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Long-term controls on continental-scale bedrock river terrace deposition from integrated clast and heavy mineral assemblage analysis: an example from the lower Orange River, Namibia

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Abstract

Establishing relationships between the long-term landscape evolution of drainage basins and the fill of sedimentary basins benefits from analysis of bedrock river terrace deposits. These fragmented detrital archives help to constrain changes in river system character and provenance during sediment transfer from continents (source) to oceans (sink). Thick diamondiferous gravel terrace deposits along the lower Orange River, southern Namibia, provide a rare opportunity to investigate controls on the incision history of a continental-scale bedrock river. Clast assemblage and heavy mineral data from seven localities permit detailed characterisation of the lower Orange River gravel terrace deposits. Two distinct fining-upward gravel terrace deposits are recognised, primarily based on mapped stratigraphic relationships (cross-cutting relationships) and strath and terrace top elevations, and secondarily on the proportion of exotic clasts, referred to as Proto Orange River deposits and Meso Orange River deposits. The older early to middle Miocene Proto Orange River gravels are thick (up to 50 m) and characterised by a dominance of Karoo Supergroup shale and sandstone clasts, whereas the younger Plio-Pleistocene Meso Orange River gravels (6-23 m thick) are characterised by more banded iron formation clasts. Mapping of the downstepping terraces indicates that the Proto gravels were deposited by a higher sinuosity river, and are strongly discordant to the modern Orange River course, whereas the Meso deposits were deposited by a lower sinuosity river. The heavy minerals present in both units comprise magnetite, garnet, amphibole, epidote and ilmenite, with rare titanite and zircon grains. The concentration of amphibole-epidote in the heavy minerals fraction increases from the Proto to the Meso deposits. The decrease in incision depths, recorded by deposit thicknesses above strath terraces, and the differences in clast character (size and roundness) and type between the two units, are ascribed to a more powerful river system during Proto-Orange River time, rather than reworking of older deposits, changes in provenance or climatic variations. In addition, from Proto- to Meso-Orange River times there was an increase in the proportion of sediments supplied from local bedrock sources, including amphibole-epidote in the heavy mineral assemblages derived from the Namaqua Metamorphic Complex. This integrated study demonstrates that clast assemblages are not a proxy for the character of the matrix, and vice versa, because they are influenced by the interplay of different controls. Therefore, an integrated approach is needed to improve prediction of placer mineral deposits in river gravels, and their distribution in coeval deposits downstream.

Key words: heavy minerals, gravel terraces, drainage basin, source-to-sink, Orange River, clast assemblage

1. Introduction

Constraining the dynamics of long-term landscape evolution requires analysis of the coeval downstream stratigraphic record (e.g., Morton, 1991; Dickinson and Gehrels, 2003; Mange and Otvos, 2005; Bhattacharya et al., 2016; Romans et al., 2016). However, environmental signals (e.g., climate, tectonic uplift) and provenance signatures are modified during sediment transfer from continents (source) to oceans (sink) through the sediment transfer zone (Romans et al., 2016). Terrace deposits within bedrock river systems provide a fragmented archive of landscape evolution in the sediment transfer zone (e.g., Bridgland and Westaway, 2008; Wegmann and Pazzaglia, 2009). Therefore, improved understanding of these records in sites dominated by erosion will help to constrain controls on long-term changes in ancient river system character and provenance, and to predict and unravel the downstream depositional record of quasi-contemporaneous marine sediments (Pazzaglia and Gardner, 1993; Aalto et al., 2008; Marsaglia et al., 2010; Kuehl et al., 2016). Analysis of clast assemblages is the most common approach used to investigate changes in provenance of fluvial gravels and to establish the denudation and evolution of drainage basins (Gibbard, 1979; Green et al., 1982; Bridgland, 1999). An alternative technique is the use of heavy minerals, because they are more physically and chemically resilient than many clasts, and may survive multiple phases of weathering and transport (Hassan, 1976; Morton, 1984, 1991; Goodbred et al., 2014).

Most drainage reconstruction studies have either used clast assemblage analysis (Gibbard, 1979; Dowdeswell et al., 1985; Bridgland, 1999; Jones, 2000; Mikesell et al., 2010) or heavy mineral assemblages (Uddin et al., 2007; Morton et al., 2011). Both techniques are problematic. Clasts derived from mechanically or chemically unstable bedrock might be preferentially degraded owing to abrasion during transport or chemically weathered post deposition (Green et al., 1980), which hinders accurate fingerprinting of source areas. Heavy mineral studies also contain inherent weaknesses (Smale and Morton, 1987; Dill, 1994; Morton and Hallsworth, 1999, 2007; Faupl et al., 2007; Uddin et al., 2007; Tsikouras et al., 2011; Wong et al., 2013; do Nascimento et al., 2015; Caracciolo et al., 2016; Krippner et al., 2016). For example, the relatively high density of heavy minerals may restrict their transport distance (Komar and Wang, 1984; Komar, 2007). Maher et al. (2007) present a rare example of combining clast assemblage and heavy minerals analysis to reconstruct a drainage capture event of the Rio Alias, southeast Spain.

In this study, we aim to integrate clast assemblage and heavy mineral signatures within a critical part of source-to-sink systems, the sediment transfer zone, where a depositional record is found within sites dominated by erosion, to provide information on sediment transport, bypass, deposition, and provenance controls. The lower Orange River, southern Namibia, was chosen because it is a rare example of a continental-scale bedrock river with a well-constrained drainage basin geology, and accessible, extensive, and unlithified gravel terrace deposits owing to the arid climate and active mining operations (Fig. 1). Furthermore, the gravel terrace deposits represent multiple cycles of degradation and aggradation, allowing investigation into changing controls through time. Finally, the coeval marine gravels offshore southern Namibia host economic diamond deposits, and therefore an improved understanding of the drainage history of the lower Orange River can feed into revised offshore exploration strategies. Specific objectives are i) to reconstruct the drainage history of the lower Orange River using two river terrace deposits, ii) to investigate extrinsic and intrinsic controls on the clast assemblage and heavy minerals assemblage, and iii) to evaluate the value of a combined approach to understanding continental-scale bedrock river evolution.

2. Geological Setting and Geomorphology

2.1 Geological Setting

The Orange River and its major tributary, the Vaal River, are the main bedrock confined rivers in a ~10^6 km^2 catchment in southern Africa (Garzanti et al., 2014). The geology exposed in the catchment is highly variable. In the east, geology comprises the Archaean Kaapvaal Craton (de Wit et al., 1992) intruded by Cretaceous and older diamondiferous kimberlites (de Wit, 1999; Shirey et al., 2001; Moore and Moore, 2004). The upper Orange River traverses rocks of the extensively eroded Permo-Carboniferous to Jurassic Karoo Supergroup (Visser, 1993; Johnson et al., 1997; Catuneanu et al., 1998, 2005; Key et al., 1998; Bangert et al., 1999). Between Noordoewer (300 km east of the Orange River mouth) and Oranjemund (Fig. 2), the lower Orange River cuts through the Mesoproterozoic Namaqua Metamorphic Complex (Thomas et al., 1994; Jacobs et al., 2008) before incising the Neoproterozoic Gariep Belt (Frimmel and Frank, 1998; Frimmel et al., 2004) close to the river mouth on the Atlantic Ocean coast (Fig. 2). The Namaqua Metamorphic Complex forms the basement of the area. The Gariep Belt, which also extends into northwestern South Africa (Fig. 2), comprises mainly metamorphosed rocks, including chert, quartzite, meta-greywacke, metapelite and metadiamictite (Frimmel et al., 1996; Frimmel and Frank, 1998; Basei et al., 2005). The mineralogy of the Namaqua Metamorphic Complex rocks and Gariep Belt rocks is summarised in Table 1. Around the Noordoewer area, the Ediacaran to early Cambrian Nama Group, a foreland basin succession (DiBenedetto and Grotzinger, 2005; Grotzinger et al., 2005; Grotzinger and Miller, 2008) caps the Namaqua Metamorphic Complex basement (Fig. 2). These rocks are possible sources of sediment in the Orange River terrace deposits. Along the lower Orange River, three distinct terrace deposits are recognised based on terrace elevation, bedrock strath level and exotic clast suite, which Jacob (2005) informally termed, in stratigraphic order, Pre-Proto Orange River, Proto Orange River, and Meso Orange River deposits. This nomenclature has been adopted in the present study. Here, we concentrate on Proto and Meso Orange River gravels in terms of the clast assemblage and heavy minerals.

Eocene marine gravel is the oldest Orange River-derived sediments on the west coast of Namibia and is preserved at 160 m above present-day sea level (Stocken, 1978). However, no equivalent Eocene-age gravel is preserved in the lower Orange River. The age of Pre-Proto Orange River deposits remains unknown. Dating of coarse grained fluvial terraces is challenging due to lack of continental biostratigraphy. The Proto Orange River suite has been dated as early to middle Miocene, using macrofauna fossils, including Lopholistriodon moruoroti found in gravel terrace deposits at Auchas and Arrisdrif of the lower Orange River (Corvinus and Hendey, 1978; Hendey, 1978; Pickford, 1987; Pickford and Senut, 2002) (Fig. 1). The Meso Orange River gravel suite has not been dated due to lack of macrofauna fossils, but is inferred to be Plio-Pleistocene (2-5 Ma) in age based on correlations with littoral beach gravel deposits (Pether, 1986).

2.2 Geomorphology

2.2.1 Regional Geomorphology

Over the last 66 Ma, the southern African landscape has been shaped by tectonics, climate and geomorphic processes (Knight and Grab, 2016a) although feedbacks produced by tectonics and climate are often difficult to isolate (Knight and Grab, 2016b). Periods of uplift and associated increased erosion in southern Africa include during the late Cretaceous (de Wit, 1999; Stevenson and McMillan, 2004; Richardson et al., 2017), the Miocene and the Pliocene (Partridge and Maud, 2000; Green et al., 2017). Alternatively, van der Beek et al. (2002) propose that the topography of the southeast African margin is a result of a thin elastic lithosphere (~10 km). Evidence of major Cretaceous uplift is recorded offshore where sediment supply rates in the Orange Basin offshore Namibia and South Africa (Rust and Summerfield, 1990; Aizawa et al., 2000; Rouby et al., 2009) and the Outeniqua Basin, offshore South Africa (Tinker et al., 2008a; Sonibare et al., 2015) show a significant increase. There is a general consensus that erosion rates have decreased from the Cretaceous to the present (Richardson et al., 2017), as shown by apatite fission track denudation (Brown et al., 1999; Tinker et al., 2008b; Wildman et al., 2015) and cosmogenic dating evidence (Fleming et al., 1999; Cockburn et al., 2000; Bierman et al., 2014).

During the Miocene, southeastern Africa underwent a maximum uplift of 250 m, almost twice that of the western subcontinent (150 m) (Partridge and Maud, 2000). This is in agreement with Hanson et al. (2009), who estimated high erosion rates for the Monastery kimberlite pipe (~1350 m) in eastern South Africa, relative to the Kimberley and Koffiefontein pipes (~850 m) in central South Africa. However, there is also a possibility that the eastern subcontinent might have already been relatively more elevated than the western subcontinent prior to uplift (Roberts and White, 2010; Richardson et al., 2016). Apatite fission track studies have estimated 2.5 to 3.5 km of land surface erosion for the late Cretaceous (Brown et al., 1999; Gallagher and Brown, 1999; Tinker et al., 2008b; Decker et al., 2013; Wildman et al., 2015; Green et al., 2017), and that the uplift events increased the erosive power of rivers in southern Africa.

The central part of southern Africa is marked by a low relief elevated central plateau (> 1000 m above mean sea level) whereas the coastal margins along the Indian and Atlantic Oceans are characterised by a high relief low elevation coastal plain (Knight and Grab, 2016a). The two are separated from each other by the Great Escarpment (Gallagher and Brown, 1999), which occurs between 50-200 km inland from the coast (Partridge and Maud, 1987, 2000; Partridge et al., 2010). In addition to uplift, rivers have also played an important role in shaping the southern African landscape. The Orange River is one of the major drainage systems in southern Africa, and with its many tributaries, has played a major role in shaping the landscape since the late Mesozoic. According to Jacob (2005), the Orange River deeply incised the landscape (between 600-1000 m deep) following Cretaceous uplift. However, contrasting views regarding the evolution and development of the Orange River fluvial system remain (Jubb, 1964; Dingle and Hendey, 1984; Skelton, 1986; de Wit, 1999; de Wit et al., 2000).

2.2.2 Geomorphology of the lower Orange River

Outcrops of both Namaqua Metamorphic Complex and Gariep Belt rocks together with the Orange River make up the main geomorphic features in the study area. The area between Noordoewer and the Orange River mouth is characterised by a low relief coastal plain and high relief inland area. High relief in the area is a product of the resistant lithologies that comprise Namaqua Metamorphic Complex rocks (Fig. 2). Ephemeral tributaries to the lower Orange River include the Gamkab River, Fish River and Boom River. From Noordoewer towards the river mouth, the palaeo-Orange River (early to middle Miocene) valley widens from 550 m to 2300 m, and its gradient decreases downstream (from 0.87 m/km to 0.38 m/km) with an average gradient of 0.69 m/km (Jacob, 2005) (Fig. 3A).

3. Methods

3.1 Terrace mapping, river profiles and gradients

Mapping flights of downward-stepping terrace surfaces on the northern and southern banks of the Orange River was performed using a handheld Global Positioning System (GPS). In the field, two stratigraphically distinct fining-upward terrace deposits were differentiated based on their bedrock strath terrace height and elevation of terrace deposit, the mapped palaeo-river course and overall geometry (Fig. 3), and where terrace deposits are in contact, their cross-cutting relationships. Given the bedrock river valley setting, the older and higher terraces (the Proto) are less continuous and more dissected than the younger terrace deposits (the Meso) that are lower in elevation (Jacob et al., 1999). In general, this elevation difference is also recorded by the height of the strath terrace, although deep scours on the Proto strath can be lower in elevation than nearby Meso strath terraces. Four planform types of terrace deposits are present in the study area: i) cut-off meander loops, ii) unpaired terraces, iii) terraces preserved downstream of the tributary input points, and iv) paired terraces. The cut-off meander loop terraces are the thickest, and are exclusive to the Proto Orange River deposits, whereas paired terraces are only observed in the Meso Orange River deposits. The gradient of the Proto and Meso Orange River terrace profiles (Fig. 3A) were calculated using the strath terrace height and terrace deposit height recorded in meters above sea level (Fig. 3A). Overall, both the Proto and Meso

Orange River terrace deposits form downstream thickening and fining wedges (Jacob et al., 1999) coincident with the widening of the Orange River valley (Fig. 3C).

3.2 Clast Analysis

Clast analysis was performed on the Orange River gravel terrace deposits flanking the modern lower Orange River on the Namibian side of the river. These are the Proto Orange River deposits and the stratigraphically younger Meso Orange River deposits. The clast assemblage analyses were undertaken in order to characterise the gravel deposits of different ages according to the assemblages of exotic clasts.

Fourteen and twelve samples were analysed from the Proto and Meso Orange River deposits, respectively. To avoid bias towards clasts that are resistant to surface weathering, the surface gravel was avoided. Sampling was completed by excavating a 2 m x 2 m area of gravel from the head of an in-channel bar deposit, which is the most stable part of a gravel bar (Li et al., 2014). After excavation, the gravel was screened on site through stacked sieves, which split the clasts into +300 mm, +200 mm, +90 mm, +40 mm, +25 mm, +16 mm, +8 mm and +3 mm. The clasts were split further with a sample splitter until the desired number of clasts was attained per size fraction. The size fraction below 3 mm was retained for heavy mineral analysis. A minimum of 50 clasts was inspected in the +300 mm, +200 mm, +90 mm, +40 mm and +25 mm size fractions, and a minimum of 100 clasts was analysed for the +16 mm, +8 mm and +3 mm size fractions. In total, 7700 and 6600 clasts were analysed for the Proto and Meso deposits, respectively. A minimum of 50 clasts was analysed for the coarse size fractions (> 25 mm) given the large volume of sample needed. Similar studies have used a minimum of 100 clasts although they have not indicated the size fractions analysed (Jones, 2000; de Carvalho Faria Lima Lopes et al., 2016).

Lithology, clast shape and clast roundness were recorded for each individual clast. Clast roundness, which is a proxy of distance travelled and lithology durability, was visually estimated using the roundness chart developed by Powers (1953).

3.3 Heavy Mineral Analysis

Heavy minerals were recovered from the smaller than 3 mm sand samples. Fourteen and twelve samples were analysed for the Proto and Meso Orange River deposits, respectively. To make the separation of heavy minerals from the light minerals and the rest of the sand more effective, the bulk samples were first sieved into 2-4 mm, 1-2 mm, 0.5-1.0 mm, 0.25-0.50 mm, 0.125-0.250 mm, 0.063-0.125 mm and below 0.063 mm size fractions using an automatic electrical sieve shaker.

Heavy minerals were separated from the rest of the sample material using a Met-Solve Analytical Table, a flowing film gravity separator which produced a heavy mineral concentrate. Only the 1-2 mm, 0.5-1.0 mm and 0.25-0.50 mm size fractions were processed on the gravity settling table for heavy mineral recovery because very few heavy mineral particles were observed in the coarser fractions. The selected size fractions were processed at 1° slope angle, 1.5 litres/minute water flow rate and 60 strokes/minute deck rocking speed. The 1-2 mm size fraction was also processed at a slope angle of 1°, as opposed to the manufacturer recommended steeper angle of 2°(Met-Solve, 2016) because even a slope angle of 1.5° proved too steep for retention of sub-rounded and rounded garnets, the dominant heavy mineral in this size fraction. A single concentrate was produced for each size fraction because the heavy minerals recovered from the gravels have overlapping densities. The heavy mineral concentrates were dried and weighed.

Heavy mineral proportions were determined by counting a minimum of 300 grains per size fraction per sample under a binocular microscope following the methodology of Dill (1998), Faupl et al. (2007), Scheneiderman and Chen (2007), Garzanti et al. (2015) and Krippner et al. (2016). The sample was reduced in volume by coning and quartering to generate a sub-sample of 300 grains per sample. This equates to a total of 4200 and 3600 grains analysed for Proto and Meso samples, respectively. Magnetite was removed using a hand magnet, and grain counts according to mineral type were undertaken on the remaining sub-sample. Minerals that could not be identified visually on the microscope were mounted on polished epoxy blocks (n = 10, 16, 2, 4, 1, for garnet, amphibole, epidote, titanite and zircon, respectively) and identified using the EDS facility of FEI Quanta FEG 650 Scanning Electron Microscope (SEM), at the University of Leeds using a 20 kV accelerating voltage and 5 nm spot size. Representative heavy mineral grains of garnet, magnetite and epidote were mounted on double sided adhesive tape attached to a metal plate for grain surface texture analysis. The grains were coated with a thin layer of iridium (~2 nm). The grains were examined in secondary image mode using the same SEM. Garnet composition was analysed with JEOL JXA8230 electron microprobe at the University of Leeds under operating parameters of 20 kV accelerating voltage, 30 nA beam current, 30 seconds on-peak count time and 15 seconds off-peak count time.

4. Results

4.1 Stratigraphic relationship and gravel terrace characterisation

Fluvial gravel terrace successions, such as those deposited on bedrock flanking the lower Orange River (Fig. 1) form during a cycle of incision then aggradation (e.g., Bridgland and Westaway, 2008). During incisional phases, the palaeo-Orange River cut into the bedrock, locally forming deep scours (10-30 m) below the bedrock strath terrace level (Figs. 3A, 4). Gravel deposition above strath terraces occurs during aggradational phases that are driven by the combined effects of decreased stream power and increased sediment supply (e.g., Blum and Törnqvist, 2000). The location of scours are coincident with the outside of meander bends, changes in bedrock lithology, structural features (e.g., joints), and tributary input points. The scours underlying the Proto Orange River deposits are deeper (average depth 10-30 m), than those beneath the younger Meso Orange River deposits, and have a maximum depth of 40-50 m (Fig. 3A). For the Meso Orange River deposits, their underlying scours show varying depths but have a maximum depth of 35-40 m (Fig. 3A). Some scours that underlie Meso deposits might have been formed during earlier phases of incision given that the Orange River is a superimposed river. Multiple cycles of bedrock incision and aggradation resulted in flights of downward-stepping strath surfaces and dissected overlying terrace deposits (Fig. 4A). Although the thickness of the terrace deposits varies between deposit sites, the Proto deposits are thicker (up to 50 m thickness) than the Meso deposits (6-23 m) (Fig. 3A). In the study area, the bedrock river valley widens downstream from 1300 m between Boom and Lorelei to 2340 m between Sendelingsdrif and the river mouth. The widening of the river valley in downstream reaches of the river has enhanced the preservation of terrace deposits (Fig. 1).

In terms of river courses, the Proto Orange River had a higher sinuosity than the Meso Orange River, which had a course similar to that of the modern Orange River (Fig. 3). The higher sinuosity Proto Orange River course is supported by the preservation of cut-off meander loop terraces (Fig. 3). The depth of incision (Figs. 3A, 4B, 4C), size of imbricated clasts (Fig. 5A) and height of coarse grained cross bedding (Fig. 5B) suggest a high energy river system during the Proto incision and aggradation cycle than during the Meso (e.g., Dott and Bourgeois, 1982).

4.2 Gravel Characterisation

The overall makeup of the gravel is a combination of both exotic and locally derived clasts, with the large cobble size fractions (> 25 mm) dominated by quartzite clasts. Exotic clasts include agate (Fig. 6A), Karoo Supergroup shales and sandstones (Fig. 6B), Karoo Supergroup basalt and banded iron formation (BIF) (Fig. 6C). These clasts are derived from the Orange River catchment area. The relative abundance of each clast in a given gravel deposit is related to the timing and geomorphic evolution of the Orange River drainage basin.

4.2.1 Clast Assemblage

Size fractions 16-25 mm, 8-16 mm and 3-8 mm are reported (Fig. 7) because these contain prominent distinctions between the stratigraphically-distinct Proto and Meso Orange River gravels in terms of key exotic clasts. The Proto Orange River gravel terrace signature is characterised by a dominance of Karoo Supergroup shales and sandstones among the exotic clasts (Figs. 6B, 7). The exotic clast suite of Meso Orange River gravels is dominated by BIF relative to other exotic clasts (Figs. 6C, 7). For example, in the 16-25 mm size fraction, Karoo Supergroup sedimentary rock clasts constitute 22% and 7% of the Proto Orange River and Meso Orange River gravels, respectively, and BIF is 6% in the Proto Orange River gravel and 10% in the Meso Orange River gravel (Fig. 7). At Auchas Major, the Meso Orange River gravel has an uncharacteristic abundance of Karoo shales and sandstones (Fig. 7).

Another feature of Meso Orange River gravel is the presence of Karoo Supergroup basalt clasts (Fig. 7), sourced from the early Jurassic (190-183 Ma) Drakensberg Flood Basalts (Duncan et al., 1997; Marsh et al., 1997; Jacob, 2005; Jourdan et al., 2007), but these are rare in the older gravels (Fig. 7). Feldspar clasts were recorded in the small size fractions (8-16 mm and 3-8 mm) in both Proto and Meso Orange River gravels (Figs. 6D, 7).

4.2.2 Clast Roundness

The lithology of a clast and the distance it travels before deposition is reflected in the degree of rounding (Lindsey et al., 2007; Miao et al., 2010). Proto Orange River gravels show a higher degree of rounding than the Meso Orange River gravels (Fig. 8). For size fractions smaller than 40 mm, clast roundness decreases exponentially with decreasing clast size in both the Proto and Meso Orange River gravels (Fig. 8).

4.3 Heavy Mineral Assemblages of the Proto Orange River and Meso Orange River gravels

The heavy minerals present in the Proto and Meso Orange River gravels are magnetite, garnet, amphibole, epidote and ilmenite. Titanite, and zircon are present in trace amounts. Figures 9 and 10 illustrate the relative abundance of individual heavy minerals within the overall heavy fraction according to locality and gravel stratigraphy. In plotting the heavy mineral assemblages, the lower density minerals amphibole (2.97-3.13 g/cm^3) and epidote (3.3-3.6 g/cm^3) have been grouped together, because they have similar chemical stabilities (Morton and Hallsworth, 2007; Andò et al., 2012). These are referred to as amphibole-epidote throughout.

The Proto Orange River gravel shows relatively higher magnetite and ilmenite contents than the Meso Orange River gravel for the 0.5-1.0 mm and 0.25-0.50 mm size fractions (Figs. 9, 10). Most of the garnets in the Proto Orange River gravel are in the coarsest size fraction such that garnet abundance decrease by more than half from the coarse (1-2 mm) to the fine size fraction (0.25-0.50 mm) (Fig. 9). In contrast, in the Meso Orange River deposits, garnet reduces gradually from the coarse to the fine size fraction (Fig. 9). For example, at Arrisdrif, garnet content reduces from an average of 89% of the total heavy mineral in the 1-2 mm size fraction to 30% in the 0.25-0.50 mm, whereas in the Meso Orange River gravel it changes from 34% to 26%, respectively (Fig. 9).

The Meso Orange River samples are characterised by a relative higher abundance of amphibole-epidote than the Proto Orange River samples (Figs. 9, 10). The distinction between the Proto and Meso Orange River deposits in terms of amphibole-epidote content is clear at Arrisdrif, Auchas Lower, Daberas, Lorelei and Boom (Figs. 9, 10). However, at Auchas Major and Sendelingsdrif, the Meso Orange River samples have low amphibole-epidote content that is similar to the Proto Orange River samples (Figs. 9, 10). In the Meso Orange River gravel, amphibole-epidote content increases downstream from Boom to Arrisdrif, whereas magnetite decreases downstream most especially for the 0.5-1.0 mm size fraction (Fig. 10B). However, neither trend is observed in the Proto Orange River gravel (Fig. 10). At Boom, for example, the average amphibole-epidote:magnetite ratio of the Meso Orange River sample is 0.3 in the 0.5-1.0 mm size fraction, whereas farther downstream at Arrisdrif it is 0.96 in the same size fraction (Fig. 10B).

There is no difference in the range of grain surface textures on garnets between the Proto and Meso Orange River gravels. Conchoidal fractures and etch pits were recorded on both units (Fig. 11A, B). Magnetite shows a much lower degree of dissolution textures compared to garnet (Fig. 11C, D). Etch pits are present but rare. Epidote shows much more extensive chemical etching relative to garnet and magnetite in both the Proto and Meso Orange River deposits (Fig. 11E, F). Saw-tooth terminations are present on Meso Orange River epidotes (Fig. 11F) but none was recorded in the Proto Orange River gravels.

4.4 Distinction between Proto and Meso Orange River deposits on basis of clast and heavy mineral assemblages

There is a clear distinction between the Proto and Meso Orange River gravels, at both clast and heavy mineral scales (Figs. 12, 13). The Proto Orange River gravel is characterised by a high percentage of Karoo shales and sandstones and low amphibole-epidote content, and the Meso Orange River gravel is characterised by high BIF and amphibole-epidote content (Figs. 12, 13). The assemblage difference in the 3-25 mm size population between the Proto and Meso is significant (at a 5% significance level). For example, there are 17 more Karoo shales and sandstones clasts, and 9 less BIF clasts, in every 100 counted between the Proto to Meso (Fig. 13). Namaqua Metamorphic Complex clasts are higher in the Meso Orange River gravels than the Proto Orange River gravels (Figs. 7, 13).

Proto and Meso Orange River garnets show similar FeO compositions but with a very narrow range (Fig. 14B, D). However, a small number of garnets from Proto Orange River deposits (n = 4) and Meso Orange River deposits (n = 2) show slightly lower FeO than the rest of the group (Fig. 14). When compared to the Namaqua Metamorphic Complex garnets (Humphreys and Van Bever Donker, 1990; Diener et al., 2013; Bial et al., 2015) the Orange River garnets are similar to the Namaqua

Metamorphic Complex garnets in both their FeO, MgO and MnO contents (Fig. 14). An exception are the low FeO, low MgO garnets that are different from the Namaqua Metamorphic Complex garnets (Fig. 14). These are similar to the Gariep Belt garnets (Diener et al., 2017) (Fig. 14B, D).

5. Discussion

5.1 Controls on clast assemblage differences

An interplay of provenance, palaeohydraulics, and reworking, influence clast assemblages in the different terrace successions. Provenance is widely invoked as a dominant control on compositional differences between sediments on a regional to local scale (e.g., Gibbard, 1979; Green et al., 1982; Bridgland, 1999; Roberts et al., 2008; Claude et al., 2017). Clast provenance can vary through time due to changes in surface exposure and availability of different rock types, or through drainage re-organisation (e.g., Mather, 2000). Re-organisation of drainage basin networks can be caused by tectonism and volcanism (e.g., Maddy et al., 2012; Richardson et al., 2016), or through drainage capture events (e.g., Mather, 2000; Maher et al., 2007) during the evolution of degradational landscapes. Periods of tectonic uplift, and increased erosion and sediment flux in southern Africa, that could have influenced the clast assemblage of the Proto and Meso Orange River deposits include during the Cretaceous (de Wit, 1999; Stevenson and McMillan, 2004; Tinker et al., 2008b; Guillocheau et al., 2012; Richardson et al., 2016, 2017), and the Miocene and Pliocene (Partridge and Maud, 2000; Green et al., 2017). A Pliocene period of uplift, which occurred after deposition of the Proto Orange River gravel, could be invoked to have driven drainage re-organisation and influenced clast assemblage differences between the Proto and Meso Orange River gravels. However, there is neither a diagnostic clast lithology in either Proto Orange River or Meso Orange River gravels, nor geomorphological evidence for drainage re-organisation reported for the Orange River catchment during this period.

Only the relative dominance of exotic clasts distinguishes the clast assemblages between the stratigraphically distinct Proto and Meso Orange River successions in the Orange River gravel deposits (Figs. 7, 12, 13). This suggests that there has not been a major change in sediment provenance available to the Orange River between the Proto and Meso periods of terrace deposition, although different lithologies have been eroded and transported during different periods. For example, the proportions of Karoo shales and sandstones suggest that the majority of the Karoo Supergroup sediments within the Orange River drainage basin were entrained by the end of Proto-Orange River times and were less available to the Orange River in Meso-Orange River times. The opposite is true for the BIF (Figs. 7, 12, 13). Although the erosion rates, and associated sedimentation rates, of the southern African landscape remain highly debated (Hawthorne, 1975; Brown et al., 1999; Gallagher and Brown, 1999; Tinker et al., 2008b; Hanson et al., 2009; Richardson et al., 2017), sedimentation rates in the Orange Basin offshore Namibia and South Africa (Rust and Summerfield, 1990; Aizawa et al., 2000; Rouby et al., 2009) and Outeniqua Basin, offshore South Africa (Tinker et al., 2008a), suggest that sediment production and deposition continued to decrease after the Cretaceous uplift event. The Proto and Meso Orange River deposits are younger than the Cretaceous, therefore tectonic uplift may have not directly influenced the clast assemblage between the two sets of deposits. However, tectonic uplift may have influenced the rate at which Karoo rocks were eroded such that most of the Karoo shales and sandstone were eroded during the Proto Orange River period and were less extensively exposed and available for transport in the Meso Orange

River period. In summary, changes in the availability of rocks exposed in the drainage basin were a more significant control on differences between the Proto and Meso Orange River deposits clast assemblages than drainage re-organisation, as evidenced by decreasing Karoo shale and sandstones rock clasts and increasing BIF from the Proto Orange River deposits to the Meso Orange River deposits (Fig. 7).

Rivers vary in their discharge capacity and power through time due to changes in channel dimensions, drainage basin area, gradient, and climate (Schumm and Lichty, 1965; Bull, 1979; Charlton, 2008; Hamers et al., 2015). This impacts their ability to erode and transport sediment of different calibre (size and density), and the clast character (Charlton, 2008). The higher degree of clast roundness in the Proto Orange River gravel relative to the Meso Orange River gravel (Fig. 8) suggests a higher sediment load and/or a higher supply of relatively abrasive quartzite (Lindsey et al., 2007; Miao et al., 2010). The thicker, and volumetrically larger Proto Orange River gravel terrace deposits (up to 50 m thick) (Figs. 3, 4A) provide evidence for a more powerful river, with higher sediment loads, during the incisional phase compared to the Meso Orange River incisional phase with thinner gravel terrace deposits (6-23 m thick) (Fig. 3A). Within the study area, there is a steeper river gradient of the Proto age Orange River (0.69 m/km) compared to the Meso age Orange River (0.60 m/km) (Fig. 3A). A steeper surface gradient would increase the power and carrying capacity of the Proto age Orange River, despite its more sinuous planform (Fig. 3C).

There is a paucity of Karoo Supergroup basalt clasts in the Proto Orange River gravel (1%) relative to the Meso Orange River gravel (3%) (Figs. 7, 13) even though they are both derived from the Drakensburg Karoo Supergroup, the youngest member of the Karoo Supergroup (Duncan et al., 1997; Marsh et al., 1997; Jourdan et al., 2007; Hanson et al., 2009), which could be expected to have been eroded relatively early in the erosional history of the drainage basin. There are two possible explanations for this difference. Firstly, a wetter and more humid climate both before and during the Proto Orange River period may have eliminated basalt preferentially through chemical weathering (Amiotte Suchet and Probst, 1993; Louvat and Allègre, 1997; Dessert et al., 2001; Malvoisin et al., 2012; Cox et al., 2016). Secondly, the majority of the basalt clasts might have been mechanically broken down during transport in the Proto Orange River period, which would explain their presence mostly in the smaller size fractions of 3-8 mm (Fig. 7C). The presence of unweathered feldspar clasts, in the Proto Orange River gravel (Fig. 6D), does not support the hypothesis of climate induced chemical weathering of basalt (Pellant, 2000; Maddy et al., 2012; Tan et al., 2017). In addition, Bluck et al. (2007) and Miller (2008) reported that arid conditions in the region were prevalent in the Eocene, based on the occurrence of thick (18 m) aeolian sandstone overlying basal marine gravel at Buntfeldschuh, an Eocene outcrop of shoreline deposits about 130 km north of the Orange River mouth. Therefore, the Proto and Meso deposits were exposed to similar arid conditions. Evidence from incision rates and clast roundness suggests that the Proto-Orange River was a higher energy environment than the Meso-Orange River sedimentary system, and one in which basalt clasts would be preferentially mechanically degraded (Figs. 3A, 8). However, the garnet composition data suggest that the heavy minerals are sourced locally from the Namaqua Metamorphic Complex and Gariep Belt (Fig. 14). Therefore the heavy mineral anomalies that have been liberated from the mechanical disintegration of catchment area derived Karoo basalts could not be established. However, this does not exclude that some heavy minerals could be derived from higher in the catchment.

The Proto Orange River and older deposits were incised by the Meso Orange River system, and were available to be reworked and incorporated into the Meso Orange River deposits. Locally, downstream reworking of older deposits can be an important process as suggested by the uncharacteristic abundance of Karoo Supergroup shales and sandstones in the Auchas Major Meso deposit (Fig. 7). However, in general, the absence of significant reworking of the Proto Orange River deposits is striking (Fig. 7). The lack of evidence for extensive reworking is possibly because the Orange River evolved to a straighter planform during the Meso period (Fig. 3C), such that the Proto Orange River gravel terraces are well preserved because they are largely situated outside the influence of the Meso Orange River course. The decrease in clast roundness from the Proto to the Meso Orange River deposits also suggests minimal reworking and downstream redeposition of older deposits within the study area (Fig. 8).

5.2 Controls on mineralogy of heavy mineral assemblages

Physical sorting, mechanical breakdown, and dissolution by chemical weathering influence the preservation of heavy mineral assemblages (Morton and Hallsworth, 2007; Weibel and Friis, 2007). The distance a heavy mineral grain travels before deposition depends both on its density and size (Komar and Wang, 1984).

Amphibole-epidote shows significant changes in proportion between the Proto and Meso Orange River deposits (Figs. 9, 10, 12, 13). Amphibole-epidotes are sourced from the local Namaqua Metamorphic Complex rocks (Botha and Grobler, 1979; Bailie et al., 2010) (Table 1) on the basis that high amphibole-epidote proportions coincide with high amounts of Namaqua Metamorphic Complex clasts (Fig. 13). In addition, the similarity in composition of the detrital Orange River garnets and the Namaqua

Metamorphic Complex garnets, and to a lesser extent the Gariep Belt garnets, constrain the provenance of the majority of the detrital heavy minerals in the Orange River gravels to these rocks (Fig. 14). Among the trace minerals in the Orange River gravels (titanite and zircon), titanite has been reported in the Gariep Belt rocks (Frimmel et al., 1996; Frimmel and Frank, 1998) (Table 1) but not in Namaqua Metamorphic Complex rocks. Therefore, titanite provides evidence for a contribution of Gariep Belt rocks to the lower Orange River gravels. Commonly, amphibole is absent in buried sediment owing to its chemical instability at depths greater than 600 m (Morton, 1984; von Eynatten and Gaupp, 1999; Mange and Morton, 2007). Epidote also has similar diagenetic stability to amphibole persisting only to 1100 m (Morton and Hallsworth, 2007). However, loss of amphibole and epidote due to dissolution alone cannot explain their relatively low abundance in the Proto Orange River deposits that have a maximum thickness of 50 m (Jacob, 2005) (Fig. 3A) and a thin sand cover (<2 m). Furthermore, chemical weathering is considered unlikely given the presence of unweathered feldspar (Fig. 6D). The increase of amphibole-epidote content from the Proto to Meso Orange River deposits (Figs. 9, 10, 13) could be influenced by the interpreted decrease in river energy that increased the preservation potential of mechanically weaker and softer minerals, such as amphibole and epidote. This is supported by garnet showing conchoidal fractures that are produced by mechanical processes (Velbel et al., 2007), although conchoidal fractures are also present on garnets from the Meso Orange River deposits (Fig. 11A). Therefore, the dominant control on the increase in the proportion of amphibole-epidote (Figs. 9, 10, 13) is interpreted to be a consequence of the larger influx of Namaqua Metamorphic Complex-derived material (Fig. 15).

The downstream decrease of magnetite and increase of amphibole-epidote between Boom and Arrisdrif in the Meso Orange River gravel (Fig. 10) coincides with the downstream decrease in gravel grain size and increase in sand content for both Proto and Meso Orange River deposits. Given that both magnetite and amphibole-epidote were liberated from Namaqua Metamorphic Complex and Gariep Belt rocks (Fig. 14), their different downstream changes in concentrations may be controlled by density, of 5.2 g/cm^3 and 2.97-3.13 g/cm^3, respectively (Pellant, 2000) where more magnetite is retained in the upstream deposits. This trend also suggests that there is no further addition of Namaqua Metamorphic Complex material to the Orange River downstream of Boom. The low abundance of amphibole-epidote in the Auchas Major Meso Orange River sample (Figs. 9, 13) coincides with a high abundance of Karoo Supergroup shale and sandstone (Fig. 7), which are characteristic features of the Proto Orange River deposits. This suggests that reworking of the Proto Orange River gravel affected the clast and heavy mineral assemblages by diluting the amphibole-epidote content of the sand sized fractions at this location. The lack of reworking of the older Proto Orange River deposits is also evident in mineral surface textures, because the magnitude of chemical dissolution (e.g., etch pits) increases with decreasing mineral stability from magnetite and garnet to epidote in both the older and younger deposits. A large percentage of the garnets in the Proto Orange River gravel are relatively coarse (1-2 mm) (Fig. 9A) whereas the fine grained garnets (0.5-1.0 mm and 0.25-0.50 mm) appear to be much less common (Fig. 9B, 9C), presumably removed by higher water energy in the Proto period and transported offshore. Imbricated clasts in the Proto Orange River gravel attest to a high energy bedload-dominated river system (e.g., Ashley et al., 1988; Wittenberg, 2002) (Fig. 5A).

This study has established that the Proto and Meso Orange River deposits are not only distinguishable from each other at clast scale, but also at a heavy mineral scale (Figs. 7, 9, 13). The integrated clast and heavy mineral assemblage of the Orange River deposits can therefore be used to understand the distribution and timing of the deposition of the coeval marine gravels in response to the evolving depositional and erosional phases of the Orange River. A good understanding of the stratigraphic record of the gravels, in terms of age of deposition and sediment distribution patterns for marine deposits, is important for better resource exploitation and improved sampling and resource exploration techniques.

5.3 Implications for river terrace deposits analysis

The clast assemblage of the Proto and Meso Orange River gravel terrace deposits is controlled by catchment-scale processes (Fig. 15). In contrast, differences in the heavy mineral assemblages between the two gravels (Figs. 9, 13) is influenced by local controls, such as the availability of Namaqua Metamorphic Complex rocks to the Orange River and the lower preservation potential of amphibole-epidote. This implies that extrinsic controls on clast assemblage and intrinsic controls on heavy mineral assemblage of the Orange River gravels need to be considered in evaluation of terrace deposits of other bedrock river systems globally. The sand size fraction and the clasts can be derived from different sources such that they carry different provenance signatures and reflect different transport histories. This is likely to be a similar scenario in other continental-scale bedrock rivers. Therefore, prediction of the nature of the fine size fraction on the basis of clast provenance alone is problematic. Mechanically weaker rocks such as basalt may be lost. Therefore, using clast assemblages to reconstruct the drainage history of high energy river systems should take into account the possibility of loss of mechanically weaker clasts. Bridgland (1999) used clast analysis to reconstruct the drainage evolution of the Thames River, England, and argued that tributaries have been re-organised over its history and that the river has diverted its course in response to middle Pleistocene glaciation based on evidence from changes in the composition of clasts. However, chalk is an important rock type exhumed in the Thames drainage basin. Therefore reconstructing palaeo-tributaries that have drained solely through chalk on the basis of clast assemblage alone is problematic in this case, because chalk is mechanically weak. Through clast analysis of late Quaternary sediments, Jones (2000) noted a downstream decrease of granite clasts in the Pineta Basin, Spain, and attributed it to mechanical breakdown. If these Pineta Basin sediments were deposited by a higher energy river system, the granite clasts might have been broken down and their signature lost. In such cases, an integrated analysis of clast assemblages and heavy mineral assemblages would be a better approach because heavy minerals would have survived mechanical breakdown and retained the source signature. Therefore, the heavy mineral assemblage technique is a useful tool for studying drainage basin evolution in areas where rivers and their associated tributaries drain areas whose geology is dominated by mechanically weaker rock types. Studies that use clast analysis to deduce provenance of sediments make an implicit assumption that sand sized sediments are from the same source as the clasts (e.g., Bridgland, 1999; Mikesell et al., 2010).

This study has shown that assessment of the controls on clast and heavy mineral assemblages needs to be treated separately due to the differences in density that affect the preservation and behaviour of pebble sized clasts and sand sized heavy minerals. However, despite different factors controlling the clast assemblage and heavy mineral contents of the lower Orange River deposits, the Proto and Meso

Orange River deposits differ in terms of both clasts and heavy minerals (Figs. 7, 9, 13).

6. Conclusions

We have integrated clast and heavy mineral assemblages to investigate the character and controls in downstepping flights of bedrock-confined river terrace deposits formed during multiple incision and aggradation cycles by the palaeo-Orange River. The stratigraphic decrease in terrace deposit thickness, and clast character, between the Proto and Meso Orange River deposits is linked to a more powerful river system during Proto times driven by a changing drainage basin geomorphology, rather than reworking of older deposits or changes in provenance. Local reworking of Proto Orange River gravel into younger deposits (Auchas Major) is evidenced by a significant increase in Karoo Supergroup sedimentary clasts, and decrease in amphibole-epidote content, in the Meso Orange River gravel. The decrease in incision depths, and sediment transport from Proto to Meso Orange River deposits was accompanied by an increase in the proportion of sediments supplied to the river from local lithologies, including an increase of amphibole-epidote in the heavy mineral assemblages sourced from the Namaqua Metamorphic Complex rocks. This study indicates that clast assemblage analysis should not be uncritically used as a proxy for the character of the matrix and vice versa. An integrated approach in analysis of these important but fragmented archives in source-to-sink studies is recommended when evaluating the controls on drainage basin evolution, and to improve prediction of heavy minerals and placer minerals in time equivalent deposits in downstream sedimentary basins.

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Table 1. Mineralogy of the Namaqua Metamorphic Complex and Gariep Belt rocks.

Source Lithology	Heavy Minerals Density > 2.8 g/cm^3	Other Minerals	References
Namaqua Metamorphic Complex	Garnet Amphibole Epidote Spinel Pyroxene Ilmenite Magnetite Sillimanite Zoisite	Plagioclase Feldspar Biotite Cordierite Chlorite	Botha and Grobler (1979) Waters (1989) Robb et al. (1999) Diener et al. (2013) Bial et al. (2015)
Gariep Belt	Amphibole Epidote Ilmenite Titanite	Biotite Plagioclase	Frimmel et al. (1996) Frimmel and Frank (1998)

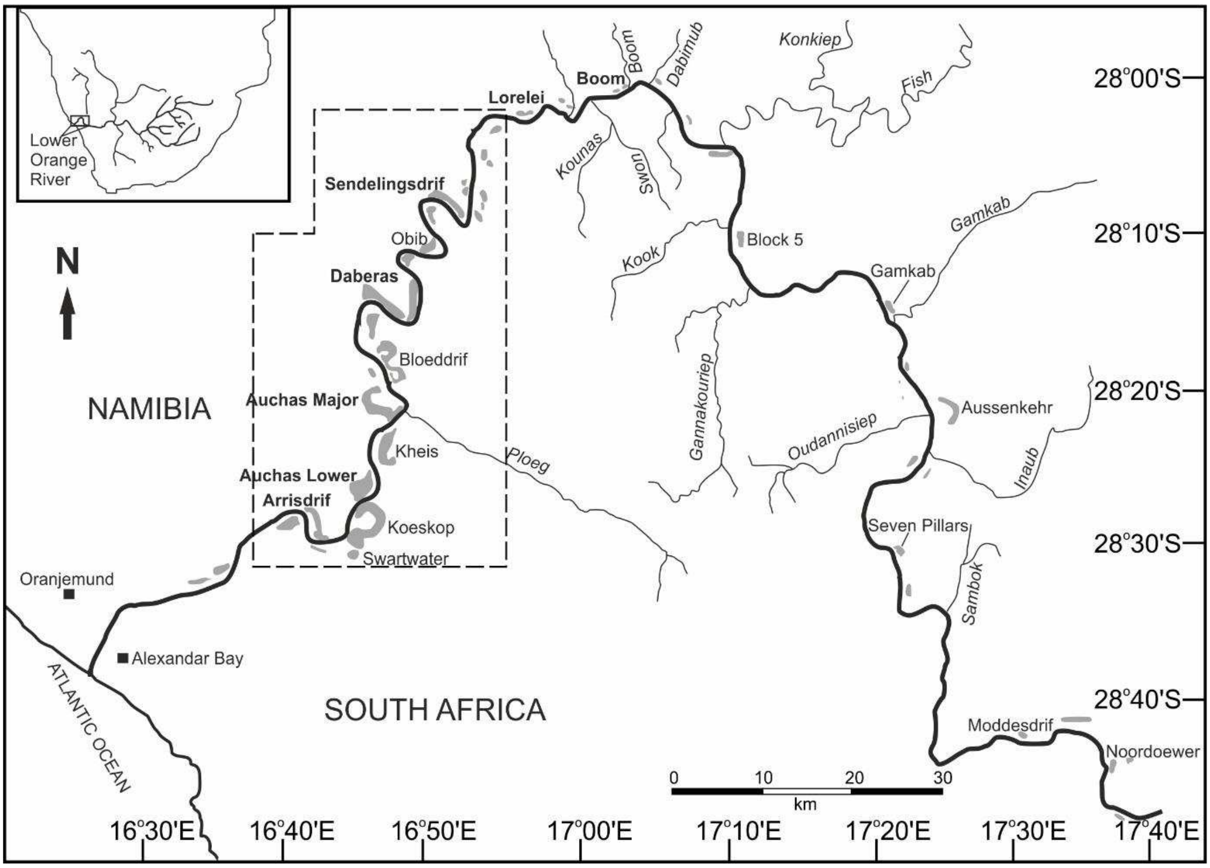

Fig. 1. Study area with distribution of gravel terrace deposits (grey colour) along the lower Orange River. Deposits analysed in this study are marked in bold. Modified from Jacob et al. (1999). The boxed area with the broken line represent the area shown in Figure 3C.

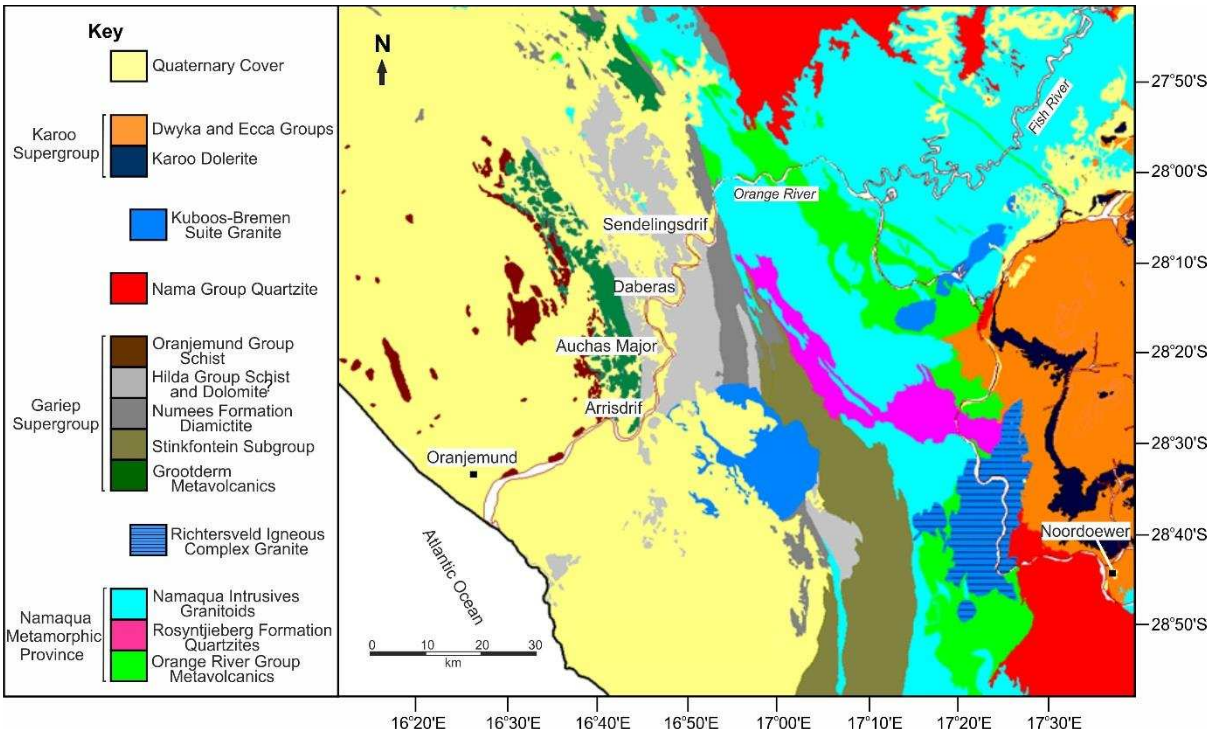

Fig. 2. Simplified geology of the lower Orange River. Locations of Sendelingsdrif, Daberas, Auchas and Arrisdrif deposits are indicated for reference. Namibia GIS-based data obtained from the Geological Survey of Namibia. South African data after de Villiers and Sohnge (1959).

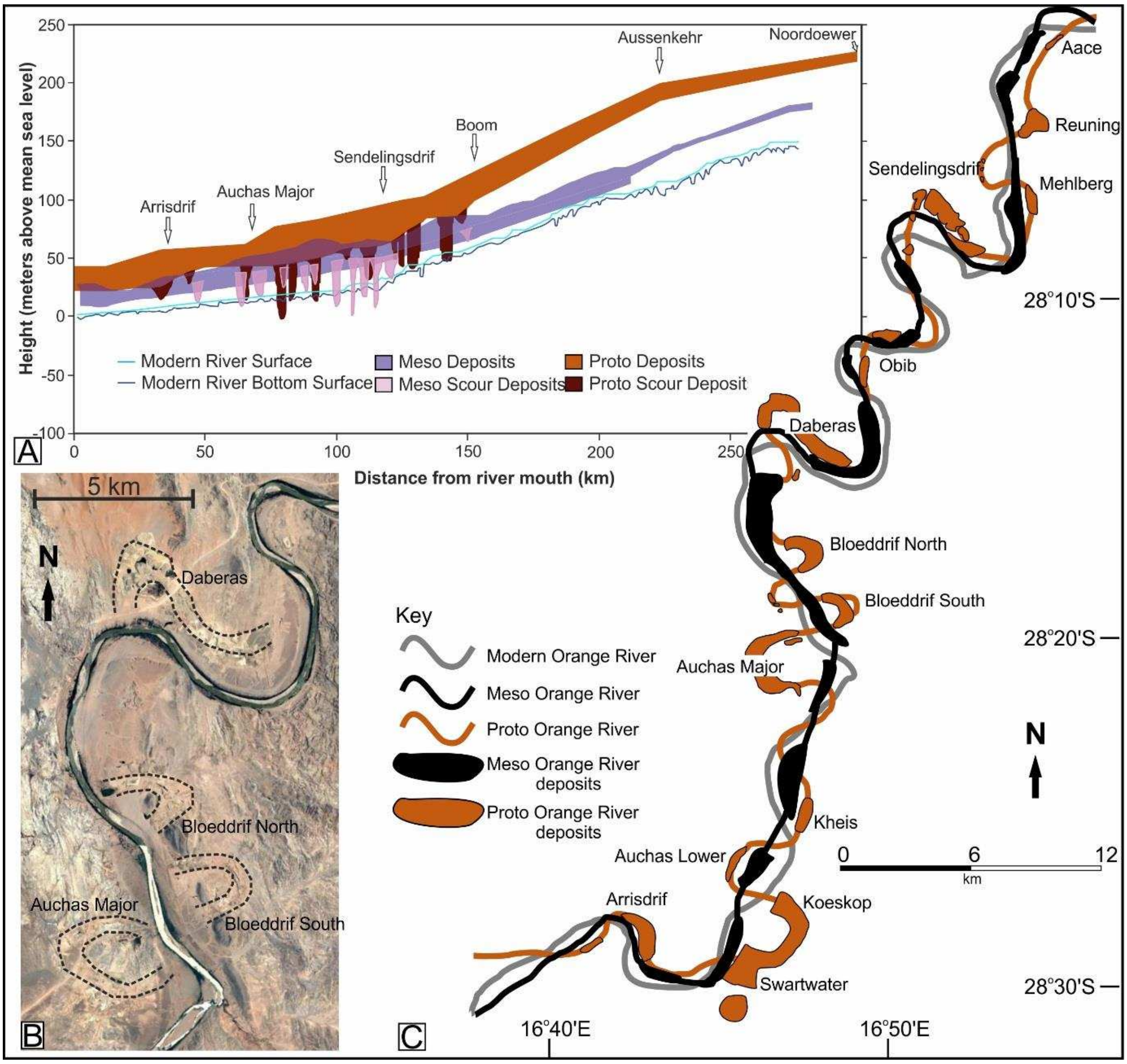

Fig. 3. (A) Proto Orange River and Meso Orange River profiles relative to the modern Orange River profile. (B) Google Earth image of the Proto terraces between Auchas Major and Daberas (C) Proto Orange River, Meso Orange River and modern Orange River courses. Figures A and C modified from Jacob (2005).

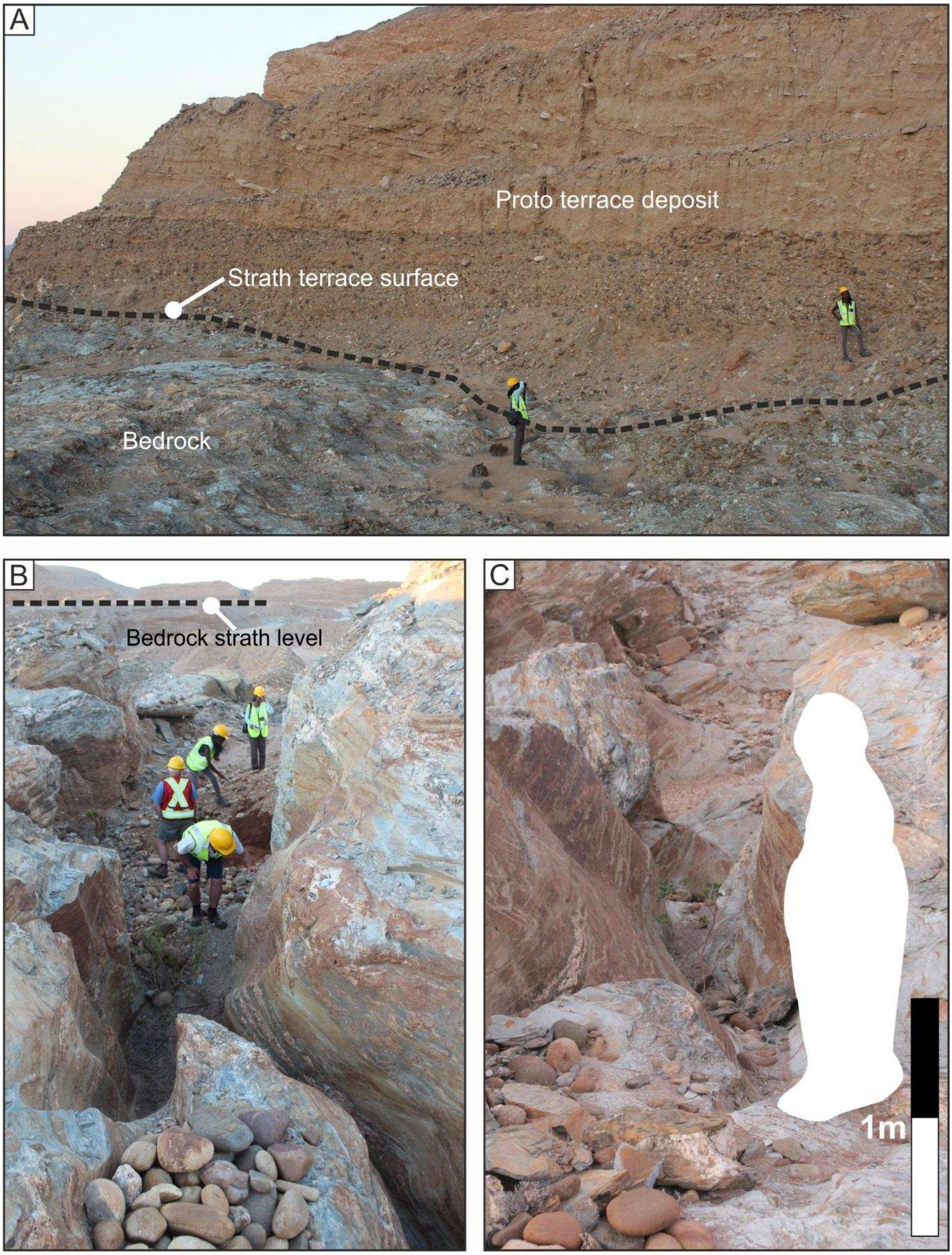

Fig. 4. (A) Representative photograph of the thick Proto Orange River terrace deposit at Auchas deposit. Photograph taken looking southeast. (B) and (C) Photograph of deep scours cut into bedrock below the bedrock strath level at Auchas deposit. Note the smooth walls of the scours formed by abrasion.

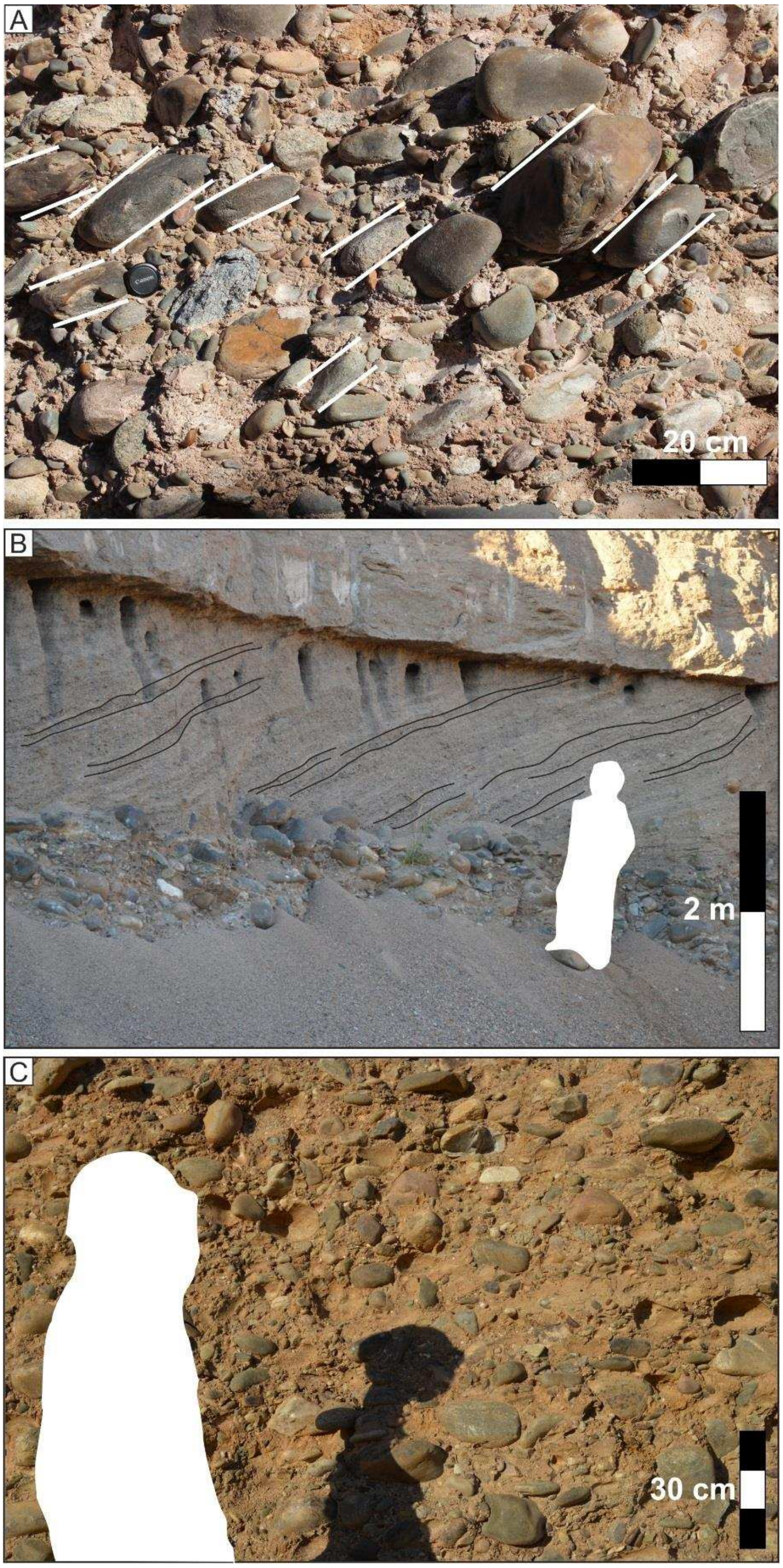

Fig. 5. Imbricated clasts (marked by white lines) (A) and coarse cross bedding (B) as seen in Proto Orange River unit and above Meso Orange River unit, respectively. (C) Meso Orange River gravel.

Fig. 6. Agate (A), Karoo sedimentary rocks (B) and banded iron formation (C) clasts that comprise the exotic clast suite of the Orange River derived gravels. (D) Fresh non-weathered feldspar clasts from Proto Orange River gravel, Daberas deposit.

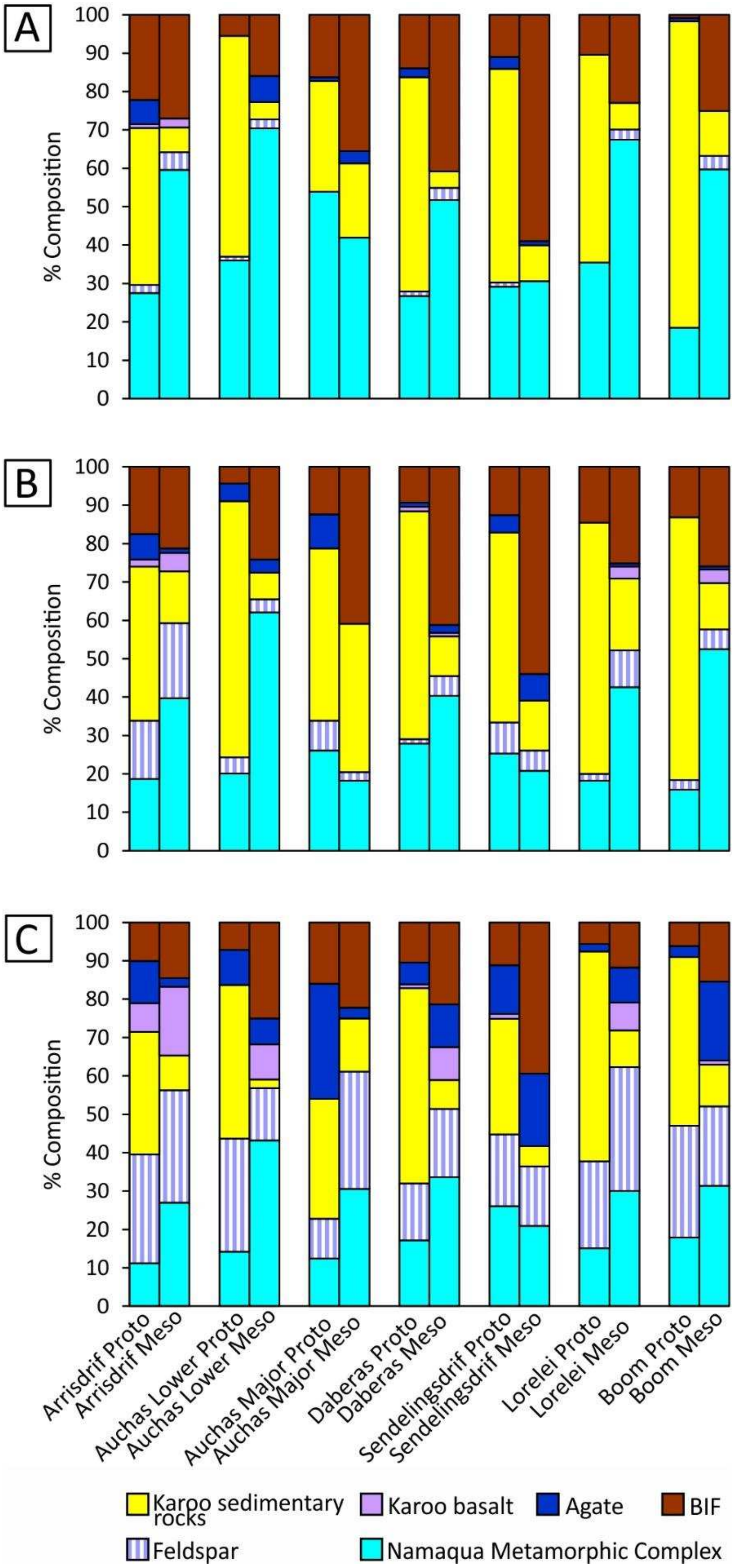

Fig. 7. Clast assemblage of Proto and Meso Orange River gravels for size fractions (A) 16-25 mm, (B) 8-16 mm and (C) 3-8 mm for different locations along the river. Data from Jacob (2005).

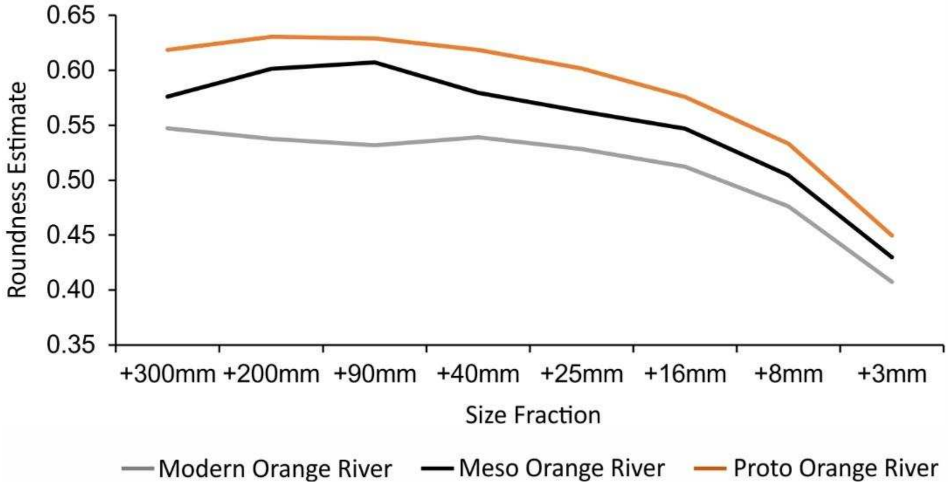

Fig. 8. Clast roundness of the Proto and Meso Orange River gravels. Modern Orange River data is included for comparison. Size fraction are +300 mm, +200 mm, +90 mm, +40 mm, +25 mm, +16 mm, +8 mm and +3 mm. All data from Jacob (2005).

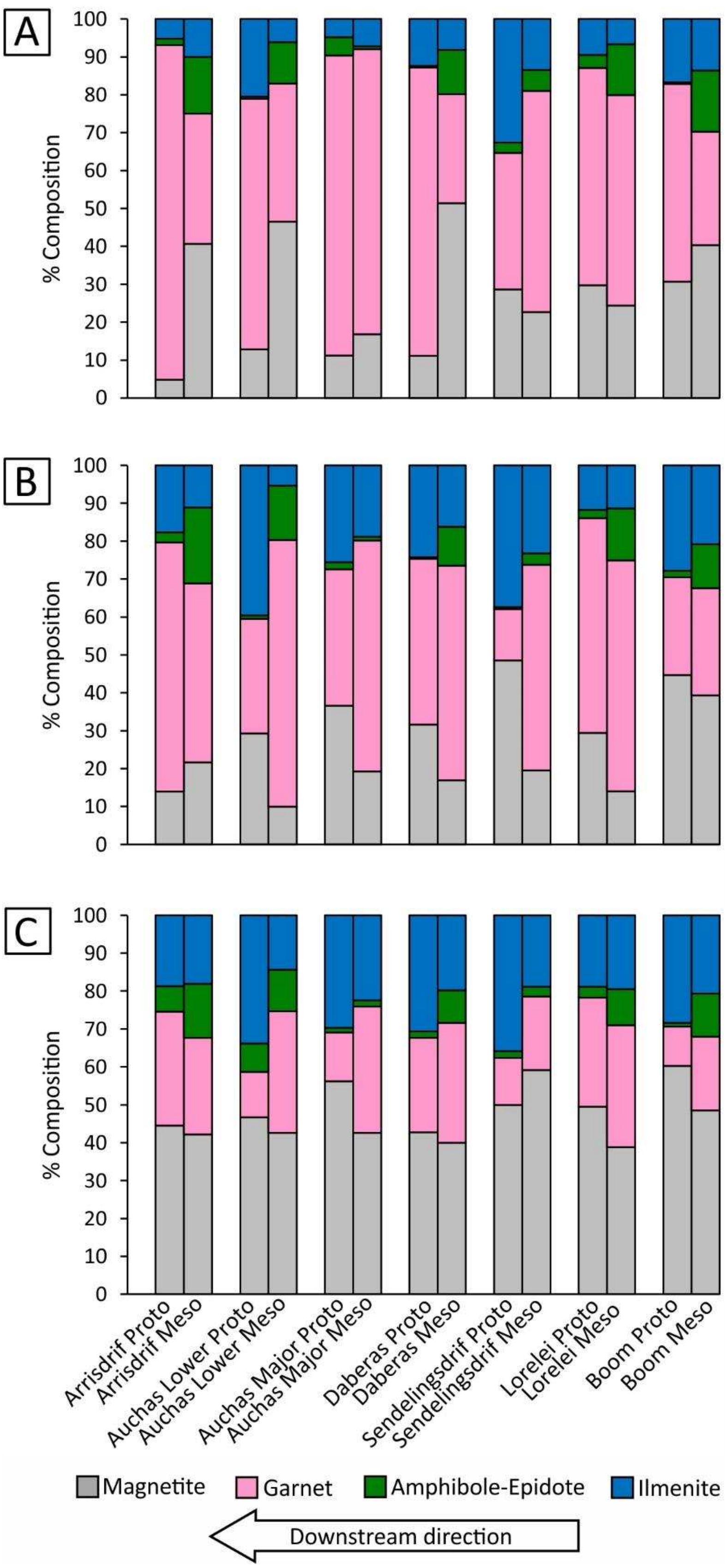

Fig. 9. Heavy mineral assemblage of Proto and Meso Orange River deposits for size fractions (A) 1-2 mm, (B) 0.5-1.0 mm and (C) 0.25-0.50 mm for different locations along the river.

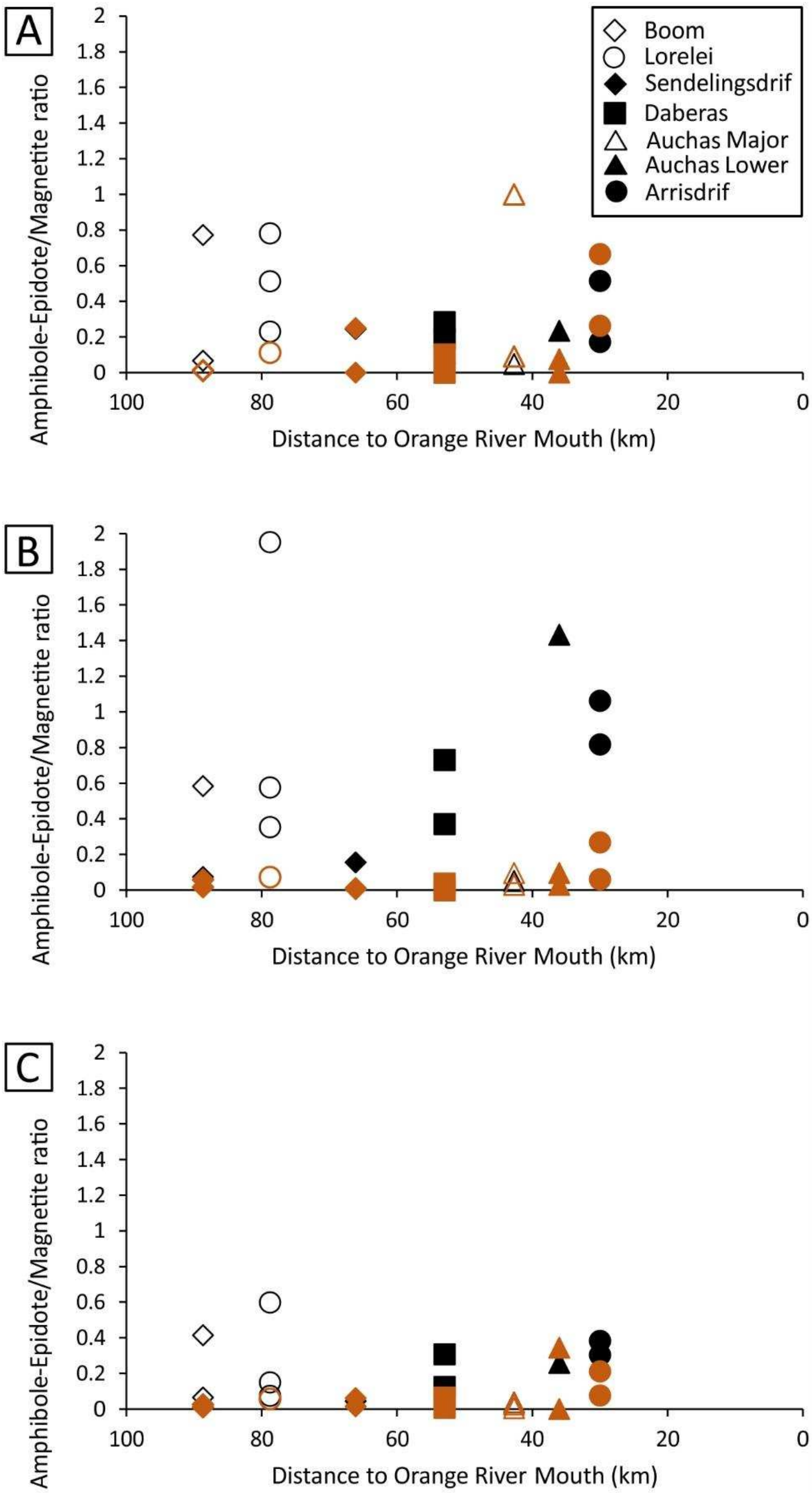

Fig. 10. Downstream change in amphibole-epidote/magnetite ratio from Boom to Arrisdrif for the Proto Orange River gravel (orange symbols) and Meso Orange River gravel (black symbols). (A) 1-2 mm, (B) 0.5-1.0 mm and (C) 0.25-0.50 mm size fractions.

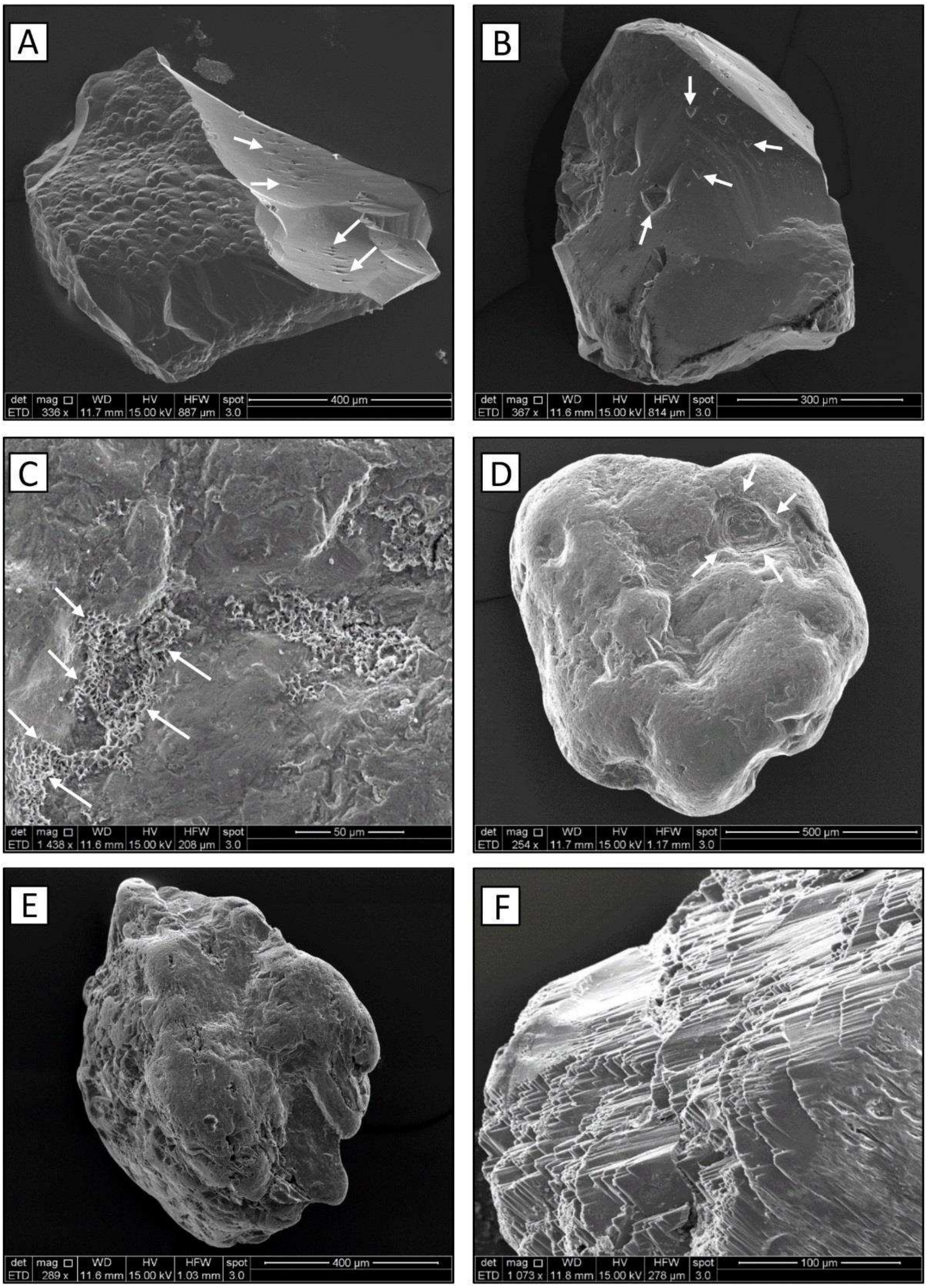

Fig. 11. SEM images of mineral grains from the Orange River. (A) Etch pits on conchoidally fractured surface (arrows) on garnet from Proto Orange River Sendelingsdrif deposit. (B) Euhedral etch pits on garnet from Boom Meso Orange River deposit. (C) Honeycomb dissolution texture on magnetite (arrows) from Proto Orange River Arrisdrif deposit. (D) Large dissolution pit (arrows) on magnetite from Meso Orange River Sendelingsdrif deposit. (E) Irregular etching on epidote from Proto Orange River Auchas Major deposit. (F) Saw-tooth terminations on epidote from Meso Orange River Arrisdrif deposit.

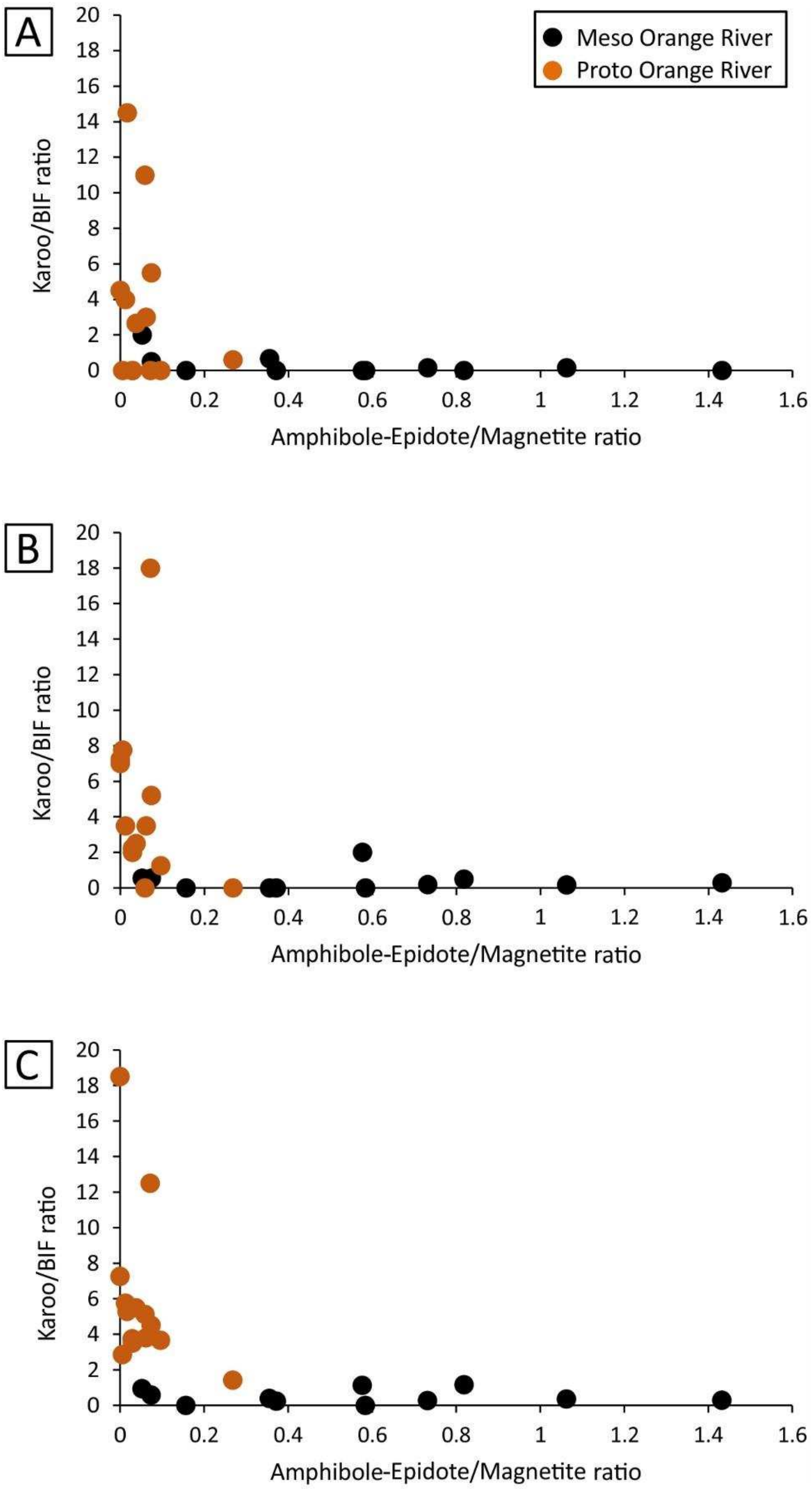

Fig. 12. Clast assemblage and heavy mineral assemblage variations between Proto Orange River and Meso Orange River gravel. Heavy mineral assemblage data is from 0.5-1.0 mm size fraction whereas clast assemblage data is for (A) 16-25 mm, (B) 8-16 mm and (C) 3-8 mm size fractions.

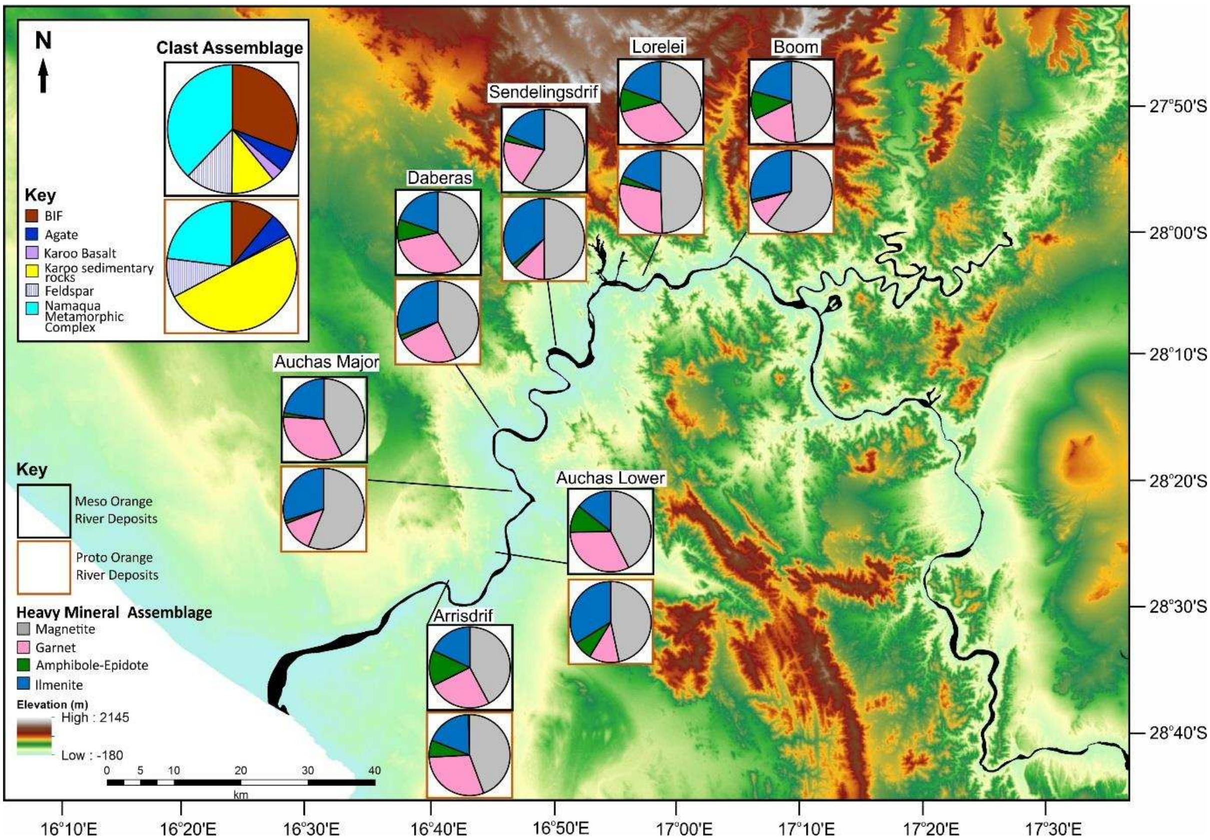

Fig. 13. Clast assemblage (inset) and heavy mineral assemblage of Proto and Meso Orange River deposits. Size fractions are 3-25 mm and 0.25-0.50 mm for clast and heavy mineral assemblage data, respectively. Clast assemblage and elevation data after Jacob (2005) and Jarvis et al. (2008), respectively.

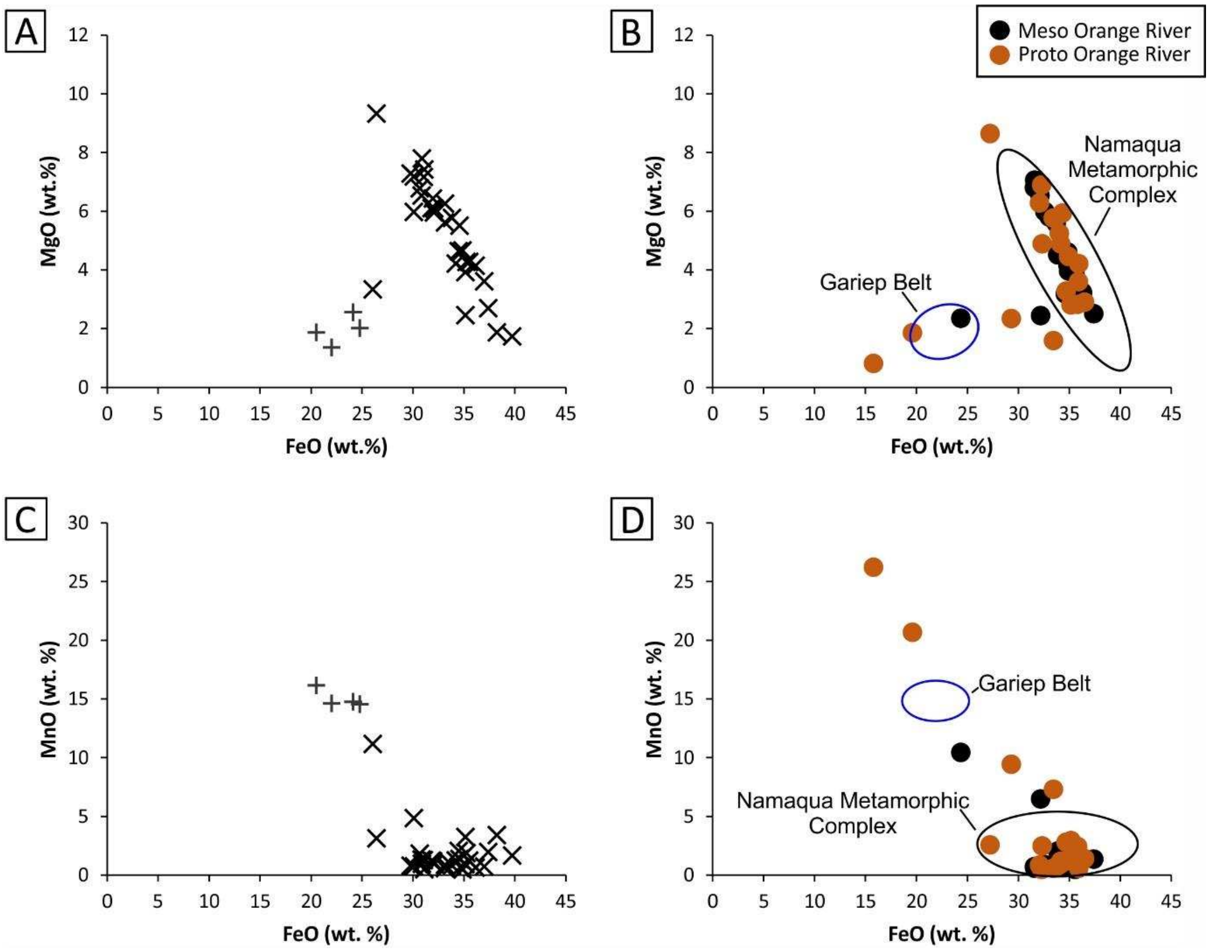

Fig. 14. Comparison of Orange River garnets with garnet composition of Namaqua Metamorphic Complex and Gariep Belt. (A) Garnet compositions in MgO versus FeO from the Namaqua Metamorphic Complex (Humphreys and Van Bever Donker, 1990; Cornell et al., 1992; Diener et al., 2013; Bial et al., 2015) and Gariep Belt garnets (Diener et al., 2017). (B) Data for Proto and Meso Orange River garnets. (C) MgO versus FeO from the Namaqua Metamorphic Complex and Gariep Belt (D) Data for Proto and Meso Orange River garnets.

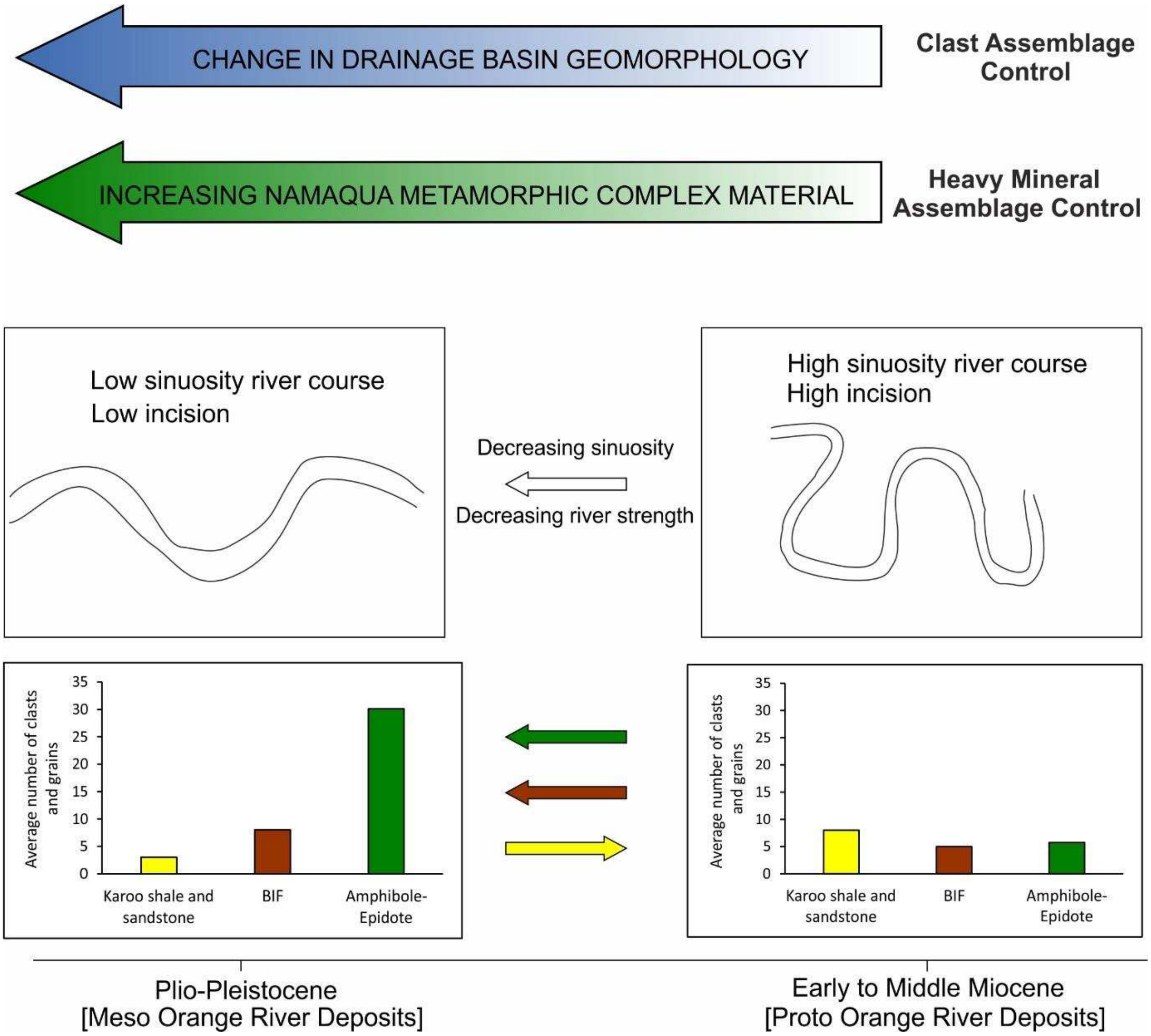

Fig. 15. Synthesis on major changes in clast and heavy mineral assemblage of the Orange River deposits, and the interpreted controls. Yellow, brown and green arrows point in the direction of increase.