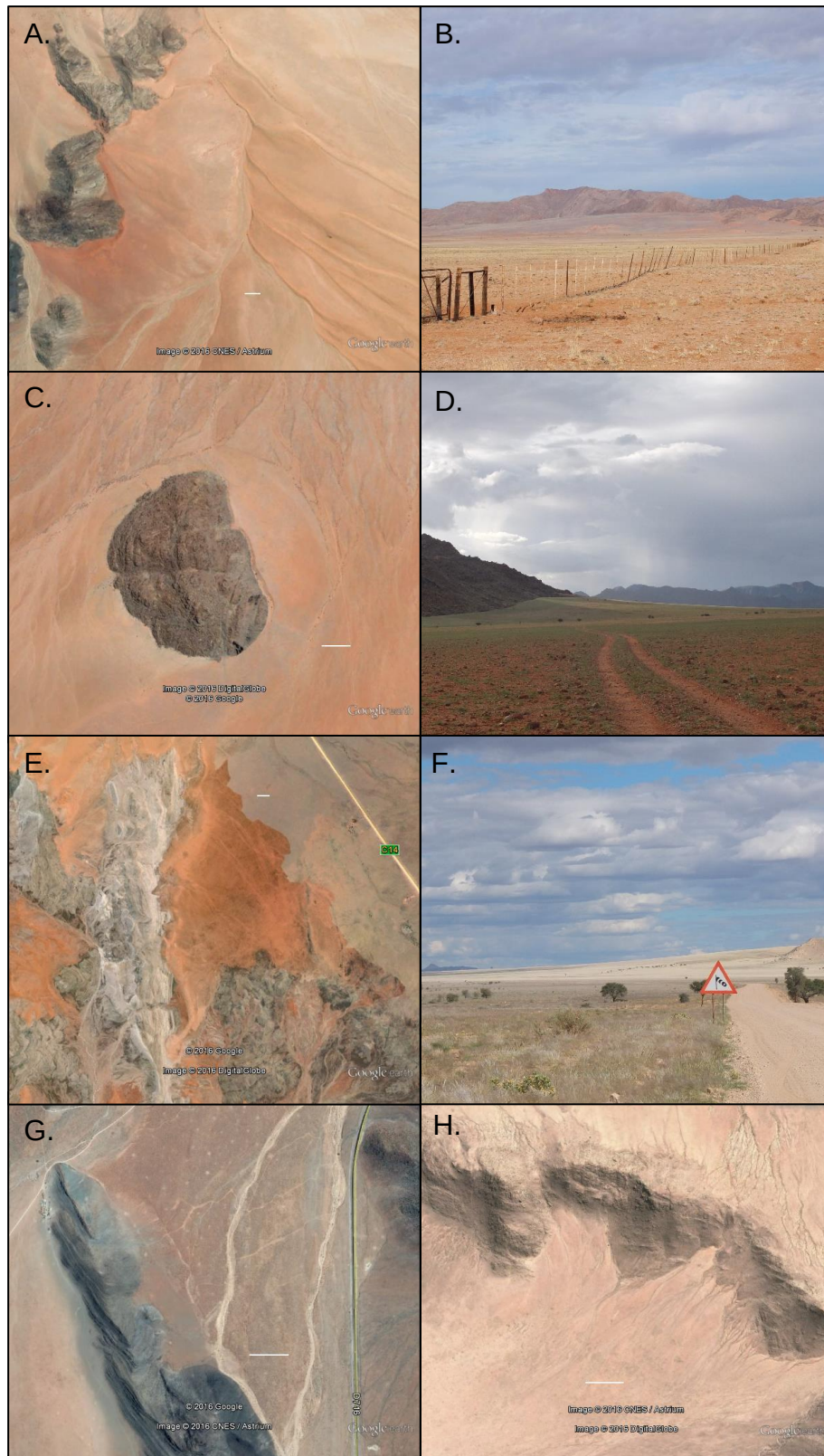


Figure 3-5. Examples of sand ramps identified using Google Earth™ and confirmed in the field (A-D; A and C are Google Earth™ images, B and D are the respective features from the ground). Example of a sand ramp identified in the field but not initially found using Google Earth™ (E & F; E is the Google Earth™ image, F is the field view). Examples of features identified using Google Earth™ but found to be thin sand drapes (G & H).

Fieldwork was conducted in two seasons: April 2013 and April 2014. During the first field season the previously studied Aus sand ramps (Bertram, 2003) and the Klipkraal Sands (Marker and Holmes, 1993; Thomas et al., 2002) were sampled and logged. Potential sand ramps in south west Namibia were assessed at ground level. During this field trip several potential sand ramps were shown to be thin sand drapes whilst some previously overlooked features were found to be large sand ramps from the ground (Figure 3-5). Some additional potential features were investigated in central South Africa but these were found not to be sand ramps.

Based on the information gained in the field in 2013 the region 23°-28°S and 15°-19°E was focused on for the second field season. The sand ramp database was updated, removing unlikely sand ramps, and those outside the region, and identifying more potential study sites. Sand ramp identification within 23°-28°S and 15°-19°E was relatively straightforward due to the simple geomorphology of inselbergs within a plain whilst low-vegetation cover made sand accumulations easy to identify. During the second field trip nine additional Namibian sand ramps were sampled and the revised database was ground-truthed leading to further updates. Gullies were a good predictor of sand ramp accumulations but the relationship was not perfect. Overall, ~70% of the final sand ramp database were verified in the field. Those that were not verified had clear gullies and morphology similar to confirmed sand ramps.

Sampled sand ramps were chosen to reflect a range of sizes, sediment compositions and morphologies, although selection was also tempered by accessibility with five previously selected sand ramps not studied due to access problems. Apart from Neuhof-1 the studied sand ramps were gullied, providing natural exposures to examine the sediments. The location of the studied Namibian sand ramps and their schematic representation is displayed in Figure 5-4.

3.2.2 SAMPLING STRATEGY

The sampling strategy was modified at each site depending on the sand ramp. Where available, samples were taken from natural gully exposures. In these instances, the face was cleaned and sediments were examined and logged. Samples were taken to bracket changes in sedimentary features (e.g. stone layers) or, when sediments were relatively homogenous, at a resolution of ~1 m. However, for large sand ramps, with reduced accessibility (e.g. Samara) sampling resolution was reduced. At Neuhof-1 pits were hand dug to the stable depth of ~2 m to allow sediments to be examined. Samples were taken bracketing stone layers or at a ~0.5 m resolution. When sand ramps had associated overlying dunes these were also sampled to assess variation across the feature. Sampling was conducted to a maximum depth of 5 m using a Dormer™ portable auger adapted for OSL sampling. Samples were taken at ~1 m. The depth of sampling was limited by poor sand cohesion, with only the top ~2 m sampled at most sites.

Sampling was conducted using opaque black tubes to avoid light exposure. Once back in the lab, the potentially light exposed ends were removed and used for sediment analysis whilst the central portion was used for OSL dating. Not all extracted samples were dated; at Aus, Sandkop, Samara and Japkop the old age and low precision of the measured samples indicated no additional chronological information could be derived from the remaining samples.

The morphology of all sampled sand ramps was mapped in the field. Google Earth™ was used to record sand ramp orientation, size and shape, the presence of dunes and gullies and the proximity and direction to potential sediment sources (river channels, dunes) for all sand ramps in the database.

3.3 LABORATORY ANALYSIS

3.3.1 SEDIMENT ANALYSIS

A small number of sediment analyses were undertaken to provide additional context and some palaeoenvironmental insights (e.g. heterogeneity, likely depositional mode, environmental stability) for the samples taken for OSL.

The Munsell colour of samples was derived from air-dried sediment. Organic and carbonate content was calculated using sequential loss-on-ignition (LOI; Heiri et al., 2001). Percentage organic content is calculated using Equation 3-6 and Equation 3-7 multiplied by 1.36 represents the proportional weight of carbonate (Bengtsson and Enell, 1986; Heiri et al., 2001).

Equation 3-6:

$$LOI_{550} = \frac{DW_{105} - DW_{550}}{DW_{105}} * 100$$

Equation 3-7:

$$LOI_{950} = \frac{DW_{550} - DW_{950}}{DW_{105}} * 100$$

DW₁₀₅ is the dry weight of the sample after ~12 hours at 105°C, **DW₅₅₀** is the dry weight of the sample after 4 hours combustion at 550°C and **DW₉₅₀** is the dry sample weight after 2 hours combustion at 950°C. LOI provides a good approximation of organic and carbonate content but can be influenced by laboratory conditions (Ghabbour et al., 2014; Heiri et al., 2001).

Grain-size analyses were conducted on a Malvern Mastersizer2000 laser granulometer. Statistics were calculated using the GRADISTAT program (Blott and Pye, 2001). Tests were conducted to assess the most appropriate method for grain-size analysis (Section 3.3.2). Following this wet dispersion with no pre-treatment was selected as the most representative and time-efficient method.

3.3.2 COMPARISON OF DIFFERENT GRAIN-SIZE TECHNIQUES

The results of wet and dry dispersions with no pre-treatment were compared across 13 samples of varying sedimentary properties. Six of these samples were also measured using wet dispersion with pre-treatments to remove salts, carbonates and organics. Details of the methods are displayed in Table 3-1.

Table 3-1. Different tested laser granulometry methods

Technique	Method
Dry dispersion	Sample sieved to <2 mm, lightly crushed to remove aggregates and riffled to the appropriate quantity. Measured using a dry dispersion unit.
Wet dispersion (no pre-treatment)	As above but with the addition of 20 s of ultrasonic dissolution and measurement was conducted using a wet dispersion unit.
Wet dispersion (salts and carbonates removed)	As above but prior to measurement the salts were removed in solution until electric conductivity was 1 dS m ⁻¹ or less and a treatment of 1M sodium acetate and 1M sodium hydroxide was used to remove carbonates. Samples were dispersed using a solution of sodium hexametaphosphate.
Wet dispersion (salts, carbonates and organics removed)	As above but prior to measurement samples were heated to 400°C for 4 hours to remove organics

Dry unit measurements had a higher mean and median grain-size and contained proportionally more sand and fewer fines (Figure 3-6). Subsequently, sorting was generally better than samples measured using the wet unit. These differences are likely to be the result of aggregates not being dispersed. This is most pronounced in KAR13/3/1. Microscopy shows this sample to contain multiple loosely aggregated 'peds' of finer material which are likely to be post-depositional. Additionally, a significant proportion of fine material was left behind on the unit and not measured. Therefore, dry dispersion is not thought to fully represent the source material.

Results obtained using the various wet preparation techniques yielded very similar results. Mean and median grain-size show a marginal increase after pre-treatment (Figure 3-6B). The proportion of clay increased from 1.8% to 4.6% after the removal of salts and carbonate whilst the proportion of coarse silts decreased

from 21% to 10% (Figure 3-6A). These results most likely reflect the increased likelihood of material being accidentally removed by pouring during pre-treatment and has been observed by others (e.g. Leighton, 2013). Alternatively, the decrease in silts and the increase in clays may reflect the breakdown of aggregates. However, this would be expected to lead to a reduction in mean and median grain-sizes. The removal of organics led to a decrease in clay and an increase in sand, probably as a result of clays being “fired” during ashing. As initial conductivity measurements suggest relatively low salt measurements and LOI results suggest all samples contain <1% carbonate and <4% organics (Table 5-3 and Table 6-1) wet-dispersion with no pre-treatment was chosen as the most appropriate method.

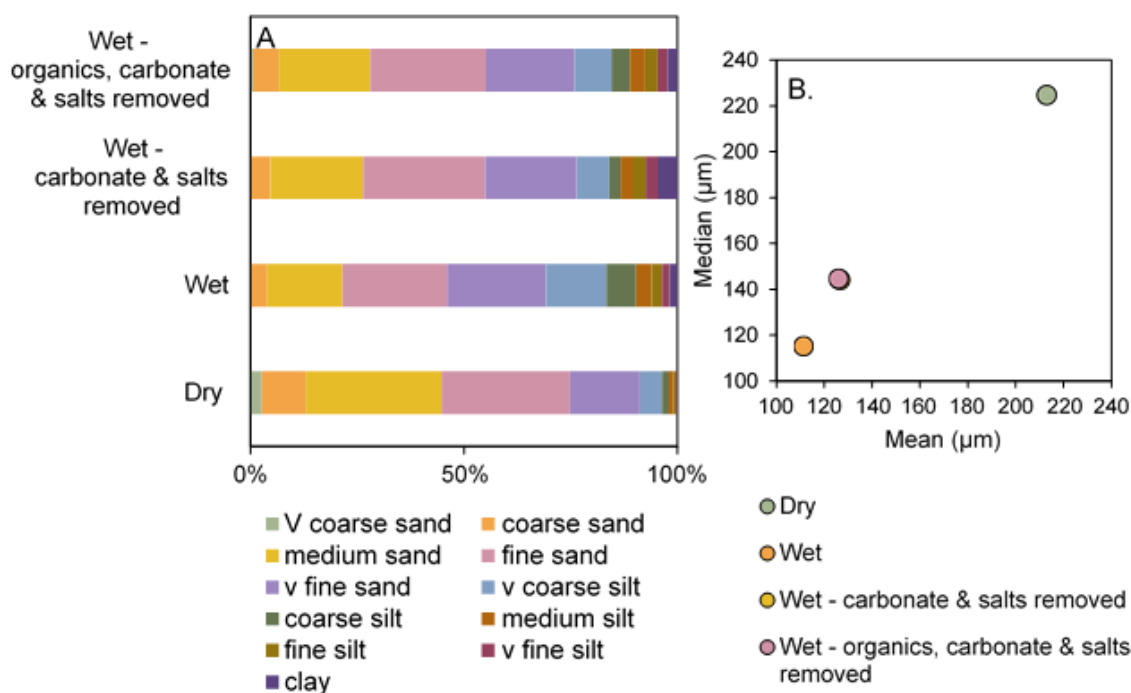


Figure 3-6. Average grain-size composition (A) and mean and median grain-sizes (B) of samples measured using different technique. All measurements were conducted on the same subset of samples.

3.3.3 HEAVY MINERAL ANALYSIS

Heavy mineral analysis was conducted on ten samples by Marta Padoan at the Laboratory for Provenance Studies, Università di Milano-Bicocca, Italy. Methods are described in Chapter 5. For four samples (AUS13/1/1, AUS13/1/11, AUS13/3/1, AUS13/6/1) two grain size fractions were measured (90-125 μm and 180-250 μm for the first three samples, <180 μm and >250 μm for sample AUS13/6/1). Results indicated a slight difference between the size fractions (Figure 3-7). However, both reflected local lithology, and variation was most likely due to original grain-size in the source geology and settling-equivalence (E. Garzanti, pers. comms.). Therefore, results from the different grain-sizes were combined for analysis in Chapter 5

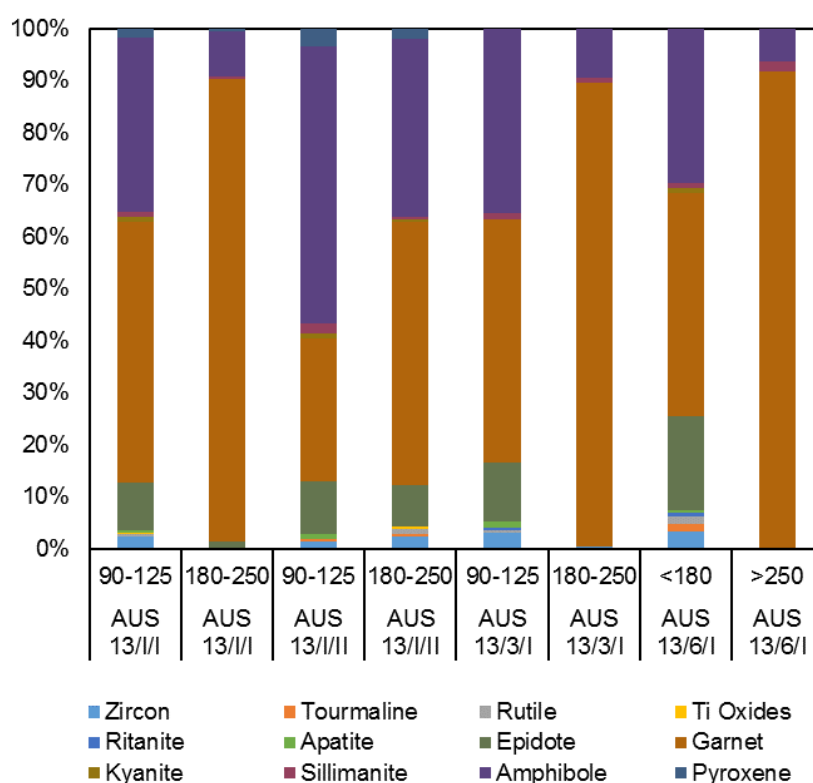


Figure 3-7. Heavy mineral assemblages of Aus samples of different grain-sizes.

3.3.4 OSL SAMPLE PREPARATION

All laboratory work to isolate the quartz fraction was conducted under subdued orange light (585 nm) at the Oxford Luminescence Dating Laboratory (OLDLAB).

The protocol used to isolate the quartz fraction is displayed in Figure 3-8. Carbonate was removed by treatment with 32% hydrochloric acid (HCl) until the reaction stopped (~2-12 hours). 30% hydrogen peroxide (H₂O₂) was used to removed organics. Samples were treated until the end of the reaction (~2-14 days). Samples were wet sieved to the desired size fraction. For KAR13 the model size fraction was used. For all other samples 90-125 µm was used for SA dating and the 180-210 µm size fraction was also prepared in case it was needed for SG dating or sensitivity experiments (section 5.3.1). Heavy liquid density separation using sodium polytungstate was used to remove heavy minerals (2.72 gcm⁻³) and feldspars (2.58 gcm⁻³). Samples were etched on a shaker for 60 minutes in 40% hydrofluoric acid (HF) to remove any remaining feldspars and the alpha-irradiated outer rind of the quartz grains. Samples were treated with 32% HCl for 12 hours on a shaker to remove any fluorides produced during the reaction with HF and re-sieved to remove broken down feldspars. Samples were then dried and stored for measurement. Pilot IR-depletion ratio measurements (Duller, 2003 section 3.3.6) indicated 14 samples contained persistent feldspar contamination. These samples were further treated with hexafluorosilicic acid (H₂SiF₆) for 7-14 days. HCL treatment and re-sieving was repeated.

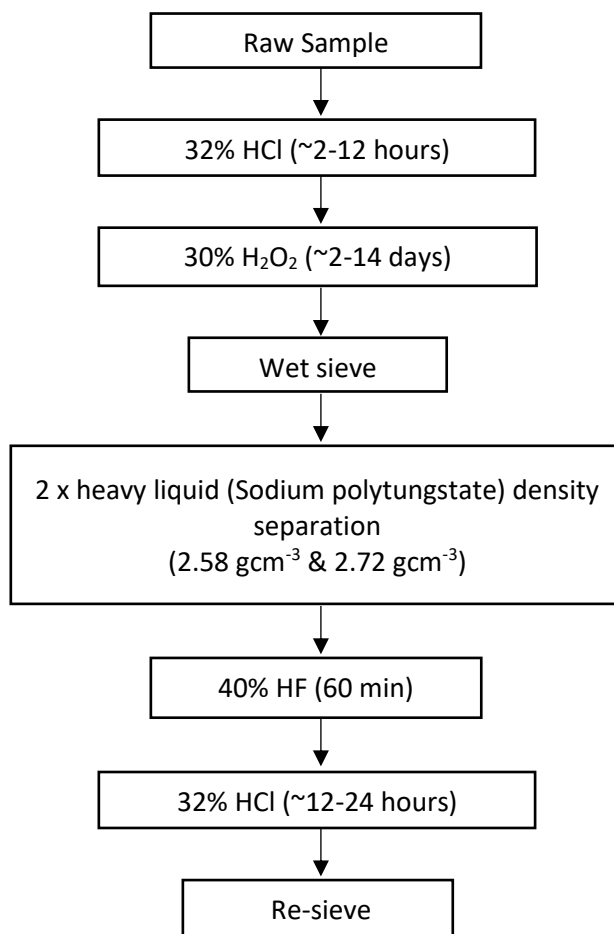


Figure 3-8. Standard protocol used to isolate the quartz fraction

SA samples were mounted onto aluminium discs using silkospray® and, unless otherwise stated, 2mm masks. Small aliquots were used as a compromise between signal strength and detecting any heterogeneity (Duller, 2008). SG discs were mounted by sprinkling 180-210 µm sediment into specially designed discs containing 10x10 300 µm diameter holes (Bøtter-Jensen et al., 2003). To minimise cross-talk SA and SG discs were loaded onto the carousel alternately.

Due to time constraints, only minimal SG analysis could be undertaken. This work looked at the properties of single grains at high doses (see section 4.4.3) and of

overdispersed samples (section 8.3). The results of these analyses suggest that SG dating could potentially increase dating accuracy and precision, particularly for older samples (Chapter 4). Unfortunately, as these findings were made towards the end of the project there was not time to act upon these results. Therefore, dates in this thesis are derived from multi-grain single aliquots.

3.3.5 EQUIPMENT

The majority of OSL measurements were conducted at the OLDLAB. A small number were conducted at the luminescence laboratory at Royal Holloway University of London (RHUL). Measurement took place on one of six Risø TL/OSL-DA-15 automated readers (Bøtter-Jensen et al., 2003) (“Atlas”, “Bacchus”, “Cyclops” & “Delphi”; OLDLAB, “Risø-B” & “Risø-E”; RHUL). Delphi is fitted with a green (525 nm) LED array, all other machines are fitted with blue (470 nm) LED arrays. These LEDs emit at 16 mW/cm², 28 mW/cm², 31 mW/cm², 40 mW/cm², 80 mW/cm² and 76 mW/cm² (order as listed initially above). IR stimulation was conducted using 870 nm LED arrays on all readers except Cyclops which is fitted with a solid state 1W IR laser (830 ± 10 nm). ⁹⁰Sr/⁹⁰Y beta sources were used for all irradiation. These were calibrated to the National Physical Laboratory, Teddington Hotspot 800 ⁶⁰Co γ-source using Risø calibration quartz (Armitage and Bailey, 2005). Calibration at Oxford was conducted in May 2014. Dose rates varied between machines from ~0.07 to ~0.13 Gy/s. OSL was detected using EMI9235QA photomultiplier tubes (PMT) fitted with 7.5 mm of Hoya U-340 filters. The filters transmit at (250-390 nm) and prevent back scatter from stimulation reaching the PMT whilst detecting the UV-A to violet

(~360-420 nm) quartz OSL emissions (Bøtter-Jensen et al., 2003). Single grain measurements were conducted on Cyclops using a 10 mW green (532 nm) laser diode.

3.3.6 SAR MEASUREMENT PROTOCOL

D_e measurements were conducted using a modified single aliquot regeneration (SAR) protocol (Table 3-2). The SAR protocol (Murray and Wintle, 2000, 2003) corrects for aliquot specific sensitivity changes (variation in counts per given dose, caused by heating, irradiation, and signal resetting) using a constant test dose (T_x), measured after each natural or regenerative dose (L_x). For quartz, pre-heats of 160 -300°C for 10s are conducted prior to L_x and T_x measurements to deplete thermally unstable traps. For most samples a PH1 (heating before measurement of L_x) of 260°C for 10 s and a PH2 (heating before measurement of T_x) of 220°C for 10 s deplete unstable traps whilst avoiding thermal transfer. However, the response to heating is sample specific and pre-heat plateau tests (section 3.3.8) are routinely performed to ensure D_e is not dependent on preheat temperature (Wintle and Murray, 2006). A hot bleach (OSL stimulation at 280°C for 10s after T_x measurement) is included to reduce recuperation (transfer of charge) between measurement cycles (Murray and Wintle, 2003).

Table 3-2. Modified SAR protocol used for De measurement. *PH 1 temperatures were 200°C and 240°C for Units 1 & 2 of the Klipkraal sands respectively. **Step 3 is only included during the measurement of the IR-depletion point.

Step	Operation	Observed
1	Beta irradiation (zero for natural)	
2	Preheat at *260°C for 10s	
3**	IRSL stimulation at 50°C for 50s	
4	OSL stimulation at 130°C for 40s (100s)	Lx
5	~15 Gy beta irradiation	
6	Preheat at 220°C for 10s	
7	OSL stimulation at 130°C for 40s (100s)	Tx
8	OSL stimulation at 280°C for 40s (100s)	
9	Return to 1	

The SAR protocol includes three internal reliability tests: The recycling ratio, 0-dose point and the IR-depletion ratio (Duller, 2003). The recycling ratio ensures sensitivity change is being accurately corrected for throughout the protocol using one or two repeat measurements (usually the smallest dose, sometimes coupled with a high dose) conducted towards the end of the measurement cycle. In this thesis aliquots were rejected if repeats were >10% of unity (Murray and Wintle, 2000). The measurement of a 0-dose point ensures charge is not being thermally transferred from light insensitive to light sensitive traps. Aliquots in this thesis were rejected if the response to a 0 Gy given dose (which should only produce noise) was >5% of the natural signal (Murray and Wintle, 2000). The IR-depletion ratio tests for feldspar contamination using a repeat dose point with an additional IRSL bleach (50 s at 50°C) before measurement of L_x. As quartz is not stimulated by IR at low temperatures, feldspar contamination is considered significant if the IR-point is depleted by >10% compared to the original measurement (Duller, 2003). Therefore, in this thesis, aliquots with IR measurements >10% of unity were rejected.

D_e s were determined by interpolating the natural dose (L_n/T_n) values onto dose response curves (DRCs) constructed from laboratory irradiated (L_x/T_x) values. DRCs were fitted with a single saturating exponential (SSE), saturating exponential plus linear (SEPL) or double saturating exponential (DSE) curve. Chapter 4 provides a detailed discussion on the shape of the DRC and its impact on D_e accuracy. A late background, averaged over the final 10 s of measurement, was subtracted from initial counts. Integration limits were chosen to reflect a constant power, equivalent to the fast component bleaching rate of 19 mJ.cm^{-2} of blue (470 nm) stimulation, between machines (i.e. the first 0.2 s (“Risø-B” and “Riso-E”), 0.5 s (“Cyclops”), 0.7 s (“Bacchus”) or 1.2 s (“Atlas”) for blue LEDs or the first 3.5 s for green LEDs (“Delphi”).

Uncertainties associated with individual L_x/T_x values are derived from Poisson counting statistics combined in quadrature with a 2.5% measurement uncertainty. Overall D_e errors are calculated using Monte-Carlo curve fitting (Duller, 2007). Data processing was conducted in Analyst v.4.14.6 (Duller, 2013).

3.3.7 FAST RATIO AND IR RATIO

Aliquots which failed the fast ratio (values <20 ; Durcan and Duller, 2011) were rejected from subsequent analysis. As the signal from feldspar is significantly slower than the fast-component of quartz (McKeever et al., 1997) aliquots which fail the IR-depletion ratio (Duller, 2003) are also likely to fail the fast ratio. Figure 3-9 demonstrates that this is typically the case for the 590 aliquots measured in this thesis. Some aliquots do fail the IR-depletion ratio but not the fast-ratio but in

these cases the aliquots either have dim signals wherein noise obscures the value or also demonstrate poor recycling. Therefore, in cases when the IR-depletion ratio was not available (e.g. reduced DRTs; section 3.3.10) the fast ratio was also used as a proxy for feldspar contamination.

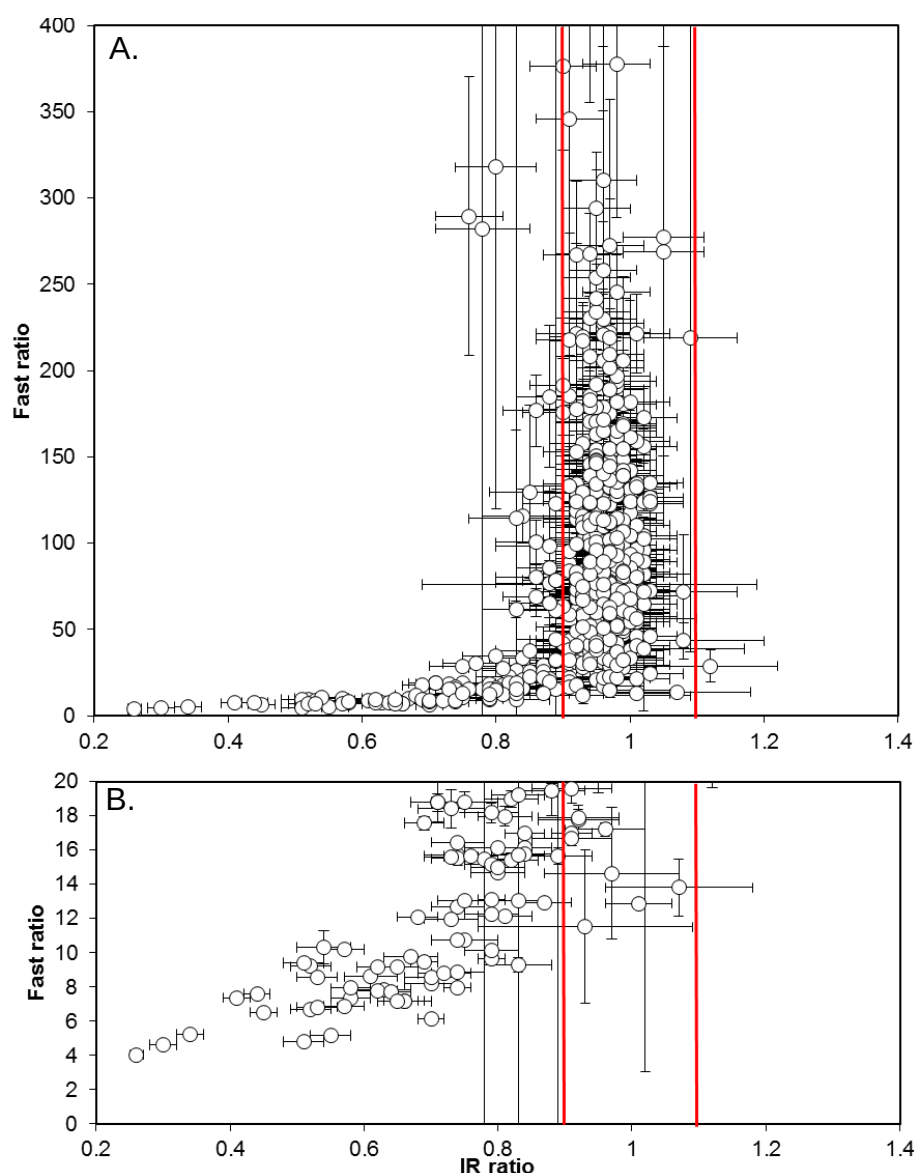


Figure 3-9. Relationship between the fast ratio and the IR-depletion ratio. A) All studied aliquots. B). Samples with fast ratios below 20. Red lines mark 10% of unity with the IR-depletion ratio.

3.3.8 PREHEAT TESTS

To ensure no thermal transfer (transfer of charge from light insensitive to light sensitive traps as a result of heating) is occurring, and to select the optimal preheat (PH) temperatures, dose recovery preheat tests were conducted on a subset of five Namibian samples. The dose recovery method is described in section 0. Subsets of 4-5 aliquots were measured using different PH1 temperatures ranging from 180-300°C. All samples performed well using a PH1 at 260°C for 10 s and a PH 2 at 220°C for 10 s (Figure 3-10). These conditions were therefore applied to D_e measurements of the Namibian samples.

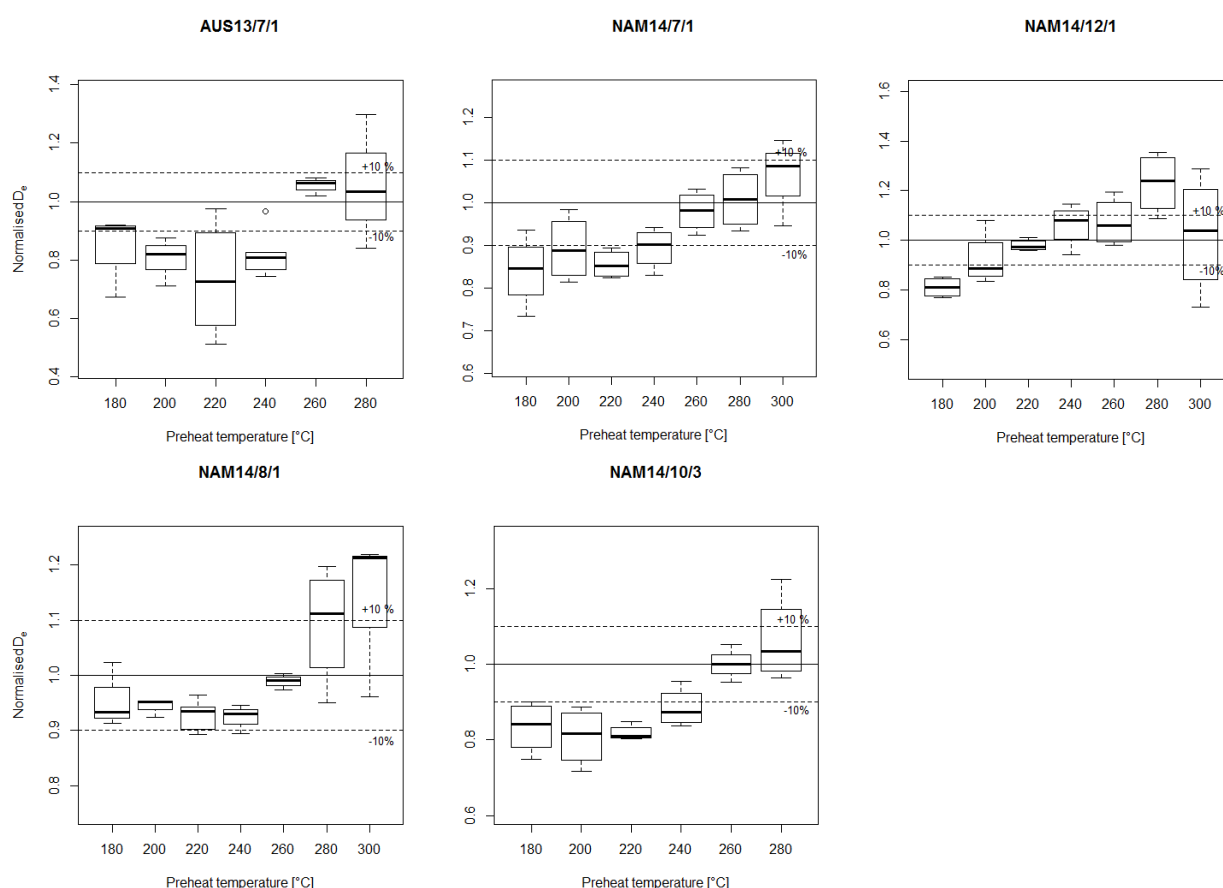


Figure 3-10. Results of PH tests for Namibian samples. Only aliquots which pass rejection criteria are included. All categories contain at least 3 aliquots. 300°C measurements are not included in figures when <2 aliquots passed rejection criteria

Table 3-3 Summary of preheat tests conducted on a subset of samples from the Klipkraal Sands. # is the number of accepted aliquots, total measured are displayed in brackets.

Unit	Sample	Temp (°C)	#	Cause of rejection	Recovered dose (normalised)
1	KAR13/1/1	160	2 (3)	Dim signal*1	1.17 ± 0.1
1	KAR13/1/1	180	2 (3)	Dim signal*1	1.08 ± 0.1
1	KAR13/1/1	200	2 (3)	Dim signal*1	1.13 ± 0.1
1	KAR13/1/1	220	3 (3)		1.14 ± 0.1
1	KAR13/1/1	240	3 (3)		1.05 ± 0.01
1	KAR13/1/1	260	3 (3)		1.11 ± 0.05
1	KAR13/4/1	120	5 (5)		1.06 ± 0.03
1	KAR13/4/1	140	5 (5)		1.04 ± 0.02
1	KAR13/4/1	160	5 (5)		1.05 ± 0.03
1	KAR13/4/1	200	5 (5)		1.01 ± 0.07
1	KAR13/4/1	240	5 (5)		0.98 ± 0.03
1	KAR13/4/1	280	3 (5)	Poor recycling *1	1.03 ± 0.02
2	KAR13/1/2	180	3 (3)		0.93 ± 0.10
2	KAR13/1/2	200	2 (3)	IR ratio *1	0.87 ± 0.06
2	KAR13/1/2	220	3 (3)		0.85 ± 0.04
2	KAR13/1/2	240	9 (11)	Dim signal *2, IR ratio *1	0.75 ± 0.01
2	KAR13/1/2	250	7 (8)	Dim signal *1	0.85 ± 0.12
2	KAR13/1/2	260	5 (11)	Dim signal *4, Poor recycling *3, IR ratio *2	0.94 ± 0.12
2	KAR13/1/2	280	3 (3)		1.05 ± 0.14
3	KAR13/1/4	200	3 (3)		0.78 ± 0.08
3	KAR13/1/4	220	3 (3)		0.89 ± 0.07
3	KAR13/1/4	240	8 (11)	Dim signal *1, Poor recycling *1, IR ratio *1	0.95 ± 0.09
3	KAR13/1/4	250	5 (8)	Poor recycling *3	1.11 ± 0.07
3	KAR13/1/4	260	6 (11)	Poor recycling *5	0.98 ± 0.09
3	KAR13/1/4	280	2 (3)	Recuperation *1	0.95 ± 0.07
3	KAR13/1/4	300	0 (3)	Recuperation *3	n/a
3	KAR13/2/1	200	3 (3)		0.71 ± 0.05
3	KAR13/2/1	220	3 (3)		0.76 ± 0.02
3	KAR13/2/1	240	10 (11)	Dim signal *1	0.90 ± 0.05
3	KAR13/2/1	250	3 (8)	Dim signal *3, IR ratio *2	0.99 ± 0.14
3	KAR13/2/1	260	6 (11)	Dim signal *2, Poor recycling *2, IR ratio *1	1.03 ± 0.16
3	KAR13/2/1	280	0 (3)	Poor recycling *2, IR ratio *1	n/a
3	KAR13/2/1	300	0 (3)	Poor recycling *2, IR ratio *1	n/a
3	KAR13/2/3	200	3 (3)		0.88 ± 0.13
3	KAR13/2/3	220	3 (3)		0.93 ± 0.07
3	KAR13/2/3	240	11 (11)		0.95 ± 0.09
3	KAR13/2/3	250	7 (8)	Poor recycling *1	0.91 ± 0.10
3	KAR13/2/3	260	6 (11)	Dim signal *3, Poor recycling *1, IR ratio *1	0.85 ± 0.21
3	KAR13/2/3	280	2 (3)	IR ratio *1	0.73 ± 0.09
3	KAR13/2/3	300	1 (3)	Poor recycling *2	0.41

Preheat behaviour of the samples from the Klipkraal Sands was more complex. Following PH and dose recovery tests, samples from Unit 1 were measured at 200°C for 10 s for both PH1 and PH2. Samples from Unit 2 were measured at 260°C and 220°C for 10s for PH1 and PH2 respectively and samples from Unit 3 were measured at 240°C and 220°C for 10s for PH1 and PH2 respectively. PH1 values were determined using PH tests whilst PH2 values were kept at 220°C or

as close as possible without being larger than the PH1 temperature (Wintle and Murray, 2006). The results are discussed in detail in section 6.3.2. and a summary is shown in Table 3-3.

3.3.9 DOSE RECOVERY TESTS

Sensitivity change during the heating and measurement of the natural dose cannot be assessed internally within the SAR protocol. Dose recovery tests (DRTs) are routinely used to ensure the test dose is accurately correcting for these changes and to test the maximum precision of the SAR protocol (Roberts et al., 1999; Wallinga et al., 2000; Wintle and Murray, 2006).

Table 3-4. Results of dose recovery tests. # is the number of aliquots accepted out of the total measured. O.D. is the overdispersion from CAM calculations.

Sample	#	Given Dose	Mean	Ratio	CAM	O.D	Ratio
NAM14/5/2	8/10	37.6	36.3 ± 2.3	0.97	37.0 ± 1.1	3 ± 1	0.99
NAM14/5/4	9/10	300	282 ± 30	0.94	277 ± 13	0 ± ∞	0.92
NAM14/7/1	12/12	37.8	36.8 ± 0.9	0.97	36.8 ± 0.5	0 ± ∞	0.97
NAM14/8/1	12/12	3.44	3.5 ± 0.2	1.01	3.5 ± 0.1	0 ± ∞	1.01
NAM13/10/3	10/12	55.2	59.2 ± 9.4	1.07	58.5 ± 2.7	13 ± 1	1.06
NAM14/12/1	10/12	207	195 ± 47	0.94	193 ± 15	20 ± 3	0.93
NAM14/12/7	10/10	300	316 ± 14	1.06	316 ± 11	0 ± ∞	1.05
AUS13/3/1	8/10	37.5	36.7 ± 0.9	0.98	36.6 ± 1.1	0 ± ∞	0.98
AUS13/3/6	4/5	10.5	12.3 ± 1.0	1.17	11.7 ± 0.5	0 ± ∞	1.11
AUS13/6/1	5/5	10.5	10.1 ± 0.3	0.96	10.4 ± 0.3	0 ± ∞	0.98
AUS13/1/1	9/10	500	494 ± 103	0.99	483 ± 34	16 ± 3	0.97
AUS13/7/1	12/12	10	8.8 ± 0.8	0.88	8.73 ± 0.3	9 ± 1	0.87
KAR13/1/2	12/12	5.07	4.5 ± 0.4	0.88	4.47 ± 0.1	6 ± 1	0.88
KAR13/1/4	9/12	25	24.5 ± 2.4	0.98	24.2 ± 0.8	8 ± 1	0.97
KAR13/2/1	11/12	50	49.6 ± 4.9	0.99	49.6 ± 1.3	6 ± 1	0.99
KAR13/2/4	11/12	150	136 ± 23	0.91	131 ± 5.6	11 ± 1	0.87
KAR13/3/1	12/12	11.6	11.4 ± 0.7	0.96	11.1 ± 0.2	0 ± ∞	0.95

DRTs were performed on a subset of samples. Previously unmeasured aliquots were twice bleached with blue (or green) light for 1000 s at 20°C with a 10,000 s pause between bleaches to allow the 110 °C trap to empty. Samples were irradiated with a known laboratory dose. This dose was then treated as the ‘natural’ dose during SAR protocol measurements. All samples could recover the given dose within 15% but precision varied between samples (Table 3-4).

3.3.10 REDUCED DOSE RECOVERY TESTS

Conducting DRTs at high doses is time-consuming and, as such, is commonly forgone in favour of measuring lower dose samples. To circumvent the long measurement times associated with high dose DRTs, the use and reliability of ‘reduced DRTs’ using only two dose points (one equal to the given dose and the other up to 200 Gy above and below this value) was explored. Results from full DRTs from ten aliquots each of three samples were used to simulate ‘reduced DRTs’. The double saturating exponential (DSE) equations used to fit full DRT data were used to calculate simulated L_x/T_x values at 25, 50, 100, 150 & 200 Gy above and below the given dose for each aliquot. To test the most appropriate dose points to use, ten DRCs were constructed for each aliquot using a SSE fitted through the origin (0,0), the measured L_x/T_x values equivalent to the given dose, and one of the simulated L_x/T_x values. The value of the second dose point influenced the accuracy of D_e estimation (Figure 3-11). D_e estimates for a 300 Gy given dose were most accurate (within 2% of unity with full DRTs) when the second ‘reduced DRT’ dose point was 80-133% of the given dose. For a 500 Gy

given dose, L_n/T_n values were most accurate (within 5% of unity) when the second dose point was between 110-130% of the given dose (Figure 3-11). This suggests that reduced DRTs with two dose points, one at the given dose and one between 110-130% of this value can be used as a time saving proxy for full DRTs. However, this method was explored towards the end of laboratory work and has not yet been applied to any samples.

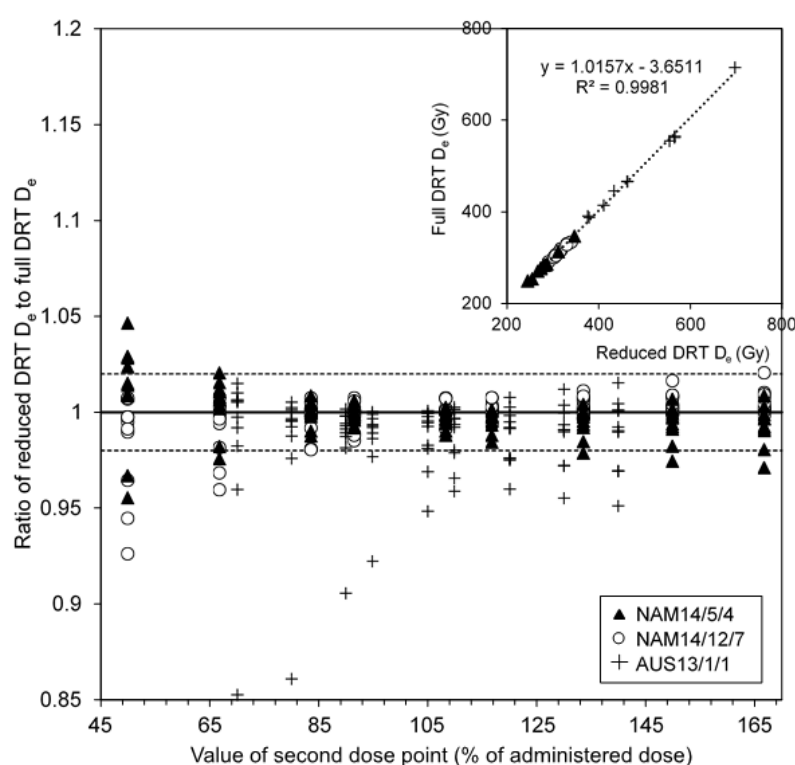


Figure 3-11. Agreement between results of “two dose point reduced DRTs” and full DRTs at varying values of the second dose point of the “reduced DRTs”. Solid black line indicates unity; dashed lines represent 2% of unity. NAM14/5/4 and NAM14/12/7 were given a 300 Gy given dose. AUS13/1/1 was given a 500 Gy given dose. Inset demonstrates the 1:1 relationship between methods when the second “reduced DRT” dose point is 100Gy higher than (120-133% of) the given dose. Dotted line represents the trend line (equation printed above).

3.3.11 MEASURING YOUNG SAMPLES

Samples KAR13/1/1 and KAR13/4/1 were shown to be both very young (~ 0.5 Gy) and to have ‘dim’, insensitive, signals (i.e. low number of counts produced per given dose) making accurate measurement difficult. As these samples were assumed to be associated with human land disturbance, obtaining an accurate date was important. Therefore, means to increase the signal to noise ratio were explored. All subsequent measurements of these samples were conducted on ‘Bacchus’ – the machine with the highest intrinsic signal to noise ratio. Tests conducted on 12 aliquots of KAR13/4/1, showed that, although above the optimum operating conditions of the PMT, a high voltage of 8.5 (1350 V) produced a good signal to noise ratio (Figure 3-12). Therefore, this setting was used for all subsequent measurements of these samples.

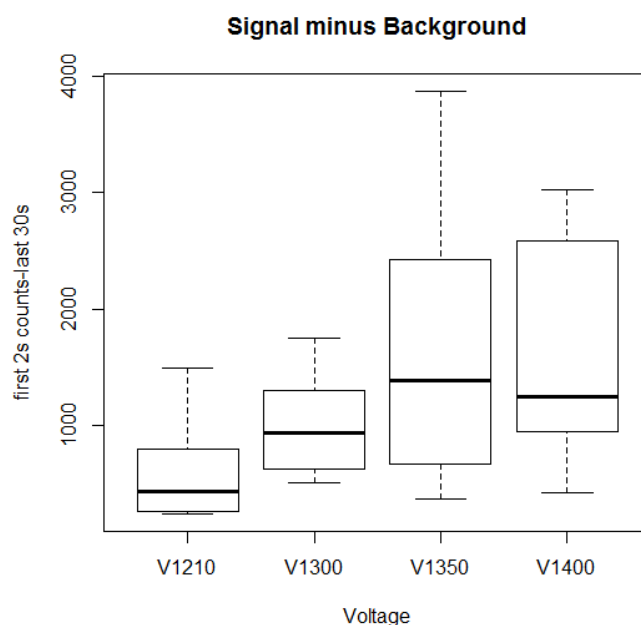


Figure 3-12. Signal of 12 aliquots after a 5 Gy given dose measured using different PMT high voltage settings. The background measurement (total counts measured in the last 30 s divided by 15) was subtracted from the total counts measured in the first 2 s (signal). A long background was used to remove the effect of ‘spiking’ which accompanied measurements above 1350V.

3.3.12 OSL SENSITIVITY TESTS.

OSL sensitivity (counts per given dose) of quartz has been shown to be dependent on environmental history with sensitivity increasing with repeated burial and bleaching cycles (Fitzsimmons et al., 2010; Moska and Murray, 2006; Pietsch et al., 2008). Therefore, sensitivity may be used as a geomorphic provenancing tool. The use of this technique is explored in Chapter 5. The measurement procedure is described in detail in section 5.3.1.

3.3.13 CALCULATING THE ENVIRONMENTAL DOSE

Radionuclide concentrations were measured using Inductively Coupled Plasma Mass Spectrometry/Optical Emission Spectrometry (ICP-MS/ICP-OES) for all samples and using in-situ gamma spectrometry for a limited number of samples. These were used to estimate the beta and gamma dose rates. Alpha contribution was assumed negligible following the 60 min HF etch.

ICP-MS

ICP-MS/OES analyses were performed at SUERC, East Kilbride. Before analysis ~200g of sediment was homogenised using a Teema™ Mill. From this, ~0.2 g was used for analysis. Dissolution was performed using triacid dissolution without perchloric acid (Figure 3-13). Standards were measured before and after each batch and after every five measurement runs (V.Olive, pers. comms.).

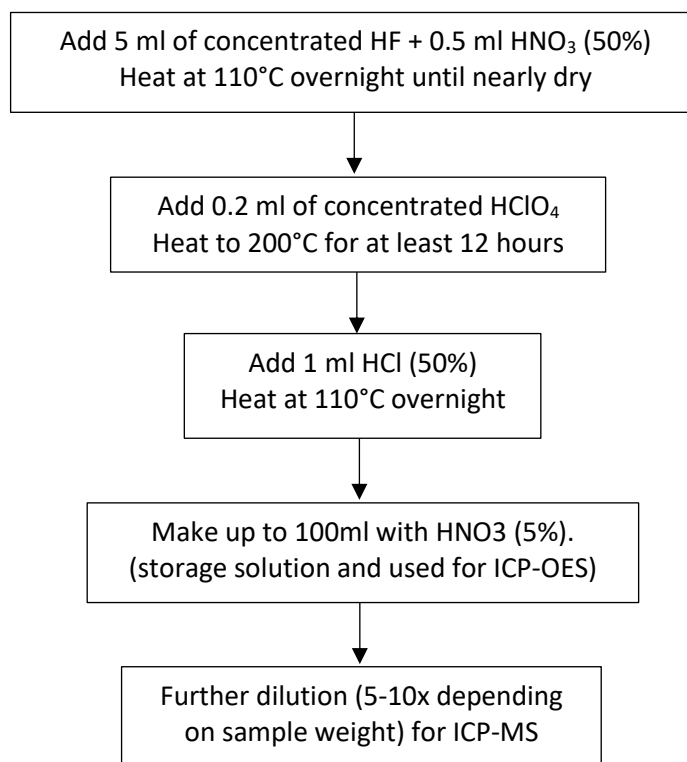


Figure 3-13. Acid Dissolution process performed at SUERC (V.Olive, pers. comms.)

ICP-MS/OES repeats and error calculation

Repeat ICP-MS/OES measurements were conducted on 11 samples. There is reasonable agreement between repeat measurements but variation is larger than expected from analytical errors alone and repeatability varies between samples (Figure 3-14). Therefore, for samples without repeat measurements, uncertainties associated with radionuclide concentrations were calculated based on the relative variance between repeats ($\pm 20\%$, $\pm 28\%$ and $\pm 9\%$ for U, Th and K respectively). For samples with repeat measurements, uncertainties were quoted as the standard deviation between repeats.

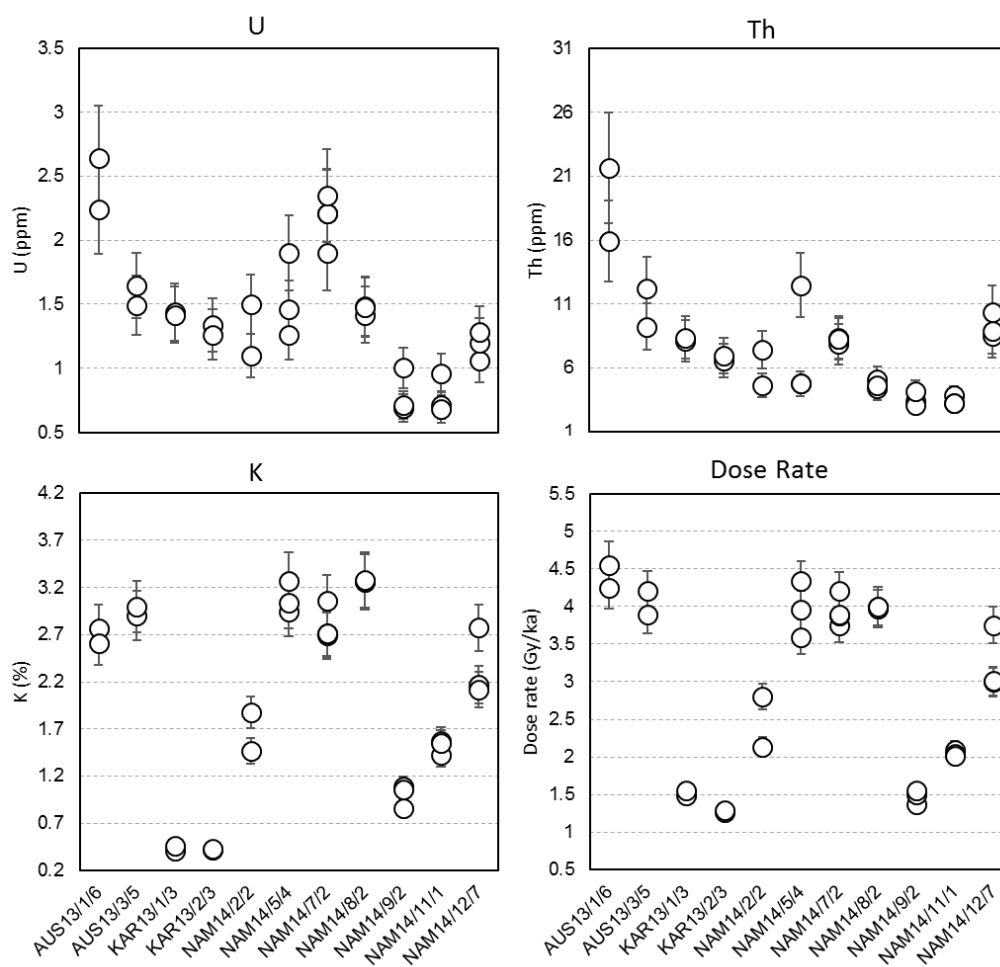


Figure 3-14. Radionuclide concentrations and dose rates measures from repeat ICP measurements and gamma spectrometry.

Source of variation

Preusser and Kasper (2001) found relative standard deviations of up to 9% for K and 5% for Th and U when comparing a series of repeat samples within and between laboratories. Beyond analytical errors, variation between ICP results is likely to reflect sample heterogeneity. Despite attempts to homogenise the sediments, the small sample size used for ICP-MS means that some variation is likely. Leighton (2013) found an average of ~4% relative standard deviation on K concentrations, 35% for Th and 10% for U. These are not particularly consistent

with the 9%, 28% and 20% for K, Th and U found in this study. However, the samples from Leighton (2013) represent homogenous dune sand with relatively low U and Th concentrations and the samples themselves were not ground to a homogenous powder (C. Leighton. pers. comms.). Incomplete or inconsistent acid digestion (i.e. the incomplete breakdown of chemically resistant minerals such as zircons) may increase variation between samples. This may also explain the underestimation of U and Th concentration in comparison to gamma spectrometry for some samples (Figure 3-15) (Preusser and Kasper, 2001). High-pressure microwave digestion is suggested as the best method to ensure complete and consistent digestion (Preusser and Kasper, 2001). Variation between K measurements may reflect the similarity between ^{40}K and the inert carrier gas ^{40}Ar (Preusser and Kasper, 2001).

Gamma Spectrometry

Due to time constraints gamma spectrometry could not be applied to all samples. Where possible, measurements were taken from samples with heterogeneous sediment matrixes (e.g. above or below clast layers) and representative measurements were taken from sites with uniform heterogeneity (e.g. clasts sporadically but evenly distributed throughout the matrix). Gamma spectrometry could not be measured for augured sites and poor battery life limited the number of measurements conducted in the 2013 fieldtrip. Measurements were conducted for 45 minutes using an Ortec® DigiBASE™ and a 2" sodium iodide detector in 2013 and for 45 minutes using a Canberra digital hand-held NaI/LaBr Spectrometer (InSpector 1000™) in 2014. Radionuclide concentrations were

calculated using the window method (Table 3-5), and in-house software written by Matt Telfer but updated to incorporate errors, Liritzis et al. (2013) conversion factors for calibration and, for 2014 measurements, new calibration measurements. Calibration was conducted using the 'Oxford Blocks' (Rhodes and Schwenninger, 2007) in April 2014 for the InSpector 1000™. Doped blocks were measured for 30 minutes whilst the undoped background block was measured for 60 minutes. The background was subtracted from doped spectra and efficiency calibration was conducted following Aitken (1985). Calibration of the DigiBASE™ was conducted in 2007. Where available, gamma spectrometry is used to calculate the gamma dose whilst ICP-MS is still used for beta dose rate calculation in all samples.

Table 3-5 Peak energy of radionuclides and energy windows used for measurement.

Isotope	Peak (MeV)	Typical Peak window (MeV)
Potassium-40	1.46	1.38-1.53
Bismuth -214 (U series)	1.76	1.69-1.84
Thallium-208 (Th series)	2.61	2.46-2.76

Comparison between gamma and ICP-MS/OES

Radionuclide concentrations calculated from gamma spectrometry using the window method are in reasonable agreement with the results of ICP-MS/OES (Figure 3-15). For most samples results agree within errors but variation of up to 70%, 75% and 35% is observed for K, Th and U based on central values; average variation is ~26%, 27% and 12% respectively. Total gamma dose rates are within errors for both methods for all samples except NAM14/12/1 and AUS13/1/1. Discrepancy between the two methods may indicate that the small subsample used for ICP-MS/OES is not representative of the total gamma dose received by

the samples (Bailey et al., 2003b). This is particularly relevant in the heterogeneous matrix of many of the sand ramps. A possible cause of heterogeneity for some of the measured samples (e.g. AUS13/1/1, NAM14/3/1, NAM14/5/4) may be calcrete formation which can either reduce dose rates via attenuation (e.g. Nathan and Mauz, 2008) or increased dose rate due to the uptake of mobile U (Grine et al., 2007). Variation between ICP-MS/OES and gamma spectrometry may also be caused by incomplete acid digestion during ICP measurement or, if a large variation in U values is observed, may suggest U-series disequilibrium (Prescott and Hutton, 1995). Variation in this study is most likely caused by matrix heterogeneity or incomplete acid digestion with the impact of disequilibrium assumed minimal.

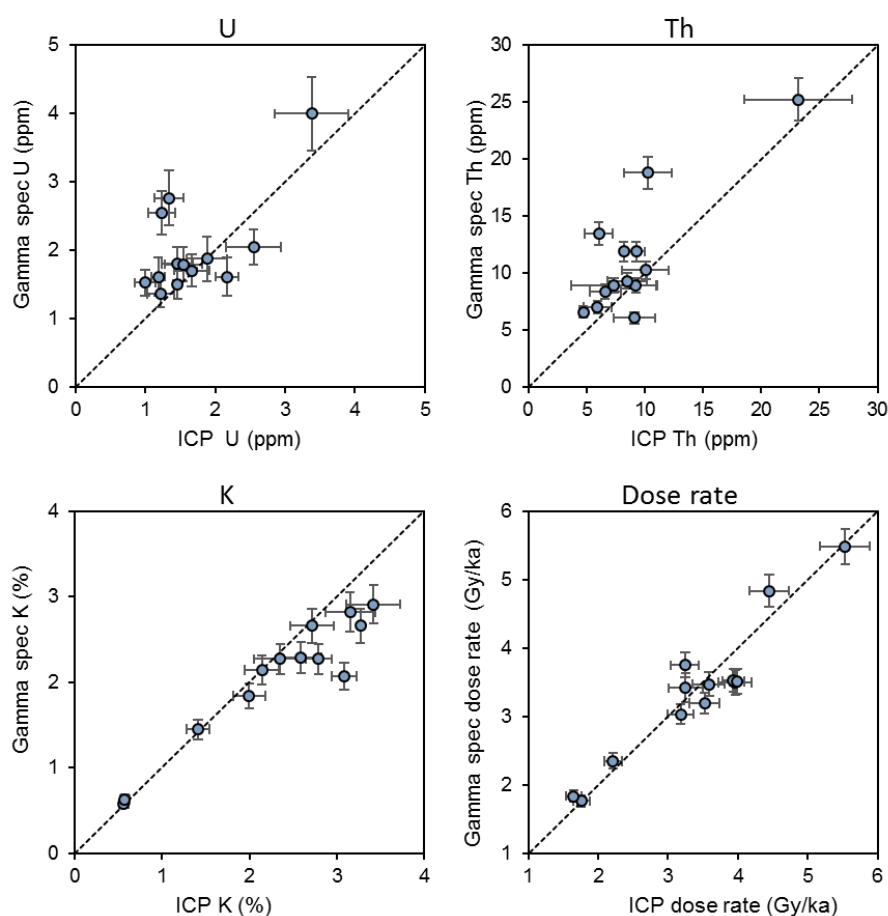


Figure 3-15. Comparison between radio nuclide and total dose rates measured using gamma spectrometry and ICP-MS/OES. For samples with repeat ICP results, the mean and standard deviation are plotted. Dashed line represents a 1:1 relationship.

Dose rate calculation

To minimise unnoticed calculation errors, final dose rate calculation was conducted using DRAC (Dose Rate and Age Calculator for trapped charge dating; Durcan et al., 2015). Encouragingly, personal calculations showed good agreement with the results of DRAC. Based on the aridity of the modern climate, values used in previous studies in the region (e.g. Bristow et al., 2005, 2007; Stone et al., 2010; Stone and Thomas, 2008) and the position of samples (either free draining slopes with evidence of fluvial activity or dunes) a water value of $5 \pm 3\%$ was used for all Namibian samples. Although the climate at Klipkraal is presently wetter than Namibia, a water value of $5 \pm 5\%$ was used for dose rate determination. This is based on the position of the sediments on a free draining hillslope and regional climate reconstructions which suggest fluctuating but typically dry conditions over the past 45 ka (Chevalier and Chase, 2015, 2016). Radioactivity was calculated using the Liritzis et al. (2013) conversion factors. Beta attenuation was corrected for using Guérin et al. (2012) and the etch depth values of Bell (1979). Cosmic dose was calculated using the formula of Prescott and Hutton (1994), incorporating, overburden depth and density, altitude, latitude and longitude. For 80% of samples the cosmic dose contributes $\leq 10\%$ of total dose and total contribution is $< 25\%$ for all samples. Therefore, variation due to overburden history is presumed negligible and cosmic dose was assumed to be constant.

Dose rate from radionuclides

For the majority of samples c.55-65% of the total dose is derived from the beta dose, c.35-45% from gamma and <10% from the cosmic dose. The low K concentrations at Klipkraal means the beta dose only constitutes c.40% of the total dose at this site. For all sites barring Klipkraal, ^{40}K contributes 70-90% of the total beta dose, making up 55-85% of the total dose. For these sites, Th and U each contribute ~10% to the total beta dose and 25-50% and 10-20% of the gamma dose respectively. Total dose rates from Th and U are ~15-25% and ~7-15% respectively. These results are comparable to the ~3700 samples analysed by Ankjærgaard and Murray (2007) wherein ^{40}K contributed 40-92% of the total beta dose for 95% of the samples making up 27-65% of total dose rate. The high ^{40}K concentrations at most sites reduced the likelihood of significant beta heterogeneity caused by this radionuclide (Chauhan and Singhvi, 2011; Cunningham et al., 2011; Guérin et al., 2015; Mayya et al., 2006).

Dose rate by location

Radionuclide concentrations, and therefore dose rates, show considerable regional variation (Figure 3-16). Mejdahl (1979) suggest a typical sample comprises 12 ppm Th, 3 ppm U and 1-2% K (Dose rate~1.5-2.2 Gy/ka). High potassium levels (2-3.5%) are observed in the Neuhof, Namib Rand, Sandkop and Aus regions, leading to high dose rates of ~3-5 Gy/ka. Samples from the Aus region contain very high concentrations of Th. Low potassium is observed at Klipkraal whilst low U and Th are found at Samara. The high dose rates of

Neuhof, Namib Rand, Sandkop and Aus regions limit the reliably datable age range at these sites to <40-70 ka.

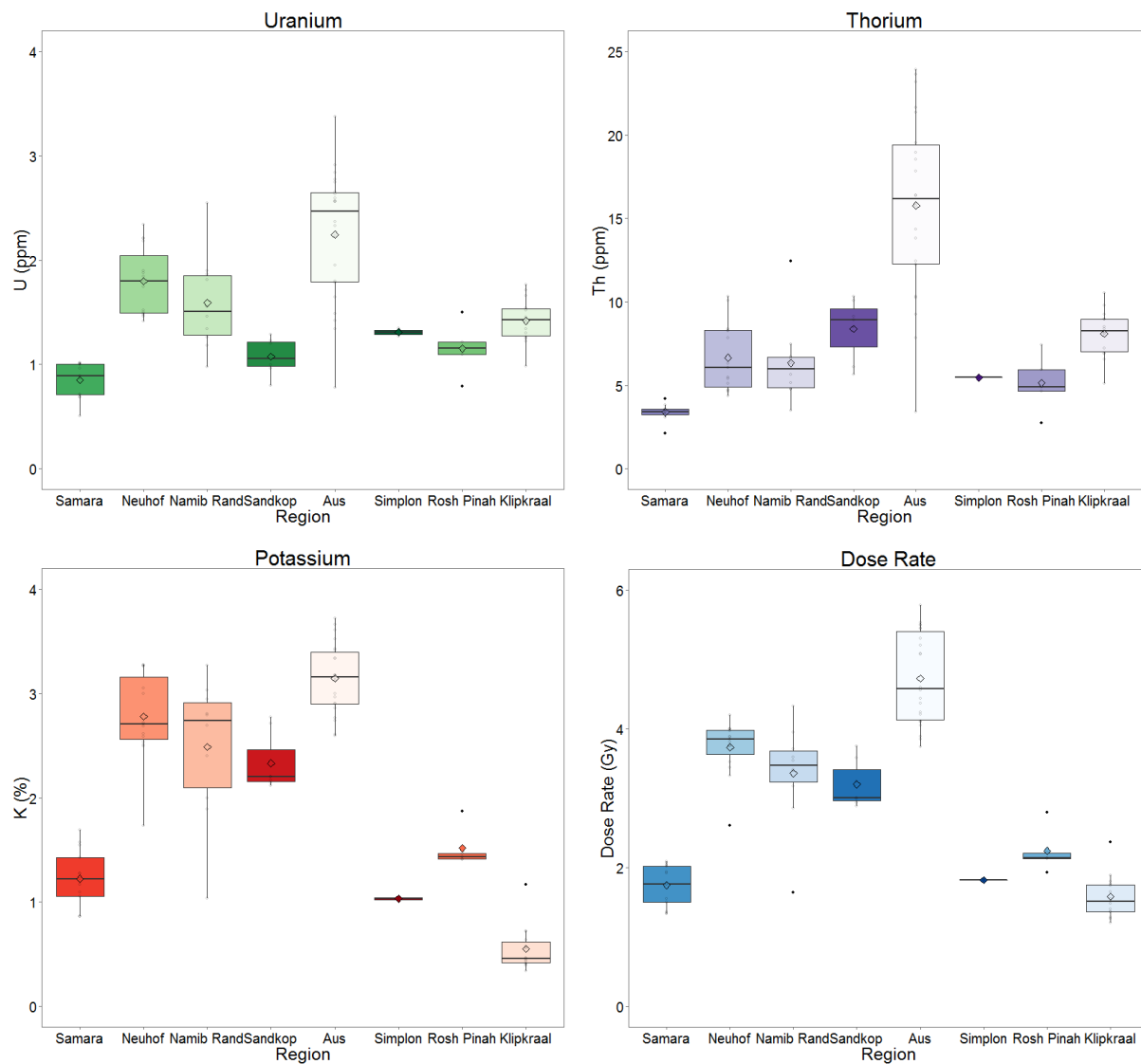


Figure 3-16. Radio nuclide concentration and dose rate by region. The Neuhof region contains samples from Neuhof 1 and Neuhof 2. Namib Rand includes Keerwader and Jagkop and Aus includes Au and Klein Aus. All other regions contain samples from a single sand ramp.

Dose rate by size fraction

For samples AUS13/1/1 and AUS13/3/1 the 90-125 and 180-250 μm grain-size fractions were not available for ICP-MS/OES analysis. Dose rates were derived from the remaining sediment and investigations were undertaken on sample AUS13/3/2 to assess how dose distribution varies with grain-size. U and Th concentrations were shown to increase with decreasing grain-sizes whilst K is highest in the 180-210 μm size fraction (Table 3-6). This distribution may be explained by the settling-equivalence principle (grains of similar density are likely to be transported and deposited together, regardless of size; Garzanti et al., 2008; Rubey, 1933). U and Th are found in heavy minerals such as zircons which, due to their high density, are most commonly found in the fine size fraction. K is most common in K-feldspars which have a similar, albeit slightly lighter, density than quartz and are common in the sand-size fraction. The highest dose rate is found in the mixed sample which contains high levels of all radionuclides. These results suggest that radionuclide concentrations measured for samples AUS13/1/1 and AUS13/3/1 may not be fully representative of the whole. Nevertheless, calculated dose rates are in good agreement with other samples from the same sections. Therefore, measured radionuclide concentrations are used but with uncertainties of $\pm 32\%$, $\pm 30\%$ and $\pm 20\%$ for U, Th and K respectively, to account for the standard deviation between size fractions.

Table 3-6. Radionuclide concentrations of different size fractions of sample AUS13/3/2.

AUS13/3/2	K (%)	U (ppm)	Th (ppm)	DR (Gy/ka)
All grains sizes	3.2 ± 0.6	3.4 ± 0.7	23.2 ± 4.6	5.4 ± 0.5
>250 μm	3.0 ± 0.6	1.3 ± 0.3	6.3 ± 1.3	3.7 ± 0.5
250-180 μm	3.5 ± 0.7	1.8 ± 0.4	11.0 ± 2.2	4.6 ± 0.6
180-125 μm	1.7 ± 0.3	2.0 ± 0.4	15.7 ± 3.2	3.3 ± 0.3
125-90 μm	2.3 ± 0.5	2.8 ± 0.6	20.6 ± 4.1	4.3 ± 0.4
<90 μm	2.1 ± 0.4	4.2 ± 0.8	28.0 ± 5.6	4.9 ± 0.4

3.3.14 AGE ESTIMATION

Sample D_e s were calculated using the Central Age Model (CAM; Galbraith et al., 1999). This assumes the true D_e is drawn from a normal distribution but external factors above measurement errors contribute to the standard deviation. CAM provides an estimate of sample D_e and overdispersion (σ ; observed scatter in values above that expected from measurement errors and machine repeatability). These are calculated based on a geometric weighted mean with a user entered σ_b (amount of variation that is expected for well bleached samples, e.g. 0.2; Cunningham and Wallinga, 2012). This age model was chosen as the most appropriate as overdispersion in sand ramp samples is most likely to arise from beta heterogeneity or sediment mixing which do not systematically bias D_e distribution whilst the use of multigrain aliquots ruled out the application of the finite mixture model (Arnold and Roberts, 2009).

CAM D_e s were calculated in R 'Luminescence' Package v. 0.4.4 (Kreutzer et al., 2012b, 2015). σ_b were calculated based on maximum log likelihoods (Kreutzer et al., 2015). Aliquots were discarded as outliers if their D_e was $>3\sigma$ from the mean. Age estimates were calculated using Equation 3-1.

Chapter 4: Testing the reliability of D_e estimates from the high-dose region of the quartz OSL dose response curve

Preliminary measurements on a number of sand ramps indicated D_e values were >300 Gy. However, contrary to expected, the dose response curves of these samples were not in saturation. The experiments in this paper were conducted to test possible causes of continued dose response curve growth at high doses and assess the reliability of age estimates obtained from these samples. Based on the findings of this study, all samples with D_e values >200 Gy are quoted as minimum ages throughout this thesis. This is further discussed in section 8.3.1

4.1 ABSTRACT (FROM PAPER SUBMITTED TO 'QUATERNARY GEOCHRONOLOGY' –
AUTHORS: A.L.K. ROWELL, R.M. BAILEY AND S.J. ARMITAGE)

Quartz OSL is known to be able to produce accurate age estimates within the range of radiocarbon dating. However, above ~150 Gy (~40-100 ka and above, depending upon the dose rate) age underestimations have been reported. At these doses the dose response curve (DRC) evolves beyond the expected single saturating exponential and extra linear or exponential components are needed to fit the data. The physical mechanism which causes these additional components to be required is unknown, and therefore the reliability of equivalent doses from these curves has been questioned. This study investigates potential causes of the extra curve components and age inaccuracies. Samples accurately recover known given doses ("dose recovery tests") and pass the IR-depletion ratio at high doses, indicating that DRC form at high doses is not the result of test dose correctible sensitivity change or of the increasing significance of minor feldspar contamination. Single grain DRC properties start to diverge between 100-200 Gy with non-monotonic growth beginning at ~200 Gy for some grains. Multi-grain aliquot DRCs are strongly influenced by the DRC properties of the constituent grains. In our dataset samples with >30% of signal from non-monotonic grains yield inaccurate D_e estimates above 100 Gy. Non-monotonic growth after 750-1000 Gy was clearly observed in the fast component L_x values of 7/8 multi-grain samples indicating neither saturation nor continued growth is the ultimate curve function under current measurement procedures. Therefore, age inaccuracies at high doses are likely to be a result of averaging single grains with varying DRC properties, particularly those displaying non-monotonic growth, and the use of inappropriate functions to describe the DRC. Above ~100 Gy single grain dating is recommended.

4.2 INTRODUCTION

Optically stimulated luminescence (OSL) dating of quartz has been shown to produce reliable age estimates within the time frame of radiocarbon dating (~50 ka) (Murray and Olley, 2002). There are also examples of OSL studies in agreement with independent age control above 50 ka (e.g. Murray & Olley 2002; Murray et al. 2008; Pawley et al. 2010). However, a substantial number of quartz OSL studies conducted on samples with D_{es} above ~150 Gy (~40-100 ka and above) report underestimates compared to independent age control (e.g. Murray et al. 2007; Lowick et al. 2010a; Lai 2010; Timar et al. 2010; Lowick & Preusser 2011; Zander & Hilgers 2013). Samples used in these studies perform well in all standard reliability checks and provide no obvious cause for the observed underestimations. This suggests that the standard quartz protocols might not be appropriate at high doses.

OSL dating is based on the indirect measurement of accumulated light-sensitive charge trapped within the quartz crystal lattice. When quartz is stimulated with visible light this charge is released and is measured as luminescence. Only electrons which recombine with electron holes at radiative recombination centres emit luminescence and thus only a portion of the trapped charge is measured. Under the simplifying assumptions of first-order kinetics, a single-component OSL signal should grow with dose (i.e. produce a dose response curve; DRC) as a single saturating exponential (SSE) (Equation 4-1; Aitken, 1998)

Equation 4-1:

$$I = I_{\max} \left(1 - e^{-\frac{D}{D_0}} \right)$$

where I is luminescence intensity after a dose (D), $I_{\max}$ is the luminescence intensity at saturation and D_0 is the dose level which characterises the DRC. However, above ~100 Gy an extra linear or exponential component is frequently needed to fit the DRC and SSEs are seldom reported (Lowick and Preusser, 2011; Murray et al., 2008; Roberts and Duller, 2004).

The mechanism responsible for the extra component is not yet fully understood. Theories include: the creation of traps during irradiation which create a constantly growing signal (Yoshida et al., 2000); the creation of other defects during irradiation causing potentially uncorrected sensitivity change (Chawla et al., 1998; Fleming, 1979; Huntley and Prescott, 2001); the measurement of signal from more than one electron trap type meaning the assumption of first order kinetics is no longer applicable (Bailey et al., 2005; Lai et al., 2008); the competition with non-radiative recombination centres creating uncorrected sensitivity change (Lowick et al. 2010b) or the averaging of many of individual grains with different saturation properties (Adamiec, 2000; Roberts et al., 1999). Alternatively, extra components may be a function of high laboratory irradiation rates (Lai, 2010). 'Natural' DRCs have been shown to conform to a SSE whilst laboratory growth curves constructed from the same samples do not (Chapot et al., 2012; Timar-Gabor et al., 2015; Timar-Gabor and Wintle, 2013). These samples pass all routinely applied rejection criteria (see below), including dose recovery tests, and produce internally consistent age estimates from DRCs which do not saturate until

high doses. However, after ~150-200 Gy the age estimates start to deviate from the expected chronology and OSL ages increasingly underestimate the expected age (Chapot et al., 2012; Timar et al., 2010; Timar-Gabor and Wintle, 2013) with even infinitely old samples only reaching 60-80% of laboratory DRC saturation (Timar-Gabor et al., 2015). Lowick, et al (2010b) suggest that there is no obvious reason why the extra component in the DRC should cause the age underestimation observed in quartz and there are some studies that provide reliable age estimates in this portion of the dose curve (e.g. Pawley et al. 2010). However, for the studies which show unexplained age underestimations in comparison to independent age control (e.g. Lai, 2010; Lowick et al., 2010a; Lowick and Preusser, 2011; Murray et al., 2007; Timar et al., 2010; Zander and Hilgers, 2013), D_e estimations all fall within this portion of the growth curve. It is therefore important to investigate the origin of the extra curve fitting component and the impact of this component on D_e estimation accuracy.

In the absence of independent age control, quartz D_e estimates have conventionally been considered to be reliable if the aliquot is not in saturation and can pass a suite of rejection criteria. Typically these criteria include the recycling ratio (a repeat laboratory dose to test adequate correction for sensitivity change) (Murray and Wintle, 2000), the IR-depletion ratio test (to ensure the absence of feldspar contamination) (Duller, 2003), and a zero dose point (to check no charge is recuperated via thermal transfer). These tests are routinely included in the standard SAR measurement protocol (Murray & Wintle 2000; 2003) and are conducted on all aliquots. In addition, many studies include one or more dose

recovery tests conducted on a subset of samples. Dose recovery tests are independent from D_e measurement and determine how accurately a known laboratory administered dose can be recovered from bleached, but previously unmeasured, aliquots. This evaluates the reliability of the SAR protocol and determines whether sensitivity changes associated with heating and measuring the natural signal are corrected for by the test dose (Roberts et al., 1999; Wallinga et al., 2000; Wintle and Murray, 2006). More rigorous investigations may determine whether aliquots are fast dominated (i.e. the signal originates from the stable fast component (Bailey et al., 1997) via the fast ratio (Durcan and Duller, 2011) or via mathematical component fitting. Additional testing may evaluate the single grain properties of a sample with high overdispersion (scatter in D_e estimate above that expected from measurement error) and oversaturation (L_n/T_n values above saturation of L_x/T_x ; (Stone and Bailey, 2012) both indicators of potential D_e estimate inaccuracies. Due to time constraints, the reliability tests described above are typically performed at low doses which may not necessarily reflect behaviour at higher doses. Alternatively, these tests may not be applicable to high-dose behaviour and sources of inaccuracy may go undetected. This study investigates DRC behaviour at high doses with the aim of answering the following four questions.

1. Do standard SAR reliability tests indicate reliability problems at high doses?
2. Does minor feldspar contamination become significant at high doses?
3. To what extent do single grain DRC properties influence multi-grain aliquots?

4. What is the form of the multi-grain, single aliquot laboratory DRC up to 8000 Gy?

4.3 METHODS

Samples were extracted, prepared and measured using the methods described in section 3.3. All experiments followed the principle of the modified SAR protocol (Murray and Wintle, 2003) using the conditions described in Table 3-2. When IR-depletion ratio data was not available the fast ratio was used as a proxy for feldspar contamination (Section 3.3.7).

4.3.1 DOSE RECOVERY TESTS

Dose recovery tests were conducted following Wintle and Murray (2006). Samples were twice bleached using blue diodes (470 nm) for 1000 s at 20°C with a 10,000 s pause between bleaches. A 300 Gy or 500 Gy laboratory administered dose was given and measurement followed the SAR protocol. The given dose was chosen to approximate the natural signal of these samples.

4.3.2 IR RATIO TESTS

Two experiments were conducted to test whether minor feldspar contamination becomes more significant at high doses. Firstly, ten aliquots of three samples which pass the IR-depletion ratio but produce >0 IRSL were measured using an extended IR-depletion ratio test. Aliquots were measured eight times at three separate regenerative doses (20, 200, 400 Gy) with every fourth measurement

cycle including an IR-bleach before L_x . Additionally, a 400 Gy IR-depletion point was included alongside the standard 20 Gy IR-depletion point in the D_e measurement of 12 aliquots of six samples.

4.3.3 SINGLE GRAIN EXPERIMENTS

To assess the impact of individual grain behaviour on multi-grain aliquot DRCs, 500 grains of three samples were measured up to ~600 Gy using the SAR protocol (Table 3-2). Grains that produced a DRC were classified into one of four categories based on DRC behaviour (Table 4-1). Fast ratios (Duller, 2012) showed no systematic difference between DRC categories (Appendix B.1) indicating DRC shape was not a product of slow or medium component contamination.

Synthetic aliquots (see Jacobs et al. 2003) were created by summing the signal of each sample. Subsequent experiments used synthetic aliquots created by summing specific grains with specific DRC properties (details in Appendix B.2). Synthetic aliquots were fitted with a saturating exponential plus linear (SEPL) function. Grains from different samples were not mixed. To assess impact of DRC shape and fit on D_e accuracy L_n/T_n values were modified so they equalled the L_x/T_x values of the synthetic aliquot at 16.5, 99, 198, 396 and 594 Gy (i.e. five different L_n/T_n values for each curve). Therefore, if curve fitting is accurate D_e estimates should be identical to the given dose. Additionally, more realistic L_n/T_n values were calculated by normalising the L_n counts of each grain to the L_x/T_x values of the grain at 16.5, 99, 198, 396 and 594 Gy before summing the modified

L_n values and dividing by the summed T_x . This incorporated noise associated with varying grain sensitivity (Appendix B.5).

Table 4-1 Different DRC properties observed in single grain samples.

DRC type	Description	Example
1	Uncertainties mask form of DRC - Similar to “Dim grains” of Stone & Bailey (2010) - Most common DRC in this study - Range of DRCs and signal intensity	
2	SSE - Commonly described as accepted single grains (e.g. Jacobs et al. 2008) - Similar to “Class-2” grains of Yoshida et al. (2000) but naturals are not necessarily in saturation	
3	Continued growth (SEPL or DSE) - Commonly described as accepted single grains (e.g. Jacobs et al. 2008) - Similar to “Class-1” grains of Yoshida et al. (2000) but D_{es} are not necessarily high	
4	Non-monotonic - Commonly observed but not investigated in detail (e.g. Roberts et al. 1999; Jacobs et al. 2003; 2008; Tribolo & Stokes 2006; Duller 2012; Demuro et al. 2013) Thought to contribute to age underestimates (Tribolo & Stokes 2006)	

4.3.4 HIGH DOSE MEASUREMENTS

Eight samples from a range of depositional contexts (see Grine et al. 2007; Wood et al. 2012; Thomas et al. 2009; Telfer & Thomas 2006; Stokes & Bailey 2003; Hammond 2005; Burrough et al. 2007) were measured up to 8000 Gy by Richard Bailey using a modified SAR protocol including a 40 s, 160°C IR bleach before each L_x measurement. OSL data were decomposed into first-order fast, medium and a bulk slow components (c.f. Bailey et al. 1997) using a standard Levenberg-Marquardt algorithm.

4.4 RESULTS

4.4.1 DO STANDARD RELIABILITY TESTS INDICATE RELIABILITY PROBLEMS AT HIGH DOSES?

Samples NAM14/12/7, NAM14/5/4 and AUS13/1/1 can recover doses (within 10%) at 300 Gy, 300 Gy and 500 Gy respectively indicating T_n is accurately correcting for sensitivity change during heating and measurement of the natural signal and therefore suggesting no reliability problems (Figure 4-1). High D_e variation is observed for the 500 Gy given dose (AUS13/1/1 – overdispersion = 16%), possibly indicating uncertainty underestimation during interpolation onto the comparatively flat curve at this dose. DRC shapes are varied but show no relationship with the ability to recover the given dose (Figure 4-2). Therefore, test-dose correctable sensitivity change is not related to continued growth in the high dose region of the DRC.

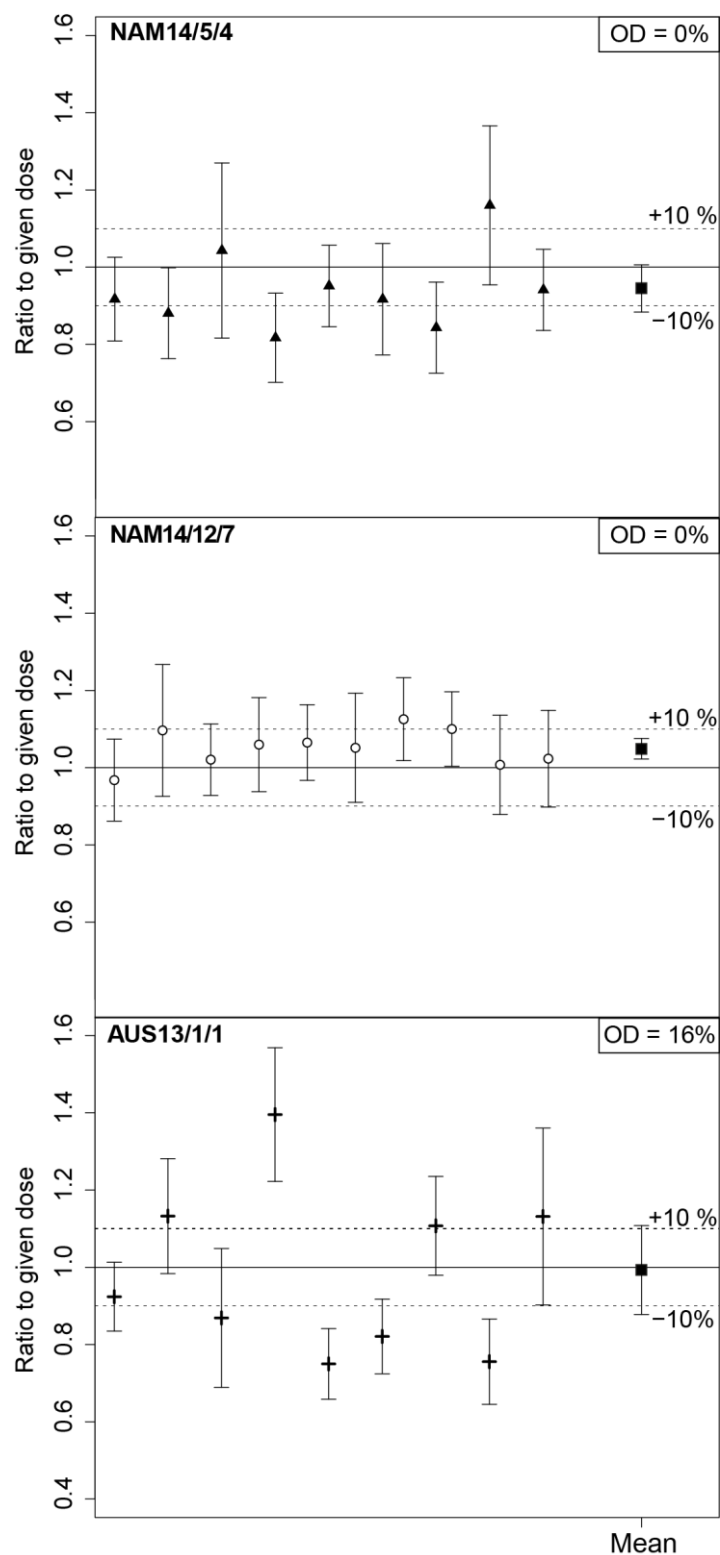


Figure 4-1. Results of multi-grain, single aliquot dose recovery tests. Samples NAM14/5/7 and NAM14/12/7 were given a 300Gy laboratory dose. Sample AUS13/1/1 was given a 500Gy dose. Black line indicates unity between given and recovered dose. Dotted lines mark 10% of unity. Black squares show the mean and standard deviation of measured aliquots. Triangles, hollow circles and pluses show individual aliquots and associated uncertainties of NAM14/5/4, NAM14/12/7 and AUS13/1/1 respectively

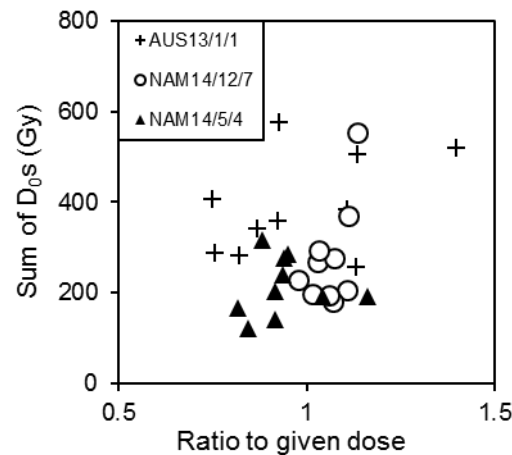


Figure 4-2. Sum of D_0 values from the double saturating exponential functions of each aliquot used in dose recovery tests plotted against the accuracy with which they recover a known given dose. No pattern is observed between curve shape and ability to recover a given dose.

4.4.2 DOES MINOR FELDSPAR CONTAMINATION BECOME SIGNIFICANT AT HIGH DOSES?

Feldspar DRCs can grow to over 1000 Gy (Thomsen et al., 2011) whilst the fast component of quartz has been reported to saturate at ~200 Gy (Wintle and Murray, 2006). Therefore, as D_0 values of impurities may be larger than for quartz, contamination may become a more significant signal contributor at higher doses. If this is the case, samples may pass the IR-depletion ratio at low doses but not at higher doses. This could explain continued DRC growth at high doses and, as feldspar exhibits anomalous fading (Spooner 1994), may explain age underestimations in some samples.

At 20 Gy all tested aliquots produced an IRSL signal above noise (150-1500 counts in the initial 1s) indicating the presence of impurities. Nevertheless, all aliquots in the repeated IR-depletion ratio test pass within errors at all doses

(based on average L_x/T_x values for both the recycling and IR-depletion ratio points). 5/72 aliquots used for D_e estimation fail the IR-depletion ratio at 20 Gy and a slightly different 5/72 fail at 400 Gy (Appendix B.3). The distribution of IR-depletion ratio results centres below unity for all doses (indicating some contamination) but the difference in IR ratio between doses is slight (Figure 4-3). No significant difference is found between groups using an ANOVA test (Appendix B.4). This suggests that minor contamination does not become significant at high doses. Therefore, contamination is not responsible for curve components above an SSE. This is corroborated by the lack of relationship between IR ratios and D_0 values, indicating D_0 is independent from impurity contribution (Figure 4-4). However, IR-depletion ratio values of the four aliquots which fail the IR ratio at both 20 Gy and 400 Gy show a decrease of ~10-20% at 400Gy compared to 20 Gy (Appendix B.3). This suggests that if a sample is already significantly contaminated the relative contribution from impurities may increase with dose.

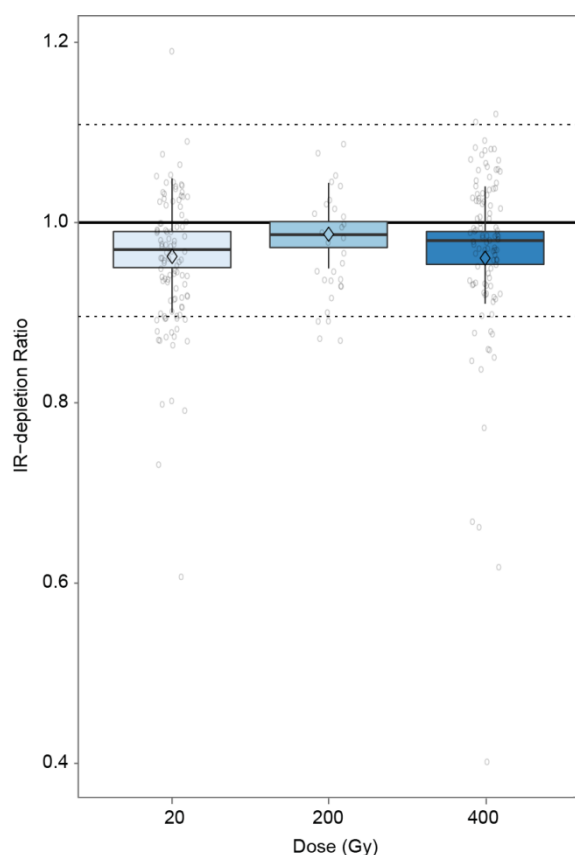


Figure 4-3 . Distribution of IR- depletion ratios by dose. Black line indicates unity between L_x/T_x values of the same dose measured with and without an IR bleach. Dashed lines mark 10% of unity. Filled box plots show the interquartile range of values, the black line shows the median and the diamond the mean. Hollow dots show individual results. The spread in these points is to give a clearer representation of distribution and does not represent any change in dose.

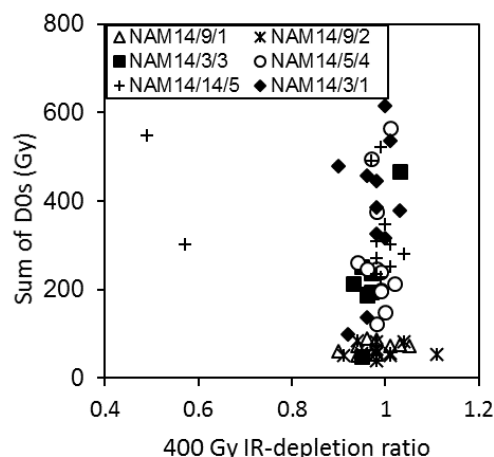


Figure 4-4. Sum of D_0 s from DSEs plotted against associated IR-depletion ratio at 400 Gy. No relationship exists between curve shape and IR-depletion ratio.

4.4.3 TO WHAT EXTENT DO SINGLE GRAIN DRC PROPERTIES INFLUENCE MULTI-GRAIN ALIQUOTS?

Single grain measurements show that >70% of the natural signal originates from <5% of the grains for all three samples (Figure 4-5A) but the proportion of signal from each DRC category is sample dependent (Figure 4-5B). Synthetic aliquots indicate that multi-grain aliquot DRCs are dependent on the grain composition with the removal or isolation of different DRC properties at the individual grain level producing significant changes in aliquot DRC shape (Figure 4-6). In this study DRC properties of constituent grains influences the accuracy of D_e estimates at and above 198 Gy (Figure 4-7). At 99 Gy and below synthetic aliquot behaviour is relatively homogenous (barring outliers- see Figure 4-7 caption) and D_e estimates are accurate (Figure 4-6 and Figure 4-7). By 198 Gy synthetic aliquot curves are significantly dispersed (Figure 4-6) with some curves producing accurate D_e estimates whilst others under- or overestimate. This difference in D_e estimate accuracy increases with curve divergence and therefore, dose (Figure 4-8 and Figure 4-9).

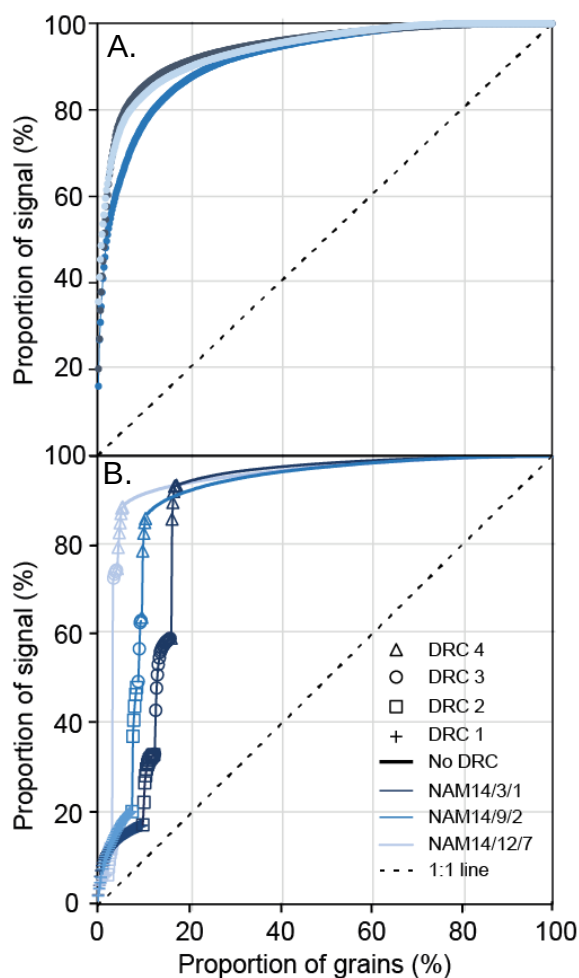


Figure 4-5. Proportion of signal from single grains. A). grains are ranked by percentage of total luminescence produced from a 99 Gy dose and are plotted cumulatively showing the signal contribution by proportion of grains for each sample. B) The data is replotted by single grain DRC type therefore showing the proportion of signal from each DRC type for each sample. Individual symbols represent individual grains.

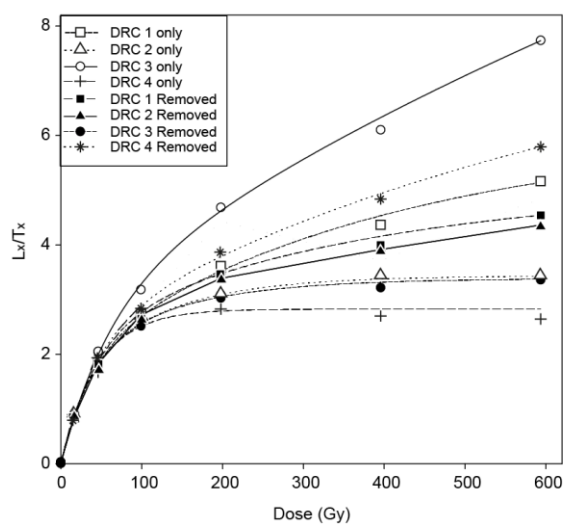


Figure 4-6 Synthetic aliquot DRCs of sample NAM14/3/1. The different curves are constructed by isolating or removing populations of grains with different DRC properties and highlight the impact of single grain properties on multi-grain DRCs.

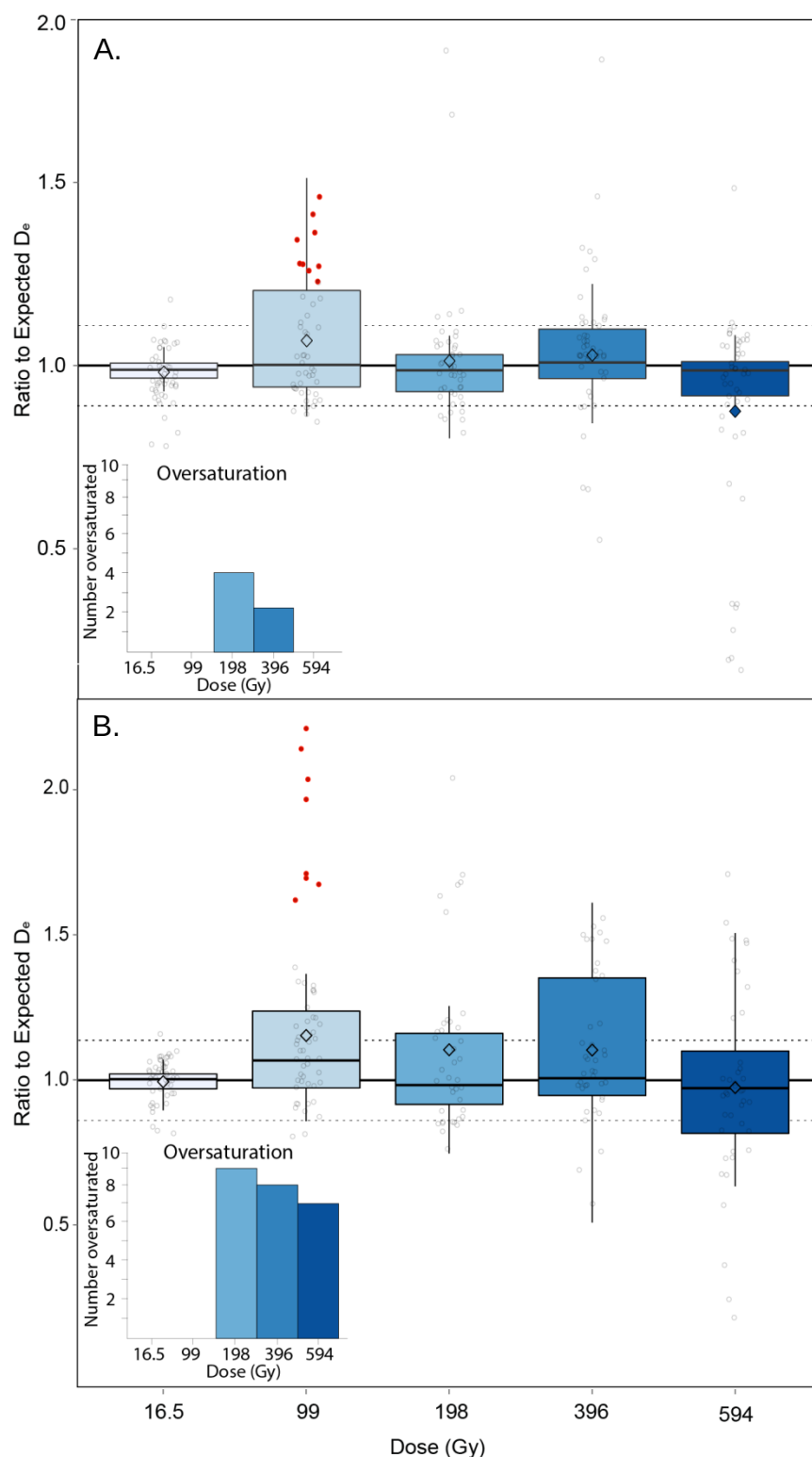


Figure 4-7 Ratio between D_e estimates of synthetic aliquots of varying composition and expected D_e . A.) L_n/T_n values of synthetic aliquots were normalised to L_x/T_x values of curve dose points at 16.5, 99, 198, 396 and 594 Gy. B.) L_n values of each grain normalised to L_x/T_x values of single grain dose point at 16.5, 99, 198, 396 and 594 Gy. Solid black line represents unity. Dashed lines mark 10% of unity. Box plots show interquartile range. Black lines represent the median and diamonds the mean. Hollow circles show individual data points. The spread in these points is to give a clearer representation of distribution and does not represent any change in dose. Insets show the number of oversaturated grains at each dose. These grains are not included in distribution box plots. Red dots represent outlier values caused by a very bright grain with an anomalous DRC at 99Gy. If these are removed the spread of values at 99Gy is greatly reduced.

The form of synthetic aliquot DRCs, and therefore associated D_e accuracy, is dependent on the relative contributions from DRC 1-4 grains (Figure 4-8 & Figure 4-9). DRC 4 (non-monotonic response) grains have the largest impact on D_e accuracy, with poor fitting and D_e inaccuracies observed in this study when $>\sim 30\%$ of the total signal originates from these grains (Figure 4-8A,B,C & Figure 4-9A,B,C). The dose at which non-monotonic DRCs reach their apex is likely to be a product of both the dose point values chosen for DRC construction and the sample measured. In this experiment, DRCs of DRC 4 grains typically peak at the 198 Gy dose point whilst SEPL fits plateau below this value. This produces apparent oversaturation at 198 Gy when $>\sim 60\%$ of the signal originates from DRC 4 grains (Figure 4-8 A,B,C). When the signal from DRC 4 grains is between $\sim 30\text{--}60\%$ the nature of the remaining grains influences DRC behaviour. When combined with $\sim 30\text{--}60\%$ of signal from DRC 2 grains D_e overestimations are observed at 198 Gy. This is because DRC 2 grains flatten the synthetic DRC and decrease L_n/T_n values, improving the fit but still producing overestimates due to the high L_n/T_n values of the DRC 4 grains (Figure 4-8B). DRC 3 grains have a positive effect on D_e accuracy with accurate D_e estimates at 198 Gy when 30-60% DRC 4 grains are combined with $>25\%$ DRC 3 grains (Figure 4-8C.). DRC 1 grains have an unpredictable influence (Figure 4-8A).

At doses of 396 Gy (not shown) and 594 Gy, D_e underestimations are observed when $>\sim 60\%$ of the signal originates from DRC 4 grains (Figure 4-9,B,C). This is because L_n/T_n values are decreased by the non-monotonic grains and interpolation fits to the first, lower value, intercept with the curve. When the signal

from DRC 4 grains contribute ~30-60% the presence of 30-60% DRC 2 grains exacerbates underestimations as L_n/T_n values are further decreased by saturated values (Figure 4-9B) whilst a significant contribution from DRC 3 grains improves the accuracy.

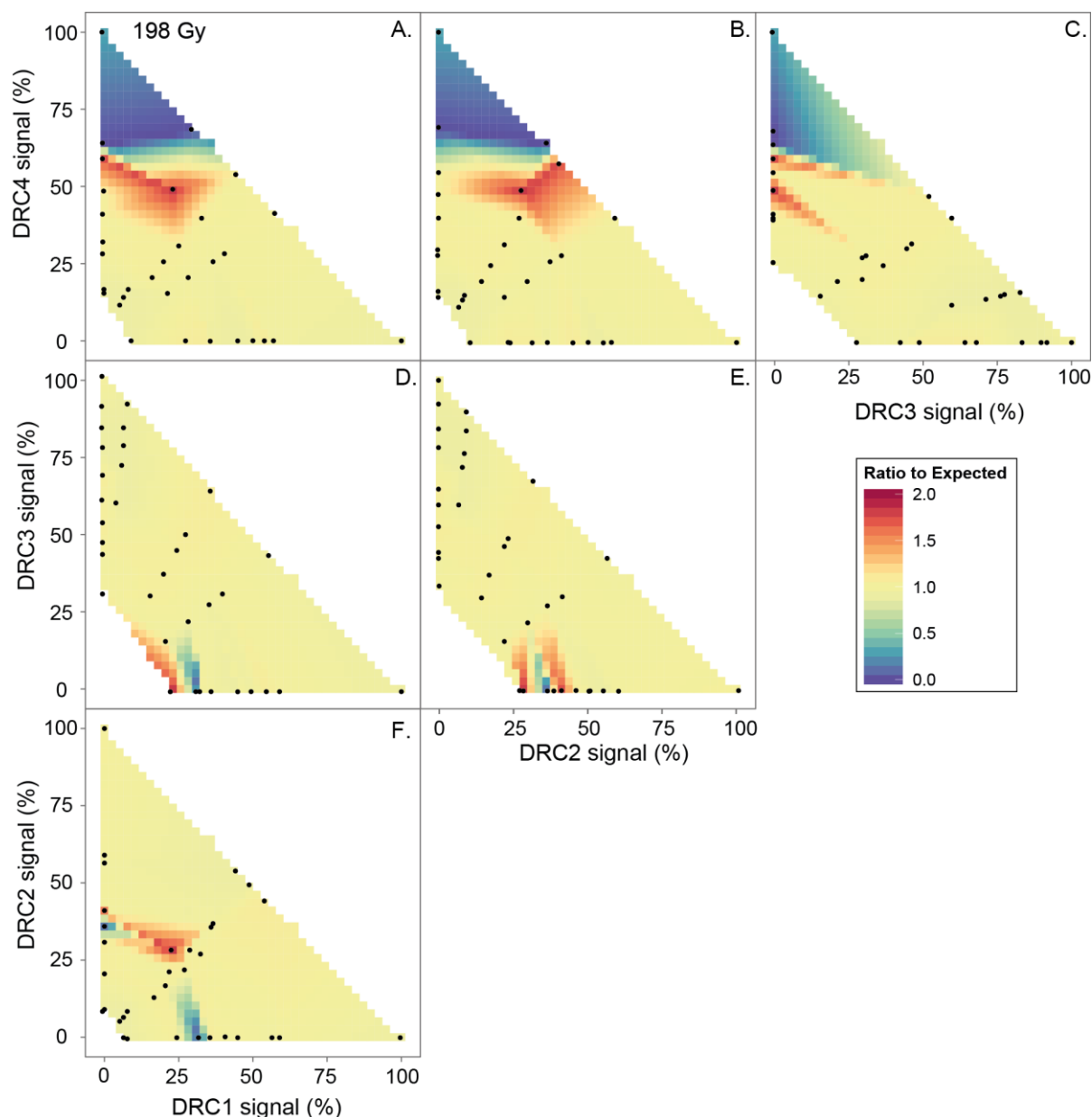


Figure 4-8. Accuracy of synthetic aliquots at 198 Gy by the proportion of signal from different grain types. L_n/T_n values of each aliquot were normalised to L_x/T_x values at 198 Gy therefore accuracy reflects the applicability of the DRC fitting and interpolation. Yellow indicates D_e estimates within 10% of the expected dose. Red indicates overestimation and blue indicates underestimations. Oversaturated synthetic aliquots were given a value of zero and are represented in dark blue. Black circles show the data points used to construct contour plots. Interpolation between points was performed in R using the “interp” function of the “akima” package (Akima and Gebhardt, 2015). As only two grain types are depicted in each plot patterns observed at low contribution from one or both grain types can be explained by the presence of other grain types.

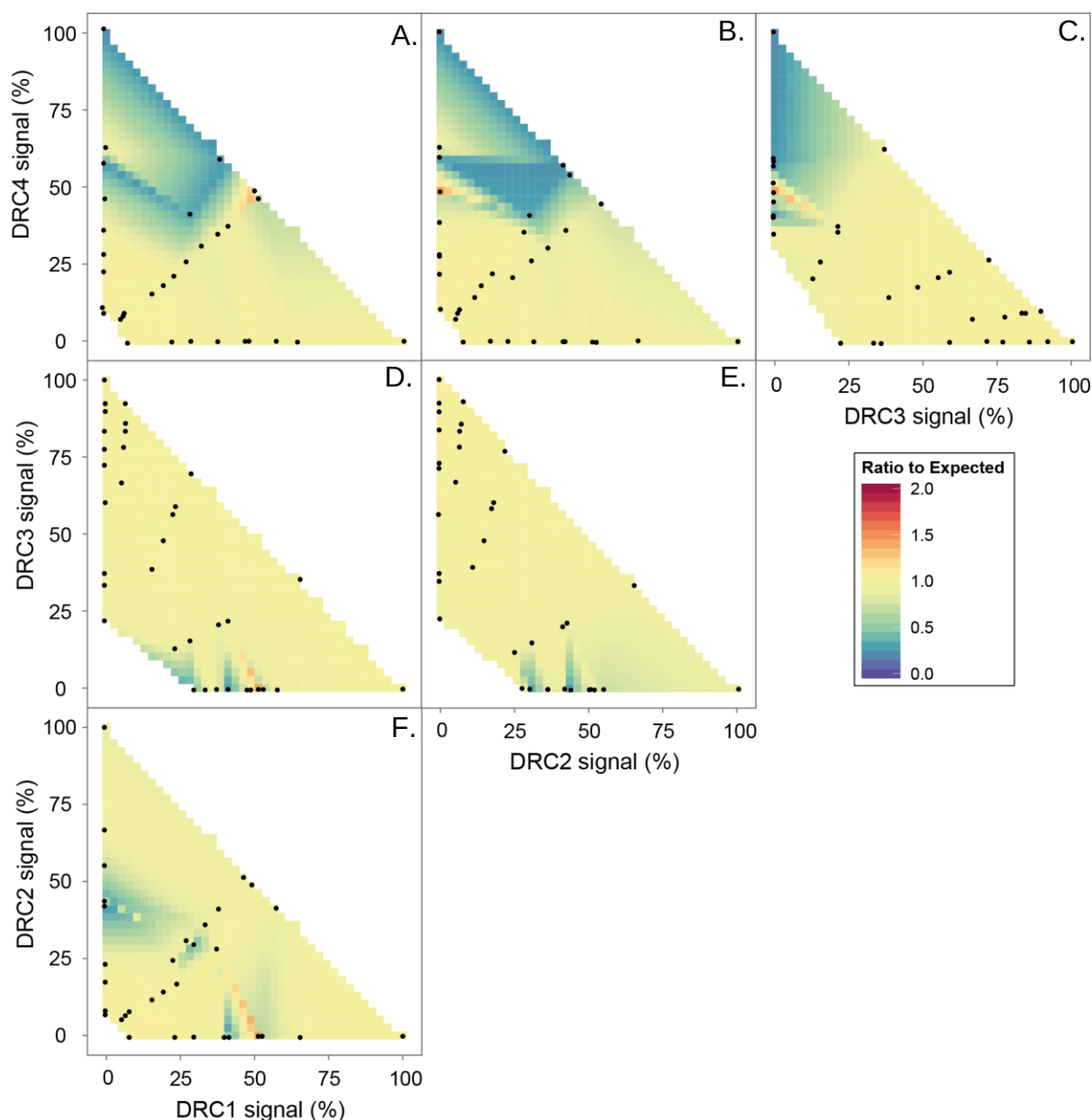


Figure 4-9. Accuracy of synthetic aliquots at 594 Gy by the proportion of signal from different grain types. L_n/T_n values of each aliquot were normalised to L_x/T_x values at 594 Gy therefore accuracy reflects the applicability of the DRC fitting and interpolation. Yellow indicates D_e estimates within 10% of the expected dose. Red indicates overestimation and blue indicates underestimations. Oversaturated synthetic aliquots were given a value of zero and are represented in dark blue. Black circles show the data points used to construct contour plots. Interpolation between points was performed in R using the “interp” function of the “akima” package (Akima and Gebhardt, 2015). As only two grain types are depicted in each plot patterns observed at low contribution from one or both grain types can be explained by the presence of other grain types.

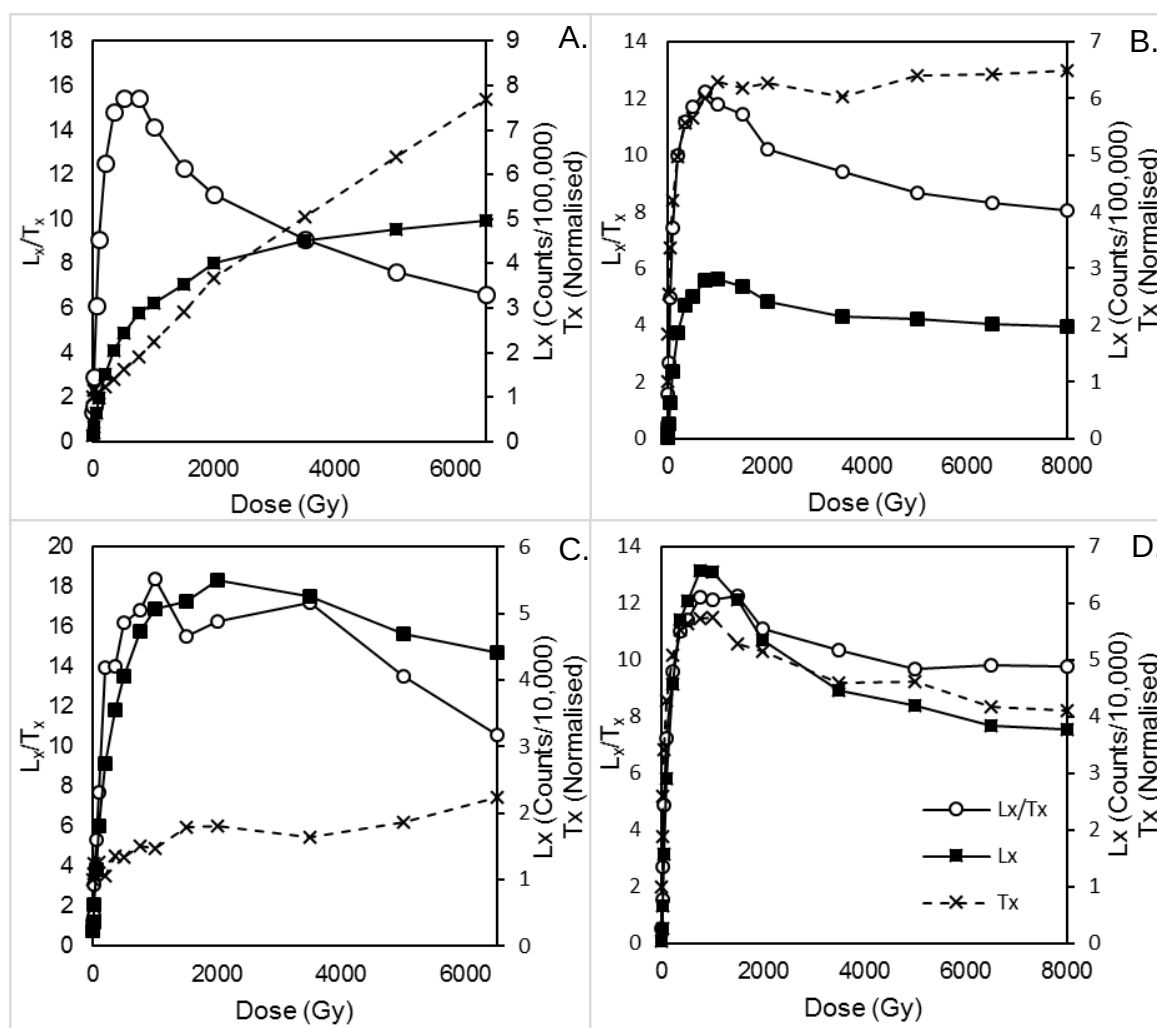


Figure 4-10. Example High dose DRCs A.&B. show bulk signals with L_x/T_x non-monotonic growth driven by increasing T_x values(A) and decreasing L_x values(B). C. &D. show fast component only curves with L_x/T_x non-monotonic growth driven by decreasing L_x values.

4.4.4 WHAT IS THE FORM OF THE MULTI-GRAIN, SINGLE ALIQUOT LABORATORY DRC UP TO 8000 Gy?

Distinct non-monotonic growth after 750-1500 Gy is found in the majority of samples but DRC shape is sample specific (Appendix B.6). Non-monotonic growth driven by either strong T_x growth with increasing dose (Figure 4-10A) or decreasing L_x values with dose (Figure 4-10B). However, when the fast component is isolated T_x values plateau and 7 out of 8 samples clearly show non-monotonic L_x values (e.g. Figure 4-10C & D). This indicates that non-monotonic growth is primarily a fast component phenomenon driven by a decrease in L_x

values at high doses but can sometimes be exacerbated by a residual build-up of the slow or medium component in T_x . As this phenomenon only appears to happen at high doses it is not observed in, and therefore cannot be corrected by, the low dose test dose.

4.5 DISCUSSION

Multiple investigations were undertaken to explore the reliability of ≥ 100 Gy D_e estimates obtained from DRC curves which show continued growth into the high dose region. As in the majority of dating applications, no independent age estimates are available for these samples. Therefore, conclusions can only be based on the various measures of internal consistency and on the presence/absence of potentially complicating factors.

4.5.1 DO STANDARD SAR RELIABILITY TESTS INDICATE RELIABILITY PROBLEMS AT HIGH DOSES?

Samples performed well in standard SAR reliability tests (Figure 4-1) indicating that the current standard SAR rejection criteria are not capable of identifying samples yielding D_e underestimates at high doses. This observation is consistent with previous studies where this effect was observed (e.g. Murray et al. 2007; Lowick et al. 2010a; Lai 2010; Timar et al. 2010; Lowick & Preusser 2011; Chapot et al. 2012; Zander & Hilgers 2013).

4.5.2 DOES MINOR FELDSPAR CONTAMINATION BECOME SIGNIFICANT AT HIGH DOSES?

High dose IR-depletion ratio tests indicate that minor feldspar contamination is not the cause of additional DRC growth beyond a SSE. (Figure 4-3). This is in agreement with Timar-Gabor et al. (2015) who show that continued growth in the high dose region for fine grains is a purely quartz based phenomenon. Furthermore, fast component only dose response curves suggest that continued growth is a feature of the fast component (Lowick et al 2010b; Chapot et al. 2012). However, the signal contribution from impurities can increase with dose if an aliquot is significantly contaminated at low doses (Appendix B.3) i.e. there is some support for the proposition that the differing D_0 properties of quartz and feldspar impact DRC construction at high doses, but only where contamination is detectable at the relatively low doses typically used to perform the IR-depletion ratio test (Duller, 2003). Consequently, the IR-depletion ratio test as currently applied is a sufficient test for feldspar contamination in high D_e samples.

4.5.3 TO WHAT EXTENT DO SINGLE GRAIN DRC PROPERTIES INFLUENCE MULTI-GRAIN ALIQUOTS?

Single-grain OSL properties are highly variable due to both extrinsic (i.e. environmental factors unrelated to the grain e.g. beta heterogeneity, partial bleaching, sediment mixing) and/or intrinsic factors (i.e. variation in the properties of the grains themselves) (Stone and Bailey, 2012). Modelled and real-world data have shown that averaging effects can create 'phantom' D_e solutions in multigrain aliquots (Arnold et al., 2012; Arnold and Roberts, 2009; Russell and Armitage,

2012) and Roberts et al. (1999) suggest that variation between single-grains, particularly in regard to saturation properties may explain age inaccuracies and the need for an extra curve fitting function when describing multi-grain aliquots at high doses. This suggestion is supported by D_e -normalised synthetic aliquots containing grains with different growth-curve forms.

In this study, DRC properties of single grains influence the accuracy of D_e estimates from synthetic aliquots at ~200 Gy and above, and there is a detrimental effect on DRC fitting when >30% of total signal originates from grains with non-monotonic dose-responses. This effect produces inaccurate D_e estimates or oversaturation. Consequently, a large population of non-monotonic grains in multigrain aliquots could explain age underestimations reported at high doses, since old grains with non-monotonic growth properties have lower than expected L_n values, reducing the D_e estimate. Additionally, a large population of non-monotonic grains decreases the accuracy of the curve fitting (Figure 4-8 and Figure 4-9). Averaging effects may explain the observation that for old sedimentary sequences, OSL yields increasing age underestimations and uncertainties with depth, but without the occurrence of age reversals (e.g. Kreutzer et al., 2012; Lai, 2010). In this scenario non-monotonic grains would dampen L_n values with increasing dose, but the monotonic population would enable L_n values to continue to grow. If >50% of total signal comes from grains which show continued growth to high doses, curve fitting is more likely to be correct and therefore D_e estimates are more likely to be accurate. These grains are also less prone to oversaturation (only 13% of DRC 3 grains were

oversaturated compared to 42% and 57% of DRC4 and DRC2 categories respectively). Therefore, fortuitous sampling of a large population of grains with favourable dose properties (i.e. continued growth or SSE before saturation) could explain why some single aliquot studies produce accurate age estimates at high doses.

4.5.4 WHAT IS THE FORM OF THE MULTI-GRAIN, SINGLE ALIQUOT LABORATORY DRC UP TO 8000 Gy?

Lowick et al. (2010b) suggest that the linear component of a SEPL function represents the start of a second saturating exponential. However, these curves were only constructed to 1000 Gy. Above 750-1500 Gy, ~90% of the multi-grain, single aliquot samples measured in this study begin to show clear non-monotonic growth. Non-monotonic growth at high doses has been frequently described in TL studies of many luminescing materials, including quartz (see Lawless et al. 2005 for review). The same response has also been modelled for OSL (Pagonis et al., 2006) and empirically observed in electron spin resonance (ESR) dating of quartz (Woda and Wagner, 2007).

The cause of non-monotonic growth is debated (Chen et al., 2015; Lawless et al., 2005; Pagonis et al., 2006; Woda and Wagner, 2007). Competition during the excitation phase produces non-monotonic growth in models of TL and OSL behaviour if a single electron trap and two recombination centres, one radiative and one non-radiative, is assumed. In this scenario, recombination occurs at high doses during excitation, decreasing the number of radiative recombination centres

during stimulation and reducing the total luminescence produced (Lawless et al., 2005; Pagonis et al., 2006). Non-monotonic growth can also be produced if two electron traps and two recombination centres are assumed. However, in this instance the shape of the DRC is very different with non-monotonic growth occurring very abruptly (Pagonis et al., 2006). The presence of non-radioactive recombination centres has been demonstrated in quartz (Bailey, 2001; Zimmerman, 1971) thus these provide a plausible explanation for the non-monotonic growth observed in quartz OSL.

However, the forms of empirical quartz electron spin resonance (ESR) DRCs do not fit these models (Woda and Wagner, 2007) and instead DRC shape is ascribed to competition between single and double electron capture. During double electron capture recombination occurs outside recombination centres, producing no signal but reducing the quantity of trapped electrons. As the likelihood of this occurring increases with dose, total luminescent recombination decreases and non-monotonic growth is produced.

Recently Chen et al. (2015) have suggested that non-monotonic growth can also be produced during the excitation phase if only one radiative centre and two electron traps are assumed. In this model, thermal fading is the primary driver, and non-monotonic growth is a function of stimulation time rather than dose. Slower dose rates are demonstrated to produce a greater degree of non-monotonic growth therefore, natural D_e estimates are expected to suffer from greater dose depletion than laboratory doses.

Irrespective of the physical mechanism responsible, non-monotonic DRC forms draw into question the reliability of age estimates at high doses. All proposed mechanisms for non-monotonic growth suggest that it is likely to be present in the natural signal, perhaps even to a greater degree than in laboratory doses (Chen et al., 2015). This is potentially supported by the natural dose response curves constructed by Chapot et al. (2012). Therefore, if natural signals exhibit non-monotonic growth then above a certain sample specific level L_n/T_n values may have multiple DRC intercepts, i.e. very different absorbed doses produce identical natural response. This, coupled with averaging effects from single grains, may explain why: 1) some studies report age underestimations at high doses (e.g. Lowick et al., 2010a; Murray et al., 2008) whilst others produce D_e estimates in line with expected ages (e.g. Murray et al., 2007; Pawley et al., 2010), 2) why infinitely old samples give D_e estimates below saturation (e.g. Timar-Gabor et al., 2015) and 3) why samples which produce age underestimates can recover a known given dose (e.g. Murray et al., 2007).

4.5.5 LIMITS OF ACCURATE DATING

The starting point and degree of non-monotonic growth is sample and grain specific, and therefore assigning a value to the dose at which D_e inaccuracy might be expected is problematic. When a DRC is fitted with an SSE, $2xD_0$ has been suggested as the threshold above which D_e estimates may be inaccurate (Wintle and Murray, 2006). However, when DRCs do not conform to a SSE this is not a meaningful threshold (Durcan, 2012). Chapot (2014) suggests $0.85(I_{NAT})$ (i.e 85% saturation of the natural growth curve) should be used as the upper dating limit

where calculable, and ~150 Gy used if I_{NAT} cannot be evaluated. DRCs of single grains in this study diverge, to the extent that D_e accuracy is affected, between 100-200 Gy. Therefore, after ~100 Gy multi-grain dating results are likely to be highly sensitive to the presence and contributions of different grain types. Above this level single grain analysis, either as a screening tool to demonstrate the absence of grains with non-monotonic DRCs (see Stone and Bailey, 2012), or as a dating method, may be appropriate.

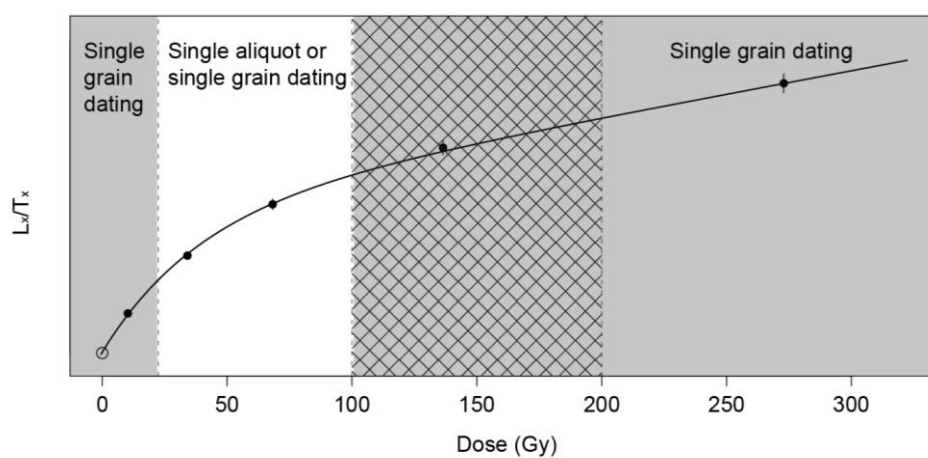


Figure 4-11 DRC regions best suited for single grain dating. Shaded areas represent doses best suited to single grain dating. The upper limit for single aliquot dating is likely to be sample specific (represented by the cross-hatched region). Samples used in this study indicate single grain dating should be used above ~100 Gy wherein components above an SSE are needed to fit the DRC.

Single grain dating is typically applied to young samples as unrepresentative grains (e.g. partially bleached grains) can be identified, increasing the accuracy of D_e estimates (e.g. Olley et al. 1999). The same is true at high doses (e.g. Jacobs et al. 2006) (Figure 4-11). Accurate single grain D_e estimates up to ~500 Gy have been recovered using unsaturated ‘super-grains’ (DRC 3 grains in this study) (Arnold et al., 2016; Demuro et al., 2015; Yoshida et al., 2000). At ~100 Gy single grain studies using standard rejection criteria can underestimate D_e estimates but accurate D_e estimates can be obtained if $D_0 \geq D_e$ and if the FR (see Duller, 2012;

Durcan and Duller, 2011) is >4 (Thomsen et al., 2016). We therefore suggest that single grain dating should be used at ~ 100 Gy and above with only grains with favourable DRC properties used for dating. If $>30\%$ of signal originates from non-monotonic grains then averaging effects are likely to produce inaccuracies in multi-grain aliquots. If $>50\%$ of signal is from DRC 3 (continually growing DRC) grains and $<30\%$ from non-monotonic grains the sample may be suitable for multi-aliquot dating. However, single grain analysis using known age samples is needed to fully investigate the relationship between DRC properties and age accuracy.

4.6 CONCLUSION

Multiple tests were carried out on naturally high-dose samples to investigate the reliability of non-saturating, high-dose quartz DRCs.

- Standard SAR reliability tests, including DRTs, did not indicate problems at high doses. Therefore, D_e inaccuracies associated with high doses may go undetected under standard measurement procedures.
- Minor feldspar contamination did not become significant at high doses indicating that continued growth in the high-dose region of the DRC is a quartz phenomenon. When samples had significant feldspar contamination, low dose IR-depletion tests adequately identified contaminated samples.

- The DRCs of single grains were highly varied, with some displaying non-monotonic growth from ~200 Gy onwards whilst others grew monotonically to at least 600 Gy. DRCs started to significantly diverge at ~100-200 Gy, above this point the proportion of different grain types in synthetic aliquots strongly influenced the accuracy of D_e estimates. Poor fitting and inaccuracies were observed when >30% of the signal originated from non-monotonic grains.
- 7/8 multi-grain, single-aliquot samples measured showed clear non-monotonic growth of the fast component above 1000 Gy.

Based on these results, the averaging of different single grain behaviours (Roberts et al., 1999) provides a plausible explanation for age inaccuracies at high doses. This is primarily driven by grain-specific, non-monotonic growth. The mechanism for non-monotonic growth is not well understood, but for our samples the effects of their presence are clear. On the individual curve level very different doses produce the same L_x/T_x values, producing two possible DRC intercepts and providing a possible explanation for age underestimations reported in the literature. Above a sample specific level (~100 Gy in this study) multi-grain aliquots with significant signal from non-monotonic grains are likely to suffer from poor fitting and the L_n value is likely to be reduced, creating D_e underestimates. Therefore, unless screening demonstrates that grains with non-monotonic DRCs are not present in a sample, single-grain analysis is the most appropriate D_e measurement method above ~100 Gy.

Part 2: Investigations and Interpretations

Chapter 5: Controls on sand ramp formation in southern Namibia

Fundamental to understanding the utility of sand ramps as palaeoenvironmental archives, this paper uses a suite of 75 sand ramps to explore the controls on sand ramp formation and the implication of this for climatic reconstruction.

5.1 ABSTRACT (FROM PAPER SUBMITTED TO 'EARTH SURFACE PROCESSES AND LANDFORMS' – AUTHORS: A.L.K. ROWELL, D.S.G. THOMAS, R.M. BAILEY, E. GARZANTI AND M. PADOAN)

Sand ramps have the potential to provide rich palaeoenvironmental information in dryland regions wherein proxy records are typically scarce. However, current knowledge of the geomorphic controls and processes of sand ramp formation is limited. This study provides a data-rich examination of the key factors controlling sand ramp formation. The location and morphology of 75 sand ramps in southern Namibia are examined. The sediments and chronologies of ten sand ramps are studied in detail using 51 OSL dates and 83 grain-size and LOI samples. Heavy mineral assemblages are used to determine the provenance of 10 samples and OSL sensitivity is used to explore geomorphic processes of 8 samples.

Sand ramp morphology can be grouped into one of four classes of increasing size and complexity and is closely linked to the available accommodation space. Heavy mineral assemblages indicate local sediment sources and all 75 studied sand ramps are within 4 km of a large ephemeral river channel or within 5.5 km of a dune field. Therefore, accommodation space and sediment supply are identified as the primary controls of sand ramp formation. Sedimentology and OSL sensitivity suggest a complex interplay of aeolian, fluvial and colluvial processes contribute to sand ramp formation with large variability observed between ramps. Three of the ten dated sand ramps have been present in the Namibian landscape for >100 ka. Eight sand ramps show episodic deposition between >75-12 ka and five show evidence of surface reworking over the past 2 ka. Environmental sensitivity is likely to be linked to the size and availability of the accommodation space. Therefore, individual sand ramps are expected to reflect local environmental conditions, recording when an abundant sediment supply coincided with available accommodation space, whilst a regional analysis of multiple sand ramps with chronometric data offers the potential to identify larger scale palaeoenvironmental controls of sediment supply.

5.2 INTRODUCTION

Studies of Quaternary environmental change in dryland regions are often hampered by a paucity of preserved proxy records (Thomas and Burrough, 2012). The arid climate inhibits the preservation of organic proxies typically used for palaeoenvironmental reconstruction, whilst ubiquitous 'geoproxy' records (e.g. dunes) can be difficult to interpret (Chase, 2009; Chase and Meadows, 2007; Stone and Thomas, 2008; Thomas and Burrough, 2012). Sand ramps are relatively common but under-investigated landforms which have the potential to provide valuable information on past sediment dynamics (Lancaster and Tchakerian, 1996). Formed when aeolian sediment is trapped against a topographic barrier which actively contributes to accumulation, sand ramps contain varying proportions of aeolian, hillslope and fluvial sediments (Bateman et al., 2012; Lancaster and Tchakerian, 1996; Thomas et al., 1997a). As such, they occupy the continuum between purely aeolian topographic features (e.g. climbing/falling dunes) and purely hillslope deposits (e.g. alluvial fans, talus cones) (Figure 5-1). The presence of the topographic barrier, which stabilises sediments and protects them from aeolian erosion (Chase, 2009), and the presence of sediments deposited under different geomorphic conditions means sand ramps have the potential to provide long and detailed palaeoenvironmental records which can be independently dated using OSL. However, relatively little is known about the geomorphic controls and processes that govern their formation and as such their current utility as palaeoenvironmental archives is limited.

The majority of investigations into the environmental significance and geomorphic controls of sand ramp formation have been conducted on single features (e.g.

Thomas et al. 1997; Bateman et al. 2012; Telfer et al. 2012). As sand ramps have been shown to display high regional variability (Lancaster and Tchakerian, 1996) it is unclear how representative these single ramp studies are. This paper presents a systematic and data-rich analysis of a suite of sand ramps from south west Namibia.

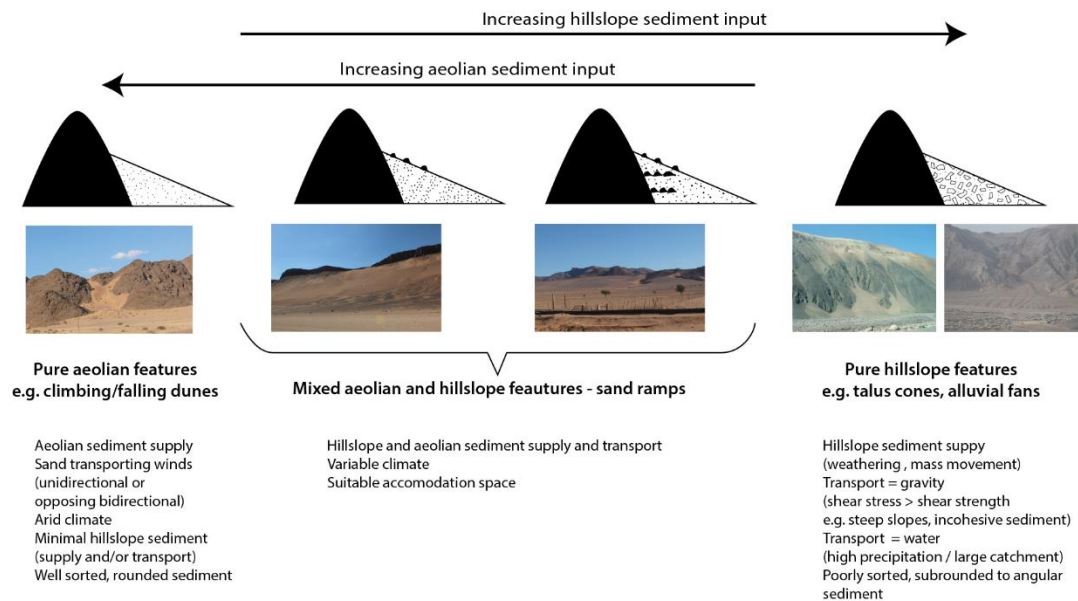


Figure 5-1. Summary of the main conditions needed for sand ramp formation and the relationship between sand ramps, and aeolian and hillslope features.

5.2.1 PREVIOUS SAND RAMP STUDIES

To date, sand ramp research has been limited. The majority of work has been conducted in the Mojave Desert, USA (Bateman et al., 2012; Lancaster and Tchakerian, 1996; Pease and Tchakerian, 2003; Rendell and Sheffer, 1996) but studies also exist from African, Middle Eastern, Asian and European and other American drylands (Berking and Schütt, 2011; Bertram, 2003; Clemmensen et al., 1997; Ellwein et al., 2015; Kumar et al., 2016; Telfer et al., 2012, 2014; Thomas et al., 1997a; Turner and Makhlof, 2002; del Valle et al., 2016). Table 5-1 summarises work to date.

Table 5-1. Summary of previous sand ramp research

Study & location	#	L (m)	T (m)	C/F	Sedimentology	Talus mantle	Gullies	Origin of sands	Chronology	Interpretation	Other
Lancaster and Tchakerian 1996, Rendell and Sheffer, 1996, Pease and Tchakerian, 2003 Mojave Desert, USA	13	100 - 2000	5-70	C&F	All ramps predominately aeolian. Varying proportions of fluvial, talus and debris flow sediments between ramps. No clear regional pattern. Aeolian sediments contain 0-5% coarse sand / gravel. Most sediments weakly cemented by calcium carbonate some evidence of palaeosols	On majority. More developed adjacent to steep slopes of metamorphic rocks. Talus adjacent to granitic rocks is mostly coarse grus	Some ramps are connected to the rock slope whilst others are separated by a trough. Thought to be associated with airflow dynamics in some circumstances and fluvial run-off in others.	Presumed local based on poor sorting. Geochemistry suggests a local fluvial/playa source for the sediments but the origin is likely to have altered slightly throughout accumulation history, possibly reflecting changes in climate dynamics.	9/13 ramps dated using MAAR quartz and feldspar TL and SAAD feldspar IRSL. 49 samples in total. Agreement between techniques is variable. Chronologies vary between ramps. Deposition between 0-45 ka. Predominantly episodic.	Different units represent discrete climatic periods. Talus accumulations within the ramp represent periods of stability. Dates suggest correlation to periods of increased fluvial activity, suggesting sediment supply is the key control on formation. Lack of Late Holocene dates are assumed a result of increased aridity.	Most relic, some currently active.
Bateman et al 2012 Soldier Mountain - Mojave Desert, USA	1	n/m	~25	C	Four sand ramp units above debris flow unit. Sand ramp units include cross-bedded Aeolian sands, discontinuous clast layers and calcrete layers. Evidence for aeolian, fluvial and colluvial deposition	Tightly packed angular clasts (1-2 clasts thick, clast-size decreases downslope. Average size 40-60 mm	Several gullies dissect the sand ramp laterally. No gully between the sand ramp and the hillslope.	Nearby (0.5km) Mojave River assumed to be main source. Also palaeolake Manix and Cayote.	6 quartz SAR OSL dates. Underlying debris flow dates to 15.0 +/- 1.0. Main sand ramp dates are between 11.6 +/- 0.8 – 9.5 +/- 0.6 ka (although likely to be 11.6 – 10.3 ka). Stone horizons formed in a maximum of 1.8 ka.	Sand ramps form quickly in a "window of opportunity" when accommodation space and sediment supply are both available. They do not represent a simple model of discrete climatic events. Stone layers do not necessarily represent former land surfaces and periods of climatic stability. Instead, they are likely formed by high-magnitude, low-frequency events with some redistribution via fluvial-aeolian creep.	Free dunes on top of ramp.
Ellwein et al 2015 Black Mesa, Arizona, USA	3	n/m	>12 m	C&F	75% Aeolian sands – well sorted with cross-bedding. Interspersed with alluvium and colluvium	no	no	Assumed to be local river channels	7 quartz SAR OSL dates. Top 6 m of sand deposited between 16.2 +/- 1.14 – 1.9 +/- 0.2 (majority before 7.9 +/- 0.55). lower >6m undated.	Associated with climatic variability under cooler, wetter, windier conditions than present.	Suggest sand ramps should be defined by morphology i.e. gently sloping wedges against a topographic barrier
Clemmensen et al 1997 Mallorca, Western Mediterranean	1	40	20	C	Predominantly aeolian dune sands interspersed with scree deposits.	no	no	Beach sands	n/a	n/a	

Study & location	#	L (m)	T (m)	C/F	Sedimentology	Talus mantle	Gullies	Origin of sands	Chronology	Interpretation	Other
Thomas et al 1997 Ardakan, Iran	1		25	C&F	Predominantly aeolian showing evidence of cross bedding interdigitated with clastic layers up to 40 cm thick. All units dip parallel to current surface. Aeolian sands well sorted, talus angular and often supported by interstitial sand	30 cm of mixed rock fragments and sand.	Dissect sand ramp from hillslope on both sides of the mountain saddle.	No modern sediment supply. Assumed ephemeral river channels to the SE under changed wind regimes.	7 MAAD and SAAD quartz samples. All 25-18 ka	Rapid accumulation in a single episode associated with the LGM. Units do not represent different climatic regimes. Wind dynamics during formation were likely different to present. Associated with cold dry conditions	Suggest sand ramps likely to have low angle surfaces whilst obstacle dunes have steep surfaces almost at the angle of repose (30-35°)
Turner and Makalouf 2002 Badia, Jordan	5	10-20	n/m	C	Primarily colluvium from adjacent sandstone. Small aeolian contribution. Colluvium deposition triggered by wind erosion.	n/m	n/m	Adjacent cliffs	Observation – rapid current accumulation	Sand ramps form rapidly from rockfall, grain flow and debris flow events. Once the sand ramp reaches a certain size it protects the hillslope from aeolian erosion and therefore ceases to grow any further. Increased proportions of rockfall sediments may indicate moisture availability as grainflows are inhibited by moisture.	Contemporary accumulation
Berking and Schütt (2011) Northern Sudan	4	200-500	n/m	C&F	Predominantly aeolian with talus mantle	Thin layer of pebbles and gravels	Dissect sand ramps from hillslope and downslope	n/m	10 Quartz SAR OSL. Dates taken from the top 1.3 m of 4 sites. Ages in stratigraphic order at each site. Ages between 0.06 ± 0.02 – 38 ± 5 ka.	Formation controlled by sediment supply. Deposition associated with aridity during the Last Glacial	
Bertram 2003 Namib Desert, Namibia	5*	280-2000	Up to 140	C	Primarily aeolian sands, discrete clastic layers and clasts sporadically within aeolian units. Evidence of root casts and palaeosols. Calcrete nodules on smaller ramps and multiple calcrete layers on large Aus ramp	Found on the majority of ramps	Dissect sand ramps from hillslope. Aus is also dissected downslope	Assumed ephemeral river channels and weathering of adjacent bedrock	3 SAR IRSL dates of 5 aliquots. Main body of Aus dated to 106.1 ± 7.9 at 3m and 104.9 ± 8.7 at 7m. Overlying dunes dated to 15.0 ± 1.1	Formation is a process of sediment mixing: sand blown upslope by aeolian processes and washed down by heavy rain which also deposits bedrock clasts. Sand ramps form under strong winds and flashy rainfall before stabilising under low intensity rainfall and weak winds. Accumulation occurred in a single phase for small ramps but over at least 8 phases of sedimentation for large Aus ramp.	Dunes overlaying some sand ramps

Study & location	#	L (m)	T (m)	C/F	Sedimentology	Talus mantle	Gullies	Origin of sands	Chronology	Interpretation	Other
Telfer et al 2012 Golden Gate National Park, South Africa	1		11	C	Predominantly well sorted fine sands with some cross-bedding (interpreted as aeolian) interdigitating with discrete clast units and lenses	no	Extensive downslope gullying, badland topography. No gully between hillslope and sand ramp	Assumed periglacial weathering of local geology	6 quartz SAR OSL dates. 45 +/- 3.0 ka – 7.6 +/- 0.6 ka. Likely episodic but dating resolution too low to be certain.	Units discontinuous between profiles of the same sand ramp. Deposition is likely to be episodic. Rapid accumulation associated with increased aeolian activity during the LGM linked to increased sediment supply from periglacial weathering and increased windiness. Holocene date is thought to represent colluvial reworking.	
Kumar et al., 2016 , Ladakh Desert, India	5	Total area: 40-2000 m ²	4-25	C	Varying layers of parallel laminated medium sand (interpreted as aeolian), hard crusts containing rock debris (interpreted as stability under wet conditions), poorly sorted coarse sand (interpreted as fluvial), parallel bedded angular gravels (Interpreted as sheet flow), parallel laminated clayey silt (interpreted as intra-dunal playa deposits).	Found at at least 3 ramps	Present but no details	Indus fluvial sediments in the pre-monsoon season. Made available by longer winters and shorter summers during the Late Pleistocene and Early Holocene.	14 Quartz SAR OSL dates. Suggest periods of deposition 25-15 ka and 12-8 ka. With a single date at ~44 ± 3 ka	Record of wet/dry climate. Aeolian deposits represent arid periods and fluvial sediments intra-dune playas and hiatuses represent wet conditions. Accumulation is not thought to occur during the most arid periods (LGM) as sand is not entrapped (i.e. high reworking and erosion). Shows good agreement with regional climate reconstructions of wet periods. No clear link to increased monsoon strength.	Overlaying dunes at one ramp.
De Valle et al., 2016 Eivissa, Western Mediterranean	1	1000 m wide		C	Calcareous aeolian deposits interbedded with colluvial deposits and palaeosols	n/m	n/m	Beach sand exposed during glacial low sea-level stands	4 fine and mix course-grain quartz OSL dates. 143 ± 8 ka (MIS 6), 100 ± 6 ka (MIS5d/c), 77 ± 5 ka and 73 ± 5 ka. (MIS 4).	Aeolian activity controlled by sediment supply. Associated with low sealevel during glacial periods. Colluvium and palaeosols develop during interglacials.	

- Number of studied sand ramps

L - Length

T – Thickness

C or F – climbing or falling.

n/m – not mentioned in text

* Unspecified number of sand ramps were observed and described. 5 sites (including complexes were studied in detail.

The studied sand ramps are highly varied in sedimentology and chronology (Table 5-1). All studied sand ramps contain a combination of aeolian and colluvial deposits, but in varying proportions, and require a suitable accommodation space (i.e. a hillslope against which aeolian and colluvial sediments can accumulate). Both climbing (up wind of a topographic barrier) and falling (downwind) sand ramps have been described. Falling ramps are typically smaller, positioned higher on mountain slopes with steeper slope angles, and with better sorted, finer sediments (Lancaster and Tchakerian, 1996). Gullies, which are evidence of fluvial activity, have been observed in many sand ramp contexts. These run parallel to the axis of the sand ramp body or are found separating the ramp from the hillslope against which it has accumulated (Lancaster and Tchakerian 1996; Thomas et al., 1997; Bertram, 2003; Bateman et al., 2012). Both gully types are assumed to post-date accumulation of the main sand ramp body (Bertram, 2003; Lancaster and Tchakerian, 1996) although an aeolian origin has been suggested for the latter type (Lancaster and Tchakerian, 1996). The origin of sediments has important implications for conditions needed for formation. Colluvial, talus and fluvial sediments must be derived locally from the adjacent topographic obstacle (Lancaster and Tchakerian, 1996; Pease and Tchakerian, 2003; Turner and Makhoul, 2002). Aeolian units are also suggested to be locally sourced from neighbouring ephemeral channels or upwind playas or dunefields (Bateman et al., 2012; Bertram, 2003). In the Mojave this is supported by geochemical fingerprinting (Pease and Tchakerian, 2003). While the overall processes of ramp accumulation (aeolian, fluvial and slope processes) are largely agreed upon, there

are components of ramp systems that are currently poorly investigated or explained.

Chronometric data are available for some sand ramps, giving clues to the timing and age of accumulation. However, some studies provide complex or conflicting data from the same location. TL and IRSL ages from sand units bracketing stone layers in Mojave ramps showed some to represent up to 10 ka breaks in aeolian unit accumulation, while others indicated no significant temporal breaks in accumulation (Rendell and Sheffer, 1996). Soldier Mountain (Mojave) was originally dated to ~10-22 ka using quartz and feldspar TL and feldspar IRSL dating (Rendell and Sheffer, 1996) and the sediments were interpreted as containing six palaeosols (Lancaster and Tchakerian, 1996). A more recent investigation of this feature, utilising SAR OSL dating, suggested rapid accumulation between ~11.6–10.3 ka (Bateman et al., 2012). No evidence of palaeosol development was found and stone layers were not associated with temporal hiatuses. Rates of modern stone movement (run-off creep) on the surface of the sand ramp could not be reconciled with the rapid emplacement of stone layers indicated from OSL dating. Therefore, deposition of this sand ramp was reinterpreted as opportunistic and event-driven with stone layers suggested to be deposited fluvially (Bateman et al., 2012).

Rapid deposition was also suggested at the Ardakan sand ramp, Iran (~25 m of sediment deposited between 25-20 ka) (Thomas et al., 1997a), and for modern sand ramps in Jordan (Turner and Makhoulouf, 2002). However, the Black Mesa,

USA and Golden Gate, South Africa, sand ramps potentially suggest episodic deposition between ~28-9 ka and ~45-7 ka respectively (Ellwein et al., 2015; Telfer et al., 2012) whilst dates from Namibian sand ramps indicate activity at >100 ka and ~15 ka (Bertram, 2003)

The above review suggests that sand ramps are complex and varied features, differing in formation history, and in the contribution from different sedimentary processes. The timescale and mechanism of sand ramp formation has important implications for palaeoenvironmental reconstructions with opportunistic, single-event driven accumulation providing limited regional information. This study aims to better understand the controls and mechanisms of sand ramp formation by conducting a regional survey on the locations of sand ramp formation and undertaking a detailed examination of the sedimentology, morphology and chronology of a morphologically varied subset of these sand ramps.

5.2.2 STUDY SITE

The study was conducted in south-west Namibia between 23°-28°S and 15°-19°E (Figure 5-2). This area encompasses the transitional zone between the Great Escarpment, where sand supply is limited, and the hyper-arid sandy coastal plain where sediment supply is high but topographic obstacles are rare. In the intermediate zone of the western foreland of the Great Escarpment both sand supply and topographic obstacles, especially inselbergs, are sufficient to lead to the presence of numerous sand ramps in a variety of accommodation spaces (Bertram, 2003). Exposed lithologies are of mixed erodibility with sandstones,

limestones and shales prevalent in the area (Mendelsohn et al., 2002). The region is currently hyper-arid to arid with a steep W-E precipitation gradient (barely above 0 to 200 mm) and high inter- and intra-annual variability (Lancaster et al., 1984; Nicholson, 2000). Towards the east precipitation is predominantly tropically derived summer rainfall associated with the Congo Air Boundary, Tropical-Temperate Troughs and the Angola Low (Eckardt et al., 2013). Rainfall becomes more spatially and temporally variable towards the west, dependent on the passing of storms from the west or rare temperate cold fronts travelling east. On the coast, and up to 100 km inland, precipitation is inhibited by the upwelling of the cold Benguela current and the subsidence of the high pressure belt. In these regions fog is the primary moisture source (Eckardt et al., 2013; Lancaster et al., 1984). Wind regimes are bimodal. Prevailing winds from the S-SW dominate for most of the year reaching maximum strength in early austral summer (September-November) frequently reaching speeds of 14 ms^{-1} at the coast but diminishing in strength towards the north and east (Lancaster, 1985). NW-NE plain-to-mountain winds are also common in austral summer (December-February) whilst E to NE katabatic, mountain-to-plain winds are frequent in austral winter (June-August) (Tyson and Seely, 1980). Occasionally, when regional pressure gradients are normal to the coast, strong easterly 'Berg' winds occur in the west. These frequently reach velocities of $14\text{-}17 \text{ ms}^{-1}$ and can account for up to 65% of total sand transport on the northern and eastern margins of the Namib Sand Sea although their strength rapidly decreases towards the west (Lancaster et al., 1984; Lancaster, 1985). The range of geological and morphological settings, E-W

precipitation gradient, and wind dynamics in the study area provide prime test conditions for assessing the controls on sand ramp formation.

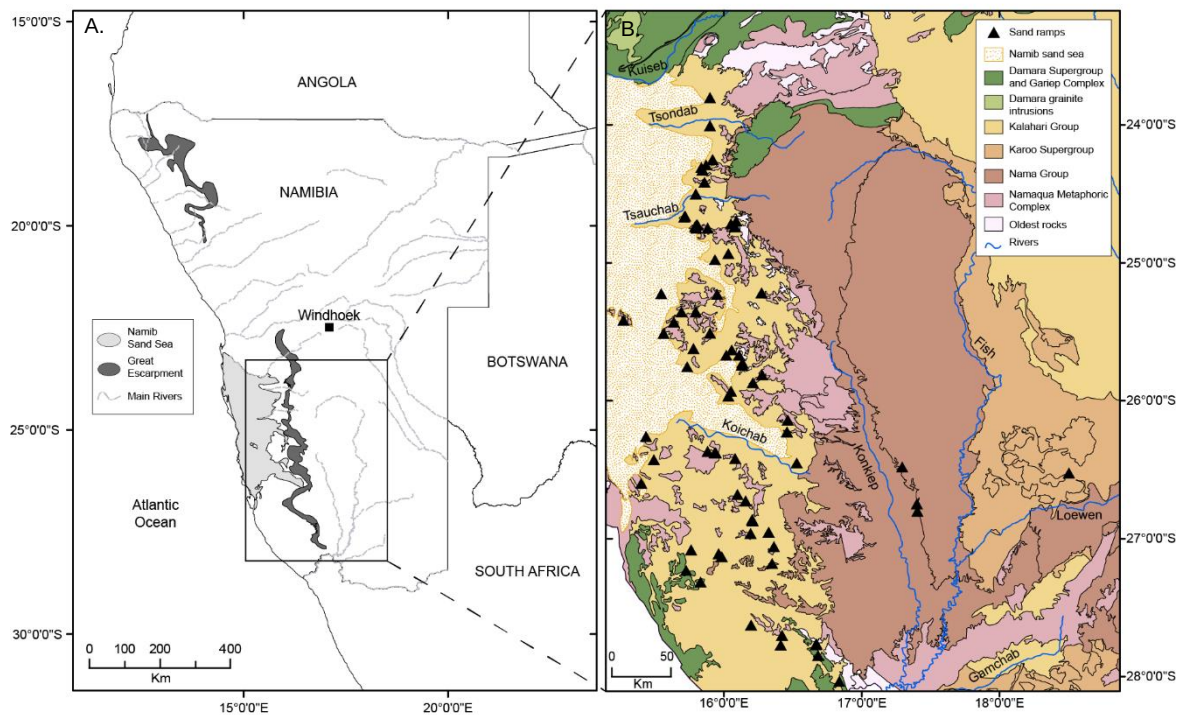


Figure 5-2. Location of study area (A) and sand ramps in geological context (B)

5.3 METHODS

The process of sand ramp identification, site selection and sampling strategy is described in section 3.2. Sediment analysis and standard OSL dating were conducted using the methods described in section 3.3. Chapter specific methods are described below.

5.3.1 SENSITIVITY TESTS

The sensitivity of quartz (counts produced per given dose) has been shown to increase significantly after heating to 500°C (e.g Bøtter-Jensen et al., 1995; Poolton et al., 2000) Additionally, quartz is sensitised in nature and the laboratory

through repeated cycles of signal accumulation and bleaching without heating (Fitzsimmons et al., 2010; Moska and Murray, 2006; Pietsch et al., 2008). Sensitivity change due to heating to 500°C has been shown to be dependent on the initial sensitivity with already sensitised quartz undergoing a subdued sensitivity increase (Moska and Murray, 2006). Therefore, grains that have undergone repeated burial and bleaching cycles in nature (e.g. aeolian sands) produce more luminescence per given dose than freshly weathered quartz (Fitzsimmons et al., 2010; Pietsch et al., 2008; Sawakuchi et al., 2011a) and are less likely to undergo very large sensitivity changes after heating (Moska and Murray, 2006). This means that the sensitivity of quartz may provide insights into the environmental history of a sample. The use of this technique, and whether different sand ramp grain-size fractions are derived from different depositional processes, was explored using sensitivity tests.

Sensitivity test were conducted following the SAR protocol and conditions described above. Aliquots were twice bleached at 20°C for 1000 s with a 10,000 s pause between bleaches. The response to a 50 Gy given dose (normalised by weight) was measured seven times. The first four measurements were conducted to measure initial sensitivity and change with dose cycle. Prior to the fifth measurement aliquots were heated to 500°C for 10 s and measured twice to assess the impact of annealing. The final measurement was an IR-depletion point (Duller, 2003). Measurements were conducted on ten medium aliquots of 90-125 µm quartz (~2000 grains; (Duller, 2008) and ten large aliquots 180-210 µm quartz (~1000 grains; Duller 2008) of eight samples. Dune samples NAM07/3/2 and

NAM07/4/12 from linear dunes of the Namib Sand Sea (see Stone et al., 2010 for details) and NAM14/10/3 from dunes overlaying the Samara sand ramp were used as examples of aeolian material from the region. These were compared with sand ramp body samples AUS13/1/1, NAM14/1/2, NAM14/7/3, NAM14/8/3 and NAM14/12/1.

5.3.2 HEAVY MINERAL ANALYSIS

Heavy minerals were separated using sodium polytungstate (density ~ 2.90 g/cm³). After mounting with Canada balsam ≥ 200 transparent heavy-mineral grains were point-counted to obtain real volume percentages (Galehouse, 1971). Heavy-mineral concentration was calculated as the weight percentage of total (HMC) and transparent (tHMC) heavy minerals relative to the bulk-sediment sample (Garzanti and Andò, 2007). Variance between sand ramps samples and samples from the Namib Sand Sea and major Namibian ephemeral rivers (see Garzanti et al., 2012 for sample details) was assessed using unscaled principal component analysis (PCA) (see Wold et al., 1987).

5.4 RESULTS

5.4.1 SAND RAMP LOCATION

In total, 75 sand ramps were identified within the study region (Figure 5-2). $\sim 90\%$ are climbing features as identified by their position in relation to topography and their morphology (Chojnacki et al., 2010; Ellwein et al., 2015; Goudie, 2013; Lancaster and Tchakerian, 1996). 74% are predominantly east facing and 17%

are west facing (Appendix C, Figure 5-3). Accumulation occurs in sheltered alcoves or along elongate mountain fronts aligned perpendicular to the wind with 95% of sand ramps accumulated against inselbergs or inselberg complexes and 5% against the edge of the Great Escarpment (Appendix C, Figure 5-3). All sand ramps are either within 4 km of a large ephemeral river channel or within 5.5 km of a dune field (Appendix C, Figure 5-3).

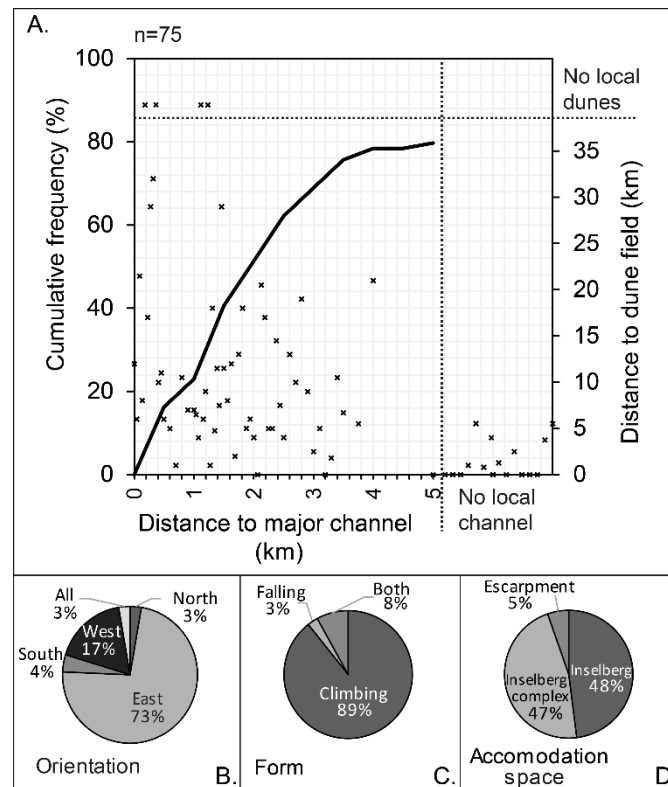


Figure 5-3. Properties of identified sand ramps. A). Distance to major channels or dune fields. Cumulative frequency of distance to major channels is displayed on the left y axis and represented by the solid black line. Distance to major channel vs. distance to dune field are displayed on the x and right y axis where each sand ramp is represented by a cross. Dashed lines delineate sand ramps with no local channel or dunes therefore only distance to the one feature is plotted, points are spaced for ease of reading. B). Orientation of the sand ramp sediments (measured head to toe). C). Form of sand ramps. D). Type of accommodation space.

5.4.2 SAND RAMP MORPHOLOGY

Field observations and aerial imagery indicated certain morphological features were common between different sand ramps. To test whether sand ramp morphology could be used as a predictor of age and accumulation history, sand

ramps were grouped into four classes based on the presence or absence of these features. Class 1 sand ramps are still attached to the adjacent hillslope, Class 2 sand ramps are separated from the hillslope by a gully, Class 3 sand ramps are separated from the hillslope by a gully and are overlaid by free-form dunes. Class 4 sand ramps are separated from the hillslope, overlaid by dunes and show extensive duricrust formation (Table 5-2). Classes 1-3 can be identified from aerial imagery but Class 4 requires analysis of the constituent sediments. Morphology was found to be correlated to size with length and width increasing with class (Table 5-2). Of the dated sand ramps; three are Class 1, two Class 2, two Class 3 and three Class 4 (Figure 5-4).

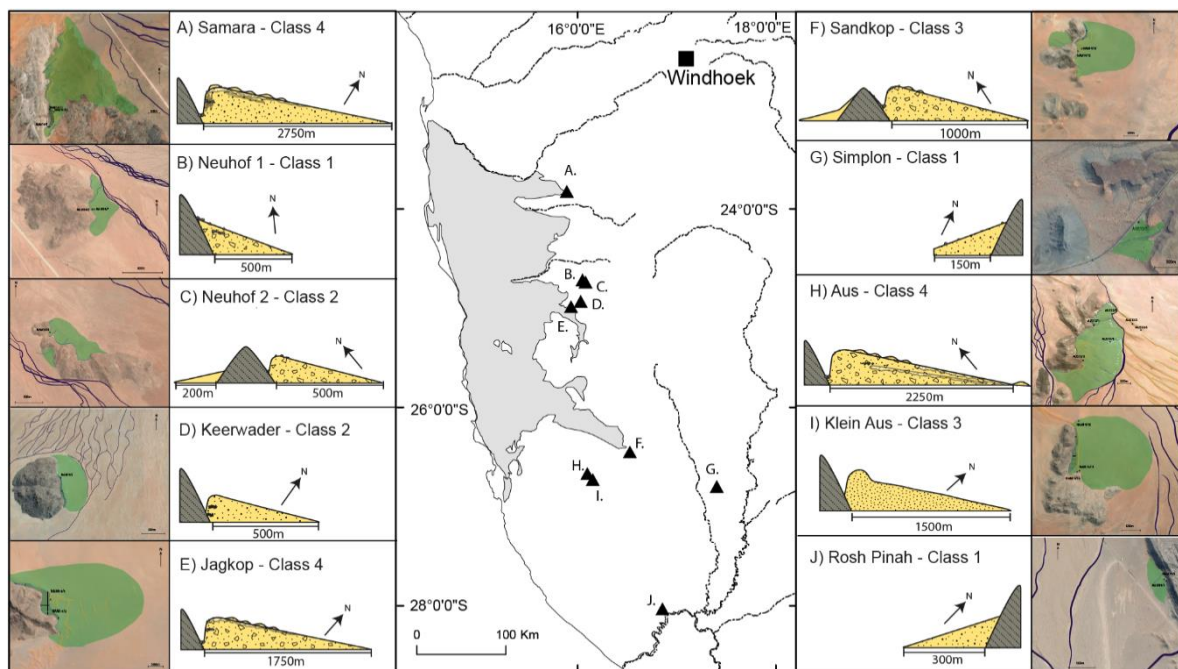
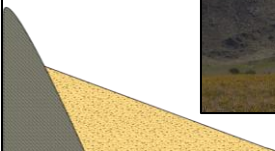
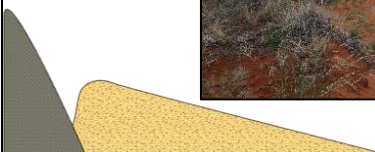
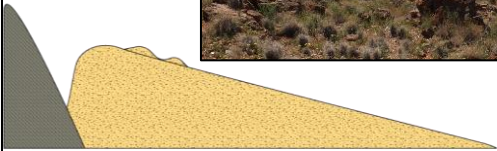
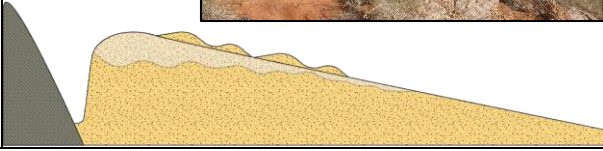


Figure 5-4. Location and morphology of dated sand ramps. Plan views are shown on the far right and far left; green shaded areas represent the extent of the sand ramps, blue lines highlight river channels.

Table 5-2 Sand ramp morphological classification

Class	Morphological description	Schematic and example	
1	Small features often with evidence of contemporary aeolian activity on surface. Sands are unconsolidated but will hold a face. The sand ramp is connected to the hillslope Length: 150-500m Width: 300-900m		
2	Medium sized features, disconnected from the hillslope by a channel or gully. Surface is often well vegetated. Sands are unconsolidated but can hold a face. Length: 300-800 m Width: 300-800 m		
3	Medium to large size features. Sand ramp is separated from the hillslope by a channel or gully. Surfaces are well vegetated and often indurated. Loose sand ridges/dunes overlay the sand ramp and show evidence of contemporary aeolian activity. Length: 1000-1500 m Width: 1000-1750 m		
4	Large complex features which imply several stages of formation. Sediments often contain duricrust layers and/or are capped by a duricrusts. Sand ramps are separated from the hillslope by a large gully and sand ridges are present on the surface of the sand ramp. The sand ramp sediments themselves are consolidated. Length: 1500-3000 m Width: 1500-3550 m		

5.4.3 SAND RAMP SEDIMENTS

Sand ramp sediments are predominantly poorly sorted sand with varying proportions of sub-rounded to angular gravels. Table 5-3 displays the grain-size properties of the ≤ 2 mm size-fraction. At Aus and Sandkop gravels are concentrated into semi-continuous lenses but with diffuse boundaries, elsewhere changes in gravel concentrations are gradual with clasts found sporadically throughout the sediment rather than concentrated into distinct layers. Samara, Jagkop and Neuhof-1 contain semi-continuous layers of large angular clasts up to 0.5 m in diameter. Duricrust formation occurs at Aus, Jagkop, and Samara. Calcrete nodules are present at Keerwader otherwise, organic and carbonate contents are typically below 1% (Table 5-3, Figure 5-5).

Simplon, Neuhof-1, Sandkop and the Class 4 sand ramps have a surface armour of angular clasts which thins towards the toe of the sand ramps. Overlying dunes are present at Class 3 and 4 sand ramps (Figure 5-4).

5.4.4 SAND RAMP AGES

The ten dated sand ramps indicate sand ramp and dune accumulation in southern Namibia occurred episodically for >100 ka with the majority of dated activity occurring between ~ 80 -12 ka (Figure 5-5 and Figure 5-6). Holocene dates are mostly obtained from dunes, near surface sediments or slumped gullies. Samples with $D_{eS} > 100$ -200 Gy are likely to be biased by individual grains with unfavourable growth curve properties (Chapter 4) and are therefore quoted as minimum ages. Details of luminescence properties can be found in Table 5-4.

Table 5-3. Individual sample sediment properties. b= brown, r = red, ry = reddish yellow, yr = yellowish red, lrb = light reddish brown. CS= coarse sand, MS = Medium sand, FS = Fine sand , VFS = Very fine sand, VCS = Very coarse silt, MW = Moderately well sorted, M = Moderately sorted, P = Poorly sorted, VP = very poorly sorted

Sample	Depth (m)	Munsell colour	Description	Mean particle size (µm)	Sorting (µm)	Skewness	Kurtosis	Coarse sand (%)	Medium sand (%)	Fine sand (%)	Silt (%)	Clay (%)	clasts >2mm (g)	Notes	LOI at 550°C (%)	Carbonate content (%)
NAM13/1/1	0.4	7.5YR 5/4 - b	MS,P	399	2.25	-0.01	0.98	39.0	32.6	26.6	1.8	0.0	0.07	Shale <5mm	0.5	1.5
NAM13/1/2	0.8	7.5YR 5/4 - b	MS,P	320	2.12	-0.04	1.09	27.1	37.1	32.8	3.1	0.0	0.08	Shale <5mm	0.5	1.3
NAM13/2/1	0.9	7.5YR 5/4 - b	MS,P	484	2.23	-0.07	0.92	49.7	28.8	20.9	0.5	0.0	0		0.5	1.8
NAM13/2/2	1.5	7.5YR 5/4 - b	MS,P	463	2.17	-0.10	1.03	47.7	31.5	19.5	1.3	0.0	0		0.5	1.4
NAM13/3/1	0.4	4 YR 5/4 - rb	MS,P	264	2.41	0.19	1.08	21.9	24.5	49.3	4.3	0.0	0.72	Angular <5mm	1.5	2.0
NAM13/3/3	1.9	5 YR 5/4 - rb	MS,P	300	2.56	0.17	0.83	31.3	20.5	44.6	3.7	0.0	0.19	<3mm	2.0	3.8
NAM13/4/0	0	5 YR 5/6 -yr	MS,M	427	1.71	-0.16	1.06	40.7	43.1	16.2	0.1	0.0	0		n/a	n/a
NAM13/4/1	1	5 YR 5/6 -yr	MS,M	418	1.70	-0.17	1.05	39.0	43.9	17.0	0.1	0.0	0		0.4	0.3
NAM13/4/2	1.8	5 YR 5/6 -yr	MS,M	372	1.76	-0.15	1.00	31.9	44.7	23.4	0.1	0.0	0		0.4	0.4
NAM14/5/1	0.5	5 YR 5/6 -yr	MS,P	363	2.61	0.05	0.78	39.3	21.2	36.9	2.6	0.0	1.02	sub rounded-sub angular <4mm	0.7	0.5
NAM14/5/2	1.1	5 YR 5/6 -yr	MS,P	409	2.47	-0.06	0.82	44.0	23.7	29.6	2.7	0.0	0.2	sub rounded-sub angular <4mm	0.6	0.5
NAM14/5/3	1.8	5 YR 5/6 -yr	MS,P	366	2.49	0.02	0.81	38.8	24.0	35.7	1.5	0.0	0.13	angular<5mm	0.6	0.5
NAM14/5/4	2.5	5 YR 5/6 -yr	MS,P	421	2.44	-0.05	0.80	44.9	23.8	31.0	0.3	0.0	1.35	angular clast <5mm, 50% calcrete	0.7	0.6
NAM14/6/1	0.4	7.5YR 5/6 - b	FS, MW	220	1.54	-0.06	0.99	1.4	37.9	58.0	2.7	0.0	0.24	Angular <5mm	0.6	0.7
NAM14/6/2	0.7	7.5YR 5/6 - b	FS, MW	226	1.51	-0.03	0.97	1.6	39.4	57.6	1.5	0.0	0.24	Angular <5mm	0.6	0.6
NAM14/6/3	1.8	7.5YR 5/6 - b	FS,M	232	1.65	-0.06	0.98	4.9	40.2	52.7	2.2	0.0	0.11	Angular <5mm	0.6	0.8
NAM14/7/1	0.25	7.5YR 5/6 - b	FS,M	239	1.72	-0.10	0.99	7.1	41.4	49.1	2.4	0.0	5.38	Angular <30mm	0.5	0.7
NAM14/7/2	0.6	7.5YR 5/6 - b	FS,M	237	1.78	-0.12	1.01	7.8	40.6	48.3	3.3	0.0	0.66	Angular <5mm	0.6	1.1

Sample	Depth (m)	Munsell colour	Description	Mean particle size (µm)	Sorting (µm)	Skewness	Kurtosis	Coarse sand (%)	Medium sand (%)	Fine sand (%)	Silt (%)	Clay (%)	clasts >2mm (g)	Notes	LOI at 550°C (%)	Carbonate content (%)
NAM14/7/3	1	7.5YR 5/6 - b	FS,M	244	1.80	-0.13	1.04	8.9	41.4	45.9	3.8	0.0	1.35	Angular <10mm	0.6	1.4
NAM14/8/1	0.15	5 YR 5/6 - yr	MS,M	336	2.00	-0.01	0.95	29.1	37.1	33.2	0.6	0.0	6.17	Angular <20mm	0.5	0.4
NAM14/8/2	0.8	5 YR 5/6 - yr	MS,M	296	1.90	-0.02	0.94	21.6	38.4	39.6	0.4	0.0	0.22	Angular <5mm	0.4	0.4
NAM14/8/3	1.6	5 YR 5/6 - yr	MS,M	318	1.97	-0.04	0.97	26.0	38.3	34.7	1.0	0.0	4.75	Angular <20mm	0.5	0.4
NAM13/9/1	0.3	5 YR 6/4 -lrb	FS,P	241	2.12	0.00	1.12	15.8	30.3	48.6	5.3	0.0	0		1.2	2.5
NAM14/9/2	7.8	5 YR 6/4 -lrb	VFS,P	120	3.18	-0.44	1.09	2.1	25.6	46.5	24.5	1.2	0		3.5	4.0
NAM14/10/0	0	7.5YR 6/6 - ry	MS,M	332	1.90	0.11	0.92	27.0	37.4	35.6	0.0	0.0	0		n/a	n/a
NAM14/10/1	1	5 YR 5/6 - yr	FS,MW	237	1.50	0.03	0.96	3.6	40.7	55.7	0.0	0.0	0		0.4	0.4
NAM14/10/2	1.8	5 YR 5/6 - yr	MS,M	276	1.69	0.10	0.99	13.9	41.8	44.3	0.0	0.0	0		0.4	0.4
NAM14/10/3	2.8	5 YR 5/6 - yr	FS,M	233	1.65	0.06	1.08	7.3	35.7	53.9	3.0	0.0	0.03	sub angular <3mm	0.6	0.6
NAM14/11/1	3.4	5 YR 6/4 -lrb	FS,P	248	2.36	-0.08	1.65	15.4	31.5	46.9	6.1	0.0	0.45	Mostly calcrete	1.0	1.8
NAM14/12/1	0.5	5 YR 5/6 - yr	MS, P	320	2.23	-0.06	1.02	29.3	33.1	33.3	4.4	0.0	49.37	Angular <60mm	0.8	1.2
NAM14/12/2	1	5 YR 5/6 - yr	MS, P	342	2.07	-0.01	0.95	30.9	35.1	32.3	1.7	0.0	7.3	Angular <20mm	0.7	0.6
NAM14/12/3	1.5	5 YR 5/6 - yr	MS, P	318	2.10	0.03	0.95	27.7	33.8	36.7	1.7	0.0	13.81	Angular <30mm	0.7	0.5
NAM14/12/4	3	5 YR 5/6 - yr	MS, P	404	2.11	-0.06	0.92	41.0	32.1	26.8	0.1	0.0	1.36	Angular <5mm	0.5	0.5
NAM14/12/5	4	5 YR 5/6 - yr	MS, P	543	2.45	-0.24	0.86	58.0	20.1	19.5	2.3	0.0	0.85	Angular <5mm	0.5	0.4
NAM14/12/6	5.7	5 YR 5/6 - yr	MS, P	500	2.53	-0.15	0.77	52.6	20.6	24.9	1.9	0.0	5.19	Angular <20mm	0.7	0.4
NAM14/12/7	6.2	5 YR 5/6 - yr	MS, P	497	2.57	-0.18	0.79	53.0	20.1	24.4	2.5	0.0	3.27	Angular <10mm	0.7	0.4
NAM14/13/1	1	5 YR 5/6 - yr	MS,M	319	1.85	-0.07	0.91	24.9	40.6	34.4	0.1	0.0	0		0.5	0.3
NAM14/13/2	1.5	5 YR 5/6 - yr	MS,M	266	1.77	0.05	0.94	14.5	38.7	46.8	0.0	0.0	0		0.5	0.3

Sample	Depth (m)	Munsell colour	Description	Mean particle size (µm)	Sorting (µm)	Skewness	Kurtosis	Coarse sand (%)	Medium sand (%)	Fine sand (%)	Silt (%)	Clay (%)	clasts >2mm (g)	Notes	LOI at 550°C (%)	Carbonate content (%)
NAM14/14/1	1	5 YR 5/6 - yr	MS, P	424	2.53	-0.22	0.97	48.4	24.6	23.2	3.7	0.0	0.15	sub angular <3mm	0.5	0.3
NAM14/14/2	1.5	5 YR 5/6 - yr	MS, P	435	2.37	-0.19	0.96	48.3	26.4	23.5	1.8	0.0	0.28	Angular <5mm	0.5	0.3
NAM14/14/3	2.5	5 YR 5/6 - yr	MS, P	414	2.38	-0.16	0.92	45.5	26.7	26.1	1.7	0.0	0.12	Angular <3mm	0.6	0.3
NAM14/14/4	3.5	5 YR 5/6 - yr	MS, P	457	2.35	-0.19	0.98	50.3	26.1	21.5	2.1	0.0	0.21	Angular <3mm	0.6	0.5
NAM14/14/5	4	5 YR 5/6 - yr	MS, P	467	2.39	-0.22	1.06	51.6	26.1	18.6	3.6	0.1	0.14	Angular <3mm	0.5	0.4
NAM14/15/1	0.5	5 YR 5/6 - yr	CS, MW	637	1.53	0.00	1.00	71.3	26.5	2.2	0.0	0.0	0		0.3	0.3
NAM14/15/2	0.8	5 YR 5/6 - yr	CS, M	709	1.63	0.02	0.95	75.3	23.6	1.0	0.0	0.0	0		0.3	0.3
NAM14/15/3	1.3	5 YR 5/6 - yr	CS, M	521	1.88	-0.11	1.12	54.4	33.1	12.0	0.4	0.0	0		0.4	0.3
AUS13/7/1	1.5	7.5YR 5/6 - sb	MS, M	328	1.80	-0.11	1.02	24.0	44.9	28.1	2.8	0.2	0		0.6	0.5
AUS13/7/2	3	7.5YR 5/6 - sb	MS, M	261	1.78	-0.05	0.99	12.7	40.9	43.6	2.8	0.0	0		0.5	0.5
AUS13/1/1	2	5 YR 5/6 - yr	FS, P	183	3.04	-0.22	1.26	13.8	24.8	45.5	14.4	0.7	20.82	Angular<2cm	1.5	1.6
AUS13/1/2	5	5 YR 5/6 - yr	FS, P	192	2.47	-0.06	0.99	13.4	25.1	50.48	10.37	0	15.19	sub angluar to angular <1cm	1.1	0.6
AUS13/1/3	6.5	5 YR 5/6 - yr	FS, P	152	3.10	-0.18	1.21	10.5	21.7	48.2	18.24	0.8	25.82	sub angluar to angular <1cm	1.2	0.6
AUS13/1/4	8	5 YR 5/6 - yr	FS, P	146	2.66	-0.08	1.29	7.1	17.7	57.4	16.4	0.1	33.06	sub angluar to angular <1cm	1.5	0.7
AUS13/1/5	8.7	5 YR 5/6 - yr	FS, P	148	2.45	-0.07	1.41	5.4	16.7	63	13	0.2	31.98	Angular<2cm	1.4	0.7
AUS13/1/6	9.2	5 YR 5/6 - yr	FS, P	143	2.32	-0.03	1.33	4.5	15.1	65.3	13.3	0.1	30.25	Angular <1cm & calcrete noduals	1.4	0.8
AUS13/1/7	10.7	5 YR 5/6 - yr	FS, P	147	2.54	-0.03	1.33	6.7	16.2	60.5	14.5	0.3	26.66	Augular <3cm	1.6	0.8
AUS13/1/8	12.7	5 YR 5/6 - yr	FS, P	155	2.28	0.01	1.19	7	17.9	62.9	11.6	0	45.4	Angular<2cm & calcrete	1.5	0.8

Sample	Depth (m)	Munsell colour	Description	Mean particle size (µm)	Sorting (µm)	Skewness	Kurtosis	Coarse sand (%)	Medium sand (%)	Fine sand (%)	Silt (%)	Clay (%)	clasts >2mm (g)	Notes	LOI at 550°C (%)	Carbonate content (%)
														noduals		
AUS13/1/9	15.7	5 YR 5/6 - yr	FS, P	131	2.88	-0.11	1.42	6.6	14.9	57.4	18.7	1.1	35.31	Augular <3cm	1.7	1.0
AUS13/1/10	18.7	5 YR 5/6 - yr	VFS,P	116	2.74	-0.27	1.55	2.9	13.6	62.6	19.6	1.2	5.26	Angular <1cm	1.8	1.1
AUS13/1/11	19.7	5 YR 5/6 - yr	VFS,P	125	2.18	-0.15	1.62	1.1	9.1	74.8	13	0.5	0		1.7	1.3
AUS13/2/1	0.8	5YR 6/6 - ry	MS, P	255	2.59	-0.18	1.31	17.3	31.2	41.1	6.3	0.7	0		0.5	0.4
AUS13/2/2	1.7	5YR 6/6 - ry	MS, P	302	2.35	-0.10	1.17	21	32.9	35.3	5	0.3	0		0.4	0.4
AUS13/3/0	0	2.5 YR 5/6 - r	MS, P	275	2.30	-0.09	0.96	20.7	30.8	39.4	4.9	0.2	0		0.6	0.4
AUS13/3/1	0.35	2.5 YR 5/6 - r	FS, P	288	2.65	0.00	0.85	8.9	23.6	52.2	13	1.1	3.36	Angular <1cm	0.8	0.6
AUS13/3/2	0.7	2.5 YR 5/6 - r	FS, P	275	2.65	0.03	0.88	7.6	23.6	51.8	14.6	1.1	0		0.8	0.6
AUS13/3/3	1.5	2.5 YR 5/6 - r	FS, P	246	3.28	-0.15	1.15	13.3	18.6	37.4	21.4	0.9	0		0.9	0.6
AUS13/3/4	2.5	2.5 YR 5/6 - r	FS, P	265	2.44	0.04	0.86	8	23.9	56	11.1	0.5	0		1.0	0.7
AUS13/3/5	3.5	2.5 YR 5/6 - r	FS, P	275	2.52	0.04	0.79	8.2	18.3	65.2	6.9	0	0		1.0	1.1
AUS13/3/6	3.9	2.5 YR 5/6 - r	MS, P	276	2.28	-0.12	0.85	22.9	30.1	39.8	3.7	0	0		0.6	0.5
AUS13/4/1	1.5	5YR 6/6 - ry	MS, P	325	2.87	-0.06	0.92	18.8	21.7	36.8	6.8	0	0		2.5	1.4
AUS13/4/2	3	5YR 6/6 - ry	MS, P	216	2.62	-0.13	1.06	16.5	26.6	44.7	10.2	0	0		1.2	0.8
AUS13/5/1	3	5 YR 5/6 - yr	FS, P	211	3.06	0.06	1.02	13.8	17.5	47.5	12.7	0	4.35	Angular <5mm	1.3	1.0
AUS13/5/2	5	5 YR 5/6 - yr	FS, P	159	2.96	-0.10	1.15	11.9	20.3	49.3	17.3	0.4	26.87	Mostly carbonate (c90%) some Angular<1cm	2.6	1.3
AUS13/5/3	6	5 YR 5/6 - yr	FS, P	116	4.09	-0.09	1.00	11.7	16.9	37	31.2	1.1	14.6	Angular<1cm and carbonate (c.50%)	1.7	0.9
AUS13/5/4	7.5	5 YR	FS,VP	170	4.01	-0.46	1.08	16.5	27.7	29.8	22.4	1.2	13.58	Angular<2cm	0.8	0.6

Sample	Depth (m)	Munsell colour	Description	Mean particle size (µm)	Sorting (µm)	Skewness	Kurtosis	Coarse sand (%)	Medium sand (%)	Fine sand (%)	Silt (%)	Clay (%)	clasts >2mm (g)	Notes	LOI at 550°C (%)	Carbonate content (%)
		5/6 - yr												and carbonate (c.50%)		
AUS13/5/5	8	5 YR 5/6 - yr	FS, P	272	2.75	-0.29	1.31	14	29.2	39.5	9.1	0.7	10.27	Angular<2cm	0.5	0.4
AUS13/5/6	9	5 YR 5/6 - yr	VCS, P	63	3.94	-0.24	0.96	1.6	11.2	41.5	43.7	2	24.1	Angular<2cm and carbonate (c.70%)	1.4	0.8
AUS13/5/7	9.5	5 YR 5/6 - yr	FS, P	218	3.59	-0.38	1.16	23.3	26.7	30.6	15.7	1	10.06	Angular<1cm and carbonate (c.20%)	1.0	0.8
AUS13/6/0	0	5 YR 5/6 - yr	MS, P	393	2.55	-0.16	0.93	28.3	25.5	26.8	4.1	0.1	0		n/a	n/a
AUS13/6/1	1	5YR 6/6 - ry	FS, P	248	2.78	-0.20	1.37	17.7	29.7	41.2	7.4	0.8	0		1.2	0.6
AUS13/6/2	2	5YR 6/6 - ry	MS, P	419	2.48	-0.08	0.87	17.6	31.1	39.4	6.4	0.6	1.88	sub rounded <5mm	0.4	0.3
AUS13/6/3	3	5YR 6/6 - ry	MS, P	456	2.61	-0.12	0.77	18.1	23.7	39.3	5.1	0.5	1.1	sub rounded <5mm	0.5	0.3
AUS13/6/4	4	5YR 6/6 - ry	MS, P	457	2.43	-0.05	0.79	14.4	28.9	44.5	4.2	0.3	0		0.4	0.4
AUS13/6/5	4.7	5YR 6/6 - ry	MS, P	286	2.04	-0.12	0.98	25.5	28.8	26.7	2.6	0.3	0.11	sub rounded <5mm	0.4	0.5

5.4.5 RELATIONSHIP BETWEEN AGE AND MORPHOLOGY

The oldest OSL ages were obtained from the sediments of Class 4 sand ramps and indicate accumulation >100 ka. The main bodies of Class 2 and 3 sand ramps, aside from Klein Aus, date to 12- >75 ka whilst all Class 1 dates are younger than 40 ka (Figure 5-5 and Figure 5-6). The erroneously young ages obtained from the gullies of Klein Aus and Rosh Pinah are thought to be due to recent slumping as sediments at both these locations struggled to hold a face. Dates from Neuhoof-1 and Rosh Pinah (Class 1) were only obtained from the top 2 m of sediment wherein they show some stratigraphic constancy with dates from Class 2 and 3 sand ramps (Figure 5-6). Near surface sediments of most dunes (Class 3 & 4) and sand ramp bodies (Class 1 & 2) show activity within the last 2 ka (Figure 5-5).

5.4.6 RELATIONSHIP BETWEEN AGE AND SEDIMENTOLOGY

Duricrust layers are associated with a hiatus in sedimentation of >c.70 ka (Jagkop, Samara) and calcrete nodules at Keerwader are associated with a hiatus of 13-20 ka (Figure 5-5). Consolidated sediments were deposited >~100 ka. The chronological significance of stone layers is site specific, at Jagkop and Samara clast layers could represent a long term hiatus in deposition whilst at Neuhoof-1 the stone layer represents a maximum hiatus of 7 ka (~4 ka weighted mean). Unconsolidated sand ramp sediments are dated to between 0.2 - >75 ka with no clear relationship between sedimentary features and accumulation rate (Figure 5-5).

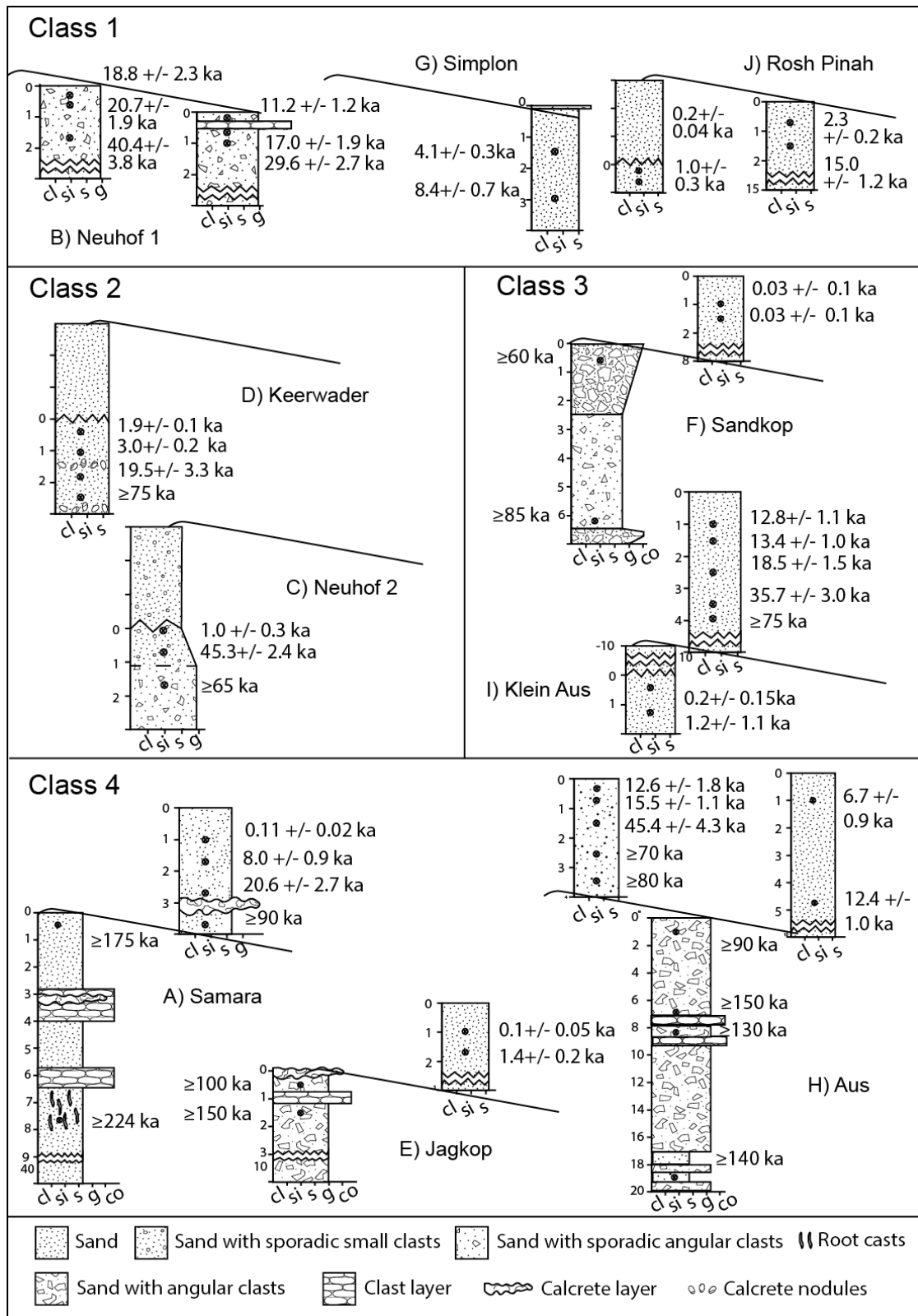


Figure 5-5. Schematic of sand ramp sediments with associated OSL ages. Sloping black lines represent the sand ramp surface. Sediment logs positioned above this were taken from dunes overlaying the sand ramp sediments. Sediment logs positioned below the line were taken from gullies at the head of the sand ramp and logs below the line in the centre were taken from gullies or pits in the centre of the sand ramp body. Width of the sediment logs represents maximum grain size (cl = clay, si = silt, s = sand, g = gravel, co = cobbles). Black circles indicate the position of OSL samples. Depth in meters is displayed on the y axis. For Neuhof 2 and Keerwader the marker above 0 represents the additional overburden of sediment which could not be sampled due to the gentle sloping side of the gully. In both cases this represents ~3m

Table 5-4. Luminescence properties. Ages calculated using CAM (Galbraith, 1999). n = number of accepted aliquots, OD = Overdispersion (*All aliquots of NAM14/13/1 & NAM14/13/2 failed the IR ratio significantly).

Site	Context	Sample	Depth (m)	n	De	OD (%)	ICP-MS			Gamma spectrometry			Beta dose (Gy/ka)	Gamma dose (Gy/ka)	Cosmic dose (Gy/ka)	Dose rate (Gy/ka)	Age (ka)	Quoted age (ka)
							U (ppm)	Th (ppm)	K (%)	U (ppm)	Th (ppm)	K (%)						
Rosh Pinah	Gully	NAM13/1/1	0.4	4	0.43 ± 0.07	-	0.79 ± 0.12	2.75 ± 0.55	1.41 ± 0.13	-	-	-	1.17 ± 0.10	0.54 ± 0.05	0.22 ± 0.02	1.93 ± 0.12	0.22 ± 0.04	
	Gully	NAM13/1/2	0.8	4	2.24 ± 0.60	-	1.22 ± 0.19	5.94 ± 1.19	1.41 ± 0.13	1.36 ± 0.20	6.97 ± 0.54	1.45 ± 0.11	1.29 ± 0.11	0.86 ± 0.05	0.19 ± 0.02	2.34 ± 0.12	0.98 ± 0.26	
	Main body	NAM13/2/1	0.9	23	4.98 ± 0.16	12 ± 1	1.16 ± 0.18	4.90 ± 0.98	1.43 ± 0.13	-	-	-	1.27 ± 0.11	0.68 ± 0.06	0.19 ± 0.02	2.15 ± 0.13	2.32 ± 0.16	
	Main body	NAM13/2/2	1.5	23	37.26 ± 1.10	14 ± 1	1.3 ± 0.20	6.05 ± 1.38	1.67 ± 0.20	-	-	-	1.48 ± 0.16	0.81 ± 0.09	0.18 ± 0.02	2.47 ± 0.18	15.00 ± 1.19	
Jagkop	Head gully	NAM13/3/1	0.4	7	311.4 ± 41.27	33 ± 6	2.55 ± 0.40	6.6 ± 1.32	2.00 ± 0.18	2.04 ± 0.26	8.35 ± 0.64	1.84 ± 0.15	1.88 ± 0.16	1.10 ± 0.07	0.26 ± 0.03	3.41 ± 0.17	99.5 ± 14.28	≥100
	Head gully	NAM13/3/3	1.9	7	249.22 ± 53.41	55 ± 12	0.98 ± 0.15	3.48 ± 0.70	1.04 ± 0.09	-	-	-	0.94 ± 0.08	0.51 ± 0.05	0.22 ± 0.02	1.67 ± 0.09	149.6 ± 33.2	≥150
	Overlaying dunes	NAM13/4/1	1	4	0.46 ± 0.16	-	1.18 ± 0.18	5.63 ± 1.13	2.81 ± 0.25	-	-	-	2.29 ± 0.20	1.04 ± 0.09	0.21 ± 0.02	3.54 ± 0.22	0.13 ± 0.05	
	Overlaying dunes	NAM13/4/2	1.8	4	4.87 ± 0.06	-	1.34 ± 0.21	5.14 ± 1.03	2.69 ± 0.24	-	-	-	2.21 ± 0.20	1.01 ± 0.09	0.19 ± 0.02	3.41 ± 0.21	1.43 ± 0.20	
Keerwader	Head gully	NAM14/5/1	0.5	27	6.58 ± 0.13	9 ± 1	1.81 ± 0.28	6.72 ± 1.34	2.40 ± 0.22	-	-	-	2.09 ± 0.18	1.07 ± 0.09	0.24 ± 0.02	3.39 ± 0.20	1.95 ± 0.12	
	Head gully	NAM14/5/2	1.1	16	8.46 ± 0.41	19 ± 2	1.86 ± 0.29	6.34 ± 1.27	1.89 ± 0.17	-	-	-	1.72 ± 0.15	0.93 ± 0.08	0.21 ± 0.02	2.86 ± 0.17	2.96 ± 0.23	
	Head gully	NAM14/5/3	1.8	23	72.5 ± 11.58	76 ± 10	1.56 ± 0.24	7.46 ± 1.49	2.80 ± 0.25	-	-	-	2.36 ± 0.21	1.17 ± 0.10	0.19 ± 0.02	3.71 ± 0.23	19.53 ± 3.35	
	Head gully	NAM14/5/4	2.5	10	301.86 ± 34.25	32 ± 5	1.54 ± 0.27	7.33 ± 3.63	3.09 ± 0.14	1.50 ± 0.21	6.57 ± 0.53	2.66 ± 0.20	2.56 ± 0.16	1.21 ± 0.08	0.17 ± 0.02	3.95 ± 0.18	78.96 ± 9.65	≥75
Neuhof 1	Main body	NAM14/6/1	0.4	18	56.49 ± 2.26	14 ± 1	1.8 ± 0.28	4.62 ± 0.92	1.73 ± 0.16	-	-	-	1.56 ± 0.13	0.81 ± 0.07	0.25 ± 0.03	3.00 ± 0.35	18.84 ± 2.31	
	Main body	NAM14/6/2	0.7	19	73.16 ± 5.99	25 ± 3	1.88 ± 0.29	6.05 ± 1.21	2.59 ± 0.23	1.79 ± 0.25	8.89 ± 0.68	2.07 ± 0.16	2.22 ± 0.19	1.17 ± 0.07	0.22 ± 0.02	3.61 ± 0.21	20.73 ± 1.86	
	Main body	NAM14/6/3	1.8	14	152.35 ± 10.75	23 ± 3	1.75 ± 0.27	10.32 ± 2.06	2.61 ± 0.24	-	-	-	2.31 ± 0.20	1.27 ± 0.12	0.19 ± 0.02	3.77 ± 0.23	40.43 ± 3.79	
	Main body	NAM14/7/1	0.25	27	42.84 ± 2.26	43 ± 4	2.18 ± 0.34	10.09 ± 2.02	2.50 ± 0.22	-	-	-	2.27 ± 0.19	1.28 ± 0.12	0.26 ± 0.03	3.81 ± 0.23	11.19 ± 1.23	
	Main body	NAM14/7/2	0.6	12	66.29 ± 6.79	34 ± 4	2.17 ± 0.16	8.19 ± 0.2	2.79 ± 0.15	2.54 ± 0.32	10.22 ± 0.78	2.14 ± 0.17	2.44 ± 0.14	1.32 ± 0.08	0.23 ± 0.02	3.99 ± 0.17	16.95 ± 1.87	
	Main body	NAM14/7/3	1	26	101.77 ± 6.91	32 ± 3	1.85 ± 0.29	6.3 ± 1.26	2.51 ± 0.23	-	-	-	2.16 ± 0.19	1.08 ± 0.09	0.21 ± 0.02	3.44 ± 0.21	29.57 ± 2.69	
Neuhof 2	Head gully	NAM14/8/1	0.15	26	3.83 ± 0.90	124 ± 19	1.5 ± 0.23	5.41 ± 1.08	3.00 ± 0.27	-	-	-	2.46 ± 0.22	1.11 ± 0.09	0.29 ± 0.03	3.86 ± 0.42	0.98 ± 0.26	
	Head gully	NAM14/8/2	0.8	15	177.39 ± 7.81	13 ± 1	1.46 ± 0.03	4.72 ± 0.3	3.27 ± 0.01	1.52 ± 0.19	6.04 ± 0.47	2.66 ± 0.20	2.64 ± 0.10	1.20 ± 0.08	0.22 ± 0.02	4.05 ± 0.13	45.32 ± 2.42	

							ICP-MS			Gamma spectrometry								
Site	Context	Sample	Depth (m)	n	De	OD (%)	U (ppm)	Th (ppm)	K (%)	U (ppm)	Th (ppm)	K (%)	Beta dose (Gy/ka)	Gamma dose (Gy/ka)	Cosmic dose (Gy/ka)	Dose rate (Gy/ka)	Age (ka)	Quoted age (ka)
Neuhof 2	Head gully	NAM14/8/3	1.6	20	221 ± 11.72	19 ± 2	1.52 ± 0.24	5.46 ± 1.09	2.54 ± 0.23	-	-	-	2.13 ± 0.19	1.01 ± 0.09	0.19 ± 0.02	3.33 ± 0.21	66.41 ± 5.41	≥65
Samara	Head gully	NAM13/9/1	0.3	11	337.27 ± 27.85	24 ± 3	1.01 ± 0.16	3.58 ± 0.72	1.27 ± 0.11	-	-	-	1.11 ± 0.10	0.57 ± 0.05	0.26 ± 0.03	1.94 ± 0.20	173.49 ± 22.75	≥170
	Head gully	NAM14/9/2	7.8	11	331.5 ± 19.15	15 ± 2	0.8 ± 0.14	3.55 ± 0.48	1.01 ± 0.10	-	-	-	0.89 ± 0.08	0.48 ± 0.04	0.10 ± 0.01	1.48 ± 0.09	224.44 ± 19.07	≥225
	Overlaying dunes	NAM14/10/1	1	7	0.19 ± 0.02	23 ± 3	0.89 ± 0.14	3.52 ± 0.70	1.17 ± 0.11	-	-	-	1.02 ± 0.09	0.53 ± 0.05	0.21 ± 0.02	1.77 ± 0.18	0.11 ± 0.02	
	Overlaying dunes	NAM14/10/2	1.8	20	19.22 ± 0.86	19 ± 1	0.5 ± 0.08	2.11 ± 0.42	1.69 ± 0.15	-	-	-	1.33 ± 0.12	0.55 ± 0.05	0.19 ± 0.02	2.42 ± 0.26	7.95 ± 0.93	
	Overlaying dunes	NAM14/10/3	2.8	20	36.47 ± 2.88	38 ± 4	0.87 ± 0.13	3.52 ± 0.70	1.22 ± 0.11	-	-	-	1.06 ± 0.09	0.54 ± 0.05	0.17 ± 0.02	1.77 ± 0.19	20.64 ± 2.73	
	Overlaying dunes	NAM14/11/1	3.4	23	184.6 ± 19.62	42 ± 5	0.79 ± 0.12	3.44 ± 0.28	1.52 ± 0.06	-	-	-	1.26 ± 0.07	0.60 ± 0.03	0.16 ± 0.02	2.02 ± 0.08	91.34 ± 10.30	≥90
Sandkop	Head gully	NAM14/12/1	0.5	22	219.37 ± 14.09	27 ± 2	1.23 ± 0.19	10.08 ± 2.02	2.14 ± 0.19	1.87 ± 0.32	13.43 ± 1.01	2.29 ± 0.18	1.90 ± 0.16	1.37 ± 0.09	0.25 ± 0.03	3.52 ± 0.19	62.67 ± 5.24	≥60
	Head gully	NAM14/12/7	6.2	25	288.14 ± 17.74	29 ± 2	1.18 ± 0.09	9.25 ± 0.80	2.35 ± 0.30	1.71 ± 0.86	11.29 ± 0.86	2.26 ± 0.18	2.03 ± 0.23	1.32 ± 0.08	0.12 ± 0.01	3.87 ± 0.24	85.24 ± 8.33	≥85
	Overlaying dunes	NAM14/13/1	1	4*	0.08 ± 0.04	-	0.8 ± 0.12	5.66 ± 1.13	2.20 ± 0.20	-	-	-	1.80 ± 0.16	0.86 ± 0.08	0.22 ± 0.02	2.88 ± 0.18	0.03 ± 0.01	
	Overlaying dunes	NAM14/13/2	1.5	4*	0.1 ± 0.03	-	0.96 ± 0.15	6.11 ± 1.22	2.20 ± 0.20	-	-	-	1.83 ± 0.16	0.90 ± 0.08	0.21 ± 0.02	2.94 ± 0.18	0.03 ± 0.01	
Klein Aus	Main body	NAM14/14/1	1	25	52.41 ± 3.28	22 ± 2	1.79 ± 0.28	10.34 ± 2.07	2.86 ± 0.26	-	-	-	2.49 ± 0.21	1.34 ± 0.12	0.22 ± 0.02	4.05 ± 0.25	12.84 ± 1.08	
	Main body	NAM14/14/2	1.5	24	72.83 ± 2.71	16 ± 1	2.66 ± 0.41	16.39 ± 3.28	3.72 ± 0.33	-	-	-	3.34 ± 0.28	1.91 ± 0.19	0.21 ± 0.02	5.45 ± 0.34	13.36 ± 0.97	
	Main body	NAM14/14/3	2.5	23	98.13 ± 5.12	24 ± 2	2.57 ± 0.40	16.39 ± 3.28	3.61 ± 0.33	-	-	-	3.25 ± 0.28	1.87 ± 0.18	0.18 ± 0.02	5.31 ± 0.33	18.50 ± 1.51	
	Main body	NAM14/14/4	3.5	22	196.8 ± 11.22	24 ± 2	2.92 ± 0.45	17.84 ± 3.57	3.66 ± 0.33	-	-	-	3.36 ± 0.28	1.99 ± 0.20	0.16 ± 0.02	5.51 ± 0.35	35.74 ± 3.03	≥35
	Main body	NAM14/14/5	4	10	345.2 ± 20.90	11 ± 2	2.33 ± 0.36	13.83 ± 2.77	2.90 ± 0.26	-	-	-	2.66 ± 0.22	1.56 ± 0.15	0.15 ± 0.02	4.37 ± 0.27	79.01 ± 6.87	≥75
	Gully	NAM14/15/1	0.5	4	0.92 ± 0.59	-	0.78 ± 0.12	3.41 ± 0.68	3.34 ± 0.30	-	-	-	2.58 ± 0.24	1.03 ± 0.09	0.25 ± 0.03	3.86 ± 0.26	0.24 ± 0.15	
	Gully	NAM14/15/3	1.3	4	4.78 ± 4.67	-	1.42 ± 0.22	7.83 ± 1.57	3.18 ± 0.29	-	-	-	2.63 ± 0.23	1.26 ± 0.11	0.21 ± 0.02	4.10 ± 0.26	1.17 ± 1.14	
Aus	Main body	AUS13/1/1	2	17	505.11 ± 34.24	24 ± 3	1.34 ± 0.21	10.27 ± 2.05	3.42 ± 0.31	2.76 ± 0.40	18.77 ± 1.40	2.91 ± 0.23	2.84 ± 0.25	1.93 ± 0.12	0.19 ± 0.02	4.96 ± 0.28	90.00 ± 7.72	≥90
	Main body	AUS13/1/4	8	6	629.48 ± 49.62	4 ± 2	1.95 ± 0.30	14.37 ± 2.87	2.74 ± 0.25	-	-	-	2.51 ± 0.21	1.51 ± 0.16	0.10 ± 0.01	4.12 ± 0.22	153.05 ± 22.27	≥150
	Main body	AUS13/1/6	9.2	5	559.3 ± 78.35	27 ± 6	2.24 ± 0.35	15.94 ± 3.19	2.77 ± 0.25	-	-	-	2.54 ± 0.21	1.62 ± 0.17	0.09 ± 0.10	4.24 ± 0.27	131.79 ± 20.31	≥130

							ICP-MS			Gamma spectrometry								
Site	Context	Sample	Depth (m)	n	De	OD (%)	U (ppm)	Th (ppm)	K (%)	U (ppm)	Th (ppm)	K (%)	Beta dose (Gy/ka)	Gamma dose (Gy/ka)	Cosmic dose (Gy/ka)	Dose rate (Gy/ka)	Age (ka)	Quoted age (ka)
Aus	Main body	AUS13/1/11	19.7	9	531.6 ± 21.85	22 ± 1	1.8 ± 0.28	12.43 ± 2.49	2.59 ± 0.23	-	-	-	2.34 ± 0.20	1.37 ± 0.14	0.04 ± 0.00	3.75 ± 0.24	141.80 ± 10.85	≥140
	Overlaying dunes	AUS13/3/1	0.35	31	64.08 ± 8.15	67 ± 7	2.6 ± 0.40	18.97 ± 3.79	3.13 ± 0.28	-	-	-	2.96 ± 0.25	1.88 ± 0.20	0.26 ± 0.03	5.09 ± 0.32	12.57 ± 1.76	
	Overlaying dunes	AUS13/3/2	0.7	47	90.02 ± 4.74	29 ± 2	3.38 ± 0.52	23.21 ± 4.64	3.15 ± 0.28	3.99 ± 0.54	25.18 ± 1.87	2.82 ± 0.23	3.15 ± 0.26	2.29 ± 0.14	0.23 ± 0.02	5.66 ± 0.03	15.50 ± 1.09	
	Overlaying dunes	AUS13/3/3	1.5	22	247.31 ± 16.66	29 ± 3	2.84 ± 0.44	23.65 ± 4.73	3.17 ± 0.29	-	-	-	3.12 ± 0.26	2.13 ± 0.24	0.20 ± 0.02	5.44 ± 0.36	45.44 ± 4.27	≥45
	Overlaying dunes	AUS13/3/4	2.5	9	344.86 ± 19.91	15 ± 2	2.37 ± 0.37	15.75 ± 3.15	2.97 ± 0.27	-	-	-	2.75 ± 0.23	1.67 ± 0.17	0.18 ± 0.02	4.60 ± 0.29	74.96 ± 6.35	≥75
	Overlaying dunes	AUS13/3/5	3.5	18	323.12 ± 12.96	8 ± 2	1.57 ± 0.08	10.73 ± 1.5	2.95 ± 0.05	-	-	-	2.54 ± 0.10	1.35 ± 0.08	0.16 ± 0.02	4.05 ± 0.13	79.76 ± 4.15	≥80
	Toe dunes	AUS13/6/1	1	11	38.58 ± 4.76	40 ± 6	2.75 ± 0.43	23.94 ± 4.79	3.52 ± 0.32	-	-	-	3.36 ± 0.28	2.21 ± 0.25	0.21 ± 0.02	5.78 ± 0.38	6.67 ± 0.93	
	Toe dunes	AUS13/6/5	4.7	14	65.31 ± 3.38	12 ± 12	2.56 ± 0.40	19.56 ± 3.91	3.34 ± 0.30	-	-	-	3.12 ± 0.26	1.95 ± 0.21	0.14 ± 0.01	5.21 ± 0.34	12.49 ± 0.98	
Simplon	Main body	AUS13/7/1	1.5	23	7.38 ± 0.46	28 ± 2	1.27 ± 0.20	5.46 ± 1.09	1.02 ± 0.09	-	-	-	1.00 ± 0.08	0.62 ± 0.06	0.20 ± 0.02	1.82 ± 0.11	4.06 ± 0.35	
	Main body	AUS13/7/2	3	24	15.39 ± 0.50	15 ± 1	1.35 ± 0.21	5.48 ± 1.10	1.05 ± 0.09	-	-	-	1.03 ± 0.09	0.64 ± 0.06	0.16 ± 0.02	1.83 ± 0.11	8.41 ± 0.56	

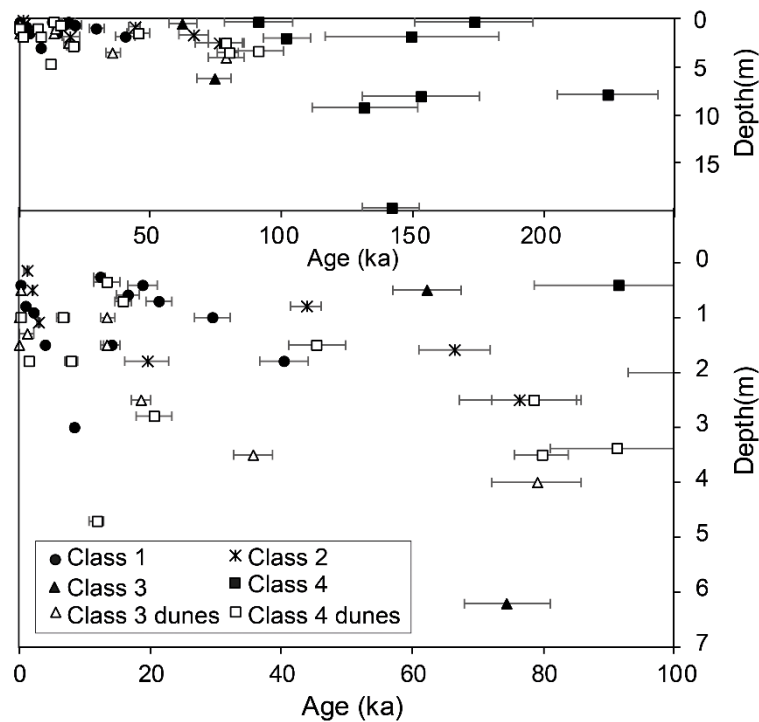


Figure 5-6. Age depth relationship of all samples displayed by morphological class. Top: all ages. Bottom : <100ka.

5.4.7 SEDIMENT ORIGIN: LUMINESCENCE SENSITIVITY

Aeolian reference samples (NAM07/3/2, NAM07/4/12 and NAM14/10/3) demonstrate that the 90-125 μm size fraction is more sensitive than the 180-210 μm fraction (Figure 5-7). This is thought to be due to an absolute increase in luminescing grains. Although different aliquot sizes were used to normalise the number of grains on discs, medium aliquots of 90-125 μm size grains are still likely to contain ~2000 grains compared to ~1000 grains in large aliquots of 180-210 μm (Duller, 2008). As ~95% of luminescence signal is typically produced by ~5% of grains (Duller et al., 2000) an increase in the number of measured grains increases the likelihood of sampling grains which produce a luminescence signal and therefore the sensitivity is likely to be higher regardless of weight. This

interpretation of the difference between aeolian size fractions is supported by their similarity in response to heating to 500°C (Figure 5-7B).

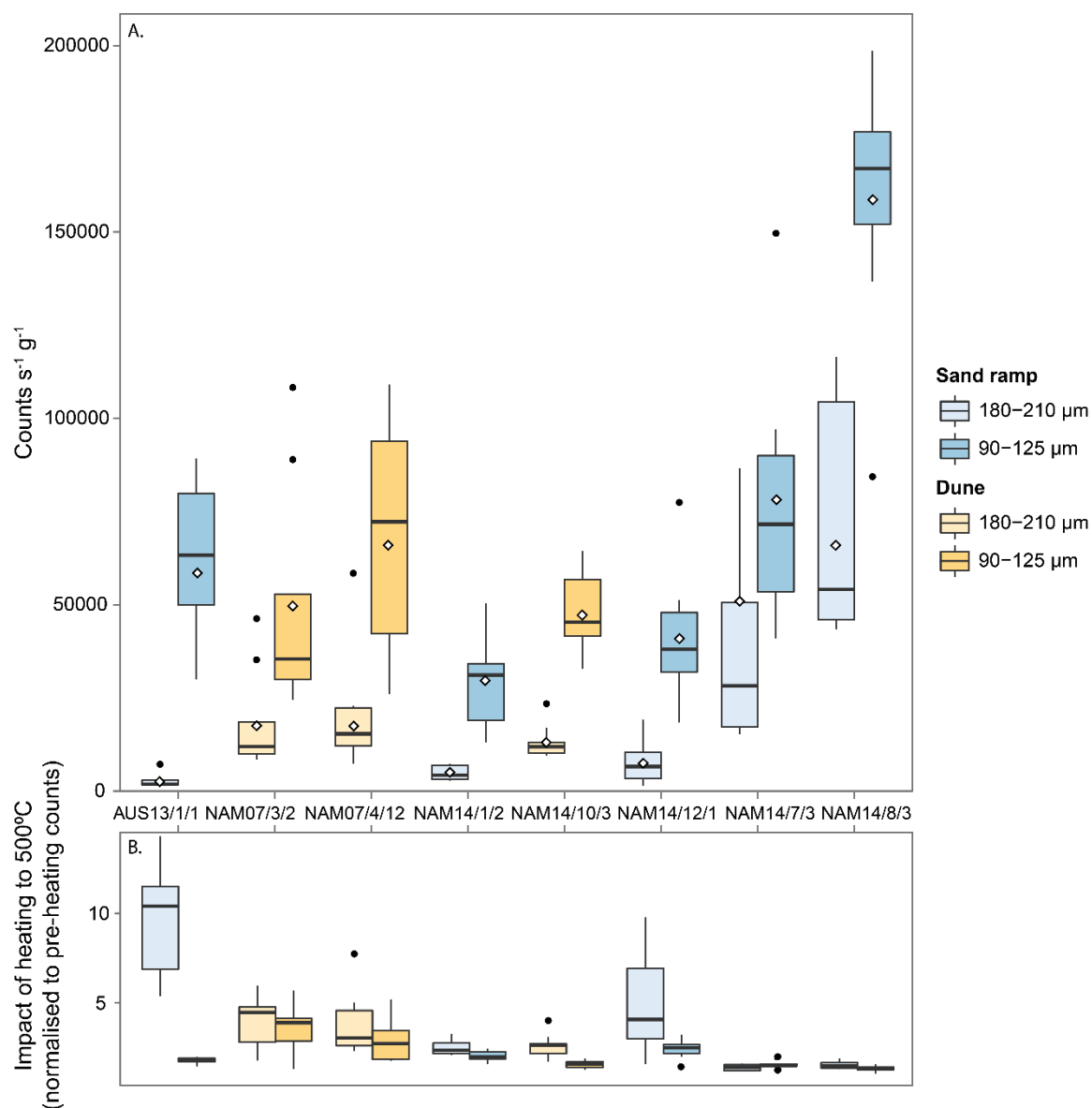


Figure 5-7. Luminescence sensitivity of different grain-sizes A.) in response to a 50 Gy given dose B) in response to a 50 Gy given dose after heating to 500°C (normalised to pre-heating counts/s/g). Box plots show the interquartile range; outliers are shown by black dots. Black lines represent the median and white diamonds the mean.

The 90-125 μm size fraction of sample AUS13/1/1 is an order of magnitude more sensitive than the 180-210 μm fraction when responding to a given dose of 50 Gy (Figure 5-7). This is well above the baseline difference observed in aeolian samples (Figure 5-7) and indicates a real difference in sensitivity between the size fractions. This is confirmed by the large sensitivity increase of the 180-210 μm size fraction after heating compared to the 90-125 μm fraction (Figure 5-7B). The observations from other sand ramps are less conclusive. A large difference in brightness between size fractions is observed for samples NAM14/8/3 and NAM14/1/2 (Figure 5-7A) but no difference is observed after heating (Figure 5-7B). Sample NAM14/7/3 showed significant feldspar contamination in the 180-210 μm size fraction. Therefore, results, which show no difference between size fractions, may be skewed by the small number of accepted aliquots. Sample NAM14/12/1 shows some difference between size fractions in both brightness and response to heating. This preliminary dataset highlights the potential of this technique as a provenancing tool but also demonstrates the variability of sand ramp sediments and the need for a larger dataset before definitive conclusions can be drawn.

5.4.8 SEDIMENT ORIGIN: HEAVY MINERAL PROVENANCE

All Aus samples contain a high garnet concentration and a secondary amphibole component (Figure 5-8). This matches the Mesoproterozoic rocks of the Namaqua Metamorphic Complex which outcrops in the Aus region and consist mainly of garnet/kyanite to garnet-sillimanite paragneisses with subordinate orthogneisses and amphibolites (Becker et al. 2006). The sediments of Neuhof-1 (NAM14/6/2,

NAM14/7/3) and Neuhof-2 (NAM14/8/3) and the ephemeral river channel at the base of Neuhof-1 (NAM14/6/R) are similar, all dominated by a very high amphibole content and a lesser epidote content (Figure 5-8). This shows a strong agreement with the Neuhof Formation described in this area (Becker et al., 2006). The presence of kyanite in NAM14/7/3 and NAM14/8/3 suggests significant contribution from locally exposed amphibole-facies metasediments. Dune (NAM14/4/2) and sand ramp (NAM14/3/2) sediments from Jagkop are similar but not identical. Both contain significant pyroxene suggesting an affinity to the sands of the Orange River and the Namib Sand Sea (Garzanti et al., 2012). However, heavy mineral concentration is low and the significant presence of amphibole, garnet, epidote and ZTR is akin to the local Sinclair Group geology. The absence of pyroxenes at Aus and Neuhof indicates zero to negligible supply of sediment from the Namib Sand Sea (Garzanti et al., 2012). Comparison of sand ramp samples to samples from the Namib Desert and the major ephemeral rivers of southern Namibia (Garzanti et al., 2012) using PCA analysis shows Neuhof and Aus are distinct from each other and from the Namib Desert and major river systems. Neuhof-1 and the associated channel are closely related (Figure 5-8). Jagkop shows similarities with the Namib desert and the major ephemeral rivers.

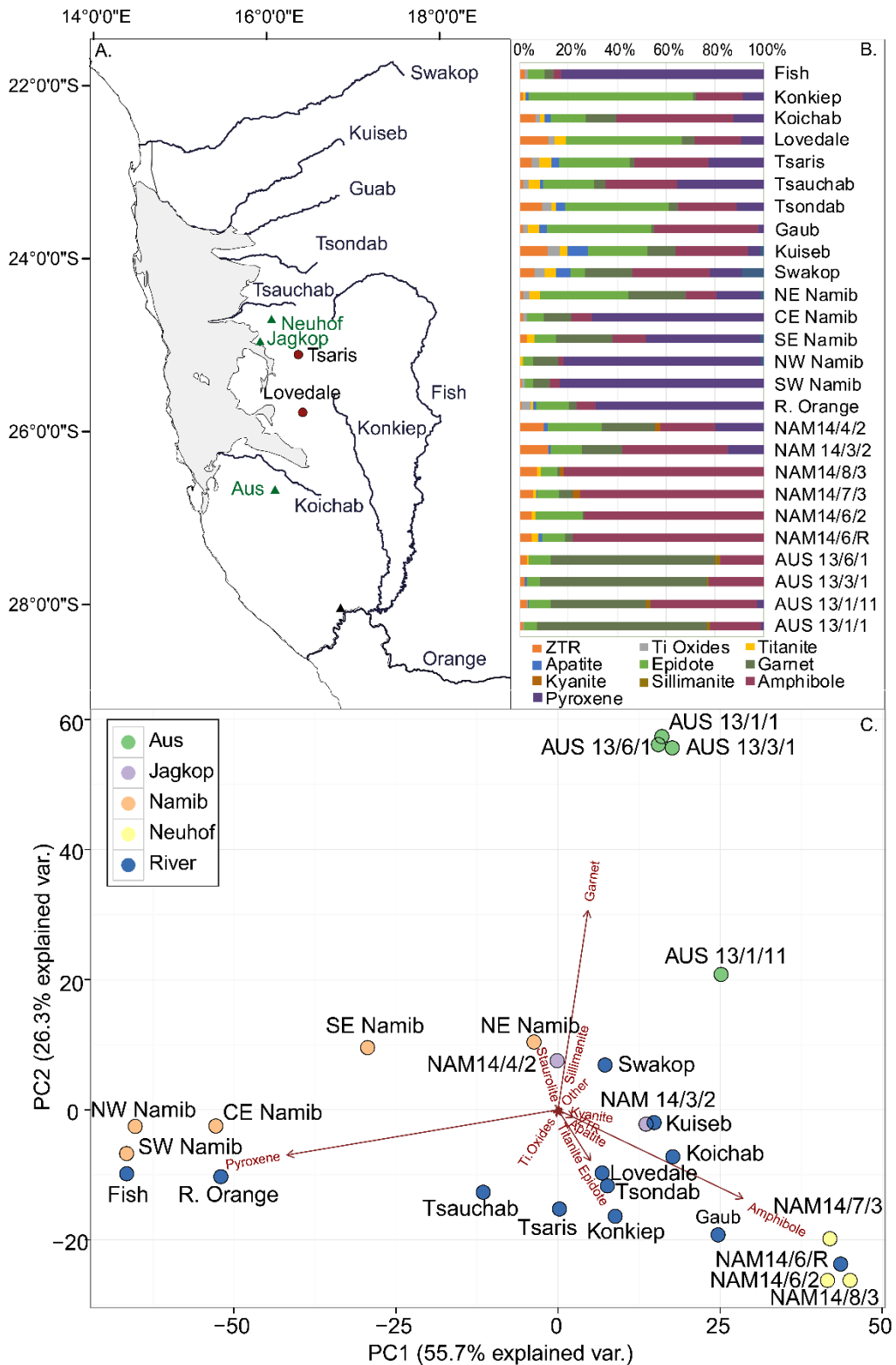


Figure 5-8. Heavy mineral properties of Aus sand ramp (AUS13/1/1, AUS13/1/11) and dunes (AUS13/3/1, AUS13/6/1), Neuhoef-2 sand ramp (NAM14/8/3), Neuhoef-1 sand ramp (NAM14/6/2, NAM14/7/3) and associated ephemeral river channel (NAM14/6/R), Jagkop sand ramp (NAM14/3/2) and dunes (NAM14/4/2) in comparison to major ephemeral rivers of southern Namibia and dunes of the Namib Desert (data from Garzanti et al., 2012). A). Sampling locations. Rivers were sampled at multiple locations along their course (see Garzanti et al. 2012). B). Heavy mineral composition by percent. C). Biplot of heavy mineral PCA results Component 1 is primarily controlled by the difference between amphibole and pyroxene and explains 56% of the variance. Component 2 is controlled by the contrast between garnet, and amphibole, epidote and pyroxene apatite and explains 26% of variance. Proximity of samples to vectors or each other indicates similarity.

5.5 DISCUSSION

5.5.1 CONTROLS ON SAND RAMP FORMATION

Data on the locations and sediments of identified sand ramps (Figure 5-2 and Figure 5-4) indicate that four factors govern sand ramp formation in southern Namibia:

1. an adequate supply of sediment with significant aeolian transport potential;
2. an appropriate accommodation space;
3. directionally-persistent winds with sufficient energy for aeolian sand transport;
4. an arid to semi-arid climate but with sufficient seasonal, or longer term, variability to promote differing geomorphic and depositional processes.

All four factors need to be met for sand ramp accumulation to occur. Factors 1 and 2 are local and determine where sand ramps form in a region where factors 3 and 4 are met. Therefore, the pattern of sand ramp (non) occurrence in southern Namibia can be explained by factors 1 and 2. All sand ramps are within 5.5 km downwind of sand sources in the form of river channels or dune fields (Appendix C, Figure 5-3). Local topographic perturbations in wind dynamics influence aeolian transport potential (Xianwan et al., 1999) and thus sand ramps do not form in accommodation spaces where winds are disrupted by topographic obstacles, despite the occurrence of available sediment sources (Appendix C, Figure 5-2). Ramps do not accumulate on larger mountain complexes with

significant run-off generation or when there are no sheltered accommodation spaces perpendicular to the wind. Thus, suitable accommodation spaces must be subject to minimal aeolian and alluvial erosion. Typically, these are small concave facets on inselbergs/small inselberg complexes or along the flanks of elongate inselbergs (Appendix C, Figure 5-2).

Over 70% of sand ramps in this study are east-facing. All of these ramps are climbing features and all morphological classes are represented. >80% of west-facing sand ramps are small, Class 1 features and ~10% are falling features. West facing Class 3 or 4 ramps are not observed (Appendix C, Figure 5-3). This suggests that seasonal easterly winds are the primary creator of sand ramps in southern Namibia whilst the prevailing SW winds also have a small influence. Wind measurements are sparse on the eastern margin of the Namib Sand Sea (Livingstone et al., 2010). Nevertheless, available data agree with the distribution of sand ramps. SW winds are capable of reaching transport velocities ~2% of the time but wind speeds rarely exceed 8 ms^{-1} . East winds only reach transport velocities ~0.5% of the time but can reach speeds of $14\text{-}17 \text{ ms}^{-1}$ and thus total sand transport potential is greater (Lancaster et al., 1984; Lancaster, 1985). The seasonality of east winds may also enhance sand ramp preservation as sediments may begin to stabilise between wind events whilst west facing sand ramps may undergo more persistent erosion. Aerial photography indicates that there are several west facing accommodation spaces that currently do not contain sand ramps.

5.5.2 TIME SCALE OF FORMATION

OSL dates confirm provisional dating by Bertram (2003) and indicate that sand ramps have been present in the Namibian landscape for over 100 ka (Figure 5-5). This demonstrates that sand ramps are not solely features of the Last Glacial and that they can act as long term sediment stores. As found by Telfer et al. (2012) and Ellwien et al. (2015) deposition is episodic for most ramps with only Simplon (Class 1) potentially representing a single, rapid phase of deposition as described by Bateman et al. (2012) and Thomas et al. (1997) (Figure 5-5). Chronologies vary between sand ramps reflecting local accommodation space availability, sediment supply and wind dynamics but common periods of activity are apparent (Figure 5-6).

5.5.3 SIGNIFICANCE OF SAND RAMP MORPHOLOGY

Similar ages are found at similar stratigraphic locations regardless of morphological class (Figure 5-5 and Figure 5-6). This indicates that the size and morphology of sand ramps is most likely to reflect the size of the accommodation space rather than representing the evolution of sand ramp complexity with age.

Akin to the sand ramps in this study, large free-form dunes are frequently superimposed by smaller dunes (Lancaster, 2009; Livingstone et al., 2010). The mechanism controlling superimposition is debated with large scale changes in wind regime and climate (Warren and Allison, 1998), seasonal or annual variation in wind (Bristow et al., 2007), landscape patterning and self-organisation (Dong et al., 2009; Ewing et al., 2006) and stabilisation of the underlying dune due to

climatic change (Dong et al., 2004) all suggested as potential drivers. There is some evidence that the size of the dune determines the quantity and complexity of superimposed features with secondary features unable to form where the primary dune is below a certain size (Al-Masrahy and Mountney, 2013; Breed and Grow, 1979). Thus, secondary dunes may simply be a function of airflow fluctuation created when the flanks of the dunes are large enough to form a planar surface (Lancaster, 1988). In this study superimposed dunes only form on sand ramps with a length or width >1 km with the number of secondary features increasing with sand ramp size (Appendix C). Therefore, for these sand ramps, superimposed dune features are likely to represent the opportunistic filling of the secondary accommodation space created by the planar flank of the ramp.

The main bodies of the sampled Class 3 and 4 sand ramps are semi-consolidated to consolidated and date to >60 ka. The superimposed dunes of Class 3 and 4 ramps and the main bodies of Class 1 and 2 ramps are unconsolidated and have episodic chronologies from present to >75 ka (Figure 5-5). This suggests that aeolian deposition becomes focused on the superimposed dunes, enabling the underlying sand ramp sediments to stabilize. Therefore, Class 3 and 4 ramps are likely to preserve initial accumulation with little reworking. The small size of Class 1 and 2 sand ramps inhibits the formation of secondary features and thus, unless protected by a talus mantle, surface sediments are vulnerable to aeolian reworking. The similarity between dune and Class 1 and 2 chronologies suggests both are responding to the same environmental forcing. Therefore, the size of the

accommodation space, and thus the size of the sand ramp, controls the accumulation history of the sand ramp.

Accommodation space may also control the formation of the gullies which separate the head of sand ramps from topographic obstacles. Steep slopes promote reverse flow and thus aeolian erosion (Tsoar 1983; Pye and Tsoar 1990; Qian et al. 2011a; Qian et al. 2011b) whilst the accommodation space also channels run-off to promote either continued erosion downslope or erosion at the head of the sand ramp. All gullies now show evidence of flow and fluvial erosion but, depending on the accommodation space, aeolian activity may have played a role in initiation at some sites.

Gullies at the head of the sand ramps protect the sediments from downslope erosion from run-off but provide a mechanism for sediment to be removed from the sand ramp via aeolian erosion and slumping, as exemplified by the erroneously young ages obtained from the gully of Klein Aus and from the shallow gully of Rosh Pinah (Figure 5-5). Therefore, if gullies have been present throughout sand ramp accumulation, episodic deposition at Keerwader and Neuhof-2 may be explained by repeated periods of erosion and subsequent accommodation space availability and accumulation. This may explain the similarity of the dune and sand ramp records.

5.5.4 SIGNIFICANCE OF SEDIMENTARY FEATURES

The consolidation of sediments and the formation of duricrust layers is associated with sand ramp stability whilst the significance of stone layers and the chronology of unconsolidated sediments is site specific (Figure 5-5). Well-developed calcretes (such as those at Samara, Jagkop and Aus) require lengthy development times under stable conditions with little sediment input (Wright, 2007). At Samara, calcrete formation is associated with a ~70 ka hiatus in accumulation whilst ages from consolidated sediments below duricrusts at Aus and Jagkop are >100 ka indicating long term sand ramp stability. Calcretes increase in complexity with time (Netterberg, 1969). Simple calcrete nodules at Keerwader are associated with a 13-20 ka hiatus, indicating medium term stability. Akin to Bateman et al. (2012) and Rendell and Sheffer (1996) buried stone layers are not always associated with long accumulation hiatuses. At Neuhof-1 the stone layer represents between ~1.6-7.1 ka (mean 4.3 ka) possibly suggesting event driven distribution (cf. Bateman et al., 2012) whilst a longer period of stability is possible for Jagkop (Figure 5-5). Unconsolidated sand ramp sediments do not necessarily represent recent activity with ages ranging between 0.2 - >75 ka with no clear relationship between sedimentary features and chronology.

There may be some relationship between the presence of a well-developed talus mantle on the surface and the underlying sand ramp accumulation history. Rapid accumulation in a single event is suggested for the sand ramps at Soldier Mountain, USA and Arkandan, Iran (Bateman et al., 2012; Thomas et al., 1997a) whilst episodic accumulation is inferred for the Black Masa sand ramps, USA and

the Golden Gate, South Africa sand ramp (Ellwein et al., 2015; Telfer et al., 2012). Both the Soldier Mountain sand ramp and the Arkandan sand ramp are capped by a talus mantle whilst the Black Masa and Golden Gate sand ramps are not, indicating the opportunistic formation of a talus layer after the initial filling of the aeolian accommodation space can protect the aeolian sediments from subsequent erosion. In this study, Rosh Pinah, Keerwader and Neuhof 2 are not covered by a talus mantle. Dates from <1 m depth indicate activity within the past 2.5 ka. Conversely dates from <1m from the sand ramps covered by talus (Simplon, Neuhof-1, Sandkop and the Class 4 sand ramps) are >10 ka suggesting the talus offers some protection against reworking. At Neuhof-1 ~12 ka of surface stability is implied whilst Neuhof-2, 4 km away, suggests activity within the last ~1000 years. However, accumulation of Neuhof-1 is episodic, probably representing the high sensitivity of this small sand ramp but also indicating the preservation of the sand ramp record is dependent on the timing of talus emplacement. Simplon may represent a single deposition event whilst Sandkop and the Class 4 sand ramps are at the upper limit of multi-grain OSL dating so accurate accumulation rates cannot be calculated.

Due to the number of processes operating on sand ramps, accumulation is unlikely to be a binary process of aeolian or colluvial deposition followed by preservation. The 90-125 μm size fraction of AUS13/1/1 produces an order of magnitude more counts $\text{s}^{-1}\text{g}^{-1}$ for a 50 Gy dose than the 180-210 μm fraction (Figure 5-7). This indicates that the 90-125 μm fraction has undergone several burial and exposure cycles (i.e. an aeolian sediment history) before being

deposited whilst the 180-210 μm fraction is more likely to be freshly weathered material (i.e. colluvium) (see Moska & Murray 2006; Pietsch et al. 2008; Fitzsimmons et al. 2010). The presence of both populations in a single sample indicates interplay between the two depositional mechanisms. This is supported by the poor sorting of the ≤ 2 mm fraction, the presence of angular clasts amongst the aeolian sands and the frequent absence of bedding structures in aeolian units (Bertram, 2003; Lancaster and Tchakerian, 1996). However, the sensitivity results from Neuhof-2, Rosh Pinah and Sandkop are less conclusive (Figure 5-7). Sensitivity has also been shown to be influenced by formation conditions of the parent material (Sawakuchi et al., 2011a, 2011b) therefore the inconclusive results of NAM14/1/2, NAM14/8/3 and NAM14/12/1 may be due to inherent characteristics of the source material. Alternatively, these samples may share the same sedimentary history across size fractions.

As colluvial and aeolian transport act in opposite directions on ramp slopes it is likely that the sediments undergo post-depositional reworking and mixing regardless of initial deposition mechanism. Therefore, depending on accumulation rate, OSL ages may represent environmental disturbance and reworking rather than the time of original sediment supply. Nevertheless these dates are likely to be accurate as surface reworking reduces the likelihood of partial bleaching (i.e. insufficient signal resetting, a concern in many colluvial environments (e.g. Fuchs & Lang 2009)) and given the temporal scale of the chronologies the difference between deposition and burial by other sediments is likely to be inconsequential in most cases.

Akin to the sand ramps of the Mojave (Pease and Tchakerian, 2003), sediment supply to the Aus and Neuhof sand ramps is of local origin with the heavy mineral assemblages reflecting the local geology (Figure 5-7, Becker et al. 2006). At Neuhof-1 the heavy mineral suite from the sand ramp and the neighboring ephemeral river channel are nearly identical, confirming the potential of the river channel as a sediment source. At Japkop the heavy mineral assemblage closely resembles the local geology but with a 15-30% contribution from the dunes of the Namib Sand Sea which is immediately to the south and east of the mountain. The majority of studied sand ramps are within 4 km of river channels or within 5.5 km of dune fields (Figure 5-3), suggesting these are the most likely sediment sources. Of the dated sand ramps all but Jagkop have an associated river channel <4 km downwind which most likely acts as the primary sediment source. The local origin of the sediments suggests that periods of colluvial and aeolian deposition both reflect local environmental conditions and if the accommodation space is available, local sediment supply is expected to be the controlling factor in sand ramp activity.

Individual sand ramps therefore reflect local conditions wherein sediment availability coincided with an available accommodation space. As sediment availability is determined by regional environmental conditions, periods of accumulation observed across multiple sand ramps are likely to reflect regional palaeoenvironmental conditions. Therefore, analysis of the sedimentology, morphology and chronology of multiple sand ramps, can identify periods of local and regional sand ramp activity (e.g. Figure 5-6). If factors controlling sediment

supply are understood, these results can then be used to reconstruct sediment dynamics, palaeoenvironmental conditions and potentially palaeoclimatic conditions, over Quaternary timescales.

5.6 CONCLUSIONS

Sand ramps are abundant along the eastern margin of the Namib Sand Sea and are present, although less common, in other locations in south west Namibia with 75 sand ramps identified in the study region (Figure 5-2). These sand ramps range from 150-3000 m in length with varying morphology and sedimentology. OSL dating indicates episodic sand ramp activity between 0.2 and >100 ka (Figure 5-6) although more recent ages may represent surface reworking. This confirms preliminary dates from Bertram (2003) and indicates that sand ramps are not solely Last Glacial features but existed in the Namibian landscape during the Last Interglacial, and likely before.

Analysing where sand ramps do and do not form across the geological and climatic gradients of the study area has allowed the conditions needed for sand ramp formation to be assessed. These controls are proposed to be:

1. an adequate supply of sediment with significant aeolian transport potential;
2. an appropriate accommodation space;
3. directionally-persistent winds with sufficient energy for aeolian sand transport;

4. an arid to semi-arid climate but with sufficient seasonal, or longer term, variability to promote differing geomorphic and depositional processes

Four classes of sand ramp morphology were identified. These closely link size to complexity and indicate that the available accommodation space, and the ultimate size a sand ramp can grow to, play a large role in the morphology of the sand ramp and potentially its sensitivity to environmental change. Sediments of large sand ramps are able to stabilise as aeolian activity is refocused on secondary dune features which are able to form on the planar surface of the sand ramp. Dunes are unable to form on smaller sand ramps, leaving the surface vulnerable to reworking unless it is protected by talus.

The stratigraphy of individual sand ramps reflect local conditions wherein sediment availability coincided with available accommodation space or unstabilised sediments were reworked. Accommodation space availability offers a purely local control on sand ramp formation whereas sediment supply provides a regional environmental picture with ephemeral rivers representing the likely sediment source for most of the studied sand ramps.

Overall, these results suggest that palaeoenvironmental and potentially palaeoclimatic data can be established if multiple sand ramps are analysed regionally. The primary record of individual sand ramps is local and is influenced by non-climatic controlling factors including: the size of the sand ramp, the availability of the accommodation space and the sediment supply. However, when

multiple sand ramps are analysed in sedimentological and morphological context, regional patterns of sediment dynamics, which are likely to reflect past regional climatic changes, can emerge .

Chapter 6: Challenges of reconstructing Quaternary environments from sand ramp sediments: a reanalysis of the Klipkraal Sands, South Africa

In light of the findings of Chapter 5 this paper re-analyses a previously studied sand ramp from South Africa. This paper explores the regional variation of controls operating on sand ramps and discusses the challenges of retrospectively identifying geomorphic processes from sediments alone. The implication for palaeoenvironmental reconstruction using sand ramps is touched upon and developed further in Chapter 8

6.1 ABSTRACT (FROM PAPER SUBMITTED TO 'GEOMORPHOLOGY' – AUTHORS:
A.L.K. ROWELL, D.S.G. THOMAS, R.M. BAILEY AND P. HOLMES)

The Klipkraal Sands is an example of a sand ramp on the north-eastern margin of the semi-arid Karoo. Thomas et al. (2002) dated this feature to 24-11 ka and interpreted it as evidence of long-distance aeolian transport during the Last Glacial Maximum (LGM). This study provides a new chronology and geomorphic interpretation for this landform.

Thirteen OSL dates indicate the Klipkraal Sands accumulated periodically between ~100-0.14 ka. Wet dispersal laser granulometry shows the material is finer and more poorly sorted than originally suggested. Therefore, the sediments are suggested to be locally sourced with a significant colluvial component and the feature is reinterpreted as showing no particular affinity to the LGM. This site demonstrates the complexities of retrospectively identifying geomorphic processes from sand ramp sediments. A complex interplay of aeolian and colluvial processes is suggested, from which clear palaeoenvironmental interpretation cannot be easily achieved. This implies that the local geomorphic processes and controls operating on sand ramps need to be established for a region before sand ramps can be fully utilised as palaeoenvironmental archives.

6.2 INTRODUCTION

The semi-arid interior of South Africa is largely not conducive to the preservation of organic-based proxies whilst vegetation cover is sufficient to preclude the formation of large-scale aeolian features (Chase and Meadows, 2007; Thomas and Burrough, 2012). Therefore, data for reconstructing Late Quaternary environmental change is based on a limited number of available records, often from atypical circumstances (e.g. high altitude records from the Drakensberg; Mills et al., 2009). Sand ramps (see Figure 3.1) have been identified in semi-arid South Africa (Telfer et al., 2012, 2014; Thomas et al., 2002) and are thought to be relatively common landforms (Telfer et al., 2014). The palaeoenvironmental interpretation of sand ramp sediments is developing (e.g. Bateman et al., 2012; Lancaster and Tchakerian, 1996; Thomas et al., 1997; Chapter 5). Therefore, if ramp formation processes and controls can be identified and understood, these features have the potential to provide valuable, dateable records of Late Quaternary environmental conditions in this understudied area.

The sand ramp, initially reported as a climbing dune (Marker and Holmes, 1993; Thomas et al., 2002), at Klipkraal was the first of these features to be studied in southern Africa (Marker and Holmes, 1993; Thomas et al., 2002). The 'Klipkraal Sands' feature was interpreted as a unique aeolian landform in a region otherwise dominated by hillslope processes (Marker and Holmes, 1993; Thomas et al., 2002) and given as evidence of reduced vegetation cover and strong anti-cyclonic conditions over the LGM and early Holocene (Thomas et al., 2002).

Recently, Telfer et al. (2012, 2014) have examined features similar to the Klipkraal Sands in the Drakensberg mountains. Like the feature at Klipkraal, the Drakensberg ramps have been interpreted as primarily aeolian in origin (Telfer et al., 2012, 2014). However, both the Klipkraal Sands and the Drakensberg features share sedimentary and morphological characteristics with a Quaternary colluvial unit, formalised as the Masotcheni Formation to the east of the Great Escarpment, and widespread in eastern, central southern Africa (Botha, 1996; Clarke et al., 2003; Temme et al., 2008; Watson et al., 1984). This casts uncertainty on the previous interpretation of the Klipkraal Sands and suggests reanalysis may provide valuable information on the controls and nature of sand ramp formation. Additionally, chronological techniques have advanced significantly over the past 15 years (Lowe and Walker, 2015) and re-dating of previously studied features can considerably improve interpretation and understanding (e.g. Bateman et al., 2012).

In light of recent research into regional features (Telfer et al., 2012, 2014) and the broader contexts of sand ramp development (Chapter 5), this study aimed to reassess the palaeoenvironmental significance of the Klipkraal Sands, using new chronometric and sedimentological data.

6.3 STUDY SITE AND METHODS

Klipkraal (31°23'35.10"S, 26°41'24.40"E, Figure 6-1) is located within the Eastern Cape Province on the north eastern margin of the Karoo biome, South Africa. It is situated at 1800 m a.s.l. on the north western flank of a ridge within the

Stormberg Mountain range. The ridge and surrounding mountains are composed of Elliot Formation mudstones capped with resistant outliers of Clarens Formation sandstone (Marker and Holmes, 1993). Presently the site sits within the southern African summer rainfall zone with >50% of mean annual precipitation (MAP) derived from the Indian Ocean monsoon. Occasional winter rainfall originates from the passing of cold fronts associated with mid-latitude cyclones (Tyson and Preston-Whyte, 2000) and brief but heavy snowfalls are not uncommon. MAP is 400-600 mm with 25-35% annual variability and relative humidity typically above 40%. Sub-freezing temperatures are frequently experienced throughout winter whilst mean summer temperatures are 22-28°C (Schulze et al., 1997). Regional wind regimes are bimodal, prevailing from the NW and SE (WASA, 2014).

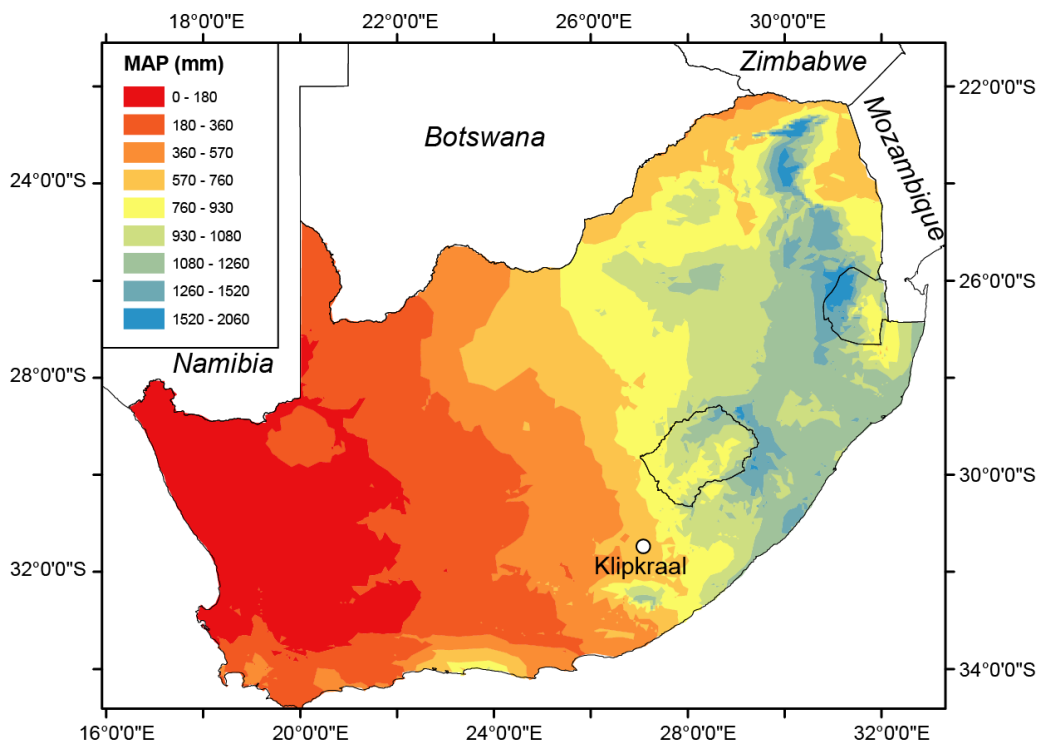


Figure 6-1. Mean annual precipitation in South Africa (1950-2000) (Hijmans et al., 2005). The location of Klipkraal is indicated.

6.3.1 FIELD SAMPLING

Sand ramp sedimentology and morphology were logged in the field. Following initial field investigations, twelve samples were taken from four sections. Sampling locations were chosen to examine the three main sedimentological units in multiple locations (Figure 6-2). Samples for OSL dating and sedimentological analysis were obtained by hammering opaque tubes into cleaned exposures (Figure 6-3). The light exposed ends were removed in the laboratory and used for particle-size analysis and loss-on-ignition (LOI) (see section 3.3 for details of sediment analysis and OSL sample preparation, measurement techniques and dose rate determination).

6.3.2 OSL DATING: EQUIVALENT DOSE DETERMINATION

Typically, 24 small aliquots of ~200-500 grains (Duller, 2008) were measured for each sample using a modified SAR protocol (Table 3-2) and the rejection criteria described in section 3.3.6.

Preheat temperatures (PH) were determined following preheat plateau tests on recovered doses from a sub-set of samples (see section 3.3.8). Samples from Unit 3 (Figure 6-2 and Figure 6-3) could recover the given dose within 10% of unity between 220-280°C (Table 3-3). However, at and above 260°C the samples increasingly failed recycling ratios and dose response curves (DRC), and could not be fitted with a single saturating exponential (SSE), a saturating exponential plus linear (SEPL) or a double saturating exponential (DSE) function. Therefore, a PH1 temperature of 240°C and PH2 of 220°C for 10 s were used for D_e

determination. The applicability of these conditions was confirmed by DRTs (Table 3-4). Samples from Unit 2 underestimated given doses by ~25% when a PH1 of 240°C was administered. However, samples recovered the given dose to between 5-12% of unity when a PH1 of 260°C was used. Therefore, samples from unit 2 were measured using a PH1 temperature of 260°C and PH2 of 220°C for 10 s. Preliminary tests indicated samples from Unit 1 were young with low sensitivity. Pre-heat tests were conducted between 120-240°C to monitor the increased likelihood of significant thermal transfer. Samples recovered doses within 10% of unity at all pre-heat temperatures (Table 3-3) and PH1 and PH2 of 200°C for 10 s were chosen for analysis to minimise the likelihood of any thermal transfer.

Individual aliquot D_e estimates were derived by interpolating the natural signal onto DRCs fitted with the most appropriate curve function (SSE, SEPL or DSE). Uncertainties were calculated using 1000 Monte Carlo fits of the curve and propagated with a 2% measurement error. Sample D_e s were calculated using the Central Age Model (CAM) (Galbraith et al., 1999).

6.4 RESULTS

6.4.1 SEDIMENTOLOGY

The sediments of the sand ramp at Klipkraal can be categorised into three distinct units (Table 6-1, Figure 6-2). Unit 1 contains moderate to poorly sorted fine sands

with no clay and minimal silt. The unit is predominantly massive but with some faint cross bedding. The contact with the underlying unit is abrupt.

Unit 2 is a dark brown palaeosol predominantly containing poorly sorted fine sands and silts. Angular clasts with long axes up to 7 cm are common. This unit is wedge shaped, up to 5m thick upslope but thinning towards the base of the sand ramp and not present in the toe sediments. The contact with the underlying unit is gradational.

Unit 3 is the thickest, representing ~70% of the total landform. Sediments are primarily poorly sorted medium sands to silts. Characteristics vary throughout the unit with lenses of consolidated fine sands and silts amid loose massive fine to medium sands, sections of horizontal sand laminations and some gravel lenses are also present. Angular clasts are rare but are found sporadically throughout the unit. Details of individual sample sediment properties can be found in Appendix D.

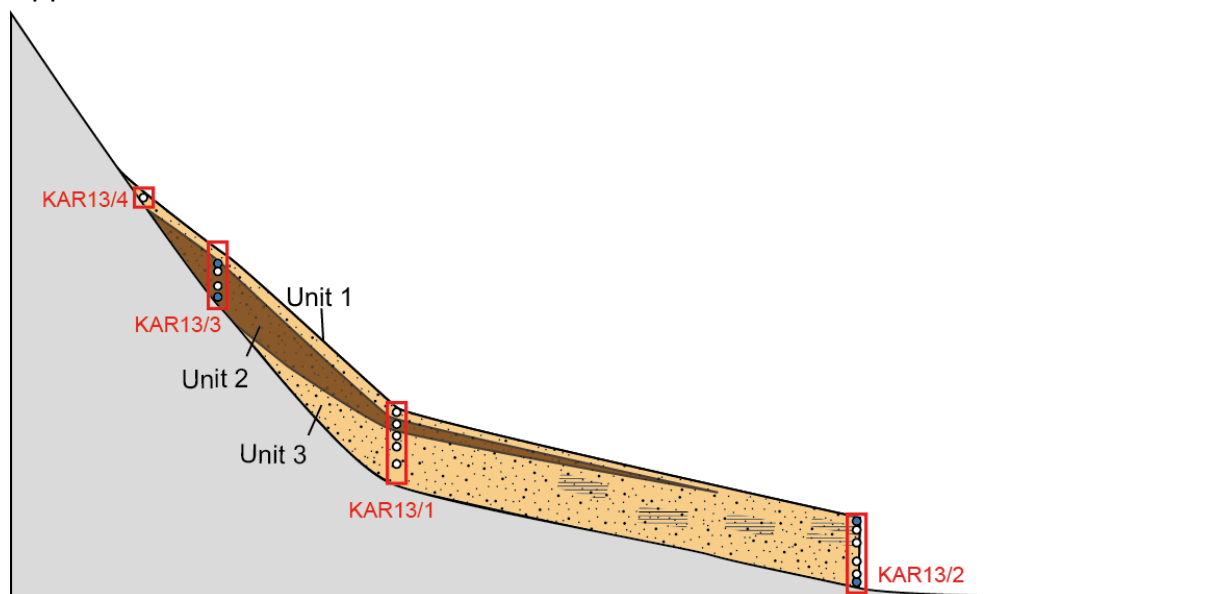


Figure 6-2. Schematic cross section of the Klipkraal sand ramp sediments. Sampling locations in this study are marked by white circles. Red boxes indicate sampling profiles (displayed in detail in Figure 4). Blue circles represent the approximate location of samples studied by Thomas et al (2002). For detail on sedimentology of the different units see Table 6-1 & Figure 6-4.

6.4.2 CHRONOLOGY

OSL data are provided in Table 6-2 and Figure 6-4 shows the final ages in their stratigraphic and sedimentological contexts. Unit 1 dates to between ~140-160 years ago (c.1853-1873 AD). Unit 2 is of Holocene age, accumulating between 3.5 ± 0.3 and 12.4 ± 1.0 ka. Unit 3 accumulated primarily during the Last Glacial, with at least two periods of significant deposition; one at 29.8 ± 3.1 ka to 35.2 ± 2.6 ka and another 86.8 ± 5.3 ka to 95.1 ± 5.3 ka and 101 ± 9 ka. Both sets of samples display apparent ages reversals but are within errors. If ages and uncertainties are assumed correct, stratigraphic relationships suggest deposition of these samples took place in maximum time periods of 32.9 to 32.6 ka and 101 ± 9 to 89.8 ka respectively. Deposition of Unit 3 is also recorded at 14.8 ± 0.9 ka (KAR13/1/4) while sample KAR13/1/3, taken from the diffuse boundary between Unit 2 and Unit 3, dates to $8.7 \text{ ka} \pm 0.5 \text{ ka}$.

Table 6-1 Sedimentological properties of the main units. Grain-size, sorting, skew, kurtosis, organic and carbonate contents are expressed as averages of the samples from these units +/- one standard deviation. Folk and Ward (1957) descriptions are based on average values.

Unit	Depth (m)	Description	Grain-size (μm)	Sorting (ϕ)	Skew (μm)	Kurtosis (μm)	Organic content (%)	Carbonate content (%)	Munsell colour
1	~0.5	Moderately to poorly sorted fine sand with sub-angular to rounded grains.	182 ± 0.5	1.0 ± 0.2 Moderately /poorly sorted	-0.17 ± 0.07 fine skewed	1.1 ± 0.1 Leptokurtic	0.9 ± 0.02	0.5 ± 0.01	7.5YR 4/4 brown
2	0.2-5.0	Sandy soil of poorly sorted silts and fine sands. Individual grains show evidence of chemical dissolution. Frequent angular sandstone clasts up to 7cm	75 ± 31	1.7 ± 0.2 poorly sorted	-0.25 ± 0.02 fine skewed	1.3 ± 0.1 Leptokurtic	2.3 ± 1.0	0.9 ± 0.3	7.5YR 3/2 dark brown
3	0-7.0	Predominantly poorly sorted sub-angular to sub-rounded fine sands with sporadic angular sandstone clast <5mm. Intermittent sections of bedding and lamination. Some evidence of palaeosol development.	120 ± 36	1.7 ± 0.3 poorly sorted	-0.21 ± 0.07 fine skewed	1.2 ± 0.2 Leptokurtic	1.1 ± 0.5	0.8 ± 0.3	5YR 4/6 Yellowish red, 7.5YR 4/6 Strong brown, 10YR 6/3 Pale brown

Table 6-2 Luminescence properties of samples. *n.* is the number of accepted aliquots D_e estimates were calculated using CAM (Galbraith et al., 1999). Dose rates were calculated using DRAC (Durcan et al., 2015).

Sample	Depth (m)	n.	De	Over-dispersion (%)	ICP-MS			Gamma Spectrometry			Beta dose (Gy/ka)	Gamma dose (Gy/ka)	Cosmic dose (Gy/ka)	Dose rate (Gy/ka)	Age (ka)
					U (ppm)	Th (ppm)	K (%)	U (ppm)	Th (ppm)	K (%)					
KAR13/1/1	0.3	17	0.19 ± 0.01	15 ± 1	1.25 ± 0.19	7.22 ± 1.44	0.34 ± 0.03	1.58 ± 0.23	8.39 ± 0.64	0.72 ± 0.07	0.51 ± 0.05	0.54 ± 0.07	0.31 ± 0.03	1.37 ± 0.09	0.14 ± 0.01
KAR13/1/2	0.6	19	6.1 ± 0.4	30 ± 2	1.66 ± 0.26	9.24 ± 1.85	0.56 ± 0.05	n/a	n/a	n/a	0.75 ± 0.06	0.72 ± 0.05	0.28 ± 0.03	1.75 ± 0.08	3.48 ± 0.26
KAR13/1/3	1.1	14	13.1 ± 0.6	21 ± 2	1.43 ± 0.01	8.24 ± 0.12	0.44 ± 0.03	n/a	n/a	n/a	0.64 ± 0.03	0.63 ± 0.02	0.25 ± 0.03	1.52 ± 0.05	8.65 ± 0.47
KAR13/1/4	1.6	12	25.5 ± 0.9	11 ± 1	1.45 ± 0.22	8.49 ± 1.70	0.57 ± 0.05	1.64 ± 0.24	8.76 ± 0.67	0.78 ± 0.07	0.74 ± 0.06	0.76 ± 0.05	0.24 ± 0.02	1.73 ± 0.08	14.76 ± 0.88
KAR13/1/5	3.2	19	180.2 ± 11.1	25 ± 2	1.54 ± 0.24	8.98 ± 1.80	0.72 ± 0.06	n/a	n/a	n/a	0.87 ± 0.07	0.74 ± 0.09	0.19 ± 0.02	1.80 ± 0.12	99.95 ± 8.95
KAR13/2/1	2.4	20	48.2 ± 1.7	15 ± 1	1.30 ± 0.20	6.89 ± 1.38	0.44 ± 0.04	n/a	n/a	n/a	0.60 ± 0.05	0.56 ± 0.07	0.21 ± 0.02	1.37 ± 0.09	35.21 ± 2.62
KAR13/2/2	3.5	19	41.8 ± 3.2	33 ± 3	1.22 ± 0.19	8.09 ± 1.62	0.46 ± 0.04	n/a	n/a	n/a	0.61 ± 0.05	0.60 ± 0.08	0.19 ± 0.02	1.40 ± 0.10	29.79 ± 3.09
KAR13/2/3	4.7	19	121.5 ± 5.9	20 ± 2	1.30 ± 0.04	6.74 ± 0.19	0.42 ± 0.01	n/a	n/a	n/a	0.57 ± 0.02	0.54 ± 0.02	0.16 ± 0.02	1.28 ± 0.03	95.14 ± 5.26
KAR13/2/4	5.5	16	151.7 ± 11.6	28 ± 3	1.52 ± 0.52	8.95 ± 1.79	0.72 ± 0.07	n/a	n/a	n/a	0.86 ± 0.07	0.74 ± 0.09	0.15 ± 0.02	1.75 ± 0.12	86.82 ± 8.80
KAR13/3/1	1	23	14.2 ± 0.2	7 ± 1	1.76 ± 0.27	9.77 ± 1.95	1.17 ± 0.11	n/a	n/a	n/a	1.23 ± 0.10	0.91 ± 0.10	0.24 ± 0.02	2.37 ± 0.15	5.99 ± 0.38
KAR13/3/2	3	24	23.3 ± 0.8	16 ± 1	1.72 ± 0.27	10.53 ± 2.11	0.64 ± 0.06	n/a	n/a	n/a	0.88 ± 0.08	0.81 ± 0.10	0.20 ± 0.02	1.89 ± 0.13	12.35 ± 0.95
KAR13/4/1	0.5	17	0.19 ± 0.02	34 ± 4	0.98 ± 0.15	5.11 ± 1.02	0.39 ± 0.04	n/a	n/a	n/a	0.48 ± 0.04	0.43 ± 0.05	0.29 ± 0.02	1.20 ± 0.07	0.16 ± 0.02

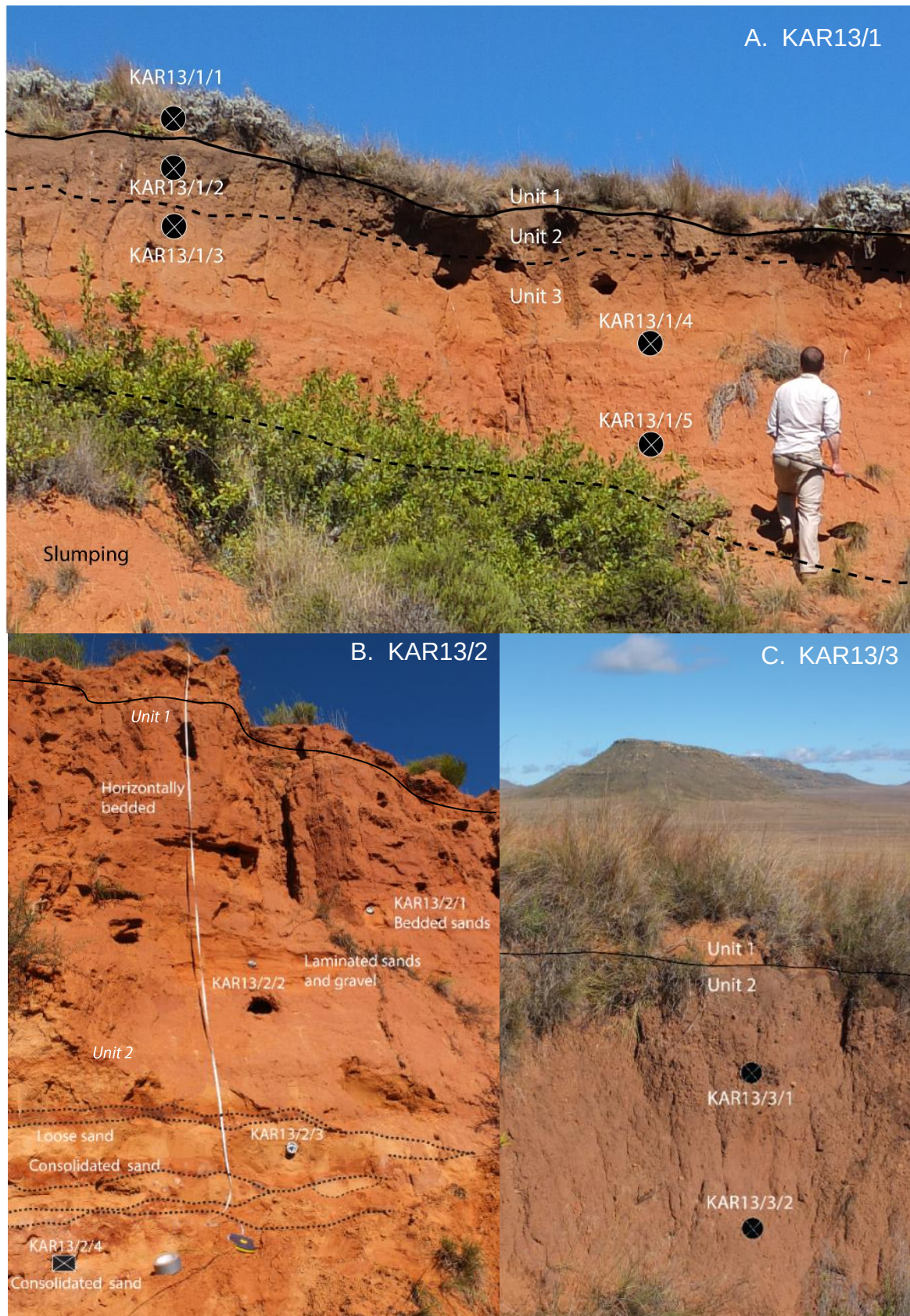


Figure 6-3. Sampling sections of the Klipkraal Sands. A). KAR13/1 B). KAR13/2 C). KAR13/3. When not in situ sampling locations are represented by a black circle or square. Circles represent tube samples and squares block samples. Annotations show sedimentological units and sedimentological properties below the unit scale. Unit properties are described in Table 6-1

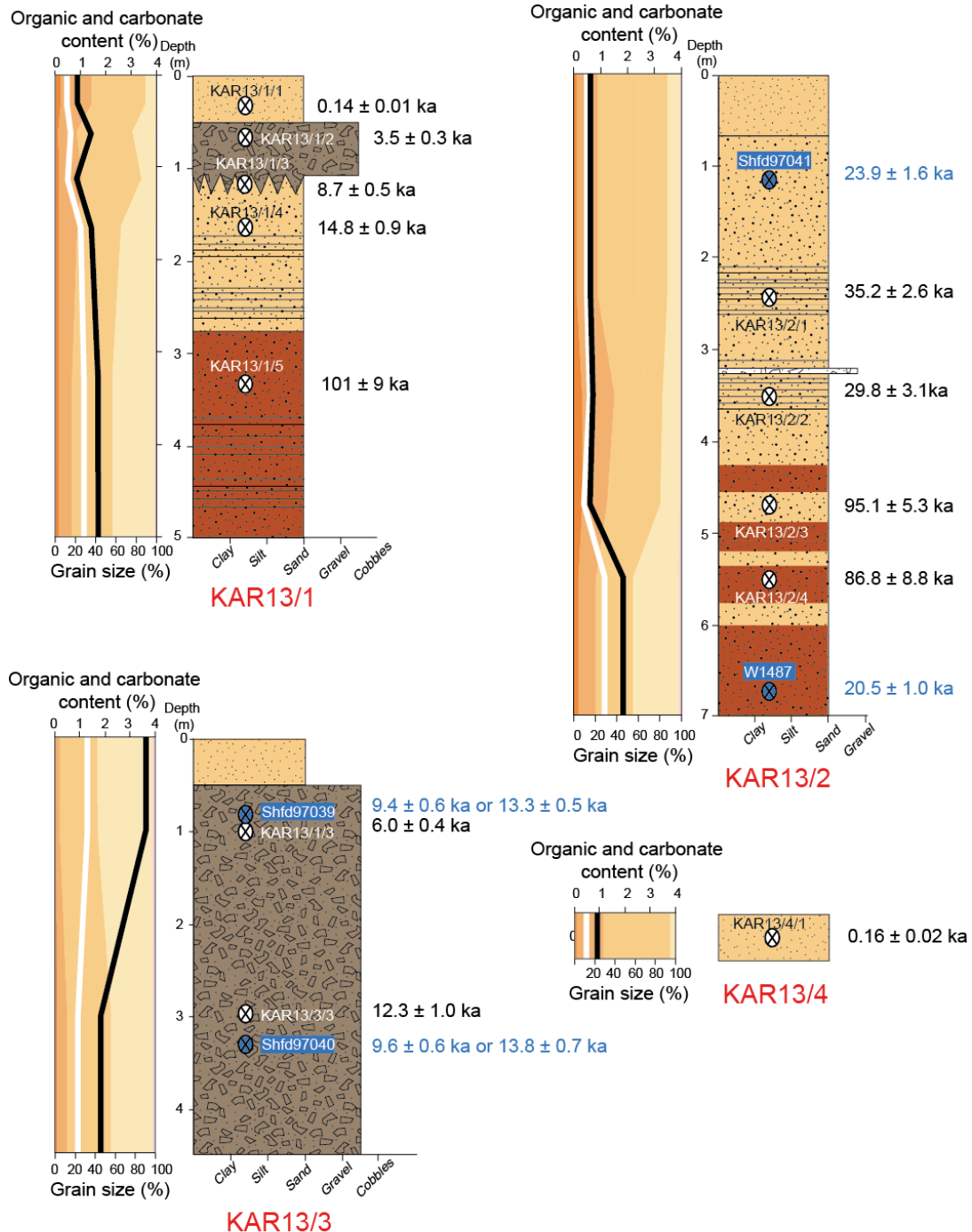


Figure 6-4 Sediment and grain size logs for the Klipkraal Sands sampling profiles. Grain size logs are to the left of sediment logs and are at the same vertical scale. Grain size for KAR13/4 is from a single sample at 0.5 m. For details on the locations of profiles and their relationship to the feature as a whole see Figure 6-2 and Figure 6-3

6.5 DISCUSSION

6.5.1 COMPARISON TO PREVIOUS WORK

Chronology

The assumed locational relationship between the samples from this study and luminescence samples reported by Thomas et al. (2002) is displayed in Figure 6-2 and Figure 6-4. Based on the site description, samples Shfd97039 and Shfd97040 are presumed to be taken from Unit 2 although no mention of a colluvial or palaeosol layer is made in Thomas et al. (2002). Ages for these samples suggest accumulation of 2.4 m of sediment between 9.6 ± 0.6 and 9.4 ± 0.6 ka (TL age) or 13.8 ± 0.7 and 13.3 ± 0.5 (GLSL age) depending on dating method. This is in approximate agreement with the ages obtained from this study (3.5 ± 0.3 ka, 6.0 ± 0.4 ka, 12.4 ± 1.0 ka) as both suggest some early Holocene accumulation (Figure 6-4). However, the chronology of Thomas et al. (2002) suggests rapid accumulation whilst more gradual accumulation over the entire Holocene is implied in this study.

A significant difference in ages is observed between the original samples taken from the toe of the sand ramp and those from this study (section KAR13/2). Thomas et al. (2002) indicate rapid deposition centred on ~22 ka (Figure 6-4). Sample Shfd97041 (TL age 23.9 ± 1.6) is in stratigraphic agreement with the ages obtained from this study (Figure 6-4). However, basal sample W1487 (TL 20.5 ± 1.0 originally from Marker and Holmes, 1993) is significantly younger than samples KAR13/2/3 and KAR13/2/4 (95.14 ± 5.26 ka and 86.82 ± 8.80 ka

respectively). Different depositional histories are derived from the two chronologies: two phases of rapid accumulation between ~24-9 ka are suggested by Thomas et al. (2002) whilst this study indicates episodic accumulation over the past ~100 ka.

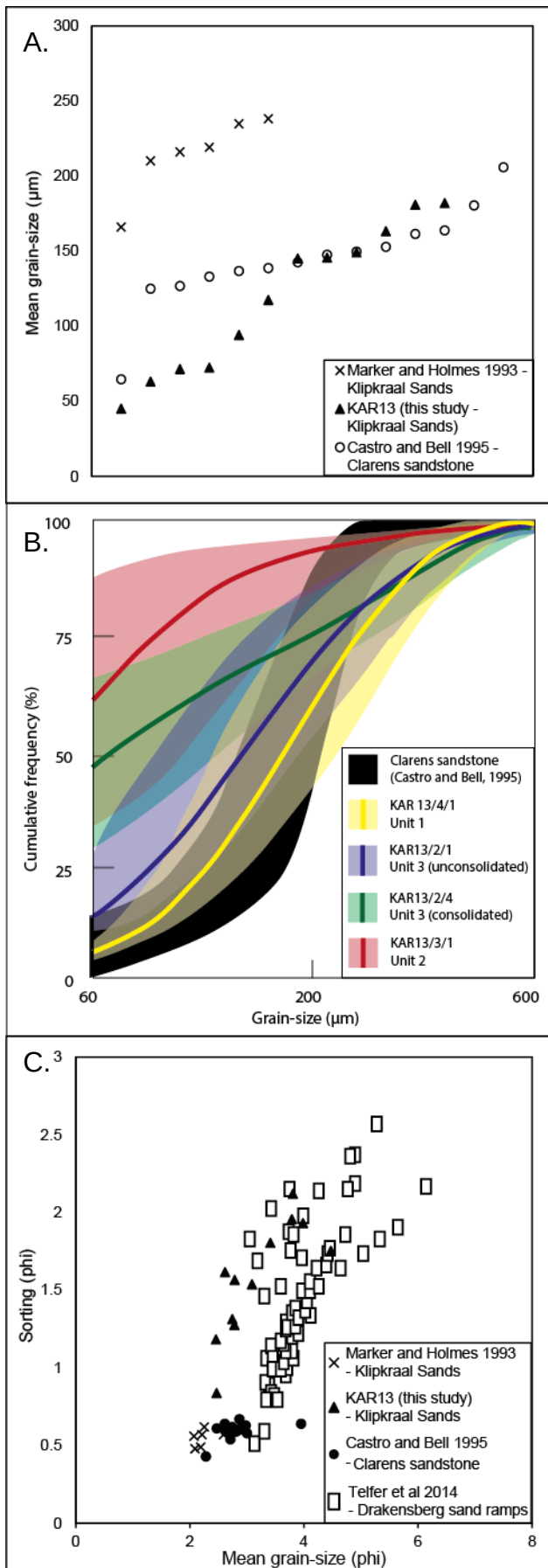
The precise sampling locations of Thomas et al. (2002) could not be re-established and thus differences between ages might reflect spatial variations in the deposition of the sand ramp. Sample W1487 is described as “basal” with no associated depth data so the precise relationship to samples KAR13/2/3 and KAR13/2/4 is unknown. Additionally, Shfd97041 and W1487 were collected during different field investigations and may originate from different sections, potentially explaining the apparent age reversal. If all ages are assumed correct this suggests a very complex accumulation history of the Klipkraal ramp, potentially indicating an old (~100 – 90 ka) core surrounded by younger (<35 ka) sands.

However, at least some of the variation is likely due to dating uncertainties. Luminescence techniques have advanced significantly since 2002 (Roberts et al., 2015) and therefore ages quoted in Thomas et al. (2002) may be associated with unaccounted measurement uncertainties. This explanation is supported by the ~4 ka age difference exhibited by samples Shfd97039 and Shfd97040 when measured using TL and GLSL.

Sediment analysis

Marker and Holmes (1993) concluded that the Klipkraal Sands are far travelled aeolian sediments, presumed to originate from the north during a period of aridity and strong aeolian activity (assumed to be the LGM based on TL dating of W1487; 20.5 ± 1.0). This conclusion was developed by Thomas et al. (2002) who suggested two periods of intense aeolian activity with reduced vegetation cover associated with strong anti-cyclonic conditions between 24-11 ka. These conclusions were primarily based on thin section analysis of Clarens Formation sandstone and on particle size analysis by dry sieving of samples from the Klipkraal Sands. These identified the Klipkraal Sands as slightly coarser than the Clarens Formation sandstone, ruling the latter out as a sediment source for the material in the sand ramp (Figure 6-5A&C; Marker and Holmes, 1993).

The grain size data in this study, obtained by laser granulometry, indicate that the Klipkraal Sands are finer than previously suggested, with mean and median grain-sizes akin to the Clarens Formation sandstone (Castro and Bell, 1995) (Figure 6-5A&B). Grain-size distributions vary between the sediments of the Klipkraal sand ramp, with a small proportion of grains larger than the Clarens Formation sandstone found in Units 1 & 3 (Figure 6-5B). Some of the differences in grain-size distributions may be a result of different analytical methods with wet dispersion laser granulometry results not directly comparable to the dry sieving used by Castro and Bell (1995) (Konert and Vandenberghe, 1997). Additionally, the samples of Clarens Formation sandstone examined by Castro and Bell (1995) originate from Lesotho and regional properties of this lithology may vary.



The overall similarity in grain-size properties of the Klipkraal Sands and the Clarens Formation sandstone means a local sediment source cannot be ruled out. Furthermore, the laser granulometry data also indicates that the Klipkraal Sands are more poorly sorted than previously suggested (Figure 6-5C). As sorting typically increases with transport distance and poor sorting is more associated with slope rather than aeolian transport (Nichols, 2009) this also lends support to the sand ramp sediment being sourced locally rather than being derived from material subjected to long-distance aeolian transport.

Figure 6-5. Grain-size properties of the Klipkraal Sands. A). Mean grain-size measurements from this study compared to the previous study by Marker and Holmes (1993) and samples of the Clarens Formation sandstone which forms the uppermost stratum in the local geology. These samples were taken from Lesotho by Castro and Bell (1995). B). Grain-size distribution of representative samples from each of the Klipkraal Sands units compared to the Clarens Formation sandstone grain-size distribution from Castro and Bell (1995). C). Comparison of grain-size and sorting between the two studies of the Klipkraal Sands (Marker and Holmes, 1993 and this one), the Clarens Formation sandstone (Castro and Bell, 1995) and the sand ramps of the Drakensberg (Telfer et al., 2014).

Human activity and sediment transport

The refined chronology presented in this study has highlighted the role of human activity impacting geomorphic processes in the region. From the mid-19th century until ~1930 the Klipkraal Sands were cultivated for wheat production (Marker and Holmes, 1993). The modern site is heavily gullied (Figure 6-6) with many of the gullies emanating from the remains of farming terraces. Severe gullying is common across the Karoo region, particularly in uplands. Formation largely occurred prior to aerial photography (1945 onwards) (Boardman, 2014; Rowntree, 2014) and is often associated with human activity and landscape disturbance #



Figure 6-6. Aerial view of the Klipkraal Sands. Dashed lines denote remains of 19th/20th century farming terraces. Sampling locations are marked by black circles. Image from Google Earth TM

Unit 1, which is better sorted than the sediments of Unit 3 (Table 6-1, Figure 6-5B, Appendix D), is most likely a product of aeolian reworking of the sand ramp sediments eroded during wheat cultivation and gully formation. It is proposed that sediments transported by water erosion to the toe of the ramp have been blown back up the slope. Therefore, OSL ages from Unit 1 represent a period (c.1863-1883) when human land practices impacted erosion and geomorphic processes in

the region. Marker and Holmes (1993) indicate that the gullies post-date the cessation of farming at Klipkraal, suggesting this unit is most likely associated with erosion due to wheat production. Additionally, the recent age of this aeolian unit suggests that, given an available sediment supply, no major shift in climatic conditions is necessary for aeolian deposition to occur at this site. However, wheat cultivation would have greatly reduced/alterd the vegetation cover and destabilised the sediment over very short time periods. Therefore, this unit only provides a limited analogue for accumulation under long-term environmental change (cf. Holmes et al., 2012).

6.5.2 COMPARISON TO REGIONAL SEDIMENTS

A Quaternary colluvium, known regionally as the Masotcheni Formation, is found throughout eastern central southern Africa with deposits described in Zimbabwe, Swaziland, Lesotho and the KwaZulu-Natal, Free State and Eastern Cape provinces of South Africa (Botha, 1996; Clarke et al., 2003; Dardis, 1990; Rienks et al., 2000; Watson et al., 1984). Sediments are a poorly sorted mixture of clay, silt, sand and gravel (Watson et al., 1984). Two units have been described: the lower Bovu colluvium predates 30 ka and is characterised by “(1) a low clay content (< 10%), (2) vermicular structure, (3) high degree of cementation, (4) incipient ferricrete development, (5) redness, and (6) a general paucity of artefacts”; The upper Mphunga colluvium is of Holocene age and characterised by “(1) high clay content (10%-45%), (2) low degree of cementation, (3) stone lags, and (4) common presence of artefacts”. (Dardis, 1990 pp.934 – citing Price-Williams et al., 1982). Temme et al. (2008 pp.285) assigned hillslope deposits to

the Masotcheni Formation based on “landscape position [low on the hillslope], the fact that sediments were deposited on bedrock stripped of regolith, the evidence of cycles of deposition, soil formation and incomplete erosion leading to colluvia with interbedded palaeosols and their current, often extreme erosion”. Given the geographical extent of the southern African semi-arid zone, it is likely that significant regional variations in the timing and characteristics of colluviation have occurred. Nevertheless, based on the above descriptions, Units 2 and 3 of the Klipkraal Sands display several colluvial properties.

Ramp-like features similar to the Klipkraal Sands have been described in the Drakensberg Mountains (Telfer et al., 2012, 2014). Based on the presence of moderately sorted coarse silt/fine sand units with upslope cross-bedding and a near absence of clasts they are interpreted as primarily aeolian (Telfer et al., 2012, 2014). However, following the definition of Temme et al (2008), Telfer et al (2014) suggest that these features could be assigned to the Mastocheni Formation, inferring that this colluvium may also contain localised aeolian contributions in favourable locations (Telfer et al., 2014). Like the Klipkraal Sands, the majority of the units described by Telfer et al. (2014) are poorly sorted and contain a significant proportion of silts and clays (Figure 6-5C) whilst some are also clast rich, akin to Unit 2 at Klipkraal.

6.5.3 IMPLICATIONS FOR SAND RAMP GEOMORPHOLOGY

Are the Klipkraal Sands colluvial or aeolian in origin?

Aeolian deposition is not necessarily associated with well-sorted, cross-bedded sands (Bateman et al., 2007b), and colluvium does not necessarily contain large clasts or possess very poor sorting (Selby, 1993). Together, this can make ascribing a depositional process to the Klipkraal Sands difficult from sedimentological characteristics alone. This is compounded by the complexity of sand ramps which, by definition, are a product of multiple processes. OSL sensitivity data from Namibian sand ramps suggests that even in aeolian dominated environments, both colluvial and aeolian sands can be present within a single sample of sand ramp sediment (section 5.4.7). This suggests a complex interplay between all erosional and depositional processes operating on sand ramps up to the sample scale.

At Klipkraal, Unit 1 is moderately to poorly sorted, with an abrupt boundary with the former land surface of Unit 2. Although predominately massive, it is thought to be aeolian in origin and associated with historical disturbance of the landscape. Unit 2 is massive with frequent sandstone clast inclusions and a relatively high clay and organic content (both ~2-3%). It is interpreted as a palaeosol formed in colluvium. Unit 3 contains some structure with laminar bedding observed in several locations and occasional cross-bedding described by Marker and Holmes (1993). Large clasts are absent from this unit but small angular clasts <1 cm in diameter are found sporadically and in some concentrated gravel layers. This suggests both aeolian and colluvial/alluvial processes are likely to have been in

operation. However, these features form a discontinuous mosaic with no clear depositional sequence. This makes the relationship between process and chronology tenuous to interpret at the current sampling resolution. Additionally, the majority of sands are massive, providing few clues as to depositional process. This provides further evidence for the complexity of sand ramp sedimentation and suggests features need to be examined individually and in a regional context before palaeoenvironmental conclusions can be drawn. Based on the current information this unit cannot be definitively classed as colluvial or aeolian. There is, however, little if any evidence for long distance aeolian transport playing a role in the development of this feature which is largely comprised of poorly sorted sandy sediments.

Comparison to other southern African sand ramps

The sand ramps of the Namib Desert are found in a region that currently receives less than 200 mm mean annual precipitation with high annual variability (Nicholson, 2000). Geology is predominantly resistant metamorphic rocks of the Sinclair and Namaqua complexes (Geological Survey of Namibia, 2004). Vegetation cover is low and loose sediment is common in the region (Chapter 5). These sand ramps are interpreted as predominately aeolian features with secondary colluvial and fluvial activity. Ramps extend from the piedmont slopes to cover up to ~80% of the hill-slope and are thought to be primarily controlled by the availability of accommodation space coupled with sediment supply. Heavy mineral assemblages confirm a local sediment source for these features (section 5.4.8).

The Klipkraal area currently receives 400-600 mm of rainfall per year and is extensively vegetated (Schulze et al., 1997). The Clarens Formation sandstone is described as relatively erodible (Castro and Bell, 1995).

The morphology of the Klipkraal ramp is similar to the “Class 1” sand ramps described in Table 5-2 (small features, no gully separating the feature from the hillslope and no secondary dunes). However, the extensive gullying present at the Klipkraal ramp has no equivalent in the Namibian sand ramps. The Klipkraal sand ramp sits low in the accommodation space, with room for further accumulation upslope. It has no obvious potential modern aeolian sediment source suggesting it is starved of aeolian sediment supply. These factors combined with the greater precipitation and erodibility of the upslope bedrock (Selby, 1993; Watson et al., 1984) indicate that colluvium is an important source material for the formation of the feature and plays a more formative role than for the Namibian sand ramps. The sands of Unit 1 indicate that aeolian activity is an influential process at Klipkraal but only when conditions are altered to enable a plentiful, mobile sediment supply.

There is some overlap in accumulation ages between the Klipkraal Sands and the sedimentologically and morphologically similar sand ramps reported from the Drakensberg (Telfer et al., 2012) with all features showing periods of rapid accumulation during the last glacial period. At Noordbrabant West deposition has been dated to between 45 ± 3 ka to 7.6 ± 0.6 ka with rapid deposition between ~18-15 ka. This is in general agreement with the dates obtained from the Klipkraal

Sands, both suggesting the majority of the features accumulated during the Last Glacial and both providing evidence for deposition ~15 ka. Deposition at 7.6 ± 0.6 ka is interpreted as colluvial, showing good agreement with the Unit 2 dates (3.5 ± 0.3 ka, 6.0 ± 0.4 ka, 12.4 ± 1.0 ka). Unlike the Klipkraal Sands, Noordbrabant West shows evidence for deposition surrounding the LGM with dates of 26 ± 1.9 , 23 ± 1.5 and 17 ± 1.1 . Telfer et al. (2012, 2014) suggest periglacially weathered material transported by rivers provides the source of aeolian sediments with decreased vegetation cover and increased wind strength enabling transport. Periglacial weathering may play a direct role in colluvium accumulation in Unit 3 of the Klipkraal sand ramp and may also contribute to aeolian accumulation, although the poor sorting of the sediments and the absence of local river channels suggests the contribution is less significant than for the Drakensberg sand ramps.

6.5.4 IMPLICATIONS FOR LATE QUATERNARY PALAEOENVIRONMENTAL RECONSTRUCTION

Ambiguity surrounding the origin of the Klipkraal Sands means that accumulation cannot be interpreted as evidence of increased aeolian activity in the Karoo region and therefore previous interpretations of reduced vegetation cover and strong anticyclonic conditions (Thomas et al., 2002) cannot be confirmed. Additionally, the paucity, high spatial variability, and low resolution of the available late Quaternary records from the region mean processes cannot be inferred from known environmental conditions. The findings of Chapter 5 suggest that periods of regional sediment availability can be identified from the primarily aeolian

Namibian sand ramp record, but only if environmental controls are known and multiple sand ramps are analysed in unison. Therefore, further work is needed, both to constrain the environmental significance and expand the number of analysed features, before sand ramps can be reliably used as a palaeoenvironmental proxy in the semi-arid region of southern Africa.

6.6 CONCLUSION

Re-analysis of the Klipkraal sand ramp feature with new chronometric and sedimentological data has necessitated a revision to the published palaeoenvironmental interpretation and accumulation history of the feature. Previous interpretations proposed a primarily aeolian accumulation history, between 22-11 ka (Marker and Holmes, 1993; Thomas et al., 2002). The improved chronology from this study indicates deposition occurred episodically within the ~100- 0.14 ka period. Additionally, grain-size analysis from this study suggests the material is finer and more poorly sorted than originally suggested by Marker and Holmes (1993). This suggests a close affinity between the ramp sediments and the local Clarens Formation sandstone which crowns the escarpment against which the ramp has accumulated. Comparison with the sediments of the Mastocheni Formation colluvium (Botha, 1996; Dardis, 1990; Temme et al., 2008) and sand ramp sediments from the Drakensberg (Telfer et al., 2012, 2014) are inconclusive. Site sedimentology suggests that deposition is due to both colluvial and aeolian processes. The latter is primarily responsible for

formation of the surface unit which has a direct relationship to enhanced sediment supply as a result of human disturbance during the 19th and 20th centuries.

This analysis highlights the difficulties that can arise in interpreting sand ramp sediments, particularly in marginal environments such as the semi-arid Karoo region. While general principles of ramp development have been established (Chapter 5), it is necessary to consider the local context and establish the local controlling factors of ramp features before they can be utilised as palaeoenvironmental archives.

Chapter 7: Fluctuating fluvial activity in southern Namibia during the late Quaternary: evidence from the sand ramp record.

Based on the understanding of controls on sand ramps discussed in Chapter 5, this paper uses the ten dated Namibian sand ramps to reconstruct palaeoenvironmental conditions in southern Namibia over the Late Quaternary. This provides a case study of the utility of sand ramps as palaeoenvironmental archives.

7.1 ABSTRACT (FROM PAPER SUBMITTED TO 'QUATERNARY SCIENCE REVIEWS' –
AUTHORS: A.L.K. ROWELL, D.S.G. THOMAS AND R.M. BAILEY)

Sand ramps have the potential to provide palaeoenvironmental information in dryland regions where proxy records are typically scarce. However, few investigations have utilised sand ramp sedimentary records in Quaternary research. This paper presents a regional palaeoenvironmental synthesis for southern Namibia based on the analysis of ten sand ramps using 53 OSL dates. Individual sand ramps provide a local record of environmental changes modulated by the availability of accommodation space. Coeval activity across multiple ramps provides a record of regional sediment dynamics.

Aeolian sediments within the sand ramps are most likely sourced from ephemeral river channels. These rivers are fed by summer rainfall whilst most sand ramp accumulation is linked to easterly winter winds. Therefore, periods of peak sand ramp accumulation most likely reflect periods of increased fluvial activity and seasonal moisture availability. Significant activity occurs at ~21-12 ka (with peaks at 21-18.5 ka and 14.5-12 ka), ~8.5-7.5 ka and ~2 ka. Some activity is also indicated 85-65 ka and 45-35 ka. Activity at ~14.5-12 ka is coeval with a climatic event recorded in multiple southern African records and possibly correlated to the Younger Dryas (YD) or Antarctic Cold Reversal (ACR). Activity at ~2 ka ~21-18.5 ka, 45-35 ka and 70-65 ka is concurrent with insolation maxima. These results indicate that sand ramps can provide records of late Quaternary environmental change and suggest the Last Glacial Maximum (LGM) and Antarctic Cold

Reversal (ACR) and/or Younger Dryas (YD) had a widespread geomorphic impact in Namibia

7.2 INTRODUCTION

Climatically, southern Namibia is situated at the confluence of sub-tropical climate systems to the north and east, the mid-latitude temperate systems to the south, and, the cold coastal current-driven hyper-aridity of the Namib Desert to the west. This region therefore represents a significant location for studying the behaviour and changing influences of these systems over the Late Quaternary. Marine-based evidence suggests significant shifts in oceanic and atmospheric conditions over the last Glacial-Interglacial cycle (e.g. Collins et al., 2011, 2014; Dupont et al., 2011; Farmer, 2005; Kohfeld et al., 2013; Little et al., 1997; Schefuß et al., 2011). Details of the terrestrial impacts of these changes is however limited by a paucity of available records and difficulties in reconciling those records that do exist (Chase and Meadows, 2007; Gasse et al., 2008).

Due to the arid and typically seasonal climate, drylands generally preserve few organic proxies for palaeoenvironmental reconstructions (Chase and Meadows, 2007). Geoproxy records, based on sedimentary and landform data, are derived from regionally widespread features, but can be difficult to interpret (see Chase, 2009; Munyikwa, 2005; Stone and Thomas, 2008; Thomas and Burrough, 2012). Until recently, Namibian drylands have proffered few Quaternary climate proxy options, with some records obtained from spatially isolated speleothems (e.g. Brook et al., 1999; Geyh and Heine, 2014; Heine and Geyh, 1984) and fluvial

sedimentary records (see Stone and Thomas, 2013 for an overview). Unlike other parts of southern Africa, the Namibian desert aeolian record has been little investigated (Livingstone et al., 2010, 2014), possibly due to its antiquity (Vermeesch et al., 2010), complex dune evolution (Bristow et al., 2005, 2007) and inaccessibility. Therefore, interpreting clear climate signals for southern Namibia over the Late Quaternary has been problematic. Recently, hyrax middens from escarpment locations have provided a breakthrough for palaeoclimatic studies in the region, providing high resolution data based on pollen (Lim et al., 2016; Scott et al., 2004) and geochemical analyses (Chase, 2009; Chase et al., 2010, 2012). Nevertheless, the spatial and temporal coverage of these records is still relatively low and the interpretation of the records complex (Chase et al., 2012)

In this paper we investigate the palaeoclimatic potential of sedimentary records preserved in sand ramps from the eastern margin of the Namib Desert. Sand ramps are topographically-controlled aeolian forms which sit at the interface between aeolian and fluvial systems (Chapter 5). Formed of aeolian, colluvial and fluvial sediments and stabilised by the adjacent hillslope, these currently under-investigated features are common landforms in the region and have the potential to generate geoproxy data that complements other emerging Quaternary records. This paper presents a palaeoenvironmental reconstruction derived from ten sand ramps. A detailed discussion of geomorphic processes and controls operating on these ramps is provided in Chapter 5. These principles provide the foundation for palaeoenvironmental reconstruction.

7.2.1 REGIONAL OVERVIEW

Southern Namibia is currently hyper-arid to arid with a strong WSW-ENE rainfall gradient. At the coast, rainfall is inhibited by the upwelling of the cold Benguela current and quasi-stationary, South Atlantic Anticyclone (SAA) (Tyson and Preston-Whyte, 2000). Mean Annual Rainfall (MAR) is <20 mm making fog the primary moisture source along the coast (Eckardt et al., 2013; Lancaster et al., 1984). Inland, towards the Great Escarpment, MAR reaches 200 mm and precipitation is predominantly sourced from tropically derived summer rainfall associated with the Congo Air Boundary, Tropical-Temperate Troughs (TTTs) and the Angola Low (Eckardt et al., 2013). Precipitation is characterized by high inter- and intra- annual variability (Lancaster et al., 1984; Nicholson, 2000). Winter rainfall occurs occasionally as a result of temperate cold fronts associated with the mid-latitude westerlies (Eckardt et al., 2013). Annual wind regimes are bimodal. Prevailing winds from the S-SW dominate for most of the year reaching maximum strength in early Austral summer (September-November) (Lancaster, 1985). E to NE katabatic, mountain-to-plain winds are frequent in Austral winter (June-August) (Tyson and Seely, 1980) and strong easterly 'Berg' winds occur occasionally in the west when regional pressure gradients are normal to the coast (Lancaster et al., 1984; Lancaster, 1985). Twelve major and numerous minor ephemeral river channels dissect the Namib desert, flowing only in periods of exceptionally high rainfall (Jacobson and Jacobson, 2013). The sand ramps in this study are found between 23°-28°S and 15°-19°E reflecting a north-south transect with little zonal variation (Figure 7-1).

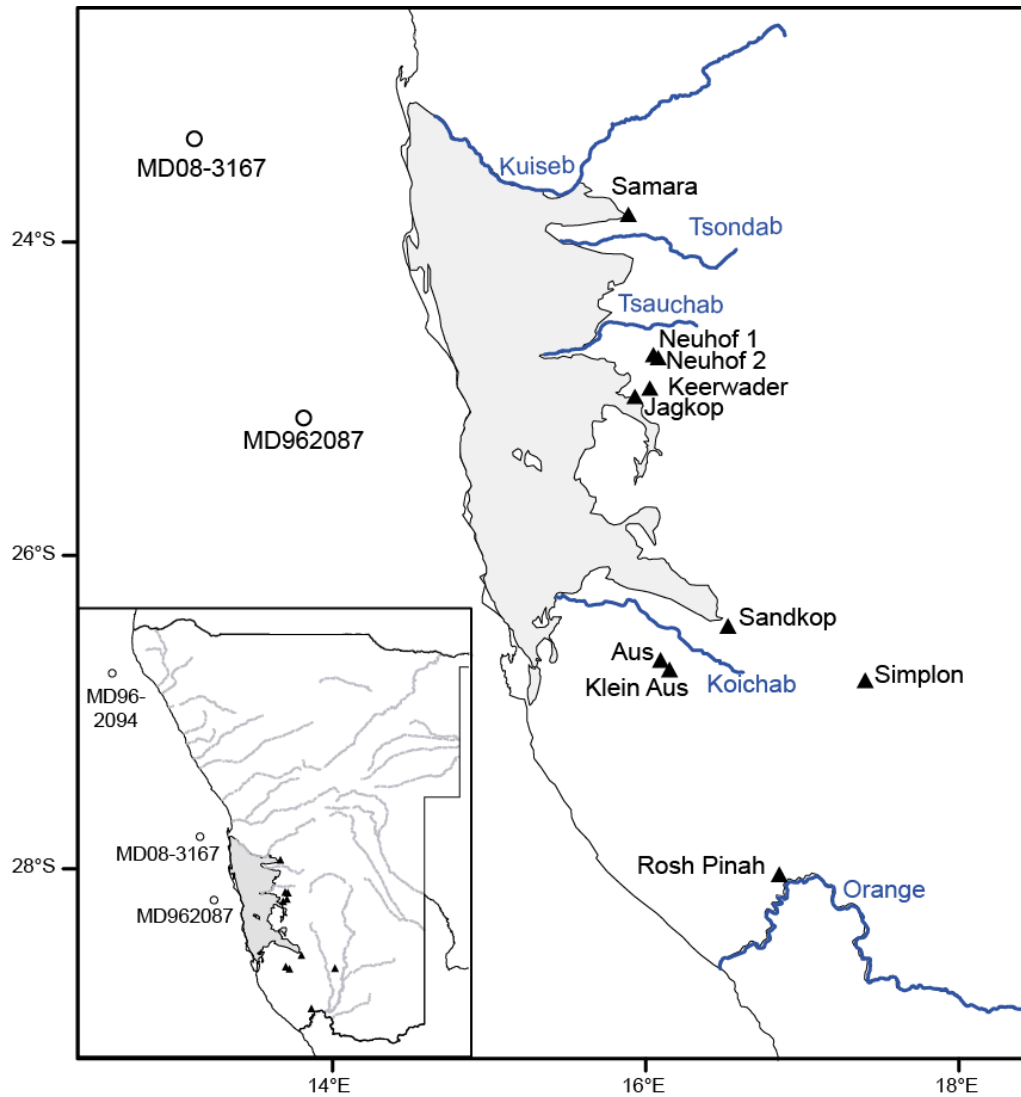


Figure 7-1. Locations of studied sand ramps and sites discussed in the text. Black triangles represent sand ramps.

Details of sand ramp identification, site selection and sampling, and the laboratory methods used to generate the results presented in this chapter are described in section 3.2 and 3.3

7.3 RESULTS

7.3.1 SEDIMENTS

Figure 5-5 provides a schematic representation of the sediments and ages of the individual sand ramps. Detailed, individual sample, sediment properties can be found in Table 5-3. Sediments of sand ramp bodies are primarily interpreted as aeolian but with clear evidence of colluvial deposition. Aeolian accumulation of both the main sand ramp bodies and the overlaying dunes are likely to reflect the same environmental forcing, with sediment availability suggested as the primary regional control on accumulation (Chapter 5). The sediments and chronologies of individual sand ramps are complex, forced by local and regional conditions (Chapter 5). The formation of duricrusts is associated with medium to long term accumulation hiatuses (13->70 ka). Stone layers are not necessarily associated with significant accumulation hiatuses and unconsolidated sediments show no clear relationship with chronology.

7.3.2 OSL AGES

OSL ages are displayed in Figure 5-5, Figure 5-6 and Figure 7-2. Details of luminescence properties can be found in Table 5-4. The dates indicate periodic accumulation over the past >100 ka. Clusters of dates exist at ~2-4 ka, ~6.5-7.5 ka, ~11-21 ka (with sub-clusters at ~12-13.5 ka, and ~18.5-20.5 ka), ~35-45 ka, and ~65-85 ka although ~65-85 ka cluster is towards the limits of OSL dating. Significant accumulation at individual sites is shown at 2-3 ka at Keerwader, 12.5-13.5 ka at Aus and Klein Aus and 78-80 ka at Aus. There is a strong correlation

between the dates obtained from Klein Aus and the dunes overlaying Aus, 7km away (Figure 7-2).

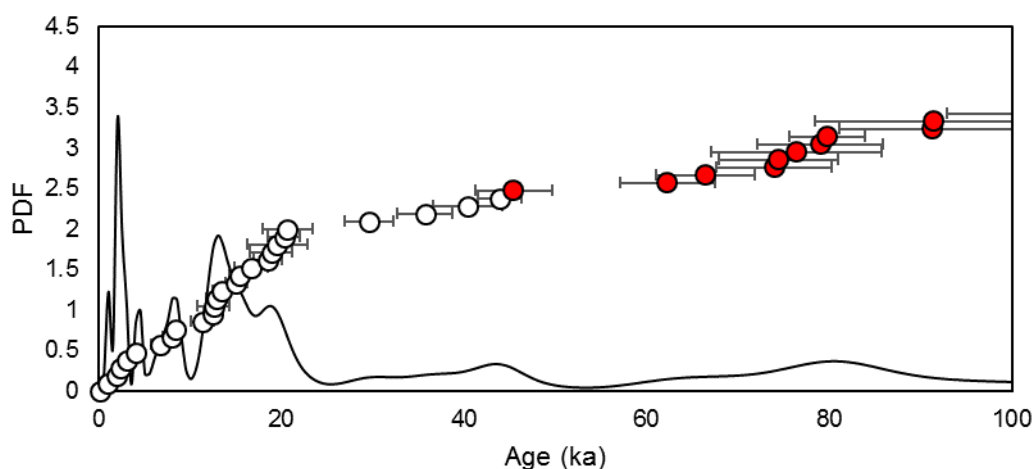


Figure 7-2. OSL ages and associated errors from sand ramp sediments (Circles). Red circles indicate dates with D_e values >200 Gy. These dates are towards the limits of reliable OSL dating. Ages >100 ka are not displayed and, due to dating uncertainties are not discussed in detail in this paper. Black line indicates the cumulative probability density function of these ages (bin width = 500 years).

7.4 DISCUSSION

7.4.1 LOCAL ENVIRONMENTAL RECORD

The complex sedimentary record of individual sand ramps provides information on local processes. At Samara, the most northerly sand ramp, ages bracketing calcrete formation display a ≥ 70 ka hiatus (20.6 ± 2.7 ka to > 90 ka). This suggests a period of stability associated with seasonal rainfall and no sediment input (Wright, 2007) during this time. However, the possibility of unpreserved accumulation and subsequent erosion means the duration or timing of stability cannot be known for certain. Samples from below surface calcrete layers at Jagkop and Aus also date to >90 ka, possibly reflecting stability associated with a

lack of available accommodation space (i.e. sediment is unable to accumulate regardless of environmental conditions).

Dates bracketing calcrete nodules at Keerwader suggest formation between 19.5 ± 3.3 ka and 3.0 ± 0.2 ka (Figure 5-5) although high overdispersion impinges the accuracy and precision of the former date (See section 8.3.3 for discussion). Coevally, large angular clasts indicate colluvial activity at Neuhoof-1 (24 km north) at some time between 16.6 ± 1.8 ka and 11.2 ± 1.2 . Meanwhile aeolian deposition is recorded at Aus, Klein Aus and Rosh Pinah (>200 km south). The similarity between Klein Aus and the dunes overlaying Aus (7 km NW) indicates a strong local environmental signal at 15.4 ± 1.1 to 12.5 ± 1.8 ka, 45.4 ± 4.2 to 35.7 ± 3.0 ka and ~80-70 ka (Figure 5-5). Minimal similarity is found between the dated records of Neuhoof-1 and Neuhoof-2, 4km apart (Figure 7-1, Figure 5-5). This most likely reflects an absence of accommodation space or reworking at Neuhoof-2. Samples NAM14/8/1 (1.0 ± 0.3 ka) and NAM14/8/2 (45.3 ± 2.4 ka) are separated by 0.65 m of sediment indicating the hiatus is not purely a product of under sampling. NAM14/8/1 is extremely overdispersed (133%) suggesting considerable sediment mixing (see section 8.3.3 for discussion).

Accumulation at individual sand ramps is dependent on sediment supply but is also modulated by a number of local factors. Accumulation can only occur if the accommodation space is available. Additionally, before gully formation, sediments may be eroded or reworked by colluvial processes. Some aeolian erosion may occur after gully formation and aeolian reworking is possible

throughout the history of accumulation (see section 5.5.3). Therefore, due to the large number of local variables, individual sand ramps are unable to provide information on regional conditions when studied in isolation. However, low resolution regional patterns can emerge when multiple sand ramps are examined in unison.

7.4.2 INTERPRETING OF OSL AGES FROM MULTIPLE SAND RAMPS

When multiple sand ramps are analysed simultaneously noise associated with local factors (e.g. accommodation space and reworking) becomes less significant and periods of common activity can be identified and analysed. As aeolian sand ramp accumulation is controlled by the interplay between accommodation space and sediment availability (see Chapter 5 for details) synchronous activity across multiple sand ramps is indicative of regional sediment dynamics. Ephemeral river channels in the vicinity of/adjacent to sand ramps are suggested to be the primary sediment source for all dated sand ramps except at Jagkop (discussed below) (see section 5.5.4 for details). Therefore, periods of sand ramp activity across multiple sites are likely to represent regional fluvial system dynamics. As wind is most likely to entrain dry sediments (Hotta et al., 1984), sand ramp accumulation may be associated with flash flooding, increased seasonal rainfall or the end phases of periods of increased moisture availability. The distribution of sand ramps (see section 5.5.1) suggest that strong easterly winter ('Berg') winds are responsible for the majority of sand ramp accumulation in southern Namibia. Collins et al. (2014) suggest that Namibian rainfall remained highly seasonal over the past 140 ka with periods of increased precipitation a result of increased

summer rainfall. This combined with the present inter- and intra-annual and spatial variability in rainfall (Nicholson, 2000; Nicholson et al., 2012a, 2012b) and the variation observed in the dated fluvial record (Stone and Thomas, 2013), suggests that rivers remained seasonal and ephemeral. Therefore, at the resolution of OSL dating, periods of sand ramp activity are most likely to be coeval with periods of increased fluvial activity with winter winds transporting the fluvial derived sediments deposited during summer.

Sediments at Jagkop accumulated prior to 100 ka and are most likely sourced from the Namib Sand Sea although the presence of ancient palaeochannels, masked or absent today, cannot be dismissed. Dated dunes overlying Jagkop are <1.5 ka with ages possibly reflecting reworking of surface sediments. However, all samples in the dataset with <10 accepted aliquots (including the Jagkop dunes) were assumed too imprecise for regional analysis. All of these samples are <2 ka and their inclusion in the dataset indicates near continuous activity during this time, most likely as a result of reworking (section 5.5.3). Additionally, due to age uncertainties (see Chapter 4) samples with D_e estimations >200 Gy are not discussed in detail. Therefore, the Jagkop dates do not influence overall conclusions despite their differing sediment source.

7.4.3 REGIONAL ENVIRONMENTAL RECORD

When amalgamating chronometric datasets on a regional scale the method of data analysis and display has a large impact on interpretation (Stone and Thomas, 2008; Thomas and Burrough, 2012). Frequency based methods such as

histograms, probability density functions (PDFs) and moving averages have frequently been applied to chronometric aeolian dune records (Chase and Thomas, 2007; Hesse, 2016; Stokes et al., 1997a; Telfer and Thomas, 2007). However, these methods can be easily biased by sampling strategy with increased frequency of dates reflecting the number of times a unit has been sampled (in various locations) rather than the environmental significance of the unit (Stone and Thomas, 2008). Bailey and Thomas (2014) and Leighton et al. (2013) suggest that analysis of accumulation rate is more representative of periods of increased aeolian activity than age frequency is, particularly as the magnitudes of events can be examined. When applied to the sand ramp record analysis of accumulation rate may also bolster the significance of dates associated with increased sediment supply as these are more likely to produce periods of increased accumulation than reworked or bioturbated sediments.

The Accumulation Rate Variability (ARV) method devised by Bailey and Thomas (2014) was experimentally applied to the sand ramp record. The results show good agreement with PDF analysis (Figure 7-3). However, no peak is present in the experimental ARV analysis at 45-35 ka. As the ARV method relates paired ages to the thickness of accumulated sediment, this is due to low accumulation rates at Klein Aus, Aus and Neuhof-2. Additionally, fully paired data is not available at Neuhof-1 during this time, meaning accumulation rates cannot be fully calculated. Accumulation 14-11.5 ka assumes significance in ARV compared to PDF data, because recognisable sediment accumulation is recorded in this period. The similarity between ARV and PDF analysis may indicate that the

sand ramp record is not biased by sampling frequency (Bailey and Thomas, 2014) although the relatively small dataset precludes definitive conclusions.

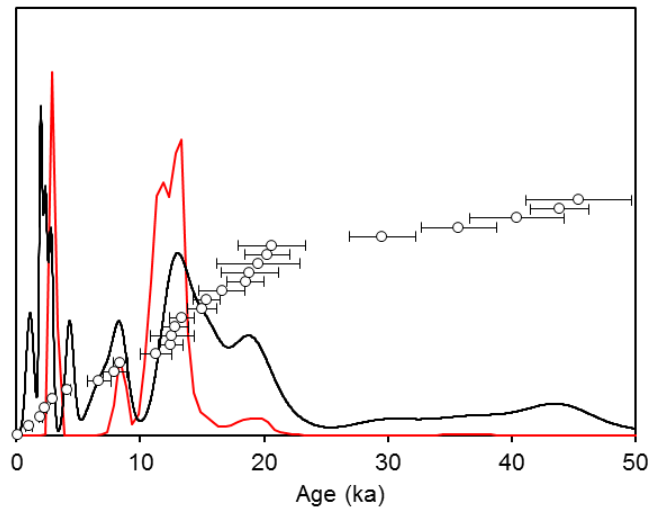


Figure 7-3. Comparison between ARV (red line) and PDF (black line) analysis over the past 50 ka. Individual dates (circles) are shown for reference.

ARV analysis has been developed primarily for the analysis of open dunefield records assuming a unitary sedimentary system that is impacted upon by uniform regional forcing factors. The system is presumed to be entirely aeolian wherein periods of activity are associated with both accumulation and erosion (Bailey and Thomas, 2014). Sand ramps are more stable than open-system dunes as, once emplaced, aeolian sediments are less likely to experience subsequent aeolian reactivation (Chase, 2009; Lancaster and Tchkarian, 1996). Therefore, hiatuses within individual sand ramp deposits are more likely to represent a cessation of sediment supply/availability or no available accommodation space than erosion (Chapter 5). Additionally, a number of local factors operate on sand ramps meaning a single regional system cannot be assumed in the manner intended in ARV analysis. Therefore, as no sampling bias is apparent from ARV analysis, PDF displayed alongside individual ages provides the most straightforward method of regional analysis for this dataset. PDF peaks are influenced by

precision, therefore their significance needs to be scaled by age (Galbraith and Roberts, 2012). Regional peaks in dated sand ramp activity are found ~2 ka, 8.5-7.5 ka and 21-11 ka, with peaks at 14.5-12 ka and 21.5-18 ka and 45-35 ka and 85-75 ka (Figure 7-2). Other peaks during the Holocene are associated with single dates and therefore assume less significance. The peak at ~85-70 ka occurs as samples are approaching OSL saturation and therefore may be biased by this.

The peak at 14.5-12 ka is strongly influenced by activity in the Aus region but also incorporates dates from Rosh Pinah and Neuhof-1 suggesting it is not solely reflecting a local signal. Accumulation at 21-18.5 ka and 45-35 ka is more regionally widespread. However, the relatively small dataset means, at present, only preliminary palaeoenvironmental conclusions can be drawn.

7.4.4 COMPARISON TO REGIONAL RECORDS

Fluvial activity in western Namibia presently consists of a suite of ephemeral rivers bracketed by the perennial Kunene and Orange Rivers to the north and south respectively. To the north of the Namib Sand Sea, river catchments are large and ephemeral rivers periodically discharge into the Atlantic Ocean with a return period of ~8 years (Lancaster et al., 2000). Towards the south, the Tsondab, Tsauchab and Koichab penetrate up to 50 km into the Namib Sand Sea before ending in terminal playas. Outcrops of fluvial silts and gravels, and lacustrine deposits indicate periods of increased river discharge over the Quaternary (Lancaster et al., 2000). Overall, OSL dated deposits from the main ephemeral rivers suggest a regionally varied discharge pattern over the late

Quaternary (Brook et al., 2006; Eitel et al., 2006; Srivastava et al., 2004, 2005, 2006; Stone et al., 2010). This is likely due to the high spatial and temporal variability of the summer rainfall which feeds these rivers (Eitel et al., 2006; Stone and Thomas, 2013).

Of the major ephemeral rivers, the Quaternary behaviour of the Tsondab, Tsauchab and Koichab is the most geographically relevant to the sand ramp record. These rivers are sourced in the Great Escarpment with changes in hydrology reflecting changes in summer rainfall (Brook et al., 2006; Stone et al., 2010; Stone and Thomas, 2013). For the Tsondab, evidence of increased flow is dated to MIS 5 (centred on 110.2 ± 3.6 ka), 16.9 ± 0.9 to 12.7 ± 0.1 ka and ~ 12.5 ka and 11.5 ± 0.5 to 10.5 ± 0.5 ka (Stone et al., 2010). This shows good agreement with the sand ramp record which indicates a peak in activity at 14.5-12 ka (Figure 7-4).

For the Tsauchab river, increased discharge is suggested at and before 24.6 ± 2.1 ka and between ~ 7 -9 ka with some additional activity indicated at 4.3 ± 0.5 ka. Minor wet phases are suggested 0.9-0.3 ka (Brook et al., 2006). These results do not correspond to the Tsondab deposits (Stone et al., 2010), possibly suggesting high regional variability or that evidence for corresponding periods of activity has been eroded or remains undated. Activity at ~ 7 -9 ka can be linked to some sand ramp activity at 8.5-7.5 ka (Figure 7-4). At the northern margin of the Namib Sand Sea Deposits of the Kuiseb river indicate two periods of increased flow centred on ~ 15 ka and ~ 5 ka (Srivastava et al., 2006) showing some agreement to the sand ramp record.

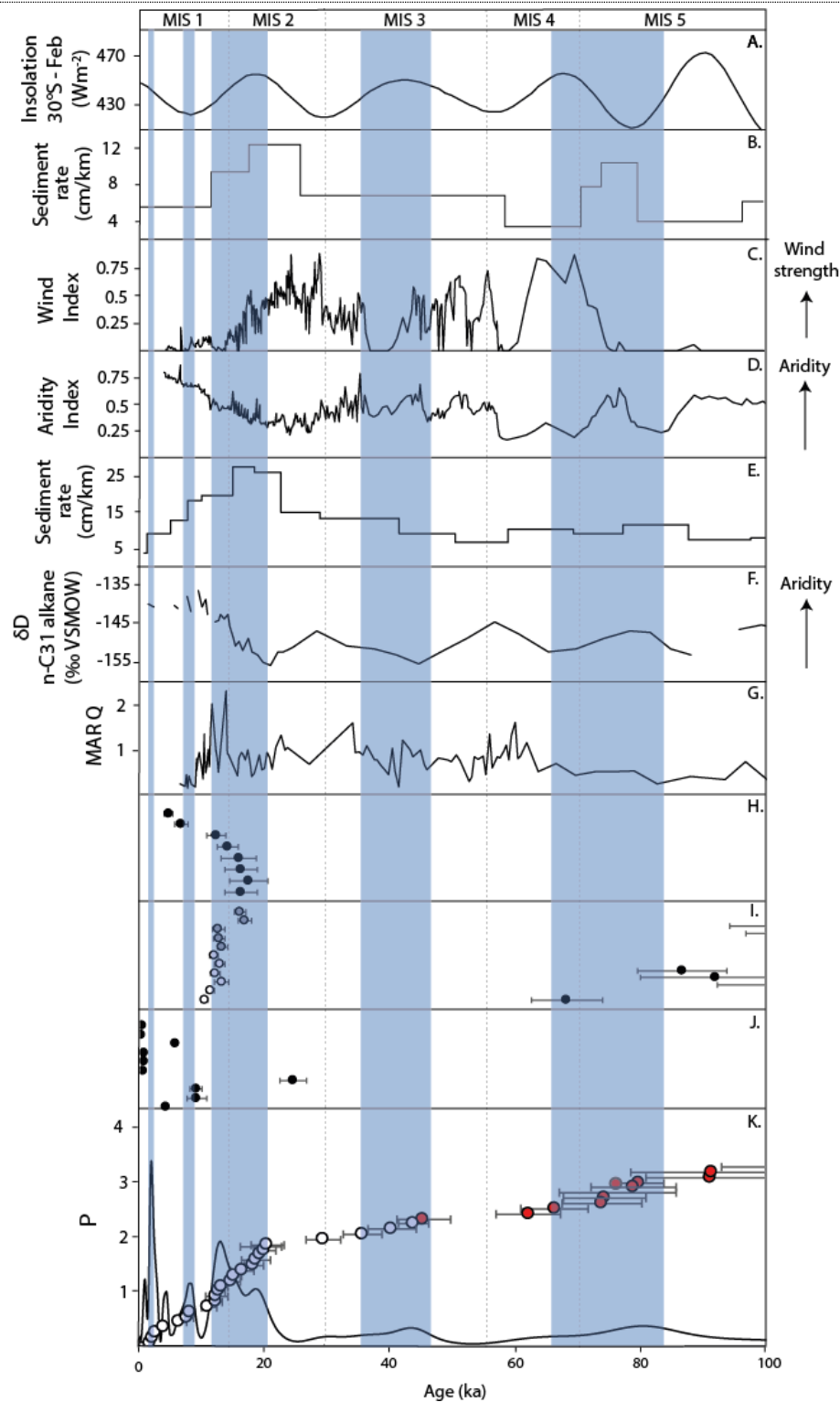


Figure 7-4. Comparison of the sand ramp record with selected Namibian palaeoclimate records. A). February insolation at 30°S (Berger and Loutre, 1991). B). Sediment accumulation rate for core MD962094. Sediments are assumed to be predominantly terrestrially derived (Stuut et al., 2002). C) and D). Wind and aridity indexes for core MD962094 derived from grain-size distributions of terrestrially derived sediments. E). Sediment accumulation rate for core MD08-3167. Sediments are assumed to be predominantly terrestrially derived (Collins et al., 2014). F). terrestrial aridity record from δD from MD08-3167 corrected for ice volume (Collins et al., 2014). G). Mass accumulation rate of quartz from MD962087 interpreted as a proxy for aridity (Pichevin et al., 2005). H). OSL dates from periods of increased flow in the Kuiseb River (Srivastava et al., 2006). I). OSL dates from evidence of increased Tsondab River flow (Stone et al., 2010). J). OSL dates from evidence of increased Tsauchab River flow (Brook et al., 2006). K). Distribution of OSL dates from sand ramps in south western Namibia.

Off shore evidence

Terrigenous sediments in marine cores off the west coast of Namibia are primarily comprised of atmospheric dust sourced from the ephemeral river valleys of western Namibia and Etosha Pan and transported by easterly Berg winds (Collins et al., 2014; Eckardt and Kuring, 2005; Stuut et al., 2002; Tlhalerwa et al., 2005; Vickery et al., 2013; Vickery and Eckardt, 2013). Down-core variability in dust concentrations reflects the changing efficiency of dust emission from these sources. As dust events from both fluvial and pan sources have been linked to ephemeral flooding (Bryant et al., 2007; Vickery and Eckardt, 2013), dust accumulation is likely associated with fluvial and lacustrine activity (Collins et al., 2014). Rates of dust accumulation in cores MD08-3167 (Collins et al., 2014) and MD962094 (Stuut et al., 2002) show some agreement with the sand ramp record (Figure 7-4) with peaks occurring at ~25-10 ka and minimums at 60-50 ka and 70-60 ka respectively. An accumulation peak at 80-70 ka, concurrent with sand ramp activity, is observed only in core MD962094. Peaks in terrestrially derived MAR Q in core MD962087 occur ~ 14-10 ka, ~ 22.5 ka, ~34 ka and ~ 60 ka (Picchevin et al., 2005) and are in reasonable agreement with the sand ramp record (Figure 7-4).

Overall, evidence for palaeoenvironmental change in Namibia shows considerable regional variability (Brook et al., 2007, 2011, Chase et al., 2009, 2010; Geyh and Heine, 2014; Gil-Romera et al., 2006, 2007; Hipondoka et al., 2014; Lancaster, 2002; Lim et al., 2016; Scott et al., 2004; Stone and Thomas, 2013). This is not unexpected given the current spatial and temporal rainfall

variability (Nicholson, 2000; Nicholson et al., 2012a, 2012b) and the differing nature of the available proxies. Nevertheless, records of fluvial activity show good agreement with the sand ramp record (Figure 7-4), supporting the conclusion that the sand ramp record is controlled by sediment supply via fluvial activity.

7.4.5 PALAEOENVIRONMENTAL SIGNIFICANCE

Periods of sand ramp accumulation coincide with distinct periods of climatic change recorded in other records from southern Africa (e.g. Chase et al., 2013, 2015; Chevalier and Chase, 2015; Holmgren et al., 2003; Kirsten et al., 2007; Lim et al., 2016; Norström et al., 2009, 2014) confirming the sensitivity and utility of the regional-scale sand ramp record as a palaeoenvironmental proxy.

Sand ramp accumulation occurred during the LGM and concurrent with a widespread regional hydroclimatic event which occurred at ~15-11 ka. Accumulation at ~2, ~20-18 ka, ~45-40 ka and ~70-65 ka occurs during insolation maxima showing reasonable agreement with the findings of Collins et al., (2014) (Figure 7-4). However, this relationship breaks down for clusters of ages at ~80-75 ka, ~14.5-12 ka and during the Early Holocene. Samples from ~80-75 ka are associated with large errors and are approaching saturation so some inaccuracies may exist whilst Holocene samples may be reflecting the preservation of smaller events which are lost in the low resolution of the older record (Bailey and Thomas, 2014). The significance of accumulation ~14.5-12 ka is discussed below.

During MIS 2 arid conditions are recorded across much of southern Africa (Chase and Meadows, 2007; Chevalier and Chase, 2015; Gasse et al., 2008 and references therein). However, there is evidence for fluvial activity in Namibia (Srivastava et al., 2004, 2005, 2006; Stone and Thomas, 2013) high lake levels at Etosha (Brook et al., 2007, 2011; Hipondoka et al., 2014) and relatively moister conditions are indicated from west coast marine cores (Collins et al., 2014; Stuut et al., 2002). This supports the moderate level of accumulation observed in the sand ramp record. Increases in moisture are possibly driven by a southward displacement of tropical moisture sources as a result of expanded northern hemisphere ice sheets coupled with southern hemisphere insolation maxima (Chiang and Bitz, 2005; Collins et al., 2014). Increases in winter rainfall due to the northward displacement of the westerly windbelt (Cockcroft et al., 1987; Van Zinderen Bakker, 1976) may also contribute to increased west coast moisture (Chase and Meadows, 2007; Shi et al., 2000) but evidence for this is limited (Gasse et al., 2008; Lim et al., 2016; Scott et al., 2004).

A Last Glacial-Interglacial Transition event (~15-11 ka, dependent on record) is observed in multiple southern African records suggesting drier (Brook et al., 2010; Chevalier and Chase, 2015; Green et al., 2015; Holmgren et al., 1994, 2003; Metwally et al., 2014; Scott, 2016; Truc et al., 2013), wetter (Kristen et al., 2007, 2010; Lim et al., 2016; Neumann et al., 2014) or rapidly fluctuating (Norström et al., 2014) conditions. This is potentially associated with the Antarctic Cold Reversal (ACR) or the Younger Dryas (YD) but for several records, including the sand ramp record, chronological resolution is not high enough to

differentiate. Mechanisms for hydrological change during this period are contested (e.g. Brook et al., 2010; Chase et al., 2015; Green et al., 2015; Holmgren et al., 2003; Truc et al., 2013). The increases in fluvial activity inferred from the sand ramp record may be driven by one or more of the following mechanisms: a southward displacement of tropical moisture responding to northern hemisphere cooling (Green et al., 2015; Johnson et al., 2002; Schefuß et al., 2011), an increase in winter precipitation (Chase and Meadows, 2007; Green et al., 2015) or increases in local convection due to zonal changes in convection (Burrough and Thomas, 2013; Lim et al., 2016; Singarayer and Burrough, 2015).

Patterns of sand ramp accumulation suggest fluctuating fluvial activity during the Holocene with peak activity at ~2 ka and 8.5-7.5 ka. This coincides with complex regional hydrological patterns observed for central southern Africa (Burrough and Thomas, 2013). Climatic fluctuations are suggested at ~8 ka in several records (Chase et al., 2015b; Chevalier and Chase, 2015; Scott et al., 2012) demonstrating the sensitivity of the sand ramp record.

Overall, periods of sand ramp activity show a coeval relationship with periods of insolation maxima and with widespread regional hydroclimatic events. This is in good agreement with Collins et al. (2014) and suggests that increases in summer rain due to direct insolation forcing exert a primary control over regional fluvial activity in southern Namibia. However, the presence of activity at 14.5-12 ka and

~8 ka suggests that this is not the only factor influencing the regional hydroclimate.

7.5 CONCLUSION

This study presents a regional application of sand ramps as palaeoenvironmental archives. Ten sand ramps were studied in detail using 53 OSL dates. Periods of increased sand ramp accumulation are interpreted to represent increased fluvial activity as supported by available fluvial records and terrestrially derived marine cores (Brook et al., 2006; Collins et al., 2014; Stone et al., 2010; Stuut et al., 2002).

Peak periods of fluvial activity are indicated at ~21-12 ka (with peaks at 21-18.5 ka and 14.5-12 ka), ~8.5-7.5 ka and ~2 ka. Some activity is also indicated 85-65 ka and 45-35 ka. Sand ramp accumulation at 14.5-12 ka is concurrent with climatic fluctuations observed in other regional records (Brook et al., 2010; Chevalier and Chase, 2015; Green et al., 2015; Holmgren et al., 1994, 2003, Kristen et al., 2007, 2010; Lim et al., 2016; Metwally et al., 2014; Neumann et al., 2014; Norström et al., 2014; Scott, 2016; Truc et al., 2013) and may be associated with the Antarctic Cold Reversal (ACR) or the Younger Dryas (YD). Activity at ~2 ka, ~21-18.5 ka, 45-35 ka and 70-65 ka are concurrent with insolation maxima although the relationship to the sand ramp record is not exact. These results confirm the utility and sensitivity of the sand ramp record as a palaeoenvironmental archive and suggest the Last Glacial Maximum (LGM) and

Antarctic Cold Reversal (ACR) and/or Younger Dryas (YD) had a widespread geomorphic impact in Namibia.

Part 3: Implications

Chapter 8: Discussion

The work presented in this thesis explores the utility of sand ramps as palaeoenvironmental archives with the aim of answering two research questions:

1. What are the main controls on sand ramp formation in southern Africa?
2. What does the sand ramp record tell us about Late Quaternary palaeoenvironments in southern Africa?

This discussion chapter draws together the findings of the four papers and discusses the significance and limitations of the investigations, as well as potential avenues for further research. The research questions are addressed individually prior to a discussion on the reliability of the age estimates which underpin all conclusions.

8.1 WHAT ARE THE MAIN CONTROLS ON SAND RAMP FORMATION IN SOUTHERN AFRICA?

Chapters 5 and 6 explored the main controls operating on the sand ramp record in southern Africa. Analysis of Google Earth™ imagery in Chapter 5 showed sand ramps to be common and regionally widespread in southern Namibia. The distribution and sedimentology of these features indicates that four factors govern sand ramp formation:

1. an adequate supply of sediment with significant aeolian transport potential;
2. an appropriate accommodation space, protected from significant erosion;

3. directionally-persistent winds with sufficient energy for aeolian sand transport;
4. an arid to semi-arid climate but with sufficient seasonal, or longer term, variability to promote differing geomorphic and depositional processes.

Factors 1,3 and 4 are dynamic, with changes over time defining the sand ramp record. At the outset factor 2 is passive; sand ramps cannot form where there is no accommodation space, but as a sand ramp accumulates the accommodation space becomes more dynamic, changing with the size and shape of the sand ramp and exerting control over the morphology and environmental sensitivity of the sand ramp (section 5.5.3). Factors 1 and 2 control local accumulation, deciding where and when sand ramps will form, whilst factors 3 and 4 determine regions conducive to sand ramp formation.

Akin to the findings of Bateman et al. (2012) for the Soldier Mountain sand ramp, the ages and geomorphology of the Namibian sand ramps suggest accumulation is controlled by sediment supply modulated by the availability of accommodation space (section 5.5.4). If accommodation space is available, sediment supply determines the timing of accumulation with both the aeolian component of the main bodies of sand ramps and the overlying dunes appearing to respond to the same climatic forcing (section 5.5.3). For the Klipkraal Sands accommodation space exerts minimal control, suggesting sediment supply is the limiting factor in formation (section 6.5.3). Sediment supply is considered the primary controlling factor for the studied American sand ramps (Ellwein et al., 2015; Lancaster and

Tchakerian, 1996) whilst climate and wind dynamics are suggested to be the dominant controls in the Ladakh Desert, India and Ardakan, Iran respectively (Kumar et al., 2016; Thomas et al., 1997a).

Factors 1-4 are broad and are likely to be regionally transferrable (e.g. Bateman et al., 2012; Kumar et al., 2016; Telfer et al., 2012). Therefore, they could be used to predict areas suitable for further sand ramp research. However, the conditions which influence factors 1-4 (e.g. climate, lithology, proximity to and nature of surrounding geomorphic features) are regionally specific, meaning the ultimate drivers of the timing and nature of sand ramp accumulation are also regionally specific (e.g. Chapter 6). For the Namibian sand ramps aeolian accumulation is linked to increased fluvial activity which is associated with summer rainfall. At Klipkraal accumulation is dominated by colluvial processes although specific controls on sediment availability and movement, other than in the historic period, are currently unclear. In other regions periglacial weathering, aridity and low vegetation cover, or any climatic process which produces sediment, have been suggested as the underlying drivers of sand ramp accumulation (Bateman et al., 2012; Kumar et al., 2016; Lancaster and Tchakerian, 1996; Pease and Tchakerian, 2003; Telfer et al., 2012, 2014). This is analogous to the alluvial fan record wherein response to climatic change is strongly dependent on the regional setting and catchment dynamics. Without strong chronological control and regional understanding, the sedimentology and stratigraphy of fans cannot be used as indicators of climatic change (Dorn, 2009; Harvey, 2011 and references therein).

Therefore, palaeoclimate, through its impact on sediment supply and wind dynamics plays a determining role in sand ramp accumulation. However, the role of palaeoclimate is regionally specific, operating alongside other regional constraints. As such, understanding of the modern climate, lithology and geomorphology of a region alongside good chronological control is needed to establish the full extent of controls operating on sand ramps. This means that palaeoenvironmental reconstructions using the sand ramp record are reliant on considerable groundwork and conclusions on the drivers of accumulation cannot be readily transferred between regions.

8.1.1 LIMITATIONS OF THE INVESTIGATION

Google Earth™ provides the most complete and easily assessable method of surveying large regions for potential sand ramps but complete identification of where sand ramps do and do not form is impossible from satellite imagery alone (see section 3.2.1). Field investigation significantly aids identification but problems of access and scale limit the scope of investigations. Therefore, knowledge of the distribution of sand ramps in southern Africa is incomplete, potentially masking information on regional controls. For example, the Klipkraal Sands, originally identified due to extensive gullying and quarrying (Marker and Holmes, 1993) is unlikely to be an isolated feature in the Karoo. It is possible that sand ramps in this region are only identifiable if heavily gullied (Telfer et al., 2014) or through extensive exploratory fieldwork. Obtaining a better understanding of the distribution of features would significantly improve understanding of the

controlling factors which operate in this region and how they compare to the Namibian sand ramps.

Retrospective identification of geomorphic processes from sediments alone can be problematic, particularly for complex features such as sand ramps. Sediment size, structure and sorting alongside the regional context provide some information but inherited properties and overlap between processes can confuse interpretation (e.g. Feng et al., 2016; Fitzsimmons et al., 2009; Kocurek and Dott, 1981; Sweeney et al., 2013). This is exemplified by the Klipkraal Sands which were previously identified as purely aeolian (Marker and Holmes, 1993; Thomas et al., 2002) but the poor sorting and affinity of the sediment to the regional lithology suggest a significant colluvial influence (section 6.5). Based on the arid environment and resistant metamorphic geology a primarily aeolian depositional history was assumed for the Namibian sand ramps, despite the absence of bedding features within the sediments. However, the sediments are typically poorly sorted with some ramps containing sporadic angular clasts throughout the sediments (Table 5-3). This may indicate a more significant colluvial component than suggested in Chapter 5 and 7 but, due to the importance of aeolian processes in the region, this should not impact significantly on interpretation.

8.1.2 FUTURE WORK AND UNANSWERED QUESTIONS

At present, the small volume of research internationally means there are many unknowns regarding the sand ramp record. The controls suggested above have only been tested for southern Africa. In addition, work on the Namibian sand

ramps focused on controls on aeolian accumulation. Extensive studies from other regions, perhaps across a climatic gradient, may be able to refine the controls on sand ramp formation; firstly by identifying the relative importance of regional variation (e.g. geology, climate, vegetation cover, proximity to geomorphic features) but also by answering questions regarding the significance of the sand ramp continuum (e.g. what are the primary factors governing the proportion of aeolian, fluvial and colluvial sediments and, if present, do sand ramps from the same region but with very different sediment compositions respond to the same forcing factors?).

Individual sand ramps have complex depositional histories (Chapter 5, Bateman et al., 2012; Kumar et al., 2016; Telfer et al., 2012). Sedimentary units can be discontinuous, with significant variation (Chapter 5, Bateman et al., 2012; Bertram, 2003). However, this, and other studies (e.g. Bateman et al., 2012; Kumar et al., 2016; Telfer et al., 2012) have focused on examining a single section or creating a composite stratigraphy from multiple sections. It is therefore unclear how spatially and temporally complex sand ramp deposition is and how representative single section studies are. Multiple cores from individual dunes suggest significant local variation which can completely alter palaeoenvironmental interpretation (e.g. Bristow et al., 2005, 2007; Stone and Thomas, 2008). An extensive study on a single sand ramp may be able to explain depositional processes and assess the scale of sampling needed to understand formation processes.

8.2 WHAT DOES THE SAND RAMP RECORD TELL US ABOUT LATE QUATERNARY PALAEOENVIRONMENTS IN SOUTHERN AFRICA?

Based on the above conclusions the utility of the sand ramp record as a palaeoenvironmental archive is dependent on:

1. Identification of controlling factors in the region.
2. A large enough sample size to differentiate regional and local forcing factors.

In South Africa, the small sample size and ambiguity surrounding the geomorphic origin of the sediments (Chapter 6) preclude definitive palaeoclimatic conclusions from being drawn.

For the Namibian sand ramps, periods of regional sand ramp accumulation are suggested to be associated with periods of regional fluvial activity (section 7.4.2). Clusters of dates and increases in accumulation rate occur at ~2 ka, 8.5-7.5 ka, 21-12 ka with peaks at 14.5-12 ka and 21-18.5 ka, 45-35 ka and 85-65 ka. These results are in reasonable agreement with records of regional fluvial activity (Collins et al., 2014; Srivastava et al., 2006; Stone et al., 2010) supporting the conclusion that fluvial activity is the main driver of sand ramp activity in Namibia. Activity at ~14.5-12 ka coincides with an LGIT event recorded in multiple southern African records whilst activity at ~2 ka ~21-18.5 ka, 45-35 ka and 70-65 ka is concurrent with insolation maxima.

These results are in reasonable agreement with other palaeoenvironmental reconstructions from the region (Collins et al., 2014; Lim et al., 2016; Stone et al., 2010) and beyond (Chase et al., 2013, 2015a; Chevalier and Chase, 2015; Holmgren et al., 2003; Kristen et al., 2007; Norström et al., 2009, 2014). This supports the validity of the sand ramp record and confirms its sensitivity as a proxy for palaeohydrological change in southern Namibia. This indicates that with a sufficient sample size and thorough ground-work, the sand ramp record can provide important palaeoenvironmental information in dryland regions, especially when examined alongside other proxies.

The correspondence between sand ramp activity and insolation maxima, and with a regional event which may be associated with the southward displacement of the ITCZ (Green et al., 2015; Johnson et al., 1997; Schefuß et al., 2011), suggests that ITCZ dynamics are a significant factor influencing regional trends in palaeohydrology over the Late Glacial. This supports conclusions that direct insolation forcing produced increased precipitation in southern Africa (Collins et al., 2014; Partridge et al., 1997) but also indicates that insolation is not the only factor influencing precipitation in southern Namibia. Changes in meridional convection driven by NH cooling (Arbuszewski et al., 2013; Chiang and Bitz, 2005; Mohtadi et al., 2014) or changes in zonal convection (Burrough and Thomas, 2013; Lim et al., 2016; Singarayer and Burrough, 2015) may also play a significant role.

8.2.1 LIMITATIONS OF THE INVESTIGATION

The above conclusions are, necessarily, based on a considerable number of assumptions. Firstly, dates are assumed to represent aeolian sand ramp accumulation rather than reworking. The complex mosaic of depositional processes operating on sand ramps (section 6.5), even within a single sample (Section 5.4.7), suggests at least some OSL ages may represent sediment stabilisation rather than first deposition. The similarity between chronologies from small, Class 1 & 2, sand ramps and the dunes overlaying Class 3 & 4 sand ramps provide further indication of sand ramp reworking (section 5.5.3).

Differentiation between primary deposition, aeolian and/or colluvial/fluvial reworking is difficult. High overdispersion can provide some indication but only if the reworking process causes uneven sediment bleaching or sediment mixing (e.g. movement of the sediment as a block, bioturbation or slumping). Sedimentology can aid the interpretation in some circumstances but an absence of sedimentary structures does not provide proof of reworking (e.g. Bateman et al., 2007). Under some circumstances (e.g. high accumulation rates, rapid stabilisation, or old samples) reworking may be inconsequential, especially when OSL precision is taken into account. In other circumstances the impact on environmental interpretation may be significant. Activity at 14.5-12 ka is associated with high accumulation rates suggesting this period represents primary deposition. However, conclusions are less certain for the other periods.

Secondly, it is assumed that dated sediment is aeolian and that aeolian sand ramp accumulation in Namibia is directly forced by fluvial activity. This assumption is based on the close proximity of many sand ramps to ephemeral river channels and the similarity of heavy mineral assemblages between a sample of the river near Neuhoof-1 and samples of the Neuhoof-1 and Neuhoof-2 sand ramps (section 5.4.8). However, other sediment sources such as direct weathering or dunes may provide some of the sediments.

Thirdly, even if an association with fluvial activity is assumed to be correct, the timing between the fluvial activity and sand ramp accumulation is unknown. The sediments need to become available (i.e. dry sufficiently so the forces of cohesion are lower than the transport capacity of the wind; Kocurek and Lancaster, 1999). As rainfall is suggested to remain seasonal, falling predominantly in the summer, over the past 140 ka (Collins et al., 2014) and easterly winter winds are likely to be responsible for the majority of sand ramp accumulation (section 5.5.1) a roughly coeval relationship between fluvial sediment transport and sand ramp accumulation is suggested. Alternatively, sand ramp accumulation may occur during dry phases succeeding wet phases. However, given the ephemeral nature of regional rivers it is unlikely that sediment availability was suspended throughout the entirety of wet phases, lending support for a coeval relationship.

Fourthly, it is assumed that the number of studied sand ramps is large enough, and the dating resolution high enough, to be representative, and that all land

forms are responding to the same primary controls and regional climatic forcing factors. As interpretation was limited to the last 100 ka (to remove issues associated with older dates – see Chapter 4) and samples with less than ten accepted aliquots were removed (see section 7.4.2), 36 ages were used for palaeoenvironmental reconstruction. Therefore, individual sand ramps can exert a significant influence on clusters of dates. For example, accumulation at 14.5-12 ka is mainly associated with the Aus region, however some accumulation at Neuhof-1 and Rosh Pinah gives this event regional credibility.

Given the current spatial and temporal rainfall variability (e.g. Nicholson, 2000; Nicholson et al., 2012, 2013) and the palaeoenvironmental record from directly dated fluvial deposits (Brook et al., 2006; Eitel et al., 2006; Srivastava et al., 2004, 2005, 2006; Stone et al., 2010) it is unlikely that ephemeral river flooding occurred in unison across the region covered by the sand ramp record. However, at the temporal resolution of the sand ramp record, these variations are likely to be smoothed out such that broad trends associated with prolonged or intense periods of increased moisture are identifiable. At present all sand ramp sites receive between ~50-150 mm MAP (Mendelsohn et al., 2002) suggesting no significant variation in climate forcing between the sites. Nevertheless, the influence of local factors on some sand ramps and rivers may be more significant than for others. Between 14.5-12 ka aeolian accumulation occurs coevally with colluvial deposition and calcrete formation possibly suggesting a different expression of the same climatic forcing or regional variation.

Overall, the good correlation between the sand ramp record and other proxies (e.g. (Collins et al., 2014; Lim et al., 2016; Stone et al., 2010) provides support for the above assumptions. Nevertheless, they highlight the uncertainties that need to be taken into consideration when utilising the sand ramp as a proxy of environmental change. The regional record is of low resolution and as such only very broad trends can be identified. More detailed local records can be obtained from individual sand ramps but local controls and forcing factors need to be taken into consideration before these results can be extrapolated to the wider region. Sand ramps are relatively common features in dryland environments and, in comparison to the dune or fluvial record, sediments remain relatively unaltered after deposition. Therefore, if sand ramp sampling is of high enough resolution, a more complete record of environmental change may be available from the sand ramp record rather than the direct dating of fluvial activity.

8.2.2 FUTURE WORK AND UNANSWERED QUESTIONS

The results in this thesis indicate that multiple sand ramps need to be analysed for regional palaeoenvironmental conclusions to be drawn. However, the optimal number of sand ramps and sampling resolution needed for representative environmental reconstructions is unknown. To investigate this, the current dataset could be expanded and statistical techniques (e.g. bootstrapping) could be employed to determine the sample size needed for robust palaeoenvironmental conclusions. Alternatively, the robustness and resolution of palaeoclimatic conclusions drawn from the sand ramp record could be tested by

investigating the sand ramps in a region with good existing palaeoenvironmental control.

8.3 RELIABILITY OF AGE ESTIMATES

The conclusions of this thesis and the answers to research questions 1 and 2 are underpinned by a reliance on the OSL dates. However, there are some common problems (e.g. high D_e values, high overdispersion, low sensitivity) which may limit the reliability of these ages.

8.3.1 HIGH-DOSE SAMPLES

Investigations into the reliability of D_e estimates at high doses (Chapter 4) suggest that the DRC properties of individual grains start to influence D_e accuracy between 100-200 Gy (Chapter 4). Non-monotonic growth has the largest impact, producing D_e underestimation, which increases with dose, when >30% of the signal originates from grains with non-monotonic DRCs. However, due to the experimental design (i.e. number and magnitude of dose points used in SG DRC construction) it is unclear where between 100-200 Gy, non-monotonic growth begins to influence the tested samples. Therefore, single grain dating, or at least screening, is suggested as the most appropriate technique above 100 Gy (Chapter 4).

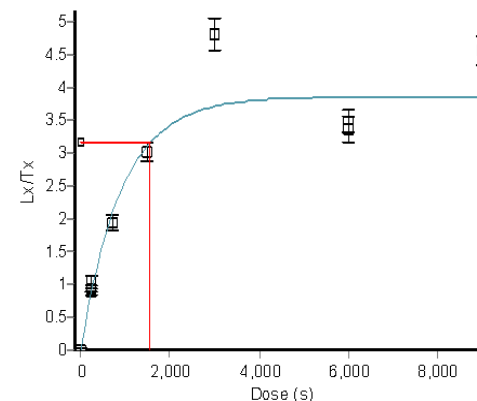
Unfortunately, these high-dose experiments were conducted after the majority of experimental work for this thesis had been completed. Therefore, time constraints prohibited SG dating of the 24 samples with D_e estimates above 100 Gy.

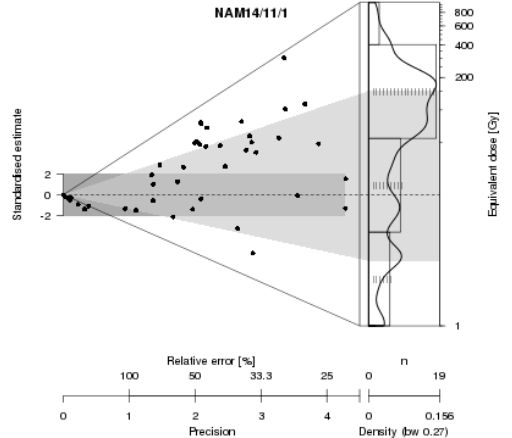
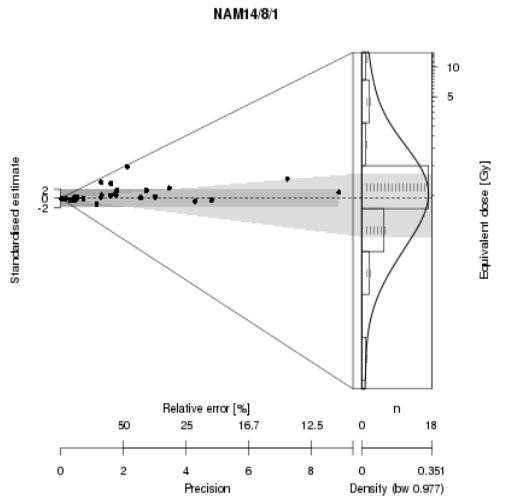
Instead, based on the findings of Chapter 4, the dates from the 16 (from a total of 62) samples with D_e values ≥ 200 Gy are treated as minimum ages, with no error bounds, throughout this thesis. Only broad geomorphic and palaeoenvironmental interpretations are drawn from these samples. 200 Gy was selected as the threshold as, although some grains had started to show diverging DRC properties by 200 Gy, the impact between 100-200 Gy was presumed minimal (Figure 4-7). Chapot (2014) propose a similar solution when interpreting high-dose samples, suggesting a 150 Gy limit for reliable D_e estimation when data on I_{NAT} is unavailable. DRC behaviour of all measured SG samples are shown in Figure 8-2 and Table 8-1. The proportion of signal from non-monotonic grains varies between samples; <35% of signal originates from non-monotonic grains for samples NAM14/3/1, NAM14/9/2, NAM14/11/1 and NAM14/12/7 whilst non-monotonic contribution reaches 50% for NAM14/8/3. This variation suggests that the impact of non-monotonic grains is likely to differ between samples. Nevertheless, as the non-monotonic behaviour of the natural dose is unknown (see section 4.5.4) even samples with relatively low non-monotonic influence are treated with caution.

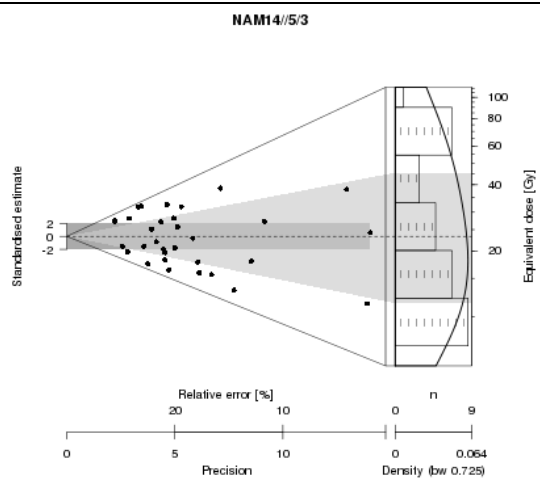
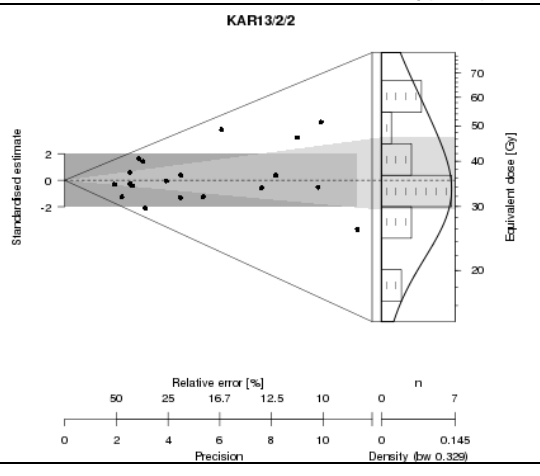
The 200 Gy threshold is an imperfect solution as, unlike single grain dating, it does not take into account variation in DRC behaviour between samples and considerable age precision is lost. Inter-aliquot variation is also not taken into account with 25 out of 62 samples containing aliquots with $D_{es} > 200$ Gy. These aliquots are still included in CAM calculations as their removal would cause greater D_e underestimations than DRC variations.

Table 8-1 Single grain properties of the subset of samples measured in this thesis.

Sample	# lum.	# DRC 3	DRC 3 age range (ka)	DRC 3 CAM age (ka)	SA CAM age (ka)	% DRC 4 sig	Other
AUS13/1/1	11/500 (2.2%)	0	n/a	n/a	n/a	n/a	Dim signal and IR depletion ratio failure of most grains suggests sample is unsuitable for dating. However, 180-210 μ m size fraction used for SG analysis may have different property to 90-125 μ m grains used for SA dating (see section 5.4.7)
NAM14/3/1	86/ 500 (12%)	7	85 $\pm$ 32 to 204 $\pm$ 50	146 $\pm$ 21	91 $\pm$ 8	25-35	Difference between DRC 3 SG age and SA age might support conclusions of non-monotonic grains causing underestimations in SA analysis. However, small sample size and no age control limits conclusions.
NAM14/9/2	52/500 (10.4%)	4	172 $\pm$ 40, 300 $\pm$ 56, 402 $\pm$ 116, and OD	269 $\pm$ 50	225 $\pm$ 19	15-22	
NAM14/12/7	28/500 (5.6%)	3	22 $\pm$ 6, 59 $\pm$ 24, 88 $\pm$ 10	50 $\pm$ 18	74 $\pm$ 7	20-30	Dominated by one very bright DRC 3 grain (88 $\pm$ 10 ka). May influence results in Paper 1.
NAM14/8/3	105/1000 (10.5%)	9	28 $\pm$ 8 to 171 $\pm$ 24 (extrap)	80 $\pm$ 15 (71 $\pm$ 14 extrap. removed)	66 $\pm$ 5	37-50	Not used in paper 1 due as different dose points were used for DRC construction
NAM14/14/5	11/500 (2.2%)	0	n/a	n/a	n/a	n/a	Bright signal (Figure 8-1). Very unusual DRC properties for 7 grains (right). Most likely measurement error but needs further investigation.



Sample	# lum.	# DRC 3	DRC 3 age range (ka)	DRC 3 CAM age (ka)	SA CAM age (ka)	% DRC 4 sig	Other
NAM14/11/1	153/1000 (15.3%)	12	2 ± 2 to 290 ± 148	42 ± 22 *53 ± 13 OD = 142 ± 19 (45 grains)	91 ± 10 OD = 42 ± 5	25-35	D_e values range from 0 to oversaturated suggesting post depositional disturbance. Sample taken from under calcrete layer but beta heterogeneity or decreased dose rates over time would not explain 0-grains. Bioturbation, sediment slumping or light exposure during sampling are suggested causes. This calls into question the reliability of SA ages which are quoted as >90 ka in this thesis. 
NAM14/8/1	53/1000 (5.3%)	n/a	n/a	*0.2 ± 0.0 OD: 29 ± 7 % (27 grains)	1.0 ± 0.3 OD = 124 ± 19	n/a	SG D_e estimates are significantly lower than highly OD SA results. 43 grains < 1 Gy although 4 > 70 Gy. This suggests post depositional disturbance possibly as a result of bioturbation (site showed evidence of termites and rootlets were present close to the sample) or slumping in the gully. Some slumping/reworking is supported by the young ages obtained from gullies at Klein Aus (0.2 ± 0.2 ka, 1.2 ± 1.2 ka) and Rosh Pinah (0.2 ± 0.04 ka, 1.0 ± 0.3 ka) although SG data is not available. Sediments presumed primarily aeolian but some colluvial influence and therefore partial bleaching cannot be ruled out 

Sample	# lum.	# DRC 3	DRC age range (ka)	DRC 3 CAM age (ka)	SA CAM age (ka)	% DRC 4 sig	Other
NAM14/5/3	146/500 pseudo-single grains (125-180 μm ; ~4-12 grains (Arnold et al., 2012))	n/a	n/a	* 6.2 ± 0.2 OD: $74 \pm 10\%$ (35 grains)	19.5 ± 3.3 OD = 76 ± 10	n/a	36 oversaturating grains excluded from D_e calculation. OD suggests SA ages for this sample may not be reliable. Stability is still suggested by the presence of calcrete nodules (Netterberg, 1969) but the duration is less certain. Calcrete nodules may have caused beta heterogeneity explaining some OD. Although, bioturbation (rootlets are present), post depositional mixing (gully location) or partial bleaching (possibly colluvial contribution) may also contribute. 
KAR13/2/2	52/1000 (5.2%)	n/a	n/a	* 34 ± 3 OD = 22 ± 3 (28 grains)	30 ± 3 OD = 33 ± 3	n/a	Good agreement between SG and SA. Relatively low sensitivity (Figure 8-1). OD is possibly a result of noise or beta heterogeneity. 

Sample	# lum.	# DRC 3	DRC 3 age range (ka)	DRC 3 CAM age (ka)	SA CAM age (ka)	% DRC 4 sig	Other
KAR13/2/3	49/500 (9.8%)	11	42 ± 14 to 234 ± 67	92 ± 12 *72 ± 7 OD: 34 ± 5 (23 grains)	95 ± 5 OD = 20 ± 2	n/a	Some grains feldspar contaminated. Relatively high DRC 3 grains possibly due to lower doses used for DRC construction (max 294 Gy). Difference between SG and SA might be due to removing oversaturating grains (Stone and Bailey, 2012). DRC 3 in good agreement with SA.

* Age estimates based on all grains which pass rejection criteria

#: number of grains which produce a DRC

DRC 3: number of DRC 3 grains which pass rejection criteria at 15%

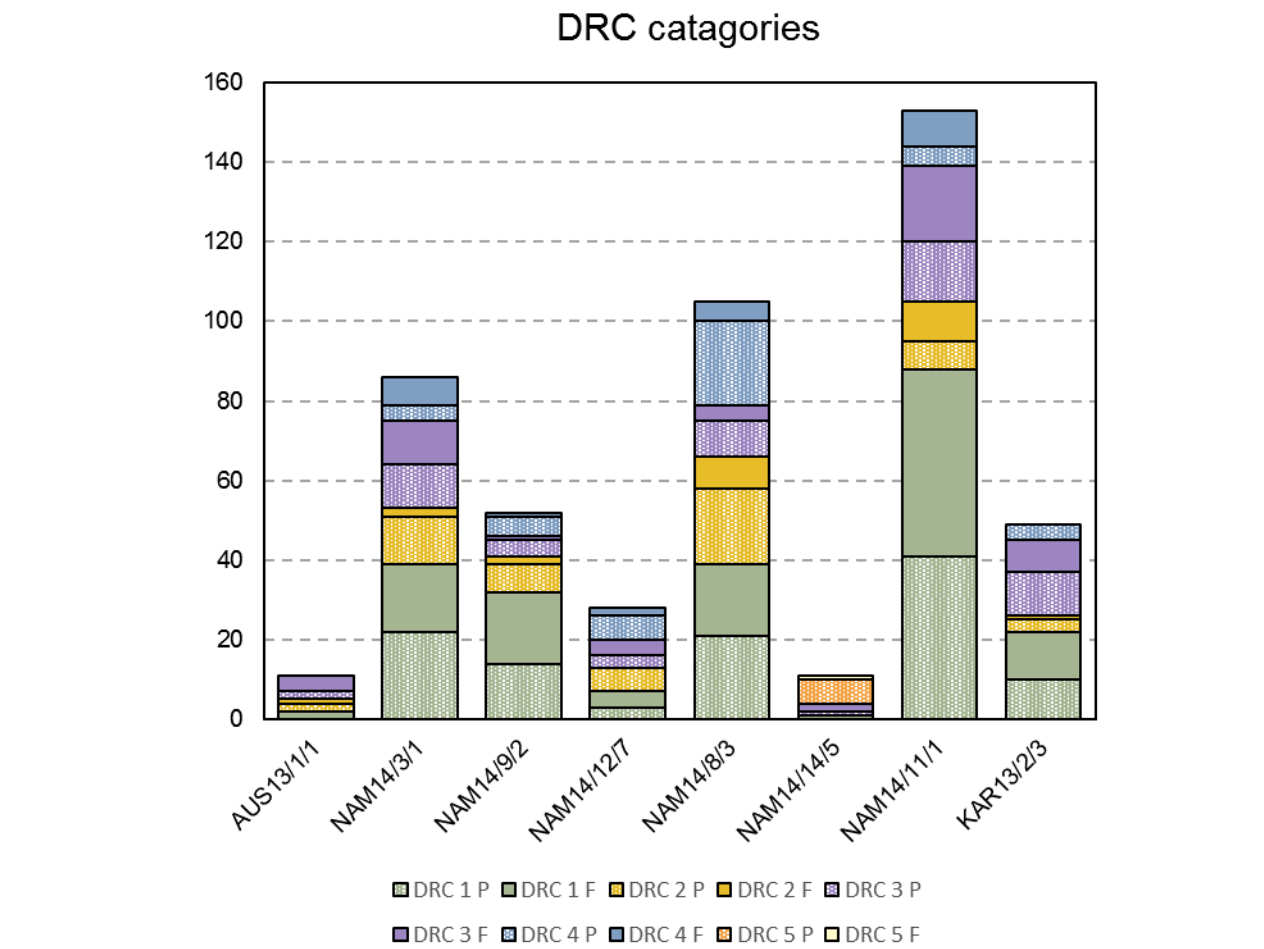


Figure 8-2. DRC properties of measured grains. Solid colour represents grains which fail rejection criteria at 15%. Hatched colour shows number of grains that passed. Table 4.1 provides explanation of the different DRC types.

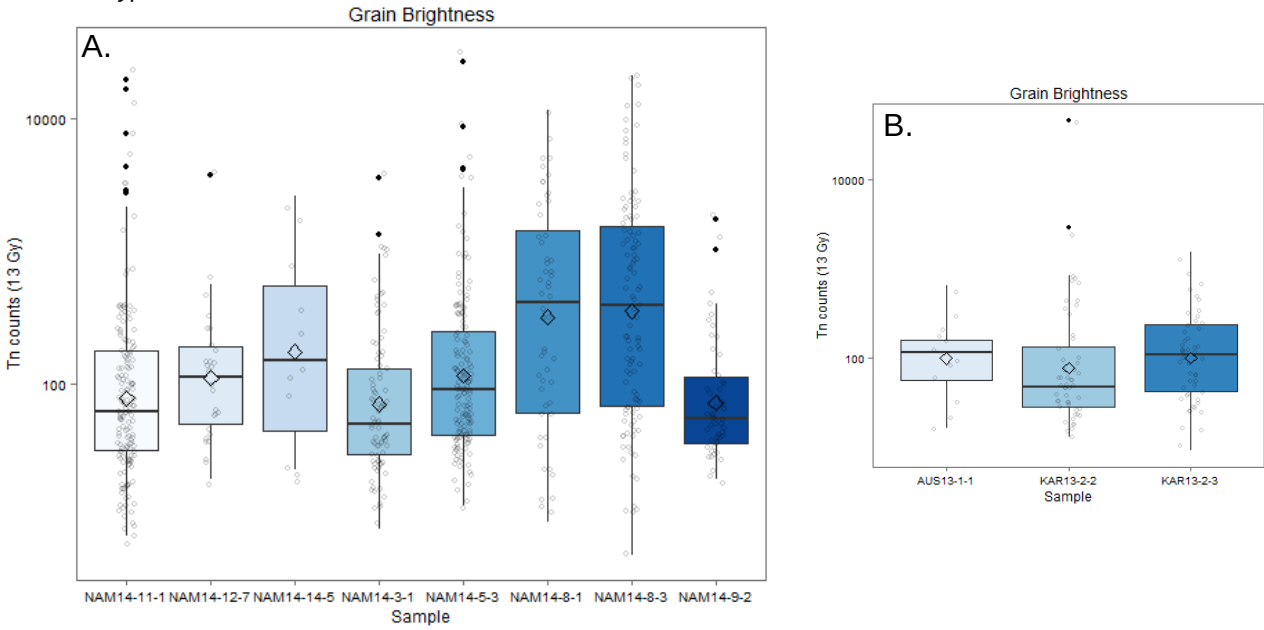


Figure 8-1 Sensitivity of luminescing SGs to a 13 Gy(A) and 19 Gy (B) Tn dose. Counts are displayed on a log scale.

8.3.2 FUTURE WORK AND UNANSWERED QUESTIONS

One significant limitation of the investigations in Chapter 4 is the lack of independent age control. As such, the conducted SG D_e accuracy tests primarily assess the accuracy of DRC fitting and the response to a laboratory dose. It is presently unclear how these findings translate to a natural dose. Non-monotonic growth is modelled within natural signals (Chen et al., 2015; Lawless et al., 2005; Pagonis et al., 2006; Woda and Wagner, 2007), perhaps even to a greater degree than in laboratory irradiated signals (Chen et al., 2015). Therefore, work is needed to characterise the behaviour of L_n . SG studies on multiple samples with independent age control could provide insights into whether DRC behaviour is the source of age underestimation and whether ‘supergrain’ dating of DRC 3 grains produces accurate results. At present, ‘supergrain’ dating appears to be a promising but time consuming technique (Arnold et al., 2016; Demuro et al., 2015; Yoshida et al., 2000). In the absence of known age control, expanding the study to cover more samples and exploring the dose response properties of SGs at various doses could provide the next steps in exploring DRC induced age underestimation.

8.3.3 OVERDISPERSION

Overdispersion values characterise the spread in D_e values above the amount expected from measurement uncertainties and machine reproducibility alone. Sources of overdispersion include: variation in luminescence properties and the response to SAR, variations between natural and laboratory responses to heating and irradiating, beta-dose heterogeneity, partial bleaching and post-depositional

mixing (Galbraith et al., 2005). The published overdispersion values from SA studies (100-1000 grains per aliquot) conducted on well-bleached quartz with no evidence of mixing range from 3 to 33% (mean 14 ± 1 %) (Arnold and Roberts, 2009).

Due to the heterogeneous nature of sand ramps some overdispersion is expected, but accurately identifying the cause is difficult. Bioturbation is common in sandy sediments but may leave no geomorphic evidence (Bateman et al., 2003a, 2007a, 2007b; Gliganic et al., 2016). The presence of colluvial sediments may increase the likelihood of beta heterogeneity or partial bleaching (Fuchs and Lang, 2009) but accurately interpreting depositional process from sediments alone can be problematic, especially as samples may be derived from multiple depositional processes (section 5.4.7). In addition, reworking and gully collapse may lead to sediment mixing. 12 samples in this study (with D_e values <200 Gy) have overdispersion values $\geq 30\%$ (Table 5-4 and Table 6-2).

The high dose rates and high $^{40}\text{K}\%$ at sites between 26.73°S to 24.72°S and 15.93°E – 16.53°E (Neuhof-1 & 2, Jagkop, Keerwader, Sandkop, Aus and Klein Aus) suggest that beta heterogeneity is less likely to be a primary contributor to SA overdispersion at these sites (Chauhan and Singhvi, 2011; Guérin et al., 2015; Mayya et al., 2006). However, for samples KAR13/4/1 (OD = 34%, K = 0.39%), KAR13/2/2 (OD = 33%, K = 0.46%), KAR13/1/2 (OD = 30%, K = 0.56%), NAM14/10/3 (OD = 38%, K = 1.2%), NAM14/11/1 (OD = 42% K = 1.5) beta heterogeneity may be significant (See section 3.1.3).

The single grain properties of samples NAM14/11/1 (OD = 42%), NAM14/8/1 (OD = 124%), NAM14/5/3 (OD = 74%) and KAR13/2/2 (OD = 33 %) are described in Table 8-1. 0-dose grains are present in samples NAM14/8/1 and NAM14/11/1 indicating post depositional mixing. Therefore, bioturbation is the most plausible cause of the observed scatter in these samples and NAM14/5/3 (Bateman et al., 2003a, 2007a, 2007b; Gliganic et al., 2016). However, as supported by the erroneously young ages obtained from gully samples at Rosh Pinah and Klein Aus, sediment slumping may also contribute to overdispersion at these sites (Table 5-4, Figure 5-5). Partial bleaching could also play a minor role in the observed scatter. Samples are presumed to be primarily aeolian but some colluvial material may also be present (section 5.4.7). $D_e(t)$ plots (Bailey, 2003) showed no relationship with D_e value, suggesting partial bleaching is not the primary cause of the scatter.

High overdispersion (32-43%) is also found at Neuhof-1, for samples NAM14/7. This may be a result of the highly heterogeneous sediment matrix, the persistent feldspar contamination found at this site, or may indicate sediment mixing or partial bleaching. These samples are associated with colluvial deposits so beta heterogeneity or partial bleaching provide a plausible explanation for overdispersion (Fuchs and Lang, 2009; Guérin et al., 2015). High overdispersion is found at the Aus dunes (AUS13/3/1 – 67%, AUS13/3/2 - 29%, AUS13/3/3 - 29%, AUS13/6/1-40%). Single grain data are not available for these samples. For sample AUS13/6/1 overdispersion may be partially explained by noise as these

aliquots exhibit very low sensitivity. At AUS13/3 bioturbation may be the most plausible explanation for overdispersion.

Fortunately, the removal of samples with high overdispersion from PDF analysis does not significantly impact overall palaeoenvironmental conclusions (Figure 8-3). Nevertheless, the ages discussed above should be treated with caution. Unfortunately, time was not available to conduct the more extensive dating, including detailed single grain studies, needed to more accurately interpret the causes of overdispersion and the most likely age estimates.

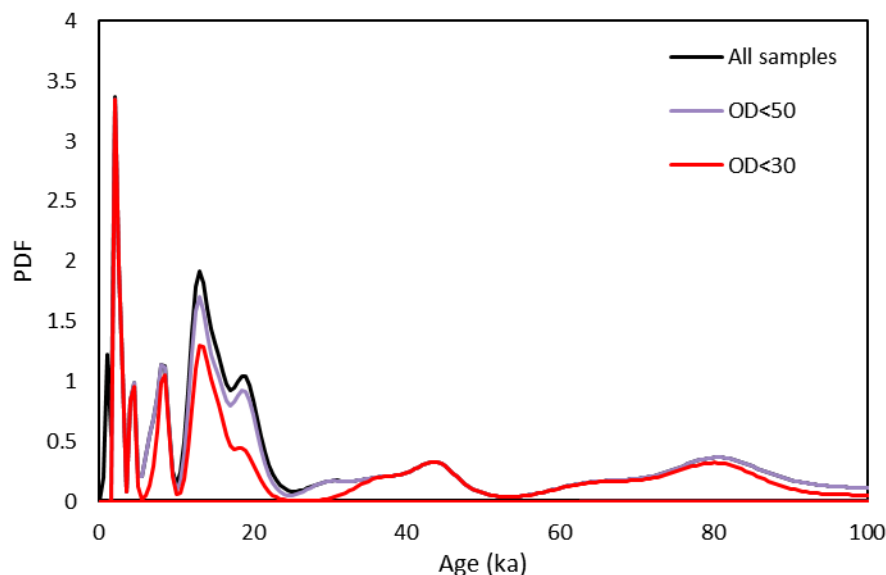


Figure 8-3. PDF analysis with different overdispersion thresholds. Black line represents all samples with >10 accepted aliquots, samples with OD values >50% (blue line) and >30% (red line) were removed from analysis. Removal of very highly dispersed samples (>50% OD) has minimal impact. Removal of moderately overdispersed (>30% OD) decreases the significance of LGM accumulation but activity at this time is still suggested.

8.3.4 MOISTURE CONTENT

Moisture content provides one of the greatest uncertainties in OSL dating (see section 3.1.2 for discussion). The comparatively low dose rates at Klipkraal means changes in water content have a big impact on calculated age. Increasing the

assumed water content from $5\% \pm 5\%$ to $10\% \pm 5\%$ increases the mean age estimate by ~80% for all samples. Such a change would indicate that the aeolian unit thought to be associated with human disturbance accumulated before the onset of farming, whilst the Unit 2 palaeosol sediments accumulated during the, regionally arid (Chevalier and Chase, 2015, 2016), LGM and Last Glacial-interglacial transition. This suggests that the applied value of $5 \pm 5\%$ is sensible. In Namibia the higher dose rates mean changes in water content are less significant. Nevertheless, these values represent an untestable uncertainty in all samples.

8.3.5 REGIONAL VARIATION

At least some of the unfavourable OSL characteristics found in this study (e.g. high dose rates, high overdispersion, limited sensitivity) may be regional specific. Telfer et al. (2012) and Bateman et al. (2012) suggest satisfactory luminescence properties indicating more favourable dating conditions in the Drakensberg Mountains, South Africa and the Mojave Desert, USA than in southern Namibia. However, Kumar et al. (2016) report high overdispersion and poor recycling in the Ladakh Desert, India, possibly as a result of a high proportion of sampled colluvial deposits, whilst the majority of sand ramp aliquots from Black Mesa, USA fail rejection criteria (Ellwein et al., 2015). This suggests that the success of sand ramp dating may be regionally specific, but also indicates that caution should be taken, particularly as unwittingly sampled colluvial deposits have the potential to impart unfavourable luminescence properties (e.g. Fuchs and Lang, 2009; Kumar et al., 2016).

Chapter 9: Conclusion

This thesis presents perhaps the largest investigation to date of sand ramp sedimentary records and addresses fundamental questions regarding the formation of these features and their utility as palaeoenvironmental archives. Sand ramps are shown to be common features in the southern Namibia area but harder to identify in semi-arid South Africa where, unless the sediments are gullied, sand ramps may go undetected. The distribution of sand ramps suggests formation is controlled by four factors. Sediment supply, and availability of accommodation space are suggested to determine the timing of accumulation, whilst wind regimes and climate control the location and the depositional mechanism. Sand ramps are dated to ~0.15 - >100 ka indicating long-term, periodic accumulation over the late Quaternary. The sediments and morphology of the sand ramps are varied, reflecting different local controls. Overall, sand ramp accumulation is suggested to be controlled by local factors modulated by regional climate.

This study presents a substantial contribution to both our understanding of sand ramp formation and Quaternary geomorphic processes in southern Namibia. The existence of sand ramps within the Namibian landscape for >100 ka sits comfortably within the 1 Ma timeframe for ongoing aeolian processes and sand residency (Garzanti et al., 2012, Vermeesch et al., 2010). Contrary to the findings of Bristow et al. (2005; 2007), the sand ramp record demonstrates that there is potential for obtaining palaeoenvironmental records which span the entirety of the late Quaternary from aeolian features in the region. The ages obtained from the suite of sand ramps (<1 ka - >100 ka) suggest that, due to their relative stability,

sand ramps may be able to provide a more complete record of late Quaternary environmental processes in southern Namibia than previously obtained River and dune records (e.g. Bristow et al., 2005; 2007; Brook et al., 2006; Srivastava et al., 2006; Stone et al., 2010). However, further work is needed to further understand the environmental forcings of the sand ramp record and a large suite of sand ramps is needed to disentangle local and regional forcing factors.

9.1 WHAT IS THE UTILITY OF THE SAND RAMP RECORD AS A PALAEOENVIRONMENTAL ARCHIVE?

Sand ramp accumulation is heavily controlled by local factors. Aeolian accumulation occurs when sediment supply coincides with accommodation space availability, whilst colluvial and fluvial deposits reflect local patterns of run-off and weathering. Controls on both aeolian and colluvial sediment supply are heavily influenced by the local and regional setting.

Due to the number of local factors in operation, individual sand ramps cannot be used in isolation to draw regional palaeoclimatic conclusions. However, individual sand ramps provide records of local conditions and when multiple sand ramps are analysed simultaneously periods of regional sand ramp activity (i.e. concurrent accumulation at multiple sites) can be identified. As aeolian sand ramp accumulation is primarily driven by sediment supply, periods of regional sand ramp activity provide information on regional sediment dynamics and therefore palaeoclimatic conditions.

However, there are several limitations to the palaeoenvironmental records that can be obtained from sand ramps:

1. Differentiating between local and regional forcing factors can be problematic, creating a large potential for misinterpretation.
2. As a complex mosaic of processes operate on sand ramps, correctly identifying the mode of deposition and its relationship to the sand ramp chronology can be difficult, once again producing a large potential for misinterpretation.
3. The combination of colluvial and aeolian processes and the potential for bioturbation and reworking can make OSL dating challenging, whilst older sediments are beyond the reliable range of the technique, limiting the scope of the investigations.
4. Palaeoenvironmental reconstructions are reliant on understanding the regional controls on formation and obtaining a large enough dataset to differentiate between local and regional forcing. Both of these processes are time consuming and labour intensive.

Overall, sand ramps do have the potential to provide low resolution regional palaeoclimatic information in dryland regions but their utility is limited by a number of caveats. Work on the sand ramp record is labour intensive and based on a number of assumptions. Regional investigations on controls need to be conducted before sand ramps can be utilised in a new area and a large number of features need to be studied to draw robust regional palaeoenvironmental

conclusions. In addition, many questions still exist regarding the processes operating on sand ramps including the relationships between chronology and sediment accumulation, and between climate and sediment accumulation, and the regional transferability of the above conclusions. Nevertheless, sand ramps are relatively common features which form in regions where very few other palaeoenvironmental proxies exist. Therefore, despite their limitations they have the potential to provide a significant contribution to palaeoenvironmental reconstructions from these regions.

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APPENDIX A: AUTHOR CONTRIBUTION

The majority of work completed in this thesis is my own however a small proportion has been contributed by the co-authors of the four submitted papers. Details of author contributions are described below.

Sampling in the field was conducted by myself (AR), Dave Thomas (DT) and Richard Bailey (RB). Peter Holmes (PH) provided access to his former study site (the Klipkraal Sands) and assisted with fieldwork at this site. The majority of sample preparation, laboratory work, dating and data analysis was conducted by AR. RB conducted and analysed the very high dose experiments conducted on eight samples (section 4.4.4). Marta Padoan prepared and analysed the fourteen heavy mineral samples presented in section 5.3.2. Edo Garzanti provided advice and guidance on the heavy mineral results. Abi Stone (AS) provided two of the samples used for aeolian sensitivity tests (section 5.3.1). These were collected and partially prepped by AS. RB wrote the script for and performed the ARV analysis described in section 7.4.2. Simon Armitage (SA) provided use of his laboratory for the experiments conducted in Chapter 4. All figures and tables were produced by AR. After discussion with RB and DT all papers were written by AR. SA and RB provided comments which significantly improved Paper 1 (Chapter 4). PH provided comments on paper 3 (Chapter 6). DT provided comments which significantly improved papers 2-4 (Chapters 5-7).

APPENDIX B: SUP. INFO. – PAPER 1 (CHAPTER 4)

B.1 SINGLE GRAIN FAST RATIOS

Duller (2012) adapted the fast ratio for application to single grains and found that grains displaying non-monotonic growth failed the fast ratio. Fast ratios were calculated following Duller (2012), Jacobs et al. (2013) and Feathers and Pagonis, (2015) for the different DRC types. No systematic difference between the categories was found (Figure B-2).

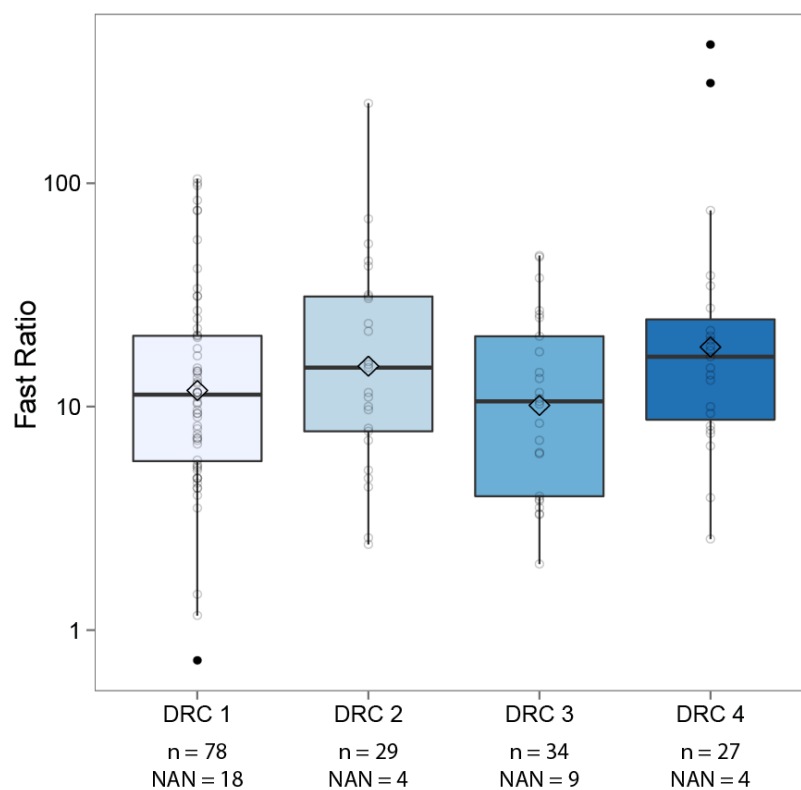


Figure B-9-1. Fast ratios of different DRC types. solid line indicates the median, diamond represents the mean. The interquartile range is represented by the box. Black dots represent outliers. n= total number of grains . NAN= number of grains the fast ratio could not be calculated for because subtraction of the background led to zero or negative medium component values.

B.2. SYNTHETIC ALIQUOT TYPES

Synthetic aliquots were created using grains of different properties. Table B-1. Presents the different categories and the number of grains used in each aliquot for each sample.

Table B-9-1 Synthetic aliquot types and number of included grains for each sample

DRC types included	Number of grains		
	NAM14/3/1	NAM14/9/2	NAM14/12/7
All grains	500	500	500
1	49	37	10
2	13	6	5
3	18	5	6
4	7	5	7
1&2	62	43	15
1&3	67	42	16
1&4	56	42	17
1&2&3	80	48	21
1&3&4	74	47	23
1&2&3&4	87	53	28
2&3	31	11	11
2&3&4	38	16	18
3&4	25	10	13

B.3. IR RATIO RESULT DETAILS

Table B-9-2 IR ratio results. The top 3 samples (NAM14/5/4, NAM14/12/7, AUS13/1/1) were measured using a repeated IR ratio test. The remaining samples were De measurements with one 20 Gy and one 400 Gy IR-ratio point.

IR ratio	20Gy average	20Gy1	20Gy2	200Gy average	200Gy1	200Gy2	400Gy average	400Gy1	400Gy2
NAM14/5/4	0.98	0.99	0.96	0.95	0.94	0.96	0.95	0.96	0.95
	1.01	1.01	1.01	0.97	0.97	0.98	1.00	1.01	0.99
	1.00	1.01	0.99	0.99	0.98	0.99	0.97	0.98	0.97
	0.99	1.03	0.96	0.98	0.98	0.99	0.97	0.97	0.97
	0.99	1.01	0.98	0.97	0.95	0.98	0.96	0.96	0.96
	1.01	1.02	0.99	0.98	0.99	0.98	0.97	0.97	0.97
	1.00	1.04	0.97	0.98	0.97	0.99	1.01	1.00	1.01
	1.00	1.03	0.98	0.99	0.99	0.99	1.01	1.00	1.02
	0.99	1.01	0.96	1.02	1.06	0.97	0.96	0.97	0.96
	0.99	1.02	0.96	0.97	0.97	0.98	0.97	0.97	0.98
NAM14/12/7	0.98	0.98	0.98	0.99	1.02	0.96	1.01	1.03	1.00
	0.96	0.96	0.96	0.98	0.98	0.98	0.92	0.92	0.93
	0.99	0.99	1.00	0.99	0.98	1.00	0.97	0.98	0.97

IR ratio	20Gy average	20Gy1	20Gy2	200Gy average	200Gy1	200Gy2	400Gy average	400Gy1	400Gy2
	0.96	0.99	0.94	1.00	0.98	1.01	1.01	1.00	1.01
	1.02	1.02	1.02	1.00	1.00	1.01	0.93	0.91	0.95
	1.05	1.04	1.06	0.97	0.99	0.95	0.95	1.01	0.90
	0.97	0.97	0.97	0.98	0.98	0.97	0.92	0.90	0.95
	0.95	0.91	0.98	0.97	0.90	1.04	1.01	0.99	1.04
	1.00	1.00	1.00	0.99	0.99	1.00	0.98	0.99	0.98
	0.97	0.97		0.96	0.96		0.93	0.93	
AUS13/1/1	0.97	0.98	0.96	1.02	0.99	1.04	0.99	0.98	1.01
	0.92	0.91	0.92	1.01	1.01	1.01	0.99	1.00	0.98
	1.00	0.99	1.01	1.02	1.00	1.03	0.97	0.94	1.00
	0.98	1.01	0.94	1.04	1.04	1.05	1.00	0.98	1.01
	1.02	1.00	1.04	1.03	1.00	1.05	0.97	0.95	0.98
	0.98	0.95	1.00	1.00	0.99	1.01	0.98	0.97	1.00
	0.98	0.97	0.99	1.00	0.98	1.02	0.97	0.96	0.98
	0.95	0.99	0.90	0.99	0.99	0.99	0.98	1.02	0.94
	0.94	0.95	0.93	0.95	0.93	0.98	0.98	1.00	0.97
	0.93	0.96	0.90	0.92	0.90	0.94	0.89	0.88	0.89
NAM14/5/4	0.94	0.94	n/a	n/a	n/a	n/a	1	n/a	1
	1	1	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	0.97	0.97	n/a	n/a	n/a	n/a	0.95	n/a	0.95
	0.93	0.93	n/a	n/a	n/a	n/a	0.97	n/a	0.97
	0.98	0.98	n/a	n/a	n/a	n/a	0.99	n/a	0.99
	0.96	0.96	n/a	n/a	n/a	n/a	0.94	n/a	0.94
	0.97	0.97	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	1.03	1.03	n/a	n/a	n/a	n/a	1.02	n/a	1.02
	0.97	0.97	n/a	n/a	n/a	n/a	0.99	n/a	0.99
	0.95	0.95	n/a	n/a	n/a	n/a	1.01	n/a	1.01
	0.97	0.97	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	0.97	0.97	n/a	n/a	n/a	n/a	0.96	n/a	0.96
	1.01	1.01	n/a	n/a	n/a	n/a	0.99	n/a	0.99
NAM14/3/1	0.95	0.95	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	1.02	1.02	n/a	n/a	n/a	n/a	0.99	n/a	0.99
	0.95	0.95	n/a	n/a	n/a	n/a	0.92	n/a	0.92
	0.81	0.81	n/a	n/a	n/a	n/a	0.92	n/a	0.92
	1.01	1.01	n/a	n/a	n/a	n/a	1.06	n/a	1.06
	1	1	n/a	n/a	n/a	n/a	0.97	n/a	0.97
	0.95	0.95	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	1	1	n/a	n/a	n/a	n/a	1	n/a	1
	0.87	0.87	n/a	n/a	n/a	n/a	0.99	n/a	0.99
	0.97	0.97	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	0.9	0.9	n/a	n/a	n/a	n/a	1.03	n/a	1.03

IR ratio	20Gy average	20Gy1	20Gy2	200Gy average	200Gy1	200Gy2	400Gy average	400Gy1	400Gy2
	0.99	0.99	n/a	n/a	n/a	n/a	0.97	n/a	0.97
NAM14/3/3	0.9	0.9	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	0.8	0.8	n/a	n/a	n/a	n/a	0.72	n/a	0.72
	0.95	0.95	n/a	n/a	n/a	n/a	0.92	n/a	0.92
	1.04	1.04	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	0.96	0.96	n/a	n/a	n/a	n/a	0.97	n/a	0.97
	0.99	0.99	n/a	n/a	n/a	n/a	0.92	n/a	0.92
	0.97	0.97	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	0.99	0.99	n/a	n/a	n/a	n/a	1.03	n/a	1.03
	0.96	0.96	n/a	n/a	n/a	n/a	0.96	n/a	0.96
	0.93	0.93	n/a	n/a	n/a	n/a	1.02	n/a	1.02
	1.12	1.12	n/a	n/a	n/a	n/a	0.92	n/a	0.92
	0.9	0.9	n/a	n/a	n/a	n/a	0.94	n/a	0.94
NAM14/9/1	0.97	0.97	n/a	n/a	n/a	n/a	0.89	n/a	0.89
	0.99	0.99	n/a	n/a	n/a	n/a	0.92	n/a	0.92
	0.95	0.95	n/a	n/a	n/a	n/a	0.88	n/a	0.88
	1.02	1.02	n/a	n/a	n/a	n/a	0.99	n/a	0.99
	0.97	0.97	n/a	n/a	n/a	n/a	0.96	n/a	0.96
	0.97	0.97	n/a	n/a	n/a	n/a	1.05	n/a	1.05
	0.99	0.99	n/a	n/a	n/a	n/a	0.97	n/a	0.97
	0.84	0.84	n/a	n/a	n/a	n/a	0.73	n/a	0.73
	0.96	0.96	n/a	n/a	n/a	n/a	0.96	n/a	0.96
	0.96	0.96	n/a	n/a	n/a	n/a	1.03	n/a	1.03
	0.96	0.96	n/a	n/a	n/a	n/a	0.94	n/a	0.94
	0.97	0.97	n/a	n/a	n/a	n/a	0.89	n/a	0.89
NAM14/9/2	0.9	0.9	n/a	n/a	n/a	n/a	0.86	n/a	0.86
	1.01	1.01	n/a	n/a	n/a	n/a	0.99	n/a	0.99
	0.98	0.98	n/a	n/a	n/a	n/a	0.94	n/a	0.94
	0.96	0.96	n/a	n/a	n/a	n/a	1.09	n/a	1.09
	0.91	0.91	n/a	n/a	n/a	n/a	1	n/a	1
	1.02	1.02	n/a	n/a	n/a	n/a	0.99	n/a	0.99
	0.92	0.92	n/a	n/a	n/a	n/a	0.91	n/a	0.91
	0.98	0.98	n/a	n/a	n/a	n/a	0.96	n/a	0.96
	0.97	0.97	n/a	n/a	n/a	n/a	0.99	n/a	0.99
	0.92	0.92	n/a	n/a	n/a	n/a	1.03	n/a	1.03
	0.95	0.95	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	0.97	0.97	n/a	n/a	n/a	n/a	0.96	n/a	0.96
NAM14/14/5	0.96	0.96	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	0.69	0.69	n/a	n/a	n/a	n/a	0.49	n/a	0.49
	0.95	0.95	n/a	n/a	n/a	n/a	1	n/a	1
	0.94	0.94	n/a	n/a	n/a	n/a	0.99	n/a	0.99

IR ratio	20Gy average	20Gy1	20Gy2	200Gy average	200Gy1	200Gy2	400Gy average	400Gy1	400Gy2
	0.93	0.93	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	0.98	0.98	n/a	n/a	n/a	n/a	1	n/a	1
	0.97	0.97	n/a	n/a	n/a	n/a	1.04	n/a	1.04
	0.97	0.97	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	0.97	0.97	n/a	n/a	n/a	n/a	1.01	n/a	1.01
	0.75	0.75	n/a	n/a	n/a	n/a	0.55	n/a	0.55
	0.96	0.96	n/a	n/a	n/a	n/a	0.98	n/a	0.98
	0.97	0.97	n/a	n/a	n/a	n/a	0.98	n/a	0.98

B.4. ANOVA RESULTS

Results of ANOVA between different IR doses. No significant difference was found between the groups. 200 Gy group appears more different due to containing fewer measurements.

	Df	Sum Sq	Mean Sq	F value	Pr(>F)	P value overall
IRallg\$Dose	2	0.0173	0.008666	2.014	0.136	(not significant)
Residuals	233	1.0027	0.004303			

Tukey multiple comparisons of means
95% family-wise confidence level

	diff	lwr	upr	p adj (p-value between groups)
200Gy-20Gy	0.024656146	-0.007445924	0.056758215	0.1679839
400Gy-20Gy	-0.001872807	-0.023434499	0.019688884	0.9771275
400Gy-200Gy	-0.026528953	-0.058631023	0.005573116	0.1273644

B.5. GRAIN BY GRAIN L_N NORMALISED GRAPHS

Results of synthetic aliquot experiments at 198 Gy (Figure B-2) and 596 Gy (Figure B-3) when the L_n counts of individual grains were normalised to the counts needed to produce the same L_x/T_x value as the 198 Gy or 596 Gy dose point. Synthetic aliquot values were created by summing the modified L_n values and dividing by the original summed T_n value. Percentage of signal from each DRC type is displayed along the axes. Yellow values are within 10% of the expected value. Blue indicates underestimation. Red indicates over estimation. Very dark

blue represents oversaturation. From left to right graphs on the x axes represent contribution (%) from DRC 1, DRC 2 and DRC 3 grains. From top to bottom graphs on the y axis represent contribution (%) from DRC 4, DRC 3 and DRC 2 grains.

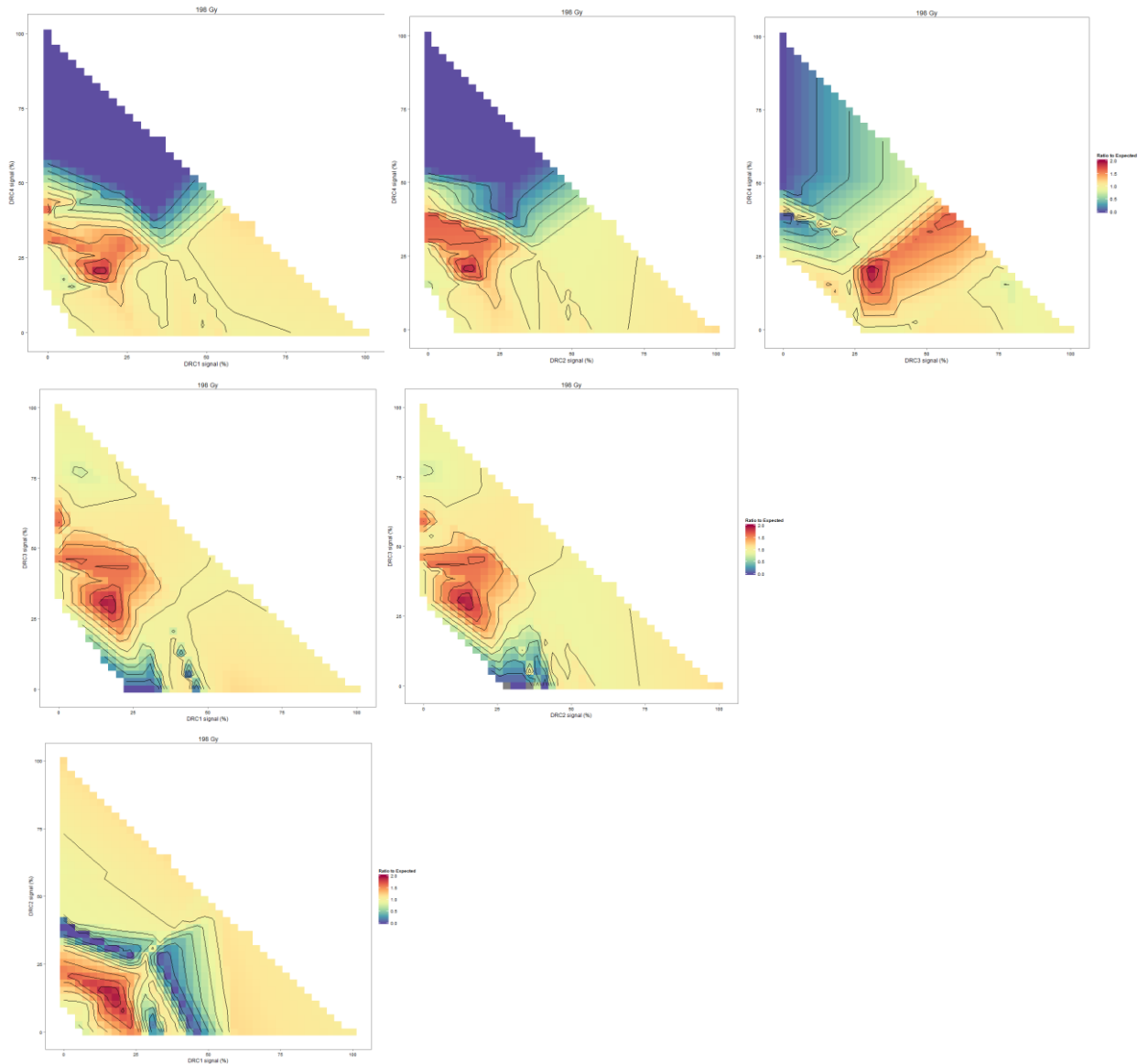


Figure B-9-2 Synthetic aliquots created from individual grains normalised to L_n counts of a 198 Gy dose.

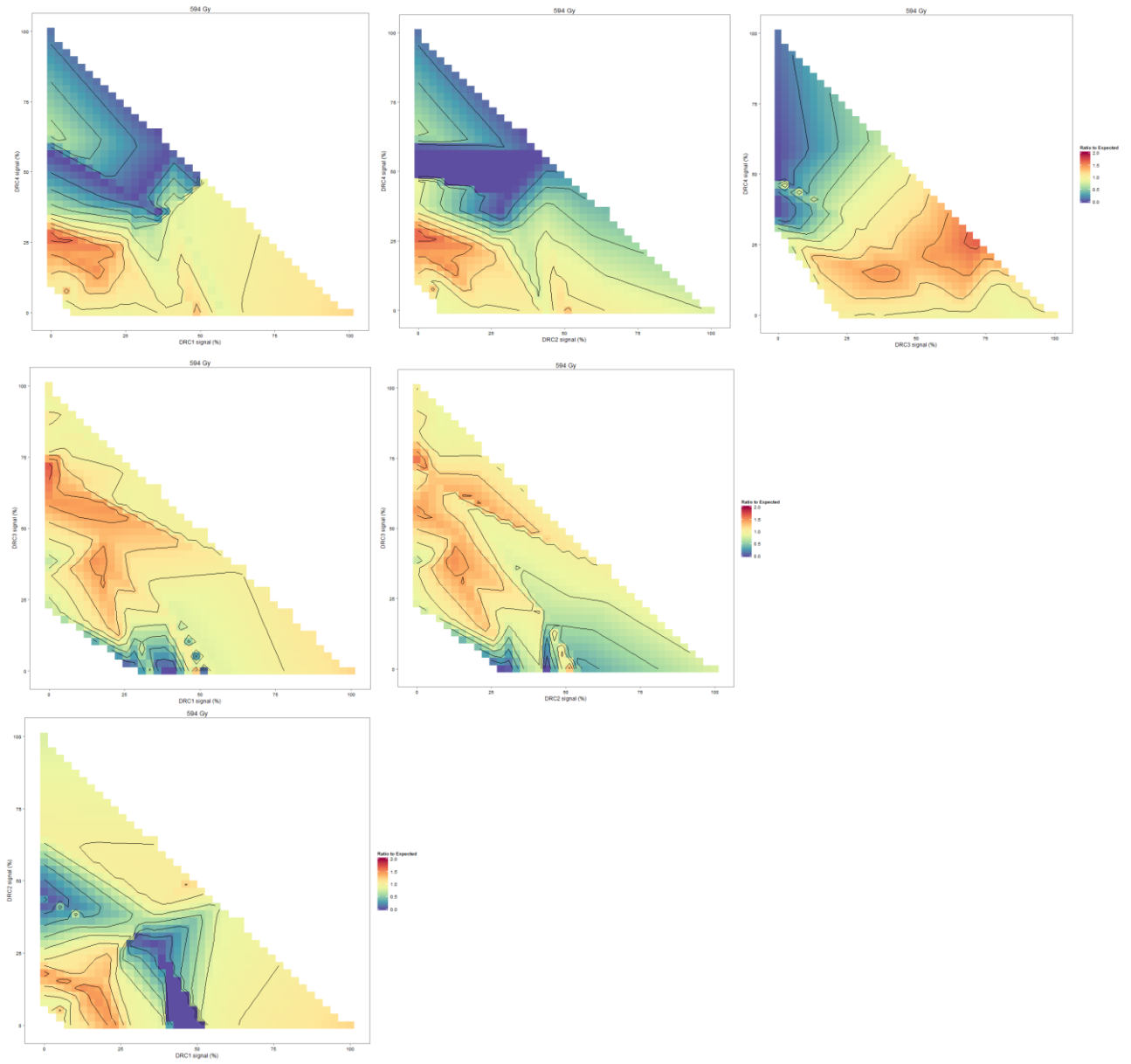


Figure B-3 Synthetic aliquots created from individual grains normalised to L_n counts of a 198 Gy dose.

B.6. HIGH DOSE DRCs

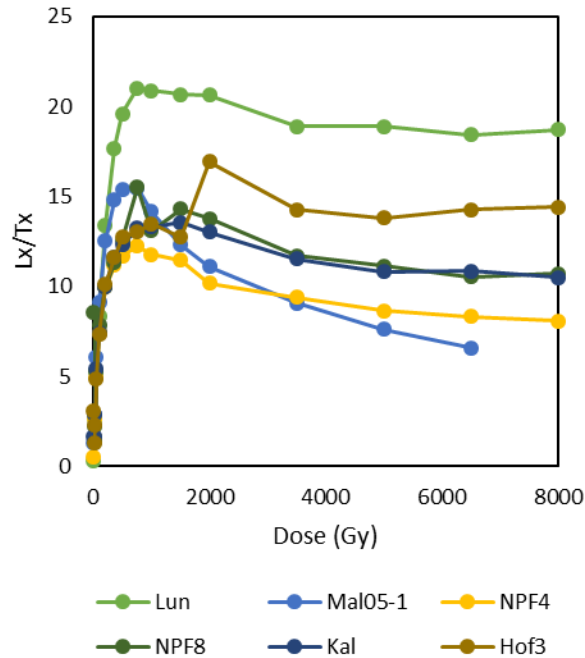


Figure B-4. High dose DRCs for the remaining six samples

Dose response curves up to 8000 Gy for the six samples not displayed Chapter 4.

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APPENDIX C: SUP. INFO. – PAPER 2 (CHAPTER 5)

Table C-1 Sand ramp Inventory

Latitude	Longitude	Orientation /Face	Climbing/ Falling	Length	Width	Class	Distance & direction to channel (m)	Distance & direction to major channel	Distance & direction to dune field	Accommodation Space	Studied
-27.7693	16.6658	East	Climbing and Falling	450	800	2	100, E	100, E	12000, W	Inselberg	-
-27.8455	16.6818	East	Climbing	250	500	1	160, E	250, E	8000, W	Inselberg	-
-26.4780	17.2925	North	Climbing	300	350	1	250, N	250, N	-	Edge of the escarpment	-
-26.5228	18.5035	East	Climbing	300	200	1	350, E	350, E	-	Edge of the escarpment	-
-18.7737	12.9956	West	Climbing	3000	3500	2	400, W	400, W	-	Edge of escarpment	-
-27.7644	16.6730	East	Climbing and Falling	300	750	1	50, E	500, E	11000, W	Inselberg	-
-18.8740	13.2689	West	Climbing	1500	900	2	500, W	500, W	-	Edge of escarpment	-
-20.5871	14.0608	East	Climbing	100	150	1	100, E	1250, NE	-	Inselberg in complex topography	-
-26.7458	17.4004	West	Climbing?	500	500	1	1300, NW	1300, NW	-	Edge of the escarpment	-
-27.6992	16.4222	East	Climbing	2000	3000	4	1500, NE	1500, NE	6000, S	Inselberg complex	-
-27.6252	16.1969	East	Climbing	500	1000	2	1100, E	2000, NW	2000, W	Inselberg complex	-
-28.0320	16.8373	West	Climbing	200	350	1	1000, W	2500, W	20500, W	Complex	Rosh Pinah
-27.0564	16.3624	East	Climbing	1000	1000	2	1500, E	2750, N	13000, SW	Inselberg complex	-
-26.8003	17.4026	West	Climbing	400	450	1	1000, W	1500, E, 2000, W	-	Edge of the escarpment	Simplon
-27.7705	16.4127	East	Climbing	600	2500	3	-	-	0, S	Inselberg	-
-27.3119	15.8297	South	Climbing	650	1000	1	-	-	0, all	Inselberg	-
-27.2275	15.7260	East	Climbing	700	700	1	-	-	0, E	Inselberg	-

Latitude	Longitude	Orientation /Face	Climbing/ Falling	Length	Width	Class	Distance & direction to channel (m)	Distance & direction to major channel	Distance & direction to dune field	Accommodation Space	Studied
										complex	
-27.1790	16.3497	West	Falling	200	150	1	-	1500, N	9000, W	Inselberg complex	-
-27.1300	15.9841	East	Climbing	1400	1500	1	-	-	0, SE	Inselberg	-
-27.1089	15.9622	East	Climbing	600	300	1	550, S	-	1000, E & S	Inselberg	-
-27.0798	15.7623	East	Climbing	200	300	1	-	-	5500, E	Inselberg	-
-22.0670	15.1579	East	Climbing	150	1000	1	150, SE	750, S	-	Inselberg	-
-25.3535	15.7952	East	Climbing	2500	3000	3	1000, SE	2500, SE	14500, W	Inselberg complex	-
-24.7470	15.8808	South	Climbing	450	550	1	200, N	3000, N	9000, W	Inselberg complex	-
-25.8103	16.2780	East	Climbing	500	1500	2	500, SE	4000, SE	21000, NW	Inselberg complex	-
-26.7259	16.1575	East	Climbing	1000	2000	3	650, E	1500, NE	1000, NW	Inselberg	Klein Aus
-22.0177	16.1532	East	Climbing	800	2000	1	150, NE	500, NE	-	Inselberg	-
-26.8608	16.2049	East	Climbing	350	900	2	800, NE	1000, NE	10500, SW	Inselberg	-
-24.4135	15.8592	North	Climbing	1500	850	3	1200, NE	1750, NE	12000, W	Inselberg complex	-
-26.5967	15.3992	East	Climbing	350	500	2	-	-	800, E	Inselberg complex	-
-25.9649	16.0323	All	Climbing and Falling	-	-	1	820, SE	2500, SE	5000, S & W	Inselberg	-
-24.7163	16.0500	East	Climbing	150	250	1	250, NE	250, NE	29000, W	Inselberg	Neuhof 1
-25.6308	16.0559	East	Climbing	3000	1500	3	300, NE	2000, NE	6000, S	Inselberg complex	-
-25.9279	16.0510	East	Climbing	1500	1200	3	1000, E	2250, NE	0, N	Inselberg complex	-
-24.9317	16.0318	East	Climbing	500	1100	2	50, E	1500, NE	11500, W	Inselberg	Keerwader
-24.2881	15.8706	East	Climbing	750	1750	3	200, SE	500, E	6000, NW	Inselberg	-
-24.7376	16.0814	East	Climbing	300	800	2	100, NE	1500, NE	29000, W	Inselberg	-
-24.2485	15.9177	East	Climbing	700	1000	3	200, N & S	1250, NE	4000, NW	Inselberg	-

Latitude	Longitude	Orientation /Face	Climbing/ Falling	Length	Width	Class	Distance & direction to channel (m)	Distance & direction to major channel	Distance & direction to dune field	Accommodation Space	Studied
-26.6779	16.0986	East	Climbing	2000	3400	4	50, E	3500, NE	0, E	Inselberg complex	Aus and Bertram (2003)
-24.7470	15.7998	East	Climbing	600	450	2	800, E	1500, E	7500, W	Inselberg complex	-
-24.9763	15.9342	East	Climbing	2500	1500	4	5000, E	5000, E	0, S & W	Inselberg	Jagkop
-25.3544	15.6919	East	Climbing	500	650	2	100, E	3000, E	10000, W	Inselberg complex	-
-26.4175	16.0816	East	Climbing	1120	900	3	800, E	-	1300, NE	Inselberg complex	-
-26.9618	16.1935	East	Climbing	2000	2000	3	1000, NE	3500, N	5000, NW	Inselberg complex	-
-25.6199	15.7780	East	Climbing	160	375	1	1200, SE	3500, SE	10500, W	Inselberg	-
-26.3757	15.9397	East	Climbing	600	900	1	1300, S & E	1500, S	18000, NE	Inselberg	-
-25.2273	15.9495	East	Climbing	1500	900	3	600, SE	1500, SE	4750, E	Inselberg complex	-
-25.5107	15.5622	East	Climbing	1150	900	3	-	-	0, All	Inselberg complex	-
-22.3611	15.2652	East	Climbing	1500	1000	3	4000, SE	4000, SE	-	Escapment+Inselberg	-
-23.8023	15.8989	East	Climbing	2500	3500	4	2000, NE	3000, E	2500, SW	Inselberg complex	Samara
-26.9523	16.3242	East	Climbing	100	500	1	500, W	2000, S, 3000, E	13000, W	Inselberg	-
-24.7237	15.8073	East	Climbing	200	400	2	-	-	5500, SW	Inselberg complex	-
-22.4498	15.2273	East	Climbing	250	350	1	150, SE	500, SE	-	Inselberg	-
-26.4493	16.5292	East	Climbing	1000	2000	3	2500, SE	3750, E,S,W	5500, NW	Inselberg	Sandkop and Bertram (2003)
-24.7373	15.7902	East	Climbing	1350	1500	3	1500, E	-	3750, S	Inselberg complex	-
-25.6678	16.1089	East	Climbing	800	2500	2	250, E	2000, N	5000, SW	Inselberg complex	-
-25.7227	16.1341	East	Climbing	200	500	1	500, NE	2500, E	5000, W	Inselberg	-
-24.3075	15.8360	East	Climbing	600	600	3	200, SE	2000, SE	4000, NW	Inselberg complex	-

Latitude	Longitude	Orientation /Face	Climbing/ Falling	Length	Width	Class	Distance & direction to channel (m)	Distance & direction to major channel	Distance & direction to dune field	Accommodation Space	Studied
-25.2226	15.5454	East	Climbing	1000	1000	1	-	-	0, All	Inselberg	-
-25.8671	16.2103	East	Climbing	900	1200	2	250, SE	250, SE	17000, W	Inselberg complex	-
-25.4143	15.2710	East	Climbing	700	1200	2	-	-	0, All	Inselberg complex	-
-25.5073	15.8986	East	Climbing	1000	3750	2	50, E	100, E	6000, E	Inselberg complex	-
-24.6603	15.7170	East	Climbing	500	400	2	850, NE	1000, SE	7000, N & S	Inselberg	-
-26.4253	15.4912	East	Climbing	750	700	1	-	-	4000, W	Inselberg complex	-
-24.3265	15.8426	East	Climbing	600	1400	2	600, SE	1250, SE	6500, NW	Inselberg complex	-
-25.4277	15.6366	East	Climbing	1500	2000	2	650, E	3500, E	6700, W	Inselberg complex	-
-25.7399	16.1328	East	Climbing	100	150	1	600, SE	600, SE	5000, W	Inselberg	-
-26.1401	16.4632	East	Climbing	450	1150	2	400, NE	400, NE	10000, SW	Inselberg complex	Bertram (2003)
-24.6956	16.0893	South	Climbing	350	200	1	50, S	300, S	32000, W	Inselberg	-
-26.8706	16.2079	West	?	200	200	1	900, SE	1750, S	8000, SW	Inselberg	-
-25.7514	15.7326	West	Climbing	200	250	1	400, W	700, W	1000, W	Inselberg	-
-26.2586	15.4353	West	Climbing	900	2000	1	-	-	0, W	Inselberg	-
-24.6677	15.7127	West	Climbing	400	400	1	350, W	2500, NW	7500, NW	Inselberg	-
-24.0052	15.8978	East	Climbing	1500	2750	2	650, E	1500, N	11500, SW	Inselberg	-
-24.7265	15.8002	West	Climbing	150	120	1	100, W	1000, N	7000, W	Inselberg complex	-
-26.2246	16.4572	All	Climbing and Falling	-	-	2	300-1000, All	3500, S	1800, W	Inselberg complex	Bertram (2003)
-24.7376	16.0781	West	Climbing	200	150	1	120, SW	120, SW	21500, SW	Inselberg	-
-26.3654	15.8809	West	Climbing	180	450	1	500, W	2500, SW	17000, N		-
-25.2178	16.2725	East	Climbing	800	1500	2	1000, NE	3000, E	19000, W	Inselberg complex	-
-26.3762	15.9331	West	Falling	250	450	1	480, W	2000, S	18000, NE		-

Latitude	Longitude	Orientation /Face	Climbing/ Falling	Length	Width	Class	Distance & direction to channel (m)	Distance & direction to major channel	Distance & direction to dune field	Accommodation Space	Studied
-25.6676	16.0159	West	Climbing	420	350	1	700, NW	-	2500, S	Inselberg	-
-24.4992	15.7949	East	Climbing	600	1750	2	50, NE	2500, S	4000, NW	Inselberg	-

APPENDIX D: SUP. INFO. – PAPER 3 (CHAPTER 6)

Table D-1 Sediment properties of the individual Klipkraal Sands samples. *b*= brown, *db* = dark brown, *sb* = strong brown, *pb*= pale brown, *yr* = yellowish red. *FS* = Fine sand, *VFS* = Very fine sand, *VCS* = Very coarse silt, *M* = Moderately sorted, *P* = Poorly sorted

Sample	Depth (m)	Munsell colour	Description	Mean particle size (µm)	Sorting (phi)	Skewness	Kurtosis	Coarse sand (%)	Medium sand (%)	Fine sand (%)	Silt (%)	Clay (%)	Clasts >2mm (g)	Notes	LOI at 550°C (%)	Carbonate content (%)
KAR13/1/1	0.3	7.5YR 4/4 - b	FS, P	182	1.18	-0.24	1.19	6	31	53	10	1	n/a	n/a	0.88	0.47
KAR13/1/2	0.6	7.5YR 3/2 - db	VFS, P	118	1.54	-0.24	1.29	3	18	55	22	2	0.05	Angular <3mm	1.41	0.61
KAR13/1/3	1.1	7.5YR 4/6 - sb	FS, P	149	1.31	-0.20	1.29	4	22	59	14	1	0.05	Angular <3mm	0.87	0.51
KAR13/1/4	1.6	5YR 4/6 - yr	VFS, P	94	1.80	-0.24	1.02	4	18	43	33	1	n/a	n/a	1.43	1.03
KAR13/1/5	3.2	7.5YR 4/6 - sb	VFS, P	73	1.96	-0.16	1.11	4	13	40	41	2	n/a	n/a	1.69	1.14
KAR13/2/1	2.4	7.5YR 5/6 - sb	FS, P	146	1.28	-0.22	1.54	3	20	65	12	1	0.17	Angular <1cm	0.62	0.48
KAR13/2/2	3.5	5YR 4/6 - yr	FS, P	163	1.62	-0.37	1.36	8	30	45	16	1	n/a	n/a	0.71	0.59
KAR13/2/3	4.7	10YR 6/3 - pb	FS, P	145	1.57	-0.17	1.19	8	23	50	18	1	0.14	Angular <5mm	0.60	0.37
KAR13/2/4	5.5	5YR 4/6 - yr	VFS, P	72	2.12	-0.15	1.04	4	16	35	42	3	n/a	n/a	1.83	1.15
KAR13/3/1	1	7.5YR 3/2 - db	VCS, P	45	1.75	-0.27	1.43	1	4	38	54	4	0.39	Angular <5mm	3.63	1.29
KAR13/3/2	3	7.5YR 3/2 - db	VCS, P	63	1.93	-0.24	1.20	2	10	44	41	3	n/a	n/a	1.82	0.91
KAR13/4/1	0.5	7.5YR 4/4 - b	FS, M	181	0.84	-0.09	1.02	2	27	65	5	0	0.45	Angular <8mm	0.92	0.49

