Late Cretaceous to Early Cenozoic Initiation of Rifting of the Windhoek Graben, Namibia

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ABSTRACT

The Windhoek Graben is a N-S trending rift in central Namibia that forms a prominent topographic feature bisecting an area of plateau uplift. It occupies a potentially crucial role in the propagation of the Late Cenozoic Southwest African Rift system regarding a possible continuation to the west of the Eiseb Rift. It is an unusual example of intra-continental rifting because it has no significant sediment fill associated with the period of active rifting, and hence the timing of rift activity nor its tectonic relevance has not hitherto been established. To constrain the age of the Windhoek Graben we examine its regional geomorphic context and its relationship to four sites of igneous activity in the central Namibian Highlands. Two of these consist of clusters of eroded phonolitic tholoid bodies that have yielded $^{40}Ar/^{39}Ar$ dates of 32 Ma and 52 Ma, respectively, that we use to bracket the age of formation of a prominent remnant land surface, termed here the P52 Surface. From previous mapping of older intrusive igneous bodies, we argue that an even older land surface is partially preserved on the highest features in the area, and this surface (termed PRS) defines an initial domally uplifted surface from which initial drainage radiated, and onto which the earliest volcanic products associated with the Graben were erupted. In particular, the strong similarity in dyke and fault orientations is used to argue for a causal connection between the earliest magmatic activity and the onset of rifting. Long range correlation of PRS into the adjacent Aranos Basin strongly suggests a Late Cretaceous age for this earliest magmatic activity and the onset of rifting, but we cannot exclude a younger origin, any time up to the Early Eocene.

1. INTRODUCTION

Our study focus is the Windhoek Graben in central Namibia (Fig.1). This relatively immature rift basin is unusual in a global context, in that it is almost devoid of a graben fill (Miller, 2008; Picart et al., 2020). As a result, there is no syn-rift stratigraphy preserved with which to frame a chronostratigraphic evolution (Rosendahl et al., 1986; Morley, 1988; Leeder and Jackson, 1993). Hence there are open questions as to the timing of graben formation, which is important to resolve the geodynamic context for rifting in this part of southern Africa (Daly et al., 2020).

The main aim of this paper is to attempt to place age constraints on the initiation of rifting of the Windhoek Graben where the lack of a preserved syn-rift succession prevents a more conventional analysis used for dating rift onset in many other rift basins. Instead, we focus on using geomorphological arguments and, more specifically, attempt to reconstruct the development of former land surfaces as proxies for chronostratigraphic datums. Our approach is based on our own primary interpretation of the present day relief of the Windhoek Graben and surrounding Namibian Highlands combined with a synthesis of previous geological and geomorphological mapping undertaken over many decades. Particularly significant, is the inclusion of a review of the post-Jurassic igneous activity in the study area, which proved to be instrumental in providing age constraints for the older, partially preserved land surfaces.

We emphasise that our focus here is specifically on the critical and unresolved question of rift initiation, rather than on subsequent rift evolution. Future work will address the rift kinematics in a broader geodynamic context. Although beyond the scope of this paper, the ultimate question of the geodynamic context for rifting can only be addressed once the timing of rifting is better constrained. This may, for example, ultimately allow recent suggestions by Daly et al. (2020) of the propagation of a through-going rift system from the East African Rift, across central Africa to Namibia to be tested, at least in the area of central Namibia.

2. PREVIOUS INTERPRETATIONS OF THE WINDHOEK GRABEN

The origin and evolution of the Windhoek Graben have received relatively little attention considering its potential significance in the wider tectonic evolution of Namibia. It was first discussed as a possible graben nearly 100 years ago. Gevers (1942) notes that the linear morphology of the ‘Windhoek Valley’ is reminiscent of the graben structure interpreted by ‘early pioneering geologists’ but then went on to argue that the lack of obvious displacement of stratigraphic markers in the basement rocks bordering the valley argued against this interpretation.

Gevers (1942) referred to N-S trending silicified breccia zones as being connected to the volcanic plugs of the Auas Mountains, and his early mapping (Gevers, 1932) of these zones included a number of small faults that strike between NW and SE in the area just south of Windhoek. It was almost 30 years later that these faults were mapped in more detail as part of a wider mapping program aimed at unravelling the structure of the Damaran complexes (Guj, 1967).

The Windhoek Graben and its near neighbour, the Swakop Graben (Fig.2), have been identified and mapped as part of the regional geological mapping of Namibia (Namibia Geological Map 1:1,000,000). This map shows prominent normal faults oriented broadly N-S extending some 100 km from south of Windhoek and bordering both sides of the prominent linear topographic depression described by Gevers (1942) (Fig.2B). The southern part of the Windhoek Graben was mapped in greater detail by Miller and Schalk (1980) and at 1:50,000 by Hoffmann and Schreiber (2011) who identified sets of NNW and NNE trending extensional faults across a 12 km wide transect (Fig. 3). These faults were identified from a combination of groundwater exploration borehole data and surface mapping, in which highly fractured and cemented fault zone breccias provided direct indications of the presence of a fault. Miller (2008) reviewed the previous mapping efforts in the area, and presented summary maps showing normal faults bordering the Windhoek Graben extending continuously for 120 km to the north of Okahandja (town). Most recently, groundwater investigations using deep-penetrating electrical resistivity surveys have added further details on the fault distribution in the area of Windhoek, including detailed petrological studies of the silicified fault breccias and a discussion of fault dips (Miller et al., 2018).

An important contribution with implications for the origin of the Windhoek Graben was a subsurface and surface mapping study made by Hartmann (1994), of the area surrounding the Otjihase Mine. This mine is located in the mountainous region 20 km NE of Windhoek (Fig. 3), where a prominent ore body associated with base metal sulfide mineralization of the Matchless Amphibolite is offset systematically downwards by a series of N-S striking, steeply west-dipping normal faults with displacements ranging from 50-230 m. The Hoffnung Fault is the largest of these with a recorded value of 230 m total displacement with a dip of c.60 degrees to the west and is associated with a prominent eroded fault plane scarp at the surface. These faults are step or terrace faults located within the footwall to and synthetic with the eastern bounding fault (Fig. 3). Offsets of the Matchless Amphibolite across the main graben bounding fault (Fig. 3B) suggests a total displacement of c. 1000 m (Miller, R, pers comm).

Previous discussion of the age of the Windhoek Graben has been limited to two brief references. Mvondo et al. (2011) briefly state that the graben was 'associated to phonolite intrusions of Paleocene'. Most recently, Picart et al. (2020) mapped planation surfaces over a large area of central Namibia. From their mapping of a surface referred to as S6, which they interpreted as forming the floor of the Windhoek Graben, they suggest that the graben was formed prior to the development of this key surface in the middle Eocene.

3. REGIONAL CONTEXT

3.1. Physiographical Setting

The Windhoek Graben lies in an arid climatic zone, where annual rainfall is < 350 mm, and this arid to semi-arid climate regime can be inferred for the area back into the Mid Cenozoic (Ward, 1988). The relative aridity also means that there is good preservation of the vestiges of former land surfaces on the elevated margins surrounding the topographic depression representing the graben floor (Gevers, 1942; Mabbutt, 1951; Martin, 1973). These have the potential to be used as datums for dating the formation of the Windhoek Graben, provided there are objective criteria to date them or at least to identify them as specific to a particular erosional episode.

The Windhoek and Swakop Grabens are located in a region known as the Namibian Highlands (Fig.2). This elevated plateau reaches altitudes above 2000 m with some notable ridges rising to nearly 2500 m just south-east of Windhoek and some 500 m above the regional land surface. The surface is deeply dissected by a diversity of drainage networks, from large valleys linked to the major rivers, to smaller, trellis-like drainage networks dissecting the flank regions of the graben, forming a deeply dissected plateau. The Namibian Highlands are bordered to the east by the extensive interior plateau region (Haddon and McCarthy, 2005), which, in the southeast, is floored by the Aranos Basin and is marked by a much smoother topography, excluding the few valleys cut by the major rivers, and with a regional elevation of between 1150 and 1200 m. To the west of the Namibian Highlands, the highly dissected land surface slopes down to the Great Escarpment (trending N-S at approximately 16ºE, Fig. 2A).

To the north of the Swakop River, the Great Escarpment ceases to have any noticeable physiographic expression, possibly as a consequence of the collective erosional action of the major rivers that rise at the continental watershed and flow into the Atlantic Ocean, namely, the Swakop, Omaruru, and Ugab rivers (Aizawa et al., 2000). A prominent dissected scarp running approximately along the trace of the Okahandja Lineament demarcates the northern limit of the Namibian Highlands. This scarp has been partly cut by the Swakop River and forms the southern valley margin of that river, but it also represents a flexure or hinge line plunging northwards, possibly exploiting the basement weaknesses. Gevers (1942) and Mabbutt (1951) both recognized this flexure and Gevers (1942) considered it to be Mid Cretaceous in age.

The southern margin of the Namibian Highlands is marked by a prominent series of ridges striking parallel to the Damaran structural grain dissected to varying degrees by south or south-east flowing rivers. These ridges have been described as relict form to a former landsurface with the characteristics of accordant summits that comprise the more resistant rock types (Gevers, 1942). The valley floors of rivers such as the Usib and Nossob cut through these ridges and descend over a distance of 200 km to the lower relief of the interior plateau to the south and east (Fig.2A). A number of partially or almost completely preserved phonolitic extrusive bodies and associated vents form two important igneous clusters referred to here as the Aris and Stalhart Clusters, described in detail in Marsh (2010). These have been dated, and since these volcanic edifices were erupted onto a land surface that is interpreted to be close to the present day valley floors of rivers rising in the Namibian Highlands, they form a principal means for dating the erosion by those rivers.

The major drainages in the region of central Namibia are arranged in a strikingly radial pattern centered on the Namibian Highlands (Fig. 2A). Six major catchment areas can be defined with irregular, spoke-like drainage divides separating them. West flowing rivers such as the Swakop and Kuiseb cross the Great Escarpment and flow into the Atlantic (Matmon et al., 2018). South and southeast flowing rivers such as the Usib and Nossob flow into the Kalahari Basin or as tributaries to join the Fish River. North flowing rivers (e.g., Omatako River) flow into the Okavango River and represent examples of interior plateau drainage. This radial pattern of drainage is clearly consequent upon the topographic elevation of the Namibian Highlands, and hence understanding the erosional history of these rivers is a critical step in reconstructing the uplift history of the region.

One of the subtler features of this radial drainage system is the pattern of tributary drainage into the Swakop River (Fig. 2B). The Swakop River is remarkably linear over much of its course, rising near the main continental drainage divide northeast of Okahandja, and maintaining a WSW-oriented course over its 360 km journey to the sea. Near its headwaters, two tributaries join the main trunk river almost orthogonally from the south, and these tributaries form prominent axial drainages along the Windhoek and Swakop Grabens. This geometrical relationship between major faulted structures and the tributaries strongly suggests consequent drainage evolution, but without any obvious signs of river capture. This further suggests that the timing of formation of tributary drainages postdates that of the structural control.

3.2. Geological Setting

The Windhoek Graben is located within the Damaran orogenic belt (Miller, 1983). The basement geology encompassing the Windhoek Graben is dominated by schists of the Swakop Group of the Neoproterozoic Damara Supergroup (e.g., Miller, 2008). These originated as shelf-slope successions deposited on the continental margin bordering a small ocean basin (Martin and Porada, 1977; Downing and Coward, 1981). These sediments were metamorphosed and deformed by closure of this ocean basin during the Late Proterozoic to Cambrian (Barnes and Sawyer, 1980; de Kock, 1992). The dominant SE vergence of this collisional orogenesis led to the formation of crustal-scale shear zones and thrusts (Fig. 4A and B) that locally emplaced fragments of the old oceanic crust (the Matchless Amphibolite) at a shallow level on the highly deformed continental margin succession (Hoffmann, 1983; Miller, 1983; Kukla, 1992). Towards the southern end of the Windhoek Graben, the original coarser facies of the continental margin succession is preserved as a series of thick quartzite units (Guj, 1967) and these form prominent ridges in the modern landscape due to their erosional resistance (Gevers, 1942) (Fig. 4C).

The shear zones and thrusts exposed at surface in central Namibia have a predominant NE-SW strike. Of these, the most prominent is the Okahandja Lineament (OKL) which extends for about 650 km along the southern limit of the central tectonostratigraphic zone (Miller, 1983) (Fig. 4A). Surface deformational evidence of late-stage, multiphase strike-slip kinematics has been documented along portions of the OKL (Downing and Coward, 1981), that is interpreted as a former prism backstop, with a near-vertical geometry extending across the full thickness of the present-day crust (Kukla, 1992) (Fig. 4B).

Further to the southeast is the more curvilinear, but broadly SW-NE trending Us Pass Lineament (UPL), which defines the southernmost limit of the Southern Tectonostratigraphic Zone of the orogen (Miller, 1983) (Fig. 4A). This complex zone of thrusting comprises a number of thrusts, and imbricate splays exposing Paleo- to Mesoproterozoic basement (Miller and Schalk, 1980; Miller, 1983). The UPL has been interpreted as a continental suture zone (Hoffmann, 1983), forming a thrust zone dipping to the NW in the subsurface (Kukla, 1992) (Fig. 4B). Intense activity of the late-stage collisional tectonics also contributed to the emplacement of large bodies of granite during the Late Precambrian to Cambrian (Barnes and Sawyer, 1980; Downing and Coward, 1981; Porada, 1989; de Kock, 1992; Gray et al., 2008).

Preservation of younger successions is limited to outliers of the Karoo Supergroup (Late Carboniferous to Early Jurassic) in the north and south of the area, and to a more substantial outcrop of the Nama Supergroup (Cambrian) in the southwestern limit of the study area (Fig. 4A). This outcrop represents the northern margin of a large foreland basin in which the Nama Group accumulated.

The later geological and geomorphological evolution of this region has been substantially influenced by the break-up of Gondwana to form the South Atlantic Ocean during the Early

Cretaceous, resulting in the formation of the Namibian Volcanic Margin (Clemson et al., 1997; Gladczenko et al., 1997). This rifting process was followed by a significant uplift and erosional event along the margin in the Late Cretaceous to Middle Eocene (Partridge and Maud, 1987; Partridge, 1998; Gallagher and Brown, 1999; Aizawa et al., 2000; Bluck et al., 2007; Burke and Gunnell, 2008). A major erosional event in the Late Cretaceous has also been inferred by Miller (2008) based on a reconstruction of the stratigraphic evolution of the Aranos Basin derived from correlation of hundreds of groundwater boreholes. The Aranos Basin is a well-defined depocenter within the broader Kalahari Basin (Fig. 4A). Miller (2008) correlated lateritized intervals of remnant Karoo to demonstrate that these Karoo sediments were highly dissected to 200-300 m below the present-day surface by an extensive river system (referred to by Miller (2008) as the Aranos River) flowing S-SE to the palaeo-Orange River, via the palaeo-Molopo River during the Late Cretaceous. The Late Cretaceous initiation of this river system, therefore, implies that a south-eastward surface topographic gradient was established in this area at this time.

The Cenozoic history of the study area is poorly understood. The area has been shaped by a number of processes including: regional-scale uplift events (Picart et al., 2020), infill of the Aranos Basin and calcrete development beneath a younger aeolian cover (Miller, 2008), igneous activity (Ferreira et al., 1979; Marsh, 2010), and fluvial incision (Gevers, 1942; Mabbutt, 1951). Uplift events cited in previous studies vary in their timing, magnitude, and origin. For example, a Late Cretaceous/Paleocene uplift event was attributed to lingering plume tails after Gondwana breakup (Nyblade and Sleep, 2003) and is supported by palaeo-topographic reconstructions of Aizawa et al. (2000), Bluck et al. (2007), and by apatite fission-track studies (Raab et al., 2002). In contrast, regional uplift of southern Africa in the Oligocene–Miocene was linked to an African superplume (Burke, 1996) or lithospheric bulge (e.g., Picart et al., 2020). A more local uplift occurred in the Bie Dome (located some 500 km north of the study area) during the Pliocene–Pleistocene, and has been linked to a secondary plume (e.g., Al-Hajri et al., 2009).

In considering any possible mantle involvement in the uplift history of the region, it is worth noting the occurrence of a number of igneous complexes in the study area. These include the Stalhart cluster of phonolites, the Aris cluster of phonolites/trachytes, and the Regenstein vent complex and associated trachyte plugs and breccia dykes (Gevers, 1932; Burger and Walraven, 1976; Ferreira et al., 1979; Marsh, 2010). The first two are dated (52 and 32 Ma, respectively), but the ages of the Regenstein vent complex and associated trachyte dykes are unconstrained due to severe alteration (Ferreira et al., 1979). The Stalhart and Aris igneous bodies provide some constraints on the rate and style of erosional processes (Marsh, 2010) but also testify to a potentially significant early to mid-Cenozoic phase of magmatism. A suite of largely phonolitic extrusive bodies dated at 46 Ma is well described from the Klinghardt Mountains (Marsh et al., 2018), on the coastal plain 520 km to the southwest of the study area. The occurrence of younger sediments is restricted to the Aranos Basin and to the coastal plain where Late Cenozoic to Recent aeolian sediments several hundred meters thick have accumulated over the highly eroded basement rocks of the coastal Damaran and Gariep orogenic belts (Ward, 1988; Haddon and McCarthy, 2005).

4. DATA AND METHODOLOGY

The paper initially focuses on reviewing published works related to the occurrence of four suites of igneous rocks and their relationship to both the graben structure and the geomorphology in order to establish key landsurfaces. We correlate key surfaces from the Windhoek area down to the neighbouring Aranos basin to the south (Fig.2), where hundreds of groundwater boreholes have proved helpful in reconstructing the age of fluvial dissection (Miller, 2008). This approach generates a chronological framework, which is used to constrain the initiation age of rifting.

We describe the regional topography and geomorphology in the immediate area surrounding the Windhoek Graben using a DEM of the study area (SRTM 30) to compute topographic profiles, identify drainage basins and characterise the surface morphology. Topographic profiles were interpreted based on previous studies that discuss the landforms occurring along specific transects. Correlations of possible palaeo-surfaces are explained in the results section. Faults were identified and extracted from published geological maps of 1:250,000 (sheet 2116 and 2216, Geological Survey of Namibia) and 1: 1,000,000 (Miller and Schalk, 1980), and 1:50,000 (Hoffmann and Schreiber, 2011), and other published sources available (Guj, 1967; Ferreira et al., 1979; Hartmann, 1994; Miller, 2008; Miller et al., 2018; Picart et al., 2020) and incorporated into a GIS framework (ArcMap 10.8© and Global Mapper v21.0©). Additional faults were interpreted based on recognizing fault scarps using standard criteria developed for rift tectonic geomorphological analysis (e.g., Gawthorpe and Leeder, 2000; Goldsworthy and Jackson, 2000). The interpretation was validated by cross-referencing the DEM with the published geological maps to identify and eliminate any linear scarps that could be attributed to lithological boundaries.

We compiled maps showing the distribution of the igneous suites utilizing a published map by Miller (2008) that is compiled from Gevers (1934), Guj (1967), Ferreira et al. (1979), Hoffmann (1983), Niku-Paalova (1997), and Hoffman and Schreiber (1998). We carried out fieldwork to verify the descriptions of all igneous suites by Gevers (1934), Ferreira et al. (1979), and Marsh (2010) to understand their genetic, spatial, and temporal relationships to the present-day surface, and additionally, to the graben.

We reconstructed the positions of two key palaeo-surfaces in the southern Windhoek area by constructing profiles across and between the dated igneous bodies. Identification of the oldest former land surface was based heavily on the previous interpretation of river superposition to form narrow gorges through the resistant ridges that occur south of Windhoek (details in following sections). Further constraints on the age of the oldest surface were derived from projections of the key reconstructed surfaces into the subsurface of the Aranos Basin along a ~400 km topographic profile (SRTM 30), taking into account the basin configuration of Miller (2008).

Recent mapping of planation surfaces has been undertaken by Picart et al. (2020) over southern Namibia, extending as far north as the Windhoek Graben. They interpreted the remnants of eight discrete surfaces, which in their interpretation are suggested to have formed at different times, sequentially from S0 (oldest) to S8 (youngest). In our study area, the oldest surfaces identified by Picart et al. (2020) (S0, S1 and S2) are not dated. S3 is argued to post-date the 'Rehoboth phonolites' and hence is of Middle Eocene age. The floor of the Windhoek Graben is mapped as surface S6, which Picart et al. (2020) dated as Early Miocene by long range correlation to the outcrops of the Middle Miocene in the Namib Desert. Relationships between these surfaces and our reconstructed surfaces are considered further in the Discussion.

5. RESULTS

5.1. Tectonic Geomorphology of The Windhoek Graben

Prominent escarpments are developed along the western and eastern margins of the Windhoek Graben, particularly in the central and northern areas (Fig.4). The mapped position of the major bounding faults of the graben coincides with the bases of both the eastern and western escarpments and, on this basis, both escarpments can be interpreted as eroded fault scarps (cf., Wallace, 1977) (Fig.5A).

In detail, the morphology of the two bounding fault scarps differs considerably. The eastern scarp conforms more classically to a highly degraded normal fault scarp, with a generalized dip of between 1.7 and 2.0 degrees up to a prominent ridge separating the scarp from the relatively undissected, gently east-dipping footwall dip slope (Fig. 3A). The topographic elevation difference of the crest of the scarp with respect to the graben floor elevation varies systematically along strike reaching a maximum value approximately midway along the length of the fault, with an elevation difference of c. 600 m. Scarp drainages are spaced every kilometer or so along strike, and exploit local heterogeneities in the basement rocks of the footwall. These short-headed rivers flow with stepped profiles down to the base of the scarp at the hanging wall cutoff (Fig. 5B). The strike of the scarp is generally NNE-SSW in the central area, but changes across a series of segment boundaries or relay structures toward the northern area, where it trends NNW-SSE. These changes of strike, and in some cases of lateral offset of the base of the scarp, allow the entire structure to be sub-divided into a number of segments, bounded by relay structures. These relay structures appear to have been breached at an early stage in the displacement history of the bounding fault and are now fully hard-linked (cf., Trudgill and Cartwright, 1994). The breached relays are loci for the more prominent scarp drainages, often with the largest catchment areas of any of the scarp drainages. The scarp drainages feed as tributaries into the main axial drainage of the graben floor, and locally grade into the profile of this axial river (Fig. 5C). However, the axial river does not incise or behead any of these tributary streams, and there is no evidence that the axial river has cut significantly downwards into the relatively smooth topography of the graben floor. Fresh triangular facets, waterfalls, or beheaded spurs are not found along the footwall scarps, suggesting there has been no significant recent displacement on the eastern bounding fault (Wallace, 1977; Hodge et al., 2020).

In contrast, the western bounding scarp is less pronounced, and more irregular than its eastern counterpart. The elevation difference is somewhat lower too, with a maximum value between footwall crest and hanging wall cut off of c. 500 m, also located approximately midway along the fault (Fig. 5A). The scarp itself is harder to differentiate as a specific feature because of the remnant topography developed along the hanging wall (Fig. 3A). It is possible that the bounding fault is itself broken down into a number of smaller step or terrace faults, as often occurs in more asymmetric half-graben in other rift systems (e.g. Rosendahl et al., 1986; Morley et al., 1990). There are a number of transverse, tributary drainages contributing to scarp degradation, but these are less well organized and more irregularly spaced than on the eastern margin. The axial river cuts across the graben floor from the eastern to the western margin and flows for a few kilometers along the base of western scarp, locally incising into the footwall before making a junction with the Swakop River beyond the northern limit of the graben (Fig. 3A).

The graben floor is relatively smooth over much of its area with a gently undulating landscape, modest stream incision of a few tens of meters, and local development of what appears to be residual relief, with basement hills rising c. 100 m above the smooth graben floor. This residual topography is most clearly seen in the north of the graben, where a range of prominent hills are oriented WSW parallel to the structural grain of the bedrock and appear to continue into the footwall area on the western margin (Fig. 3A). Basement rocks are exposed over most of the graben floor, with only very localized areas where aeolian and fluvial sediment cover of a few meters thickness is preserved. The elevation of the graben floor climbs steadily towards the southern end, where the topographic expression of the graben dies out against the high ridges that cross the area in a WSW orientation and which can be linked to the major north dipping Damaran thrusted quartzites mapped by Guj (1967) and Hoffmann and Schreiber (2011) (Fig. 3B).

5.2. Igneous Activity

As noted briefly above, there are several suites of preserved igneous bodies located within some tens of kilometers of the southern end of the Windhoek Graben, and their possible spatio-temporal connection with the development of the graben has previously been highlighted (Miller, 2008). These igneous bodies are described below in four groups based on their timing and geochemical characterization, from the pioneering work of Gevers (1932, 1934), Ferreira et al. (1979), and Marsh (2010).

5.2.1. The Stalhart Cluster of Phonolites

The Stalhart Cluster is a group of phonolitic extrusions and/or plugs that occur over an area of 180 km^2 some 50 km SW of the Windhoek Graben (Fig.6) (Marsh, 2010). The Stalhart Cluster was erupted onto a highly eroded landsurface composed of Damaran quartzite (1:250,000 geological map, sheet 2316). This eroded landsurface is sporadically covered by Quaternary sediments, deposited by south-flowing rivers that rise in the southern ridges of the Namibian Highlands (Fig. 6).

The Stalhart Cluster comprises sixteen small bodies that, in comparison with the better preserved Klinghardt phonolites, have been interpreted as eroded tholoids (Marsh, 2010). They form irregular conical hills with slightly depressed cores compared to their margins. The largest preserved body is c. 200 m high and c. 1500 m wide, but the average preserved height of the group is c.100 m with average widths of c. 1 km. The cluster was first described as a group of both phonolites and trachytes with two minor alkaline plugs (De Kock, 1934), but Marsh (2010) demonstrated from extensive sampling that, the igneous bodies are all geochemically better classed as true phonolites the exception of a single minor alkaline plug. The phonolites are aphyric (<2% phenocrysts) to porphyritic (<15% phenocrysts), where the phenocrysts range from highly altered nepheline and sanidine to minor unaltered feldspathoids, followed by microphenocrysts of aegirine. More altered phonolites and steeper dips of sheeting joint and flow banding are radially concentrated in the inner core of the circular outcrops (Marsh, 2010). A fresh nepheline sample of the Stalhart cluster was dated using the $^{40}Ar/^{39}Ar$ method, yielding a mean age of 52.6 ± 0.3 Ma (Marsh et al., 2018). The associated alkaline plug is holocrystalline, porphyritic with mainly salite phenocrysts, classified as plagioclase-bearing nephelinite (Marsh, 2010), and is more weathered than the phonolites (Miller, 2008). The Stalhart phonolites and the alkaline plug have an identical initial $^{87}Sr/^{86}Sr \approx 0.7043$ (Marsh, 2010).

The tholoid bodies were erupted as lava domes composed of viscous short-range flows onto the eroded basement surface (Marsh, 2010). This surface forms part of an extensive broad valley or amalgamated set of valleys whose topographic relationships to the Namibian Highlands can best be seen in a series of longitudinal and transverse profiles that we constructed with respect to the present-day valley network in this area (Fig. 6C). From the gently dipping surface on which the cluster was erupted, at an elevation of 1580 ± 40 m, the surface extends smoothly northwards to the slopes of the prominent WSW Damaran quartzite ridges. Profiles ii and iv of Fig.6C, were constructed to meet the southward projection of the western footwall to the Windhoek Graben. They show that this surface (referred to here as the pre-52Myr (P52) surface) is remarkably smooth, with no significant steps or scarps over the c. 60 km distance along the connecting profiles v and vi of Fig. 6C. In contrast, the transverse profiles show a subtly concave topography in a northerly direction, with more pronounced valley forms closer to the higher topography of the quartzite ridges (Fig. 6C, profiles i to iv). Taken together, these profiles show that the P52 surface can be interpreted as a preserved valley or valley system with generally south-flowing axes, and at whose southern end the Stalhart Cluster was erupted (Fig. 6A). This implies that significant incision of some hundreds of metres magnitude must have been carved by the south-flowing rivers prior to eruption of the Stalharts. This interpretation thus completely endorses the view of (Marsh, 2010) that the lava domes were emplaced onto a surface not very different in its relief from the present-day surface and have been eroded down to expose the upper parts of the conduits where the steeper dips of flow banding and sheeting joints toward the core areas can be seen exposed today.

5.2.2. The Aris Cluster of Phonolites/Trachytes

The Aris Cluster is located ~15 km south of the Windhoek Graben close to the head of the Usib Valley, at an elevation of 1770 m ± 30 m (Fig. 6 and 7A). The Aris Cluster comprises six individual phonolite bodies (Marsh, 2010), about five smaller bodies of trachyte, and a small body of trachyte breccia (Miller, 2008). Of these, the two largest bodies are Schildkrotenberg and Huguamis, composed of phonolite, and forming strikingly conical hills, and partially dissected by the south-flowing Usib River (Fig. 7A-C). Schildkrotenberg rises 290 m above the valley floor and is 1900 m wide at its base.

Topographic profiles between the Stalhart and Aris Clusters demonstrate that the elevation of the Aris Cluster is about 200 m higher up and further up the valley relative to the Stalhart cluster and is surrounded by highly dissected terrain on three sides (Fig. 6A and C). The head of the Usib River occurs 5 km north of the cluster, where a WSW striking ridge of quartzite acts as a drainage divide between the Windhoek Graben in the north and the Usib Valley in the south. The morphology of the Usib Valley opens up to the south, from the narrow valley onto which the Aris

Cluster was erupted to the more laterally extensive surface onto which the Stalharts were erupted. The valley is predominantly cut into pre-Damaran paragneiss and this basement is locally covered by a thin veneer of Quaternary sediments (Fig. 7B and 7C) (1:250,000 geological map, sheet 2216).

The nature of the phonolite outcrops (e.g., morphological shape, degree of weathering, and structural trend) is identical with the Stalhart phonolites (Marsh, 2010). The phonolites are holocrystalline fine- to medium-grained, mainly aphyric to lightly porphyritic with insignificant trachytic textures (Miller, 2008; Marsh, 2010). The groundmass of both aphyric and porphyritic phonolites is dominated by nepheline and sanidine (Marsh, 2010). The Aris phonolite composition is marked by a more homogeneous concentration of trace elements with generally lower TiO_2, Sr, Nb, and higher Na_2O, Rb, Zr, Th, Pb, and REE compared to the Stalhart phonolites (Marsh, 2010). One phonolite body was dated using a whole-rock $^{40}Ar/^{39}Ar$ method and generated an age of 32 ± 0.2 Ma (Burger and Walraven, 1976). This date has been questioned as to its accuracy by Marsh et al. (2018), who noted that the whole rock method generally lacks precision in comparison to more modern single grain techniques and suggests from comparisons in the ages of the Klinghardt Mountains that this date could easily be in error by up to 10 Ma.

The trachytes and trachyte breccia are not as extensively described as the phonolites, and their bodies are ~10 to 30% smaller than the size of the biggest phonolite (i.e., Schildkrotenberg). The formers appear as plug-like features and are mainly made of massive, chemically altered trachytes, tuffaceous trachyte, and breccia containing gneiss fragments along the margins of the body (Miller, 2008; Marsh, 2010). The latter forms a pipe-like feature and consists of trachyte breccia containing abundant xenoliths (e.g., gneiss and quartz) (Niku Paavola, 1997; Miller, 2008).

Marsh (2010) interpreted the Aris phonolites as eroded tholoids based on their morphological and petrological character, similarity to the Stalhart phonolites, and argued that they were also emplaced on a surface not very different in its relief from the present-day surface (PDS) (Fig.7D). Higher Rb, Zr, Th, Pb, REE, but lower Sr, Nb, and narrow compositional range for the trace elements indicate that the Aris phonolites were highly evolved and originated from a single magma system (Marsh, 2010). The close proximity of the bodies of trachyte and breccia are then postulated as associated members during the emplacement of mafic magmas to the surface. However, the magmatism of the Aris cluster was different from the Stalhart cluster as Aris phonolites are younger, morphologically less eroded, and geochemically more homogeneous than the Stalhart phonolites (Marsh, 2010).

Combining all the observations with the topographic profiles, we conclude that the valley erosion that pre-dated the eruption of the Stalhart Cluster and formed the pre-52Myr surface was not obviously modified by further erosion prior to the eruption of the Aris Cluster in the Oligocene (Fig. 7D). Considering that there has been relatively limited subsequent erosion of the tholoids bodies of the Aris Cluster, we also conclude that the Usib Valley was largely formed in its present configuration by the Early Eocene (i.e., prior to 52 Ma).

5.2.3. The Regenstein Vent Complex

The Regenstein vent complex is situated west of the southern end of the Windhoek Graben (Fig.6A). It differs in its topographic position from the Aris and Stalhart Clusters in that it forms an erosional depression within one of the prominent SW ridges of Damaran quartzite (Ferreira et al., 1979), with an average elevation of 2080 m or ~300 m above the valley floor location of the Aris cluster (Fig. 7C). This ridge is cut by the head of the Usib Valley at c.1900 m asl before it climbs to the far northeast of the Regenstein complex, reaching up to 2465 m asl, with an average peak elevation of about 2400 m over a 15 km length (Auas Mountains, Fig. 7A).

The outcrop of the irregular-shaped vent mostly occupies low ground, but the contact with the basement, quartzites can be traced over some 160 m of vertical relief, up to an elevation of 2120 m (Gevers, 1934) (Fig. 8). These basement quartzites have been found to be important aquifers in the region surrounding Windhoek (Miller et al., 2018). The vent complex consists of the main body with a length of 2 km and width of 1 km in an elliptical shape trending NNW-SSE and an isolated, smaller associated body located 500 m to the north (Gevers, 1934; Ferreira et al., 1979). The area is heavily faulted (Gevers, 1934). A set of NNW-SSE striking faults occur at the western margin of the vent complex, extending from the northern satellite body to the main body in the south (Ferreira et al., 1979). The faults are marked at outcrop by silicified, brecciated shear zones and the most prominent of these (Fig. 9A) runs from the center of the main body to the northern satellite body (Ferreira et al., 1979).

The Regenstein vent complex comprises distinct bodies of mainly pyroclastic breccia and marginally mafic rocks and dykes (Ferreira et al., 1979), and has not been radiometrically dated, presumably because of the high degree of hydrothermal alteration of the vent materials. The breccia is composed of fragments of highly-altered accidental clasts (mainly quartzite) and accessory clasts (phonolites, trachytes, and lapilli breccia) with clay minerals and carbonate cement ranging in diameter from 2 mm to 0.5 m. The trachytic-quartzite breccia is predominantly observed in the in the low lying outcrop areas of both vent bodies (Ferreira et al., 1979; Miller, 2008). Marginal bodies of fluidised quartzite breccia are found discontinuously along the western edge of the vents with a sharp contact with the surrounding highly-fractured Damaran quartzite. Radial fractures from the center of the vent cut through these two rock types and are filled by fluidized, well rounded quartzite fragments, testifying to the explosive nature of the eruptions resulting in the vent structure (Miller, 2008). The distribution of these pyroclastic breccias has been extended ~200 m below the present-day surface based on an exploratory borehole in the northeastern area (Ferreira et al., 1979).

A noticeable mafic body forms an irregular conical hill with a diameter of 250 m in the northwestern area of the main vent, and its location is spatially aligned with the NNW-SSE trending fault. It is composed of heterogeneous mafic phonolites showing the medium-grained, hypidiomorphic-granular texture of mafic xenoliths enclosed in a fine-grained, porphyritic matrix (Ferreira et al., 1979; Miller, 2008). Dykes are thin and radiate from the center of the vent, demonstrating a sharp contact with the breccias (Gevers, 1934; Ferreira et al., 1979). These dykes are composed of phonolite, biotite trachyte, and aegirine trachyte. The trachyte dykes cut through the dykes of phonolites (Miller, 2008, and references therein).

Emplacement of the Regenstein vent complex has been interpreted to be phreatomagmatic in repetitive cycles, and its location is thought to be structurally controlled (Ferreira et al., 1979). The interpretation of the eruptive mechanism and cyclic emplacement history is supported by an absence of essential elements in the breccias, by the silicified fault zones, by fracturing and fragmentation of the marginal quartzite, by hydrothermal alteration, and by the various textures of the pyroclastic breccias (e.g., trachyte and phonolites). Structural control is corroborated by the shape of the main vent in relation to an important cluster of normal faults identified by Ferreira et al. (1979) (Fig. 3B). This cluster of normal faults coincides approximately with the longest axis of the vent complex and the locations of the mafic body, and an anomalous trend of Pb-Zn mineralisation (Ferreira et al., 1979). Ferreira et al. (1979) proposed that the breccia was first emplaced with the phonolite and trachyte dykes, followed by the emplacement of mafic rocks that were finally altered by hydrothermal processes.

The vent complex thus represents a deeper conduit system that has been exhumed due to erosion of a land surface that existed at the time of its emplacement. From their detailed mapping, Ferreira et al. (1979) argued that this palaeo-surface was about 400 m above the present erosion level of the main vent structure (Fig. 7D). Topographic profiles from the Regenstein Vent Complex to the Aris and Stalhart Clusters demonstrate that this palaeo-surface (referred to as a pre-Regenstein surface (PRS)) onto which the phreatomagmatic products erupted, possibly as a maar-like pyroclastic cone, could not have been a continuation of the pre-52Myr surface (P52), but must instead have been at a higher level by several hundred meters (Fig. 7D). This is entirely consistent with Gevers's (1934) view that the Regenstein Vent is Late Cretaceous in age and significantly older than the Aris Cluster.

5.2.4. Dykes, Plugs, and Brecciated Fissures of the Southern Windhoek Graben area

Dykes and elongated plug-like intrusions mainly of trachyte were mapped in great detail by Gevers (1932) and Guj (1967). They found more than 160 individual intrusive bodies that are distributed mainly in the southern limit of the Windhoek Graben and in the vicinity of the WSW trending quartzite ridge (Fig. 6A). The longest dyke in this area is 12 km, and the average thickness is less than 10 m (Gevers, 1932; Miller, 2008). The dykes are almost vertical, dipping generally 75º to 80º in the southern Windhoek Graben area (Gevers, 1932). In the southeastern region of the major WSW trending quartzite ridge, the width of individual trachyte intrusions may reach up to 200 m and these are generally porphyritic in texture with vertical contacts with the basement rocks (Guj, 1967; Miller, 2008). Individual trachyte dykes intersect the crest of the ridge at an altitude of c. 2300 m, and their contacts can be traced down to c. 1900 m, implying that several hundred metres of erosion has taken place post-emplacement. In order to compare the orientation of these dykes and plugs with the orientation of minor normal faults in the southern area of the Windhoek Graben, we also measured the mapped strikes of 162 linear trachyte plugs and dykes using compilations of Gevers (1932) and Guj (1967). Their orientations range from NW to NNE strikes (Fig. 6B), and notably, this range of strikes compares very well with the strikes of the minor normal faults mapped in the southern Windhoek Graben (Fig.5D).

The dykes are predominantly trachytic and phonolitic in composition and extend from the center of small vents, and from trachyte and breccia plugs (Gevers, 1932). Sets of dykes cut across the Damaran thrust faults on the major WSW trending ridge while other groups of dykes terminate at thicker quartzite units juxtaposed across the thrust faults (Gevers, 1932; Miller, 2008). Most of the dykes are highly brecciated and are associated with siliceous impregnation zones and fissures that were brecciated and filled by siliceous material (Gevers, 1932). The dykes appear inside the core, alongside, or as an extension of the silicified fault zones in the southern Windhoek Graben and also cluster together as a dyke swarm between two sets of northerly striking normal faults (Miller, 2008). Only a few dykes cut across the silicified fault zones at a high angle (Miller, 2008). The silicified breccia dykes generally exhibit displacements of less than a few meters along their strike, but groups of fissures are both longer and wider along both margins of the southern Windhoek Graben (Gevers, 1932). Thermal springs are associated with the brecciated fissures, with the highest temperature reaching up to 79.8ºC in the eastern margin area of the southern Windhoek Graben (i.e., Pahl Spring) (Gevers, 1932).

The geometry, distribution and composition of the dykes and brecciated fissures (breccia dykes) have been taken by (Gevers, 1932; Miller, 2008) to argue that they were emplaced during intense phreatic-fumarolic activities and were spatially and temporally associated with the small vents, trachyte plugs, and the minor faults of the southern Windhoek Graben. Given that the orientations (and in some cases locations) of dykes generally correspond to the strikes of the minor normal faults in the southern area of the Windhoek Graben (Fig. 5D and Fig. 6B), this points to a direct link between the normal faulting with a generally E-W extension direction and this phase of igneous activity.

6. DISCUSSION

Two main questions arise from the preceding summary of the observations regarding the southernmost faults of the Windhoek Graben and the four sites of igneous activity, namely:

(1) What is the precise relationship between the broad swathe of normal faulting mapped in the area of the southern Windhoek Graben, the abundant trachyte and breccia dykes distributed along the major quartzite ridge, foothills, and the Regenstein Vent Complex?

(2) How can the topographic relationships between the various igneous centers be used to build a chronological framework for the initiation of the Windhoek Graben?

6.1. The Windhoek Graben and Its Relationship to the Regenstein Vent Complex

Based on the relationships between igneous bodies and normal faults (Fig. 9), we suggest that the southern limit of the Windhoek Graben abuts and terminates against the WSW trending quartzite ridge on which the Regenstein Vent Complex is situated. We furthermore argue that the graben initiated in the same time interval as the igneous activity that resulted in the Regenstein Vent Complex and the cluster of trachytes dykes, breccia dykes, and trachyte plugs and vents in this area (Fig. 9). Three main arguments favour this interpretation.

Firstly, normal faulting that is clearly linked to the development of the southern region of the Windhoek Graben is demonstrably spatially and temporally linked to the emplacement of the

Regenstein Vent Complex through the propagation of the graben-margin faults and smaller faults that displace the phreatomagmatic breccia dykes (Gevers, 1932; Ferreira et al., 1979). We essentially follow the interpretation of Miller (2008), who proposed that the west bounding fault of the Regenstein Vent Complex runs continuously to the Swakop River in the north (Fig. 3A).

Secondly, the common phreatomagmatic nature of the Regenstein Vent Complex and the trachyte dykes is a strong argument to support Gevers' (1932) arguments linking these suites of intrusive bodies. It seems unlikely that such a large number of spatially clustered phreatomagmatic products and silicified fault zones would be linked purely coincidentally. Therefore, these observations collectively argue that the initiation of the Windhoek Graben is spatially and temporal related to the igneous activity centred on the Regenstein Vent Complex.

Finally, the strikes of the >100 normal faults distributed in this southern terminus region of the Windhoek Graben have a closely comparable distribution to the strikes of the >160 dykes in the same area (Fig. 5D and 6B). Both sets of structures evidently developed in a very similar stress field, in both cases with the minimum compressive stress horizontal and oriented E-W. The simplest interpretation of this uniformity of the driving stress state for dykes and faults is that the normal faulting occurred during the same tectonic episode as the dyke intrusion. Since the dykes are linked in time and also partly spatially to the Regenstein Vent Complex, we argue that the initiation of normal faulting was also most probably linked to this igneous event. This interpretation is favoured over the alternative view that the faulting and dyke formation are completely unrelated, because the normal faults do not systematically offset the dykes, and neither do the dykes cross-cut the normal faults.

6.2. Chronological Framework for the Initiation of the Windhoek Graben

From the preceding arguments, we suggest that we can constrain the timing of initiation of the Windhoek Graben by placing the emplacement of the Regenstein Vent Complex into a chronological framework based on the correlation of remnant erosional surfaces.

The reference datum for our proposed chronological framework is defined here as the surface onto which the eruptive products from the Regenstein Vent were emplaced. Ferreira et al. (1979) estimate that the palaeo-surface intersected by the vent at the time of eruption was c.400 m above the height of the current topographic depression (Fig. 9C). They based this estimate on consideration of mapped contact morphology, drilling results and vent fill characteristics, recognizing that considerable erosion of c. 400m of the vent has occurred since emplacement. Trachyte dykes within the basement are mapped at similar elevations further east along the major quartzite ridge as shown in (Fig 7B and 9A). Making the assumption that these trachyte dykes were intruded with their upper tips close to the palaeo-surface, then we can interpolate between these two positions on the ridge to identify a conservative position for this palaeo-surface, which we refer to here as the pre-Regenstein surface, PRS (Fig. 7D and 9C).

To place any form of absolute age constraint on this surface requires long-range correlation and large assumptions in making that correlation. However, previous workers have been struck by the accordance of many of the local summits of the series of ridges that stand above the plains and valleys of the broad area flanking the Aranos Basin and the Namibian Highlands (e.g., Fig. 4C and Fig. 6C (iii, vi, viii)). Gevers (1942) argued that the accordant summit heights of the ridges must represent the remnant of a higher, older surface that forms an envelope to the present-day topography of the Namibian Highlands. He also described a spectacular example of superimposed drainage in this area, where the Usib River cuts through one of the more distal ridges (Fig. 10). He argued from the cleft in the ridge cut by the river down to its present-day level, that there must have been a gently inclined surface at or above the level of the present-day ridge crest in order for the river to cut down through the ridge rather than circumnavigate around the ridge. This surface on which the ancestral Usib evidently flowed southwards was most probably the pre-Regenstein surface (PRS), since it conforms to the simple interpolation of ridge summit heights shown in Fig. 10C and D.

Correlation of this pre-Regenstein surface (PRS) southwards from the major quartzite ridge gives a notional impression of a relatively planar older surface, but how does this surface correlate into regions of subsurface chronostratigraphic control? The closest outcrops of Mesozoic sedimentary rocks to the data point at the Usib Gorge are close to Mariental, some 150 km to the south. Projection of the PRS could justifiably be made with additional examples of superimposed drainage on some of the more distal ridges (Fig. 10B and D) and this closes the gap to less than 50 km. However, an even more well-constrained projection can be made into the subsurface of the Aranos Basin using water boreholes as the primary data points for key surfaces (Fig. 11). These boreholes calibrate the base of the Kalahari Group sediments (interpreted as Late Cretaceous in age; Miller, 2008) and also the Base of the Karoo Supergroup. The projection along this section shows that the PRS projects close to the convergence of three major unconformities, namely the Base Nama Group, Base Karoo, and Base Kalahari. We can reasonably rule out the Base Nama as being the progenitor of the PRS on the basis that it dates from the Late Precambrian to Cambrian. This leaves us with the more reasonable interpretation that the PRS is the lateral correlative of a compound unconformity surface of these two other major unconformities, i.e., a pre-Karoo surface modified by events leading into the latest Cretaceous/earliest Cenozoic. By correlation with the regional cross-sections presented by Picart et al. (2020), our PRS corresponds to their oldest surface, S0, but they did not provide a date for the formation of this surface. Similarly, we correlate our P52 surface as being approximately equivalent to the undated surfaces S1 and S2 of Picart et al. (2020).

The questions arising from this interpretation of the PRS are firstly, is there any independent support for this notional correlation, and if so, what does it imply about relative uplift of the Namibian Highlands with respect to the neighbouring areas? The interpretation that the pre-Regenstein surface (PRS) is a Late Cretaceous to Early Cenozoic modified remnant of a pre-Karoo surface is strongly supported by evidence of deep incision of the Karoo Supergroup sediments within the Aranos Basin, as documented by Miller (2008). Correlation of lateritic alteration of the incised Karoo clastic sediments led Miller (2008) to the conclusion that a south-eastward flowing river, which he termed the Aranos River, originated in the Late Cretaceous. His argument is based on meticulous comparison of depths to the top of the lateritized Karoo in water boreholes, and the preserved thickness of the laterite layer. This demonstrated that the fluvial incision occurred during the humid climatic conditions required for the development of the laterite profile, which from regional considerations he placed in the latest Cretaceous to earliest Cenozoic. He also noted that this incision most likely coincided with the onset of subsidence in the Aranos Basin, which is widely regarded as Late Cretaceous in age (Partridge and Maud, 1987; Moore, 1999). It seems entirely plausible that this Late Cretaceous origin of the Aranos River was replicated for neighbouring major drainages, including that responsible for the erosion of the Usib Valley, which we have already argued must have been almost completely incised close to its present-day profile by 52 Ma.

Extending this logic to the wider context of the Namibian Highlands, therefore, leads us to conclude that the most likely period for onset of valley development on the eastern and southern flanks of the highlands was during the Late Cretaceous. Evidence for the timing of initiation of the westerly flowing rivers like the Kuiseb and the Swakop (see Fig. 2A) is lacking at present, but Clemson et al. (1997) and Aizawa et al. (2000) both noted the development of incised valleys in the offshore corresponding to the modern river outlets, that were clearly cut within the latest Cretaceous. This would strongly suggest that well-established rivers were flowing across the site of the Great Escarpment at this time, and were probably rising on the western flank of the Namibian Highlands. If correct, this would in turn suggest that the Namibian Highlands were the locus of a radial drainage pattern dating from this time, and point to a period of pronounced doming of the pre-Regenstein surface (PRS) at this time. Similarly, Raab et al. (2002) argued from their analysis of the denudational history of central Namibia using fission track data that the major erosion and sediment transport into the offshore took place between 80Ma and 60Ma. It is not possible to be more precise in the timing, but we note that numerous previous studies have argued for a significant period of uplift of the area further south towards the Orange River in the Late Cretaceous either from Low T thermochronology or from studies of the sediment transport into the offshore basins (Aizawa et al., 2000; Jacob, 2005; Bluck et al., 2007; De Wit, 2007; Baby et al., 2020; Wildman et al., 2021).

In conclusion therefore, the evidence assembled here points to a major phase of domal uplift of the Namibian Highlands in the Late Cretaceous. This overlapped with or was coeval with the emplacement of Regenstein Vent Complex and associated trachyte dykes and hydrothermal system, and critically, with the initial fault growth of the Windhoek Graben. Based on this, it is possible to view the origins of rifting in this area as part of a geodynamic context that considerably precedes the much more recent propagation of the Central African Rift System (Daly et al. 2020), but we cannot exclude the possibility of much more recent activity of the Windhoek Graben linked to westward propagation of this younger rift system.

It is tempting to speculate that the magmatic event, the uplift, and the graben initiation are all linked to wider events along the borderlands between the continental margin and the continental interior, that completely transformed the drainage development in the greater Namibian region. Understanding these events in the broader context of the geodynamic evolution of Southern Africa should ultimately lead to a fuller understanding of why the Windhoek Graben developed at this time and in this place.

7. CONCLUSIONS

1. The initiation of rifting of the Windhoek Graben is most probably related in timing to the emplacement of the Regenstein Vent Complex and its associated suite of dykes and phreatomagmatic trachyte vein and breccia dykes.
2. Morphological correlations clearly establish that previous suggestions that the vent complex predated the eruption of the Stalhart Cluster are correct.
3. Correlation of a reconstructed palaeo-surface (PRS) with deep erosion and regolith formation in the nearby Aranos Basin gives a most likely timing of Late Cretaceous to Early Cenozoic for Regenstein and the associated trachyte dyke cluster, and this can be no younger than the 52-Ma date for the Stalhart Cluster.
4. The initiation of rifting is interpreted to coincide with the initial phase of magmatic activity in the area in the Late Cretaceous to Early Cenozoic.
5. The initiation of rifting was closely linked in timing to the formation of a localized domal uplift, responsible for the development of a radial drainage network that has persisted, albeit intermittently with climatic variations to the present day.

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FIGURES

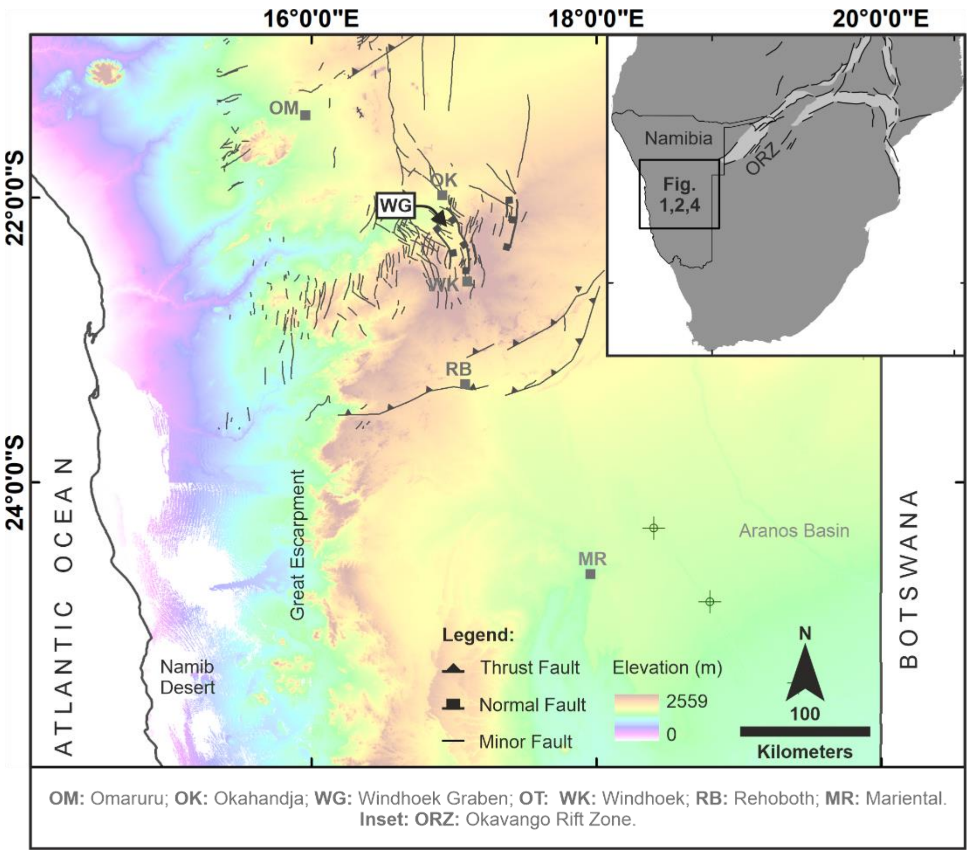
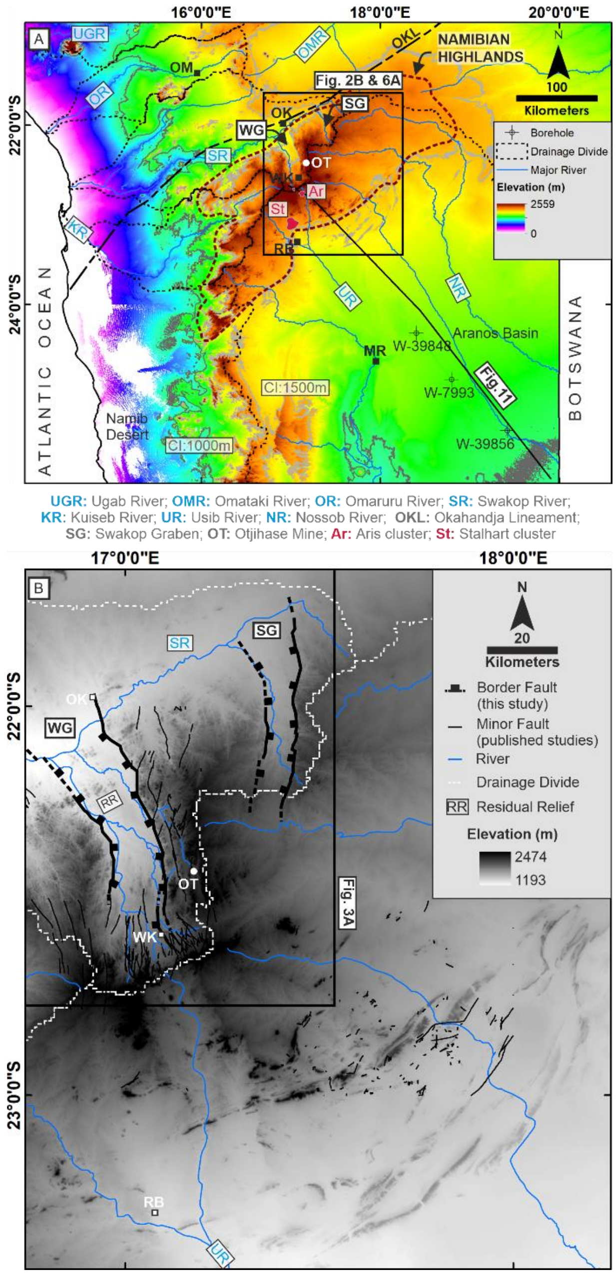


Figure 1. Regional topographic map of the study area in central Namibia overlain by a compilation of structural maps by Guj (1967), Miller and Schalk (1980), and Picart et al. (2020). Inset: Rift system in southern Africa (simplified from Morley et al. (1990), Kinabo et al. (2007), Ponte et al. (2019), and Daly et al. (2020)) and Namibia location.



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723 **Figure 2. (A)** The physiographic context of central Namibia showing Windhoek Graben (WG) and Swakop Graben (SG) at the center area and the Aranos Basin to the south. Representative boreholes are taken from Miller (2008). **(B)** Fault interpretation of Windhoek and Swakop Grabens is superimposed on the topographic map. Minor faults are taken from Hoffmann and Schreiber (2011) for Windhoek area and Miller and Schalk (1980) for regional area. Dip direction of the minor faults in southern Windhoek is adapted from Miller et al. (2018). Abbreviations are in (A).

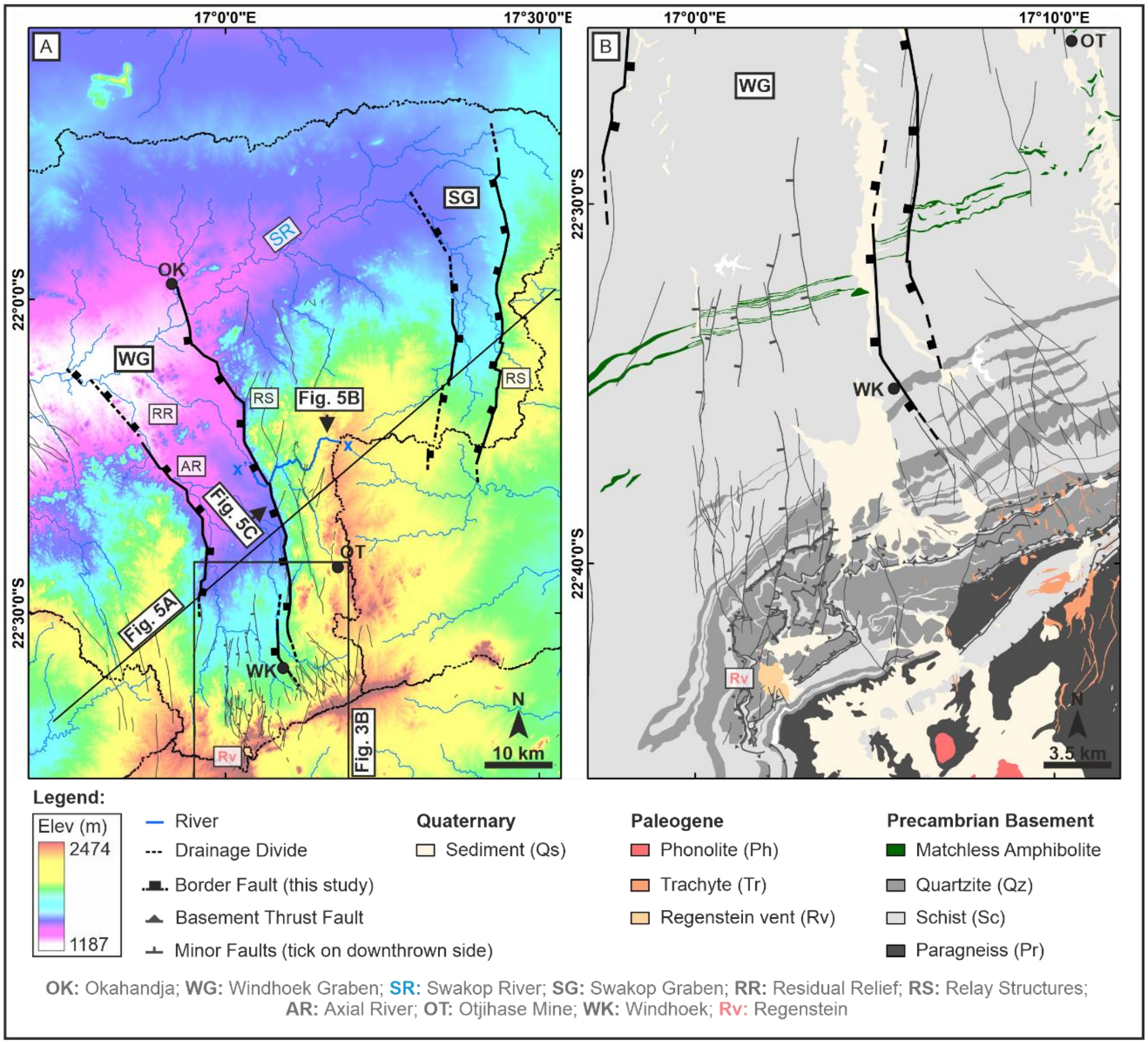

Figure 3. (A) Topographic map overlain by a structural map of the Windhoek Graben showing outcropping segmented, curvilinear normal faults and prominent escarpments along the footwall. Minor faults are taken from Hoffmann and Schreiber (2011) for Windhoek area and Miller and Schalk (1980) for regional area. Dip direction of the minor faults in southern Windhoek is adapted from Miller et al. (2018). Note: Regenstein vent complex (Rv) is located to the southwest of Windhoek city (digitized from Hoffmann and Schreiber (2011)). **(B)** 1:250,000 geological map of the area (sheet 2216, Geological Survey of Namibia) showing southern area of the Windhoek Graben. See (A) for location.

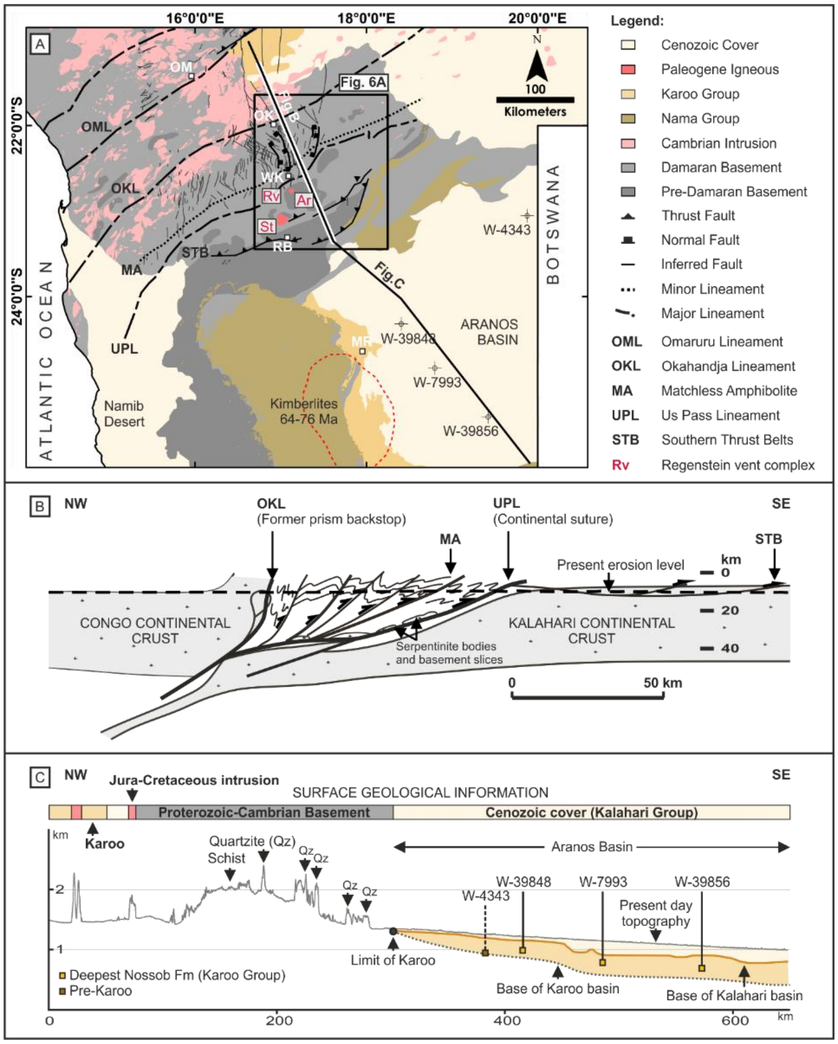

Figure 4. (A) Tectonostratigraphic zones of Damaran orogenic belt (modified from Hoffmann (1983), Miller (1983), and Miller and Schalk (1980)) overlain by Nama and Karoo Groups (Miller and Schalk, 1980), igneous bodies (Burger and Walraven, 1976; Moore et al., 2008; Marsh, 2010), N-S grabens (digitized from Picart et al., 2020), and Cenozoic cover (Miller, 2008). Representative boreholes are taken from Miller (2008). **(B)** NW-SE cross-section illustrates crustal-scale architecture (adapted from Kasch,1986; Kukla, 1992). **(C)** Regional NW-SE topographical profile depicts morphology and surface geological information of the study area. The Aranos basin-filled architecture is derived from Miller (2008). See panel (A) for locations of (B) and (C).

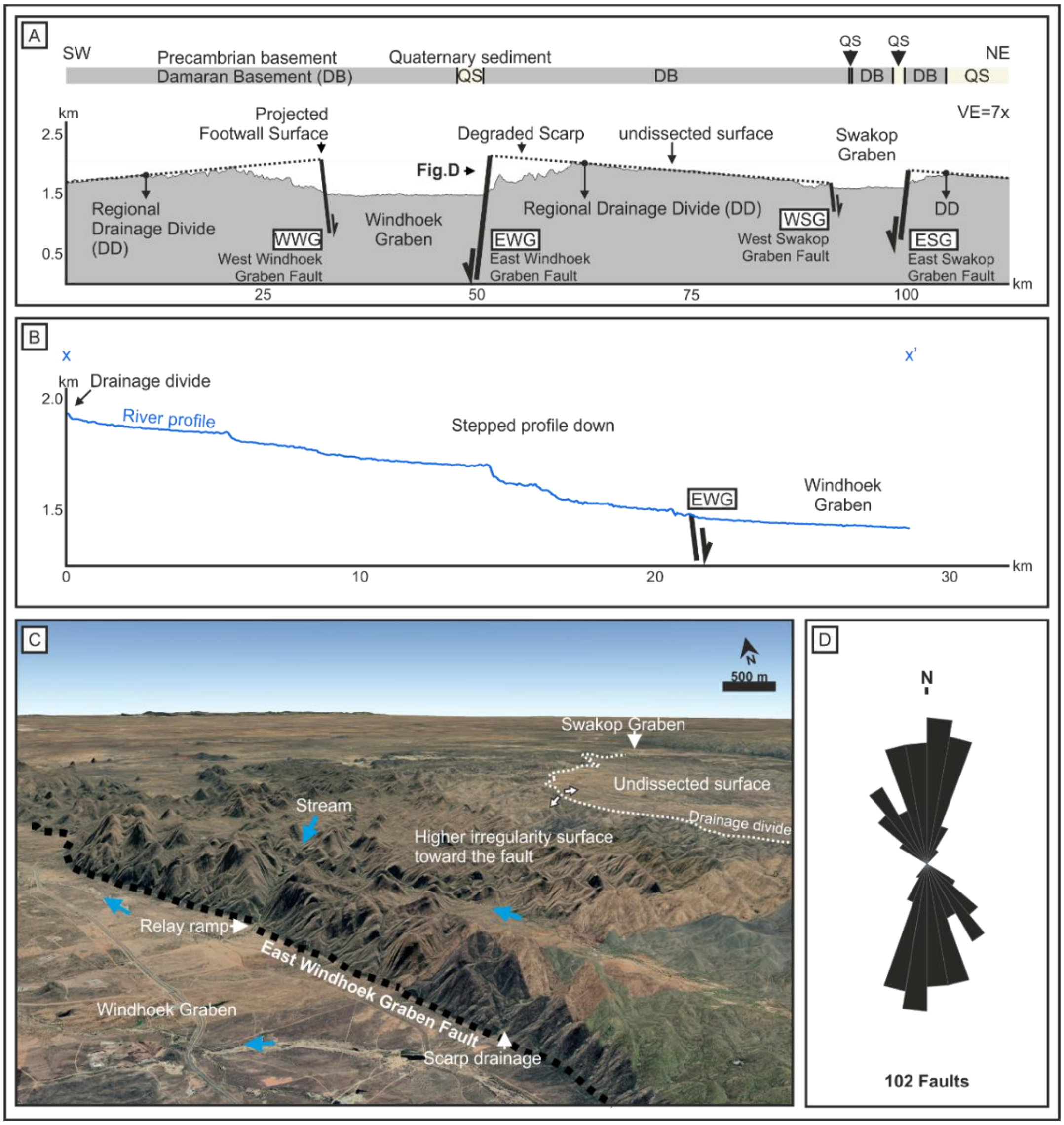

Figure 5. (A) SW-NE topographic profile of Windhoek and Swakop grabens showing the four major bounding faults. Lithology distribution is added on top of the profile (lithology taken from 1:250,000 Namibia Geological Map sheets 2116 and 2216). **(B)** A longitudinal river profile of scarp drainage (ENE-SSW) depicts a stepped-down profile. See Fig.3A for location. **(C)** An oblique aerial image shows a fault scarp morphology in the eastern margin of the Windhoek Graben. See Fig.3A for location. **(D)** Orientation of minor faults in the southern Windhoek Graben (measured from structural map presented by Hoffmann and Schreiber (2011)).

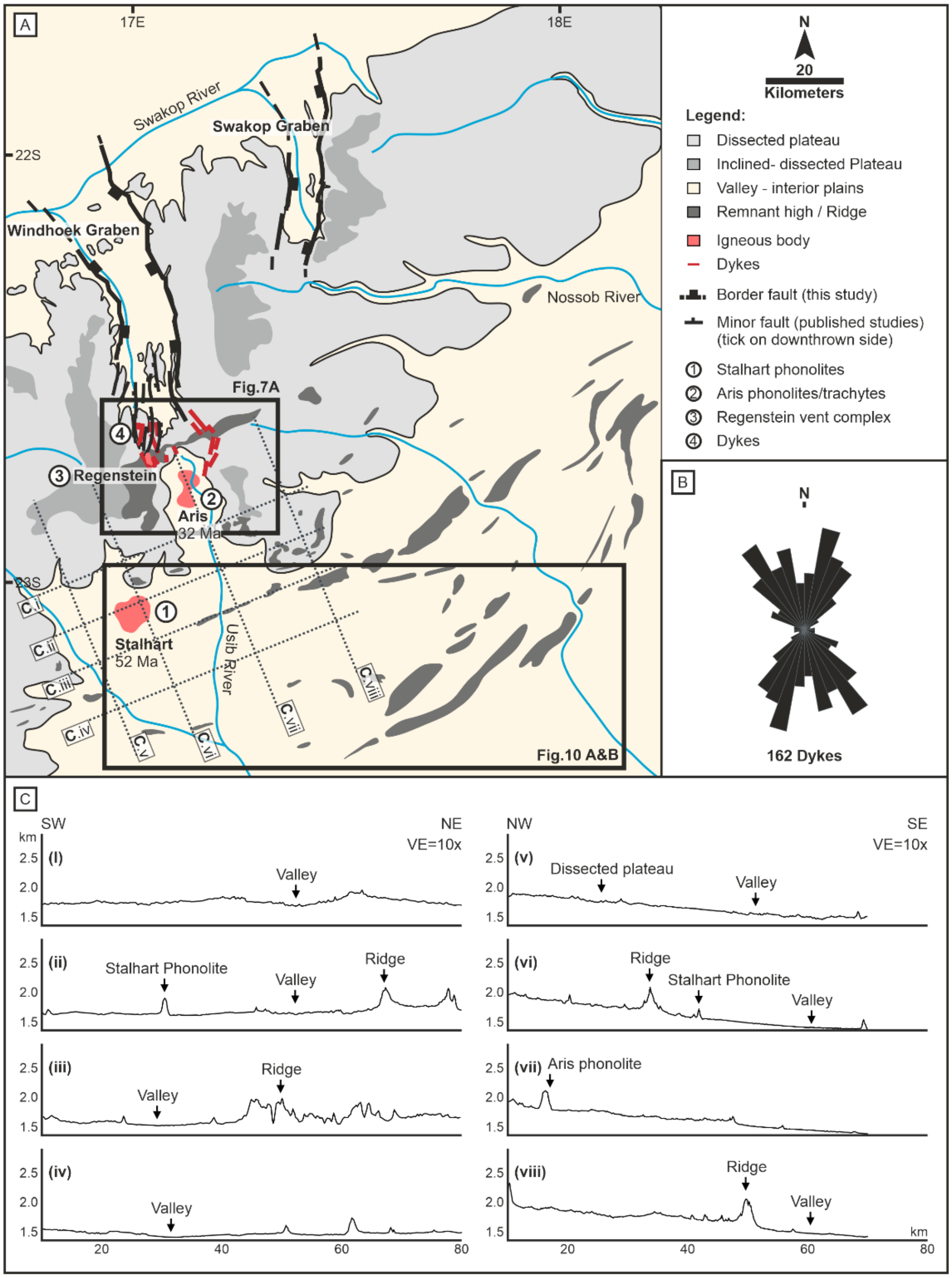

Figure 6. (A) Key faults (simplified from Fig. 3A) and igneous bodies (simplified from Gevers (1932), Ferreira et al. (1979), and Marsh (2010)) are shown superimposed on the morphological map. (B) Orientation of 162 dykes in southern Windhoek Graben (measured from maps of Miller (2008), and the references therein). (C) Transverse and longitudinal topographic profiles of the present-day valley of the Stalhart area.

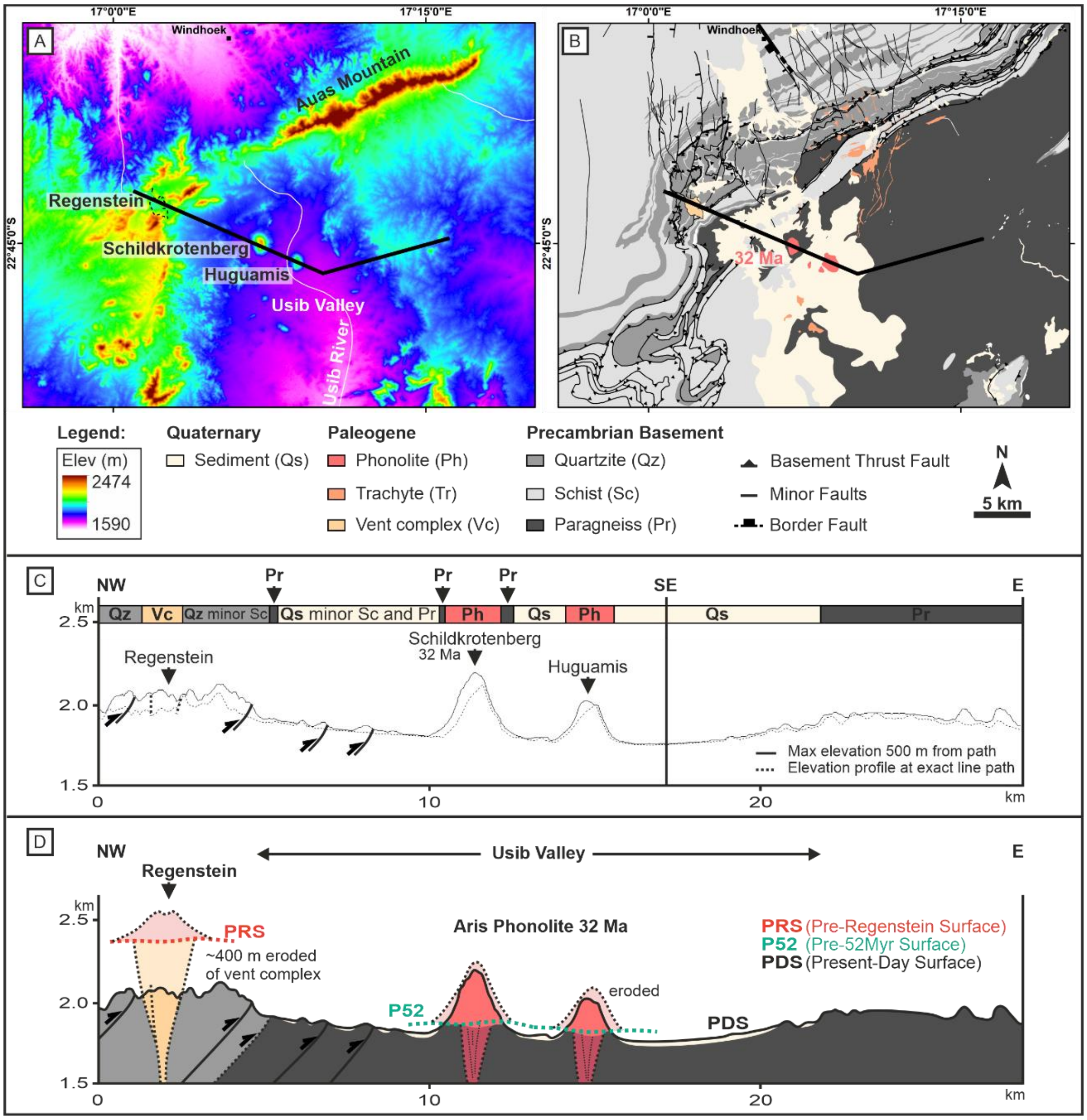

Figure 7. (A) Topographic map of vent complex (Regenstein) and Phonolites (Schildkrotenberg and Huguamis) in the southern end of the Windhoek Graben. See Fig.6A for location. **(B)** 1:250,000 geological map of the area (sheet 2216, Geological Survey of Namibia). Dips direction of the minor faults in southern Windhoek is adapted from Miller et al. (2018) **(C)** Uninterpreted NW-E topographic profile of the vent complex and phonolites. The lateral geological distribution of the 1:250,000 geological map is plotted on top of the profile. **(D)** Interpreted NW-E topographic profile.

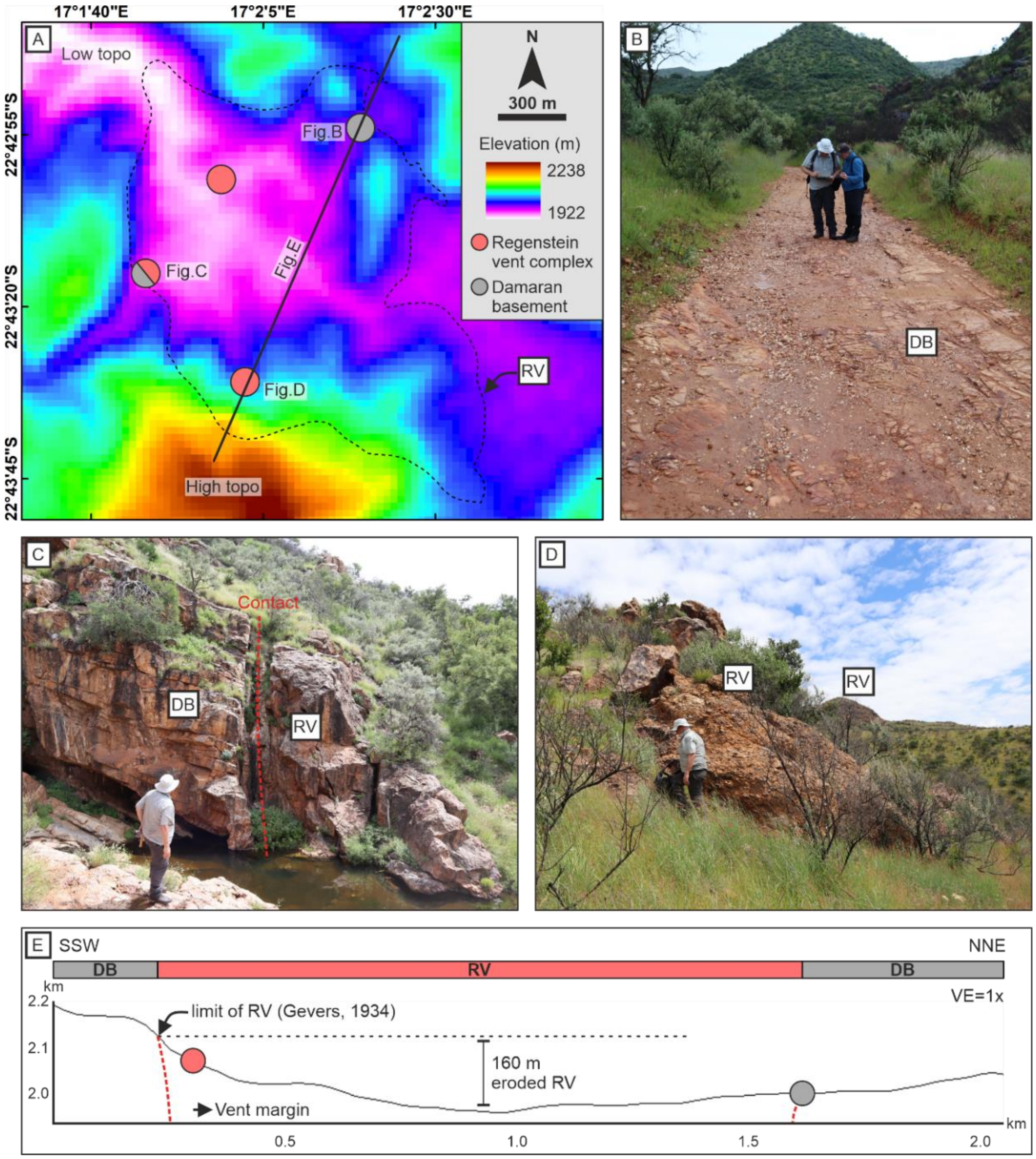

Figure 8. (A) Topographic map of the Regenstein vent complex (RV) with locations of field photographs (B-D). **(B)** Field photograph taken just outside the main vent contact at the NW margin of the Regenstein vent complex where weathered Damaran basement (DB) outcrops along stream courses. **(C)** Field photograph showing a vertical contact between Damaran Basement and Regenstein vent complex at the western margin of the vent. **(D)** Field photograph showing outcrop of vent complex found at 2060 m in the southern area. **(E)** NNW-SSE topographic profile shows ~160 m eroded of the Regenstein vent complex at the present-day surface.

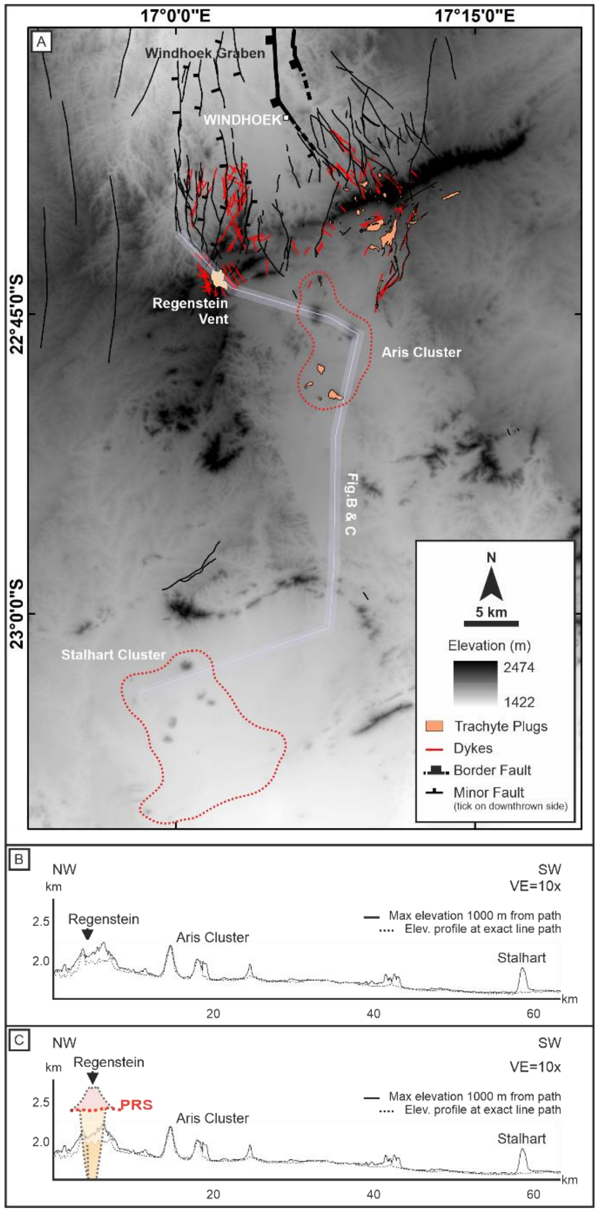

Figure 9. (A) Topographic map overlain by the southern faults of the Windhoek Graben (compiled from Miller and Schalk (1980) and Hoffmann and Schreiber (2011)), dykes, trachytes, and phonolites (compiled from Miller (2008) and the references therein), and Regenstein vent complex (digitized from Hoffmann and Schreiber (2011)). Dip direction of the minor faults in southern Windhoek is adapted from Miller et al. (2018) **(B)** Uninterpreted topographic profile from Regenstein vent complex, Aris cluster, to Stalhart cluster. **(C)** Interpreted topographic profile of (B). PRS: Pre-Regenstein Surface. See (A) for location.

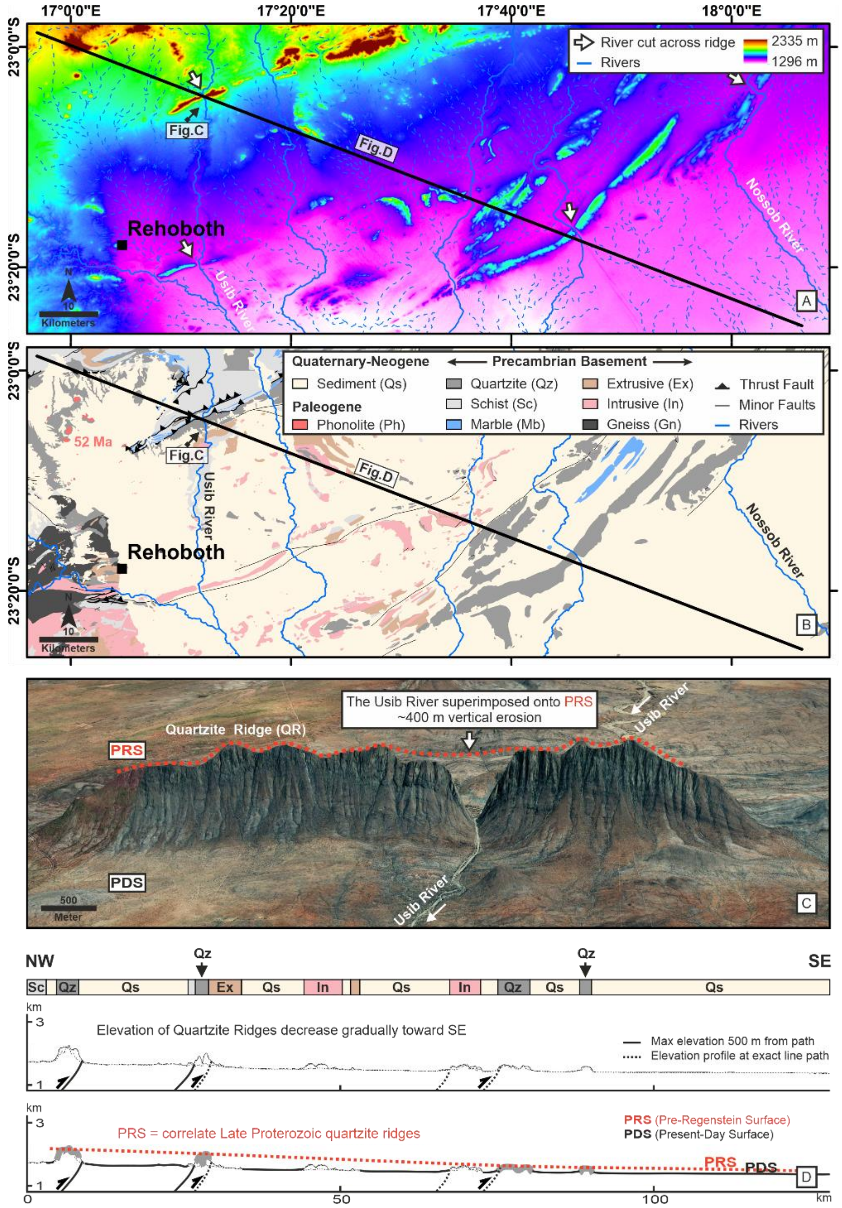

Figure 10. (A) Topographic map of Rehoboth area showing rivers that cut across WSW ridges (white arrows). See Fig.6A for location. **(B)** 1:250,000 geological map of the Rehoboth area (Sheet: 2316, Geological Survey of Namibia). Note: Stalhart phonolites are located in the NW area. Age is taken from Marsh et al. (2018). **(C)** Oblique satellite image showing the Usib River cutting across the WSW quartzite ridge. See panel B for location. **(D)** Upper: uninterpreted NW-SE topographic profile shows decreasing elevation of quartzite ridges towards the SE. Lower: interpreted section of the PRS.

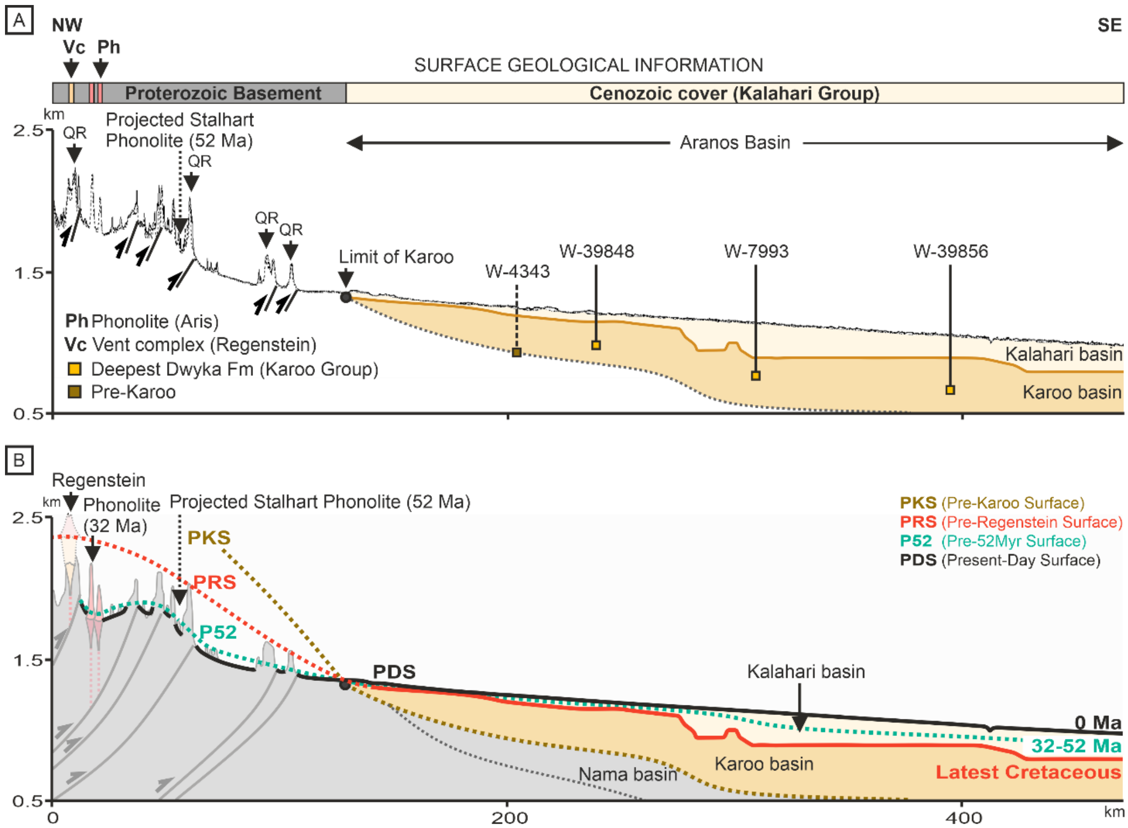

Figure 11. (A) Uninterpreted NW-SE topographic profile from the southernmost Windhoek Graben (Regenstein area) to the Aranos Basin. See Fig.2A for location. Surface geological information is added on top of the profile (Adapted from Miller and Schalk (1980) and Hoffmann and Schreiber (2011)). The bases of Kalahari and Karoo basins are drawn based on borehole information in the Aranos Basin presented by Miller (2008). **(B)** The interpreted section shows the relationship between key surfaces in the highlands and the Aranos Basin.

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