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**The mid-crustal architecture of a continental arc - a transect through the  
South Central Zone of the Pan-African Damara Belt, Namibia**



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**DECLARATION**

I, the undersigned, hereby declare that the work presented in this thesis is my own original work and that I have not previously in its entirety or in part submitted it at any other university for a degree.

Signature:.....

Date: 5 December 2009

## Abstract

The NE-trending South Central Zone of the Pan-African (ca. 550-500) Damara orogen in central Namibia exposes deeply eroded mid-crustal rocks, thought to represent the magmatic arc of the Damara orogen. Above average exposure of outcrop left unmodified by subsequent post-orogenic processes made it possible to study the internal architecture of a ca. 50km traverse, stretching from the continental suture-zone (between the Congo craton in the NW and the underplating Kalahari craton in the SE) at the Okahandja Lineament Zone, well into the leading edge of the Congo craton and into the magmatic-arc, the South Central Zone. This study considers and characterises the change in structural styles and strain intensities in rocks of the Damara Supergroup and intrusions, across the traverse between the towns of Otjimbingwe in the SE and Karibib in the NW.

In the SE of the traverse in the Okahandja Lineament Zone, steep, upright, tightly folded D2 fabrics in meta-turbidites of the Tinkas and Kuiseb Formations record bulk NW-SE shortening and steep SW extrusion of rocks. Penetrative non-coaxial fabrics imply a high-angle collision between the Congo and Kalahari cratons. This is in contrast to oblique collision described by a number of previous authors (e.g. Blaine (1977), Stanistreet et al. (1991), Tack & Bowden, 1999). A marked decrease in D2 strain, and the presence of the siliclastic basal Nosib group suggests the presence of the underlying basement rocks and thus the leading edge of the Congo craton only a few km NW of the Okahandja lineament. 8km NW of the Okahandja lineament is a km-scale NW verging F1 nappe, cored by basement gneisses and refolded into a series of bi-vergent, doubly-plunging F2 folds, the Audawib fold complex. The nappe is interpreted to have formed along a retroshear during early continental collision (syn-D1, early-D2). Tectonically overprinted basement rocks are indicative of thermal weakening, that resulted in the development of thick-skinned tectonics. Intruded mainly to the NW of and around the aforementioned nappe are the areally extensive syn-D2 Salem-type granites. Salem-type granites are shallowly intruded below the nappe and have likely detached the F1 nappe from its root. NW of the Salem-type granites lies a basement window of ca. 15km<sup>2</sup> surrounded by the lower formations of the Damara Supergroup. Sheared marbles and D1 (early D2) diorites along the basement contact indicate a shallow sheared detachment occurring just above the basement. Basement rocks (1) unaffected by Damaran (D1-D2) tectonism and (2) unconformably overlain by the Damara Supergroup are indicative of thin skin tectonics in this part of the South Central Zone, some 30km NW of the Okahandja Lineament zone. Intrusive rocks across the South Central Zone suggest that deformation in the NW ceased by 540 Ma, while deformation along the Okahandja Lineament continued until at least 520 Ma.

Along the Okahandja lineament, high angle continental collision resulted in tight, co-axial folding and lateral extrusion of rocks along the continental backstop. The introduction of numerous late-D2 granites around the Okahandja Lineament Zone (such as the massive Donkerhuk granite) resulted in thermal weakening of the crust, helping to accommodate lateral extrusion. Thermal weakening of the basement allowed the development of thick-skinned tectonics and the formation of the Audwib nappe. In the NW, cooler, more rigid crust deformed very differently to those in the SE, through shallow shearing, thin skinned tectonics. Diachronous timing of the deformation in rocks in the NW and SE of the traverse is due in part to the rheologic difference between cooler rocks in the NW that had locked up to deformation, much earlier than thermally weakened ones in the SE at the plate collision margin, where tectonic stresses were greater.

## **Uittreksel**

### Uittreksel

Die NE-strekkende Suid Sentral Sone van die Pan-Afrikaanse (ca. 550-500) Damara gordel in sentraal Namibië stel diep gëerodeerde gesteentes van die middelkors bloot wat die magmatiese boog van die Damara orogeen verteenwoordig. Goeie dagesame, ongemodifiseer deur subsekwente na-orogeenes prosesse het dit moontlik gemaak om 'n studie aan te pak van die interne argitektuur van 'n omtrent 50km opname wat strek van die kontinentale skeidings sone (tussen die Congo kraton in die NW en die onderplatende Kalahari kraton in die SE) by die Okahandja Lineament Sone, tot veroor die die voorste punt van die Congo kraton in die magmatiese-boog, die Suid Sentral Sone. Hierdie studie neem in ag en karakteriseer die verandering in struktuur styl en drukvervormings (strain) intensiteit in klippe van die Damara Supergroep, tussen die dorpie Otjimbingwe in die SE en Karibib in die NW.

In die SE van die traverse in die Okahandja lineament zone vind 'n mens styl, regop, styf gevoude D2 maaksels in die Tinkas en Kuiseb Formasies, wat bulk NW-SE verkortende en styl SW ekstrusie van rotse aandui. Deurdringende nie-coaksiale maaksels impliseer 'n hoë-hoek botsing tussen die Congo en Kalahari kratons. Dit is in teenstryding met skeefhoekige botsing wat voorgestel is deur verskeie vorige outeurs (e.g. Blaine (1977), Stanistreet et al. (1991), Tack & Bowden, 1999). 'n Vermindering in D2 drukvervorming (strain) en die teenwoordigheid van die silisiklastiese basale Nosib groep, stel die verteenwoordigheid van die onderliggende vloergesteentes voor en sodoende, dat die voorste punt van die onderliggende Congo kraton net 'n paar kilometer NW van die Okahandja Lineament ontwikkel is. 8km NW van die Okahandja Lineament is daar 'n km-skaalse NW neigende F1 dekbladvou gekern deur gneis van die vloer gesteentes en hervou is tot 'n klomp, bi-neigende,



dubbel duikende F2 plooi in wat vernoem word die Audawib vou kompleks. Die dekbladvou word geïnterpreteer om te gevorm het vooraan a retro-verkuiwing, gedurende vroeë kontinentale botsing (syn-D1, vroeë-D2). Oorverskuifde en hervervormde vloer gesteentes is 'n aanduiding van termale verswakking in die aarkors, wat gelei het tot "thick-skinned" tektoniek. Intrusiewe gesteente om, en na die NW van die Audawib dekblad vou is die' weidverspreide Salem-tipe graniet. Syn-D2 Salem-tipe graniete is vlak intrusief to onder die dekblad vou en het waarskynlik die F1 vou van sy wortel sone ontkoppel. NW van die Salem tipe graniete lê 'n 15km<sup>2</sup> grootte venster in die vloer gesteentes in omring deur die onderste formasies van die Damara Supergroep. Verskuifde marmer sowel en D1 (vroeë-D2) dioriet lae op die vloer gesteente se boonste kontak dui daarop dat hierdie boonste kontak verskeurings losmaakpunt is, 'n tektoniese kontak. Vloergesteentes (1) ongeaffekteerde deur Damara (D1-D2) tektoniek en (2) onkonformeerbare kontak met ooriggende klippe van die Damara Supergroep is 'n aanduiding van "thin-skinned" tektoniek in die gedeelte van die Suid Sentral Sone 30km NW van die Okahandja Lineament Sone. Intrusiewe gesteentes gee aanduidings dat deformatsie in die NW reeds ge-eëindig teen 540 Ma, terwyl vervorming in die Okahandja Lineament Sone nog aktief was tot minstens 520 Ma.

Oor die Okahandja Lineament het die hoe-hoekige kontinentale botsing gelei tot stywe ko-aksiale plooiing en laterale ekstrusie van gesteentes langs die kontinentale "backstop". Die intrusie van verskeie laat-D2 grantiete (soos die masiewe Donkerhuk graniet) in die Okahandja lineament sone het gelei tot termale verswakking van die kors wat gehelp het om laterale ekstrusie van klippe te akkomodeer. Termale verswakking van vloer gesteentes het gelei tot die ontwikkeling van "thick-skinned" tektoniek en die vorming van die Audawib dekblad vou. In die NW, het koeler, stewiger gesteentes anders vervorm as daardies verder suid, deur vlak skeur-verkuiwings, "thin-skinned" tectonics. Nie-samelopende vervorming in gesteentes in die NW en SE van die opname is die gevolg van die rheologiese verskil tuseen die koeler gesteentes in die NW wat vroeër bestand geraak het teen vervorm as warmer gesteentes in die SE teenaan die botsings sone, waar tektoniese druk boonop groter was.

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## **List of common acronyms:**

AMC – Abbabis Metamorphic Complex

DSG – Damara Supergroup

OLZ – Okahandja Lineament Zone

CZ – Central Zone

SCZ – South Central Zone

# Chapter 1 Introduction

## 1.1 Background and rationale

The Damara belt in central Namibia represents a deeply eroded suture recording the collision between the Congo and Kalahari Cratons during the Pan-African (ca. 550-500Ma) amalgamation of continents to form the supercontinent Gondwana (Prave, 1996; Gray et al., 2008). Within the extensive system of Pan-African belts in southern Africa, it is the only orogen that has not been affected by significant later modifications during the subsequent break-up of Gondwana. As such, the Damara belt preserves its original collisional geometry, showing well-preserved marginal foreland basins, bi-vergent fold-and-thrust belts centred around a central magmatic axis, situated on the overriding plate of the Congo Craton, and an accretionary prism, located on the underthrust plate of the Kalahari Craton (Miller, 1983; Kasch, 1983a; Coward, 1983). This renders the belt a unique area to investigate collisional processes exposed and preserved at mid- to lower crustal levels.

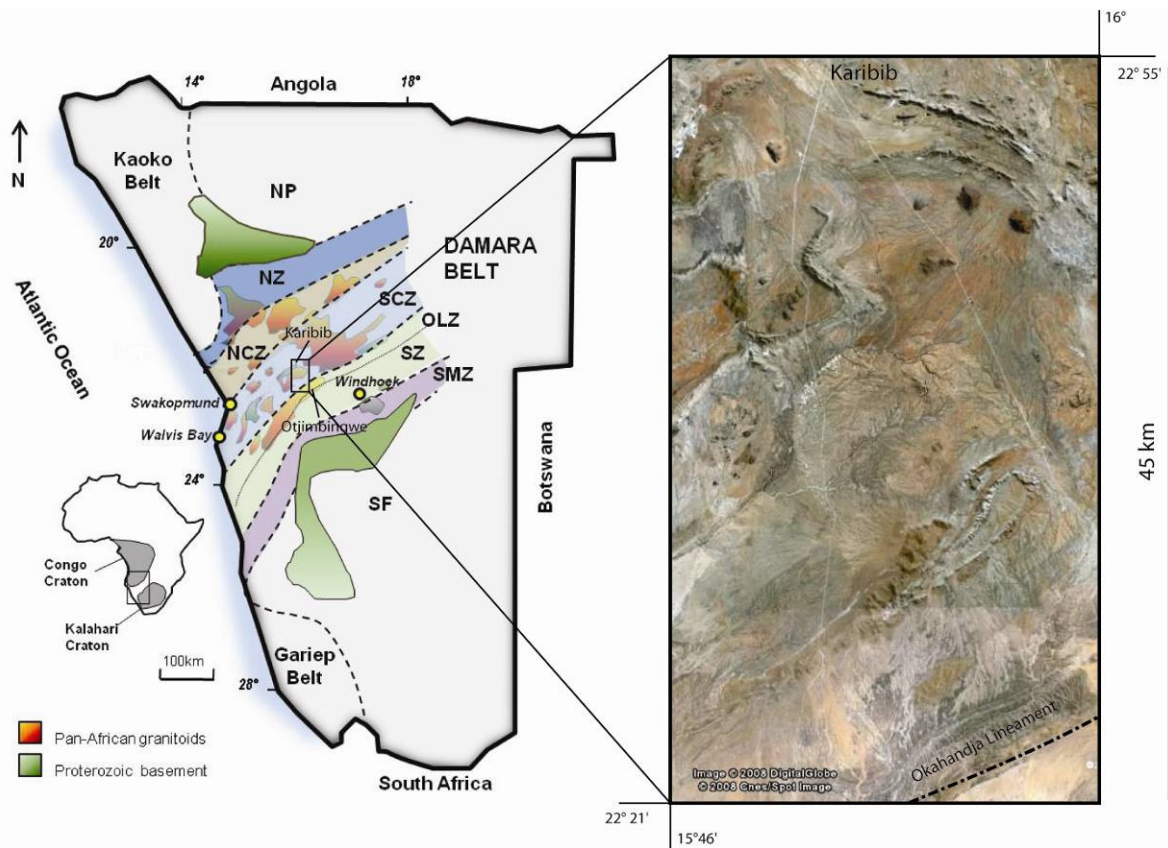
The large volume of granite intrusions in the South Central Zone (SCZ) of the Damara belt has led many workers to believe that this area represents the magmatic arc of the orogen (Miller, 1983; Kasch, 1983a; Jung, 2000; Jung et al., 2002). Basement outcrops and geochronological data indicate that this magmatic arc is developed in Mesoproterozoic basement gneisses of the Congo Craton, commonly interpreted to form the overriding plate during the convergence and collision of the two cratons. The thick sequence of high-P metaturbidites of the adjoining Southern Zone to the SE of the SCZ suggests that the SCZ represents the leading edge of the Congo Craton, situated along and directly adjacent to the main suture zone of the orogen. Metamorphic conditions in the central parts of the SCZ are not well constrained, but high-T metamorphism and mid-crustal levels are clearly suggested by regionally developed cordierite-sillimanite bearing mineral assemblages and the pervasive ductile fabrics of the rocks and thus, rocks of the SCZ of the Damara belt provide a very well-exposed and preserved window into the mid-crustal structure and architecture of the internides of a collisional orogen (Miller, 1983, 2008).

The relationship between granitic melts and regional deformation in orogenic settings tend to create a great number of feedback loops. The presence of large volumes of melt will result in strain localization and strain softening and/or dramatic strain-rate variations within the crustal column. Conversely, the crystallization of these melts to form equigranular quartzo-feldspathic plutons and batholiths typically have a strain hardening effect on the crustal column, such that

granite emplacement is commonly cited to contribute to late-orogenic stiffening of the crust (Davidson et al., 1994). Complex feedback relationships between crustal deformation and melt migration and magma intrusion often obscure the mode of melt ascent and final emplacement of granite plutons. The degree to which granite intrusion controls deformation and to what extent intrusion is controlled by deformation is often controversial (Brown & Solar 1999; Brown 2007).

A traverse across both syn- and post-tectonic granites intruded into deformed metasedimentary rocks of the Damara Supergroup provides an ideal scenario to study the interplay of regional deformation and granite plutonism. For this study a NW-trending transect was chosen that runs roughly between the towns of Karibib and Otjimbingwe, this transect cuts across strike of the SCZ down to the Okahandja Lineament Zone (OLZ), a area commonly regarded as the actual suture zone between the Kalahari and Congo Cratons (Sawyer, 1978; Barnes & Sawyer, 1980; Downing & Coward, 1981).

Much of the previous work in the SCZ has focussed on granite petrography and geochemistry (Hoffmann, 1976; Haack et al. 1982; Hawkesworth & Marlow 1983; Marlow, 1983; Mcdermott et al., 1996; Jung & Mezger, 2003). A further aspect that has been highlighted is the stratigraphic correlation of units (Smith, 1965; Miller, 1983; DeKock, 1989; Hoffmann, 1990; Badenhorst, 1992). Structural work was mainly done in the 1980's following the pioneering work of Gevers (1963), Smith (1965), Jacob, (1974) Coward (1983), Sawyer (1981), Downing & Coward (1981), Kasch, (1983a). This study attempts to document and characterize the architecture and structural evolution of the SCZ between the OLZ and the northern parts of the SCZ, a ca. 40km long transect across the SCZ down to the OLZ. Similar regional works have been undertaken by e.g. Smith (1965), Sawyer (1981) Brand (1985) and De Kock (1989) around this area, as well as the study a number of structures in the surrounding area (Smith, 1965; Downing & Coward, 1981; Sawyer, 1983; De Kock 1989; Kisters et al., 2004; Johnson 2005).



**Fig 1.1:** Simplified geological map of Namibia (Kitt, 2008) with an aerial photograph of the study area in the SCZ of the Damara belt, Namibia. The traverse/study area covers a length of about 45k across strike of the Damara belt. The traverse stretches from just south of Karibib down to the Okahandja Lineament Zone (OLZ).

## 1.2 Aims of the study:

- Documentation of the regional distribution of main formations and lithologies, thickness, facies and strain variations on a 1:25.000 scale. The detailed lithological inventory of this zone forms the basis for structural mapping aimed at documenting the crustal architecture of this zone.
- Integration of structural and lithological data into comprehensive 2D sections, aimed at developing a better understanding of the structural architecture and geological evolution of the region.
- Detailed mapping of granites that are abundant along this traverse, their relative timing and intrusive relationships in order to understand the role of granites during deformation of the middle crust.



### 1.3 Research Method

Field data was collected during fieldwork in three main field seasons including initial reconnaissance mapping in April 2008, a main three-month field season in July, August and September 2008 and a final concluding field season in March 2009.

Mapping between Karibib and Otjimbingwe was done on foot as well as using a 4X4 vehicle. *Google Earth* images as well as topographic maps of the region were used to facilitate 1:25 000 scale mapping. Sampling of rocks for petrologic studies was also done during this time.

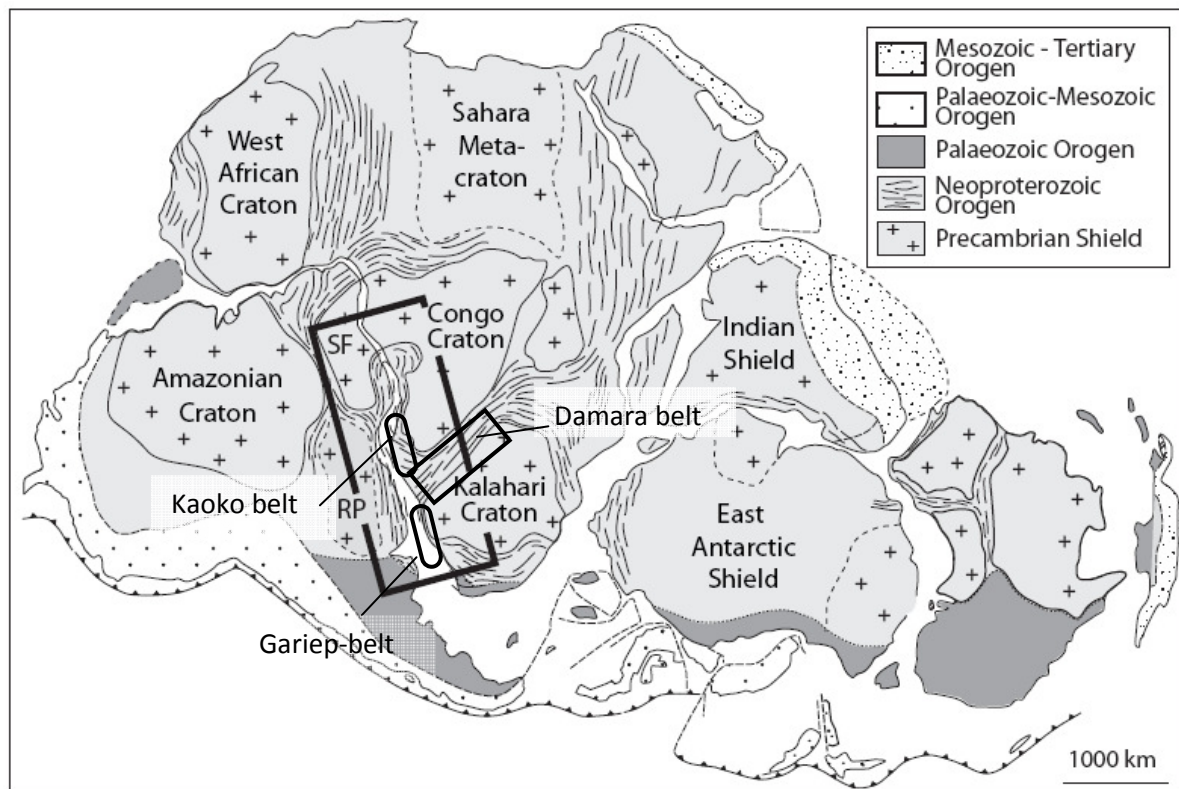
Structural trends were recorded with a *Krantz* geologic compass and georeferenced. A *Garmin Gecko* GPS was used to mark locations where measurements or observations were made. Structural measurements were taken as dip-direction and dip (for planar features) and plunge-direction and plunge (for linear features). Readings are given as planar elements and all orientation diagrams are given as equal area stereographic projections into the lower hemisphere. Maps were drawn up using ESRI® ArcMap™ 9.2 software.

The scale of mapping detail was often dependant on outcrop conditions. Traverses across and along strike were often limited to areas where outcrop provided significant exposure. Especially towards Otjimbingwe, riverbeds often provided some of the best exposure and these were consequently often studied to a much larger degree of detail.

## Chapter 2 Regional Geology

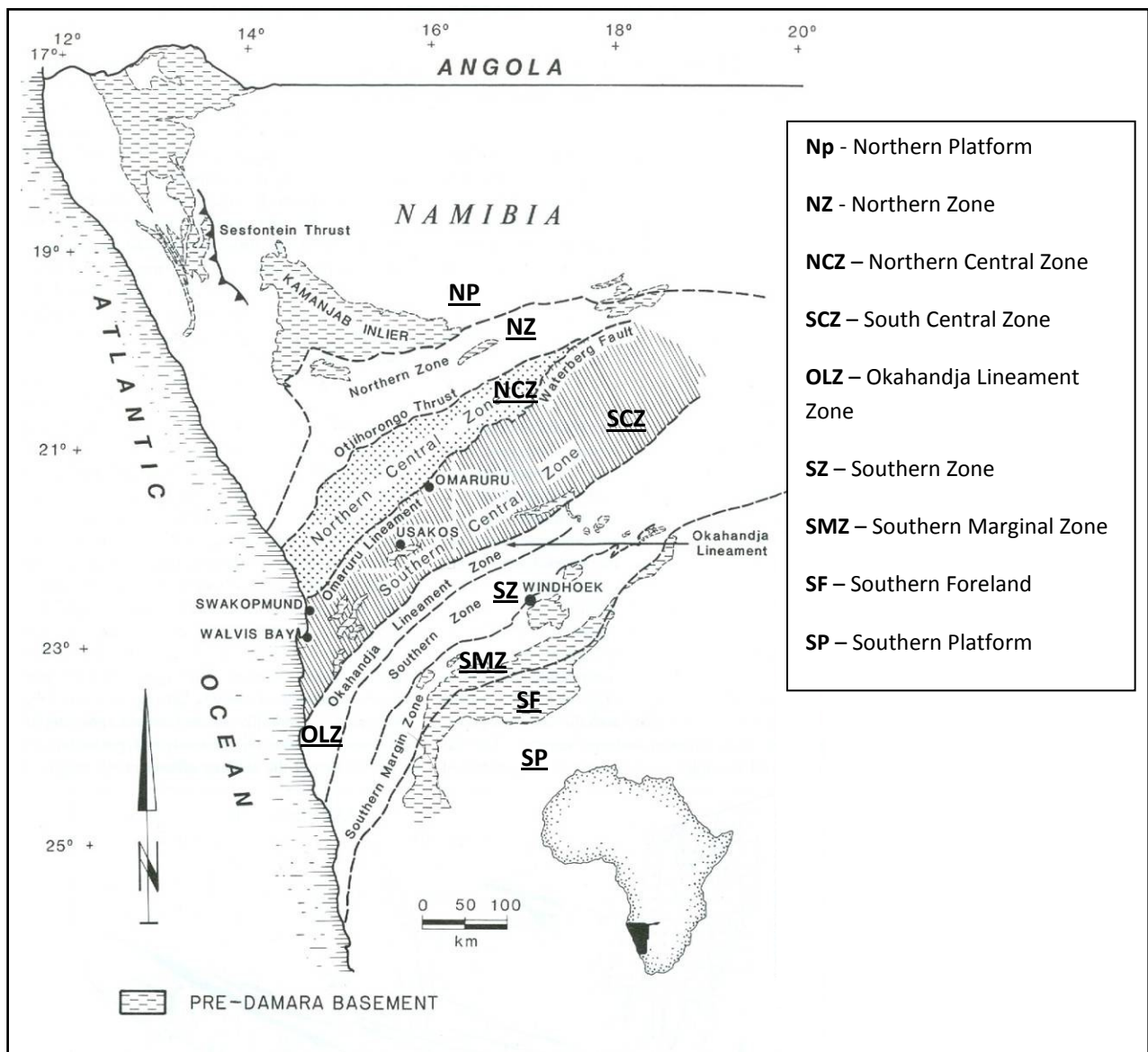
### 2.1 Introduction

The Damara orogen in central Namibia forms part of a number of Pan-African collisional belts, situated in southern Africa that formed during the amalgamation of the Gondwana supercontinent (Miller, 1983; Prave, 1996; Trompette, R., 2000; Gray et al., 2008). The Damara belt is known as the “intracratonic” or “inland” branch of the Damara orogen, having formed during the collision of the Congo and Kalahari Cratons, whereas the N-S trending “coastal-branches” of the Kaoko and Gariep belts record the collision between the African cratons and the Sao Francisco and Rio de la Plata cratons of South America (Grey et al., 2008). Recent work by e.g. Goscombe et al. (2003) and Gray et al. (2006, 2008) suggests that collisional tectonics in the Damara orogen was slightly diachronous, occurring first with the formation of the Kaoko belt that records the closure of the Adamastor ocean between southern Africa (Congo and Kalahari cratons) and the South American cratons. This was followed by the closure of the southern Adamastor ocean for the Gariep belt and concluded by the closing of the Khomas ocean, the oceanic basin between the Kalahari and Congo cratons and as a result the formation of the Damara belt (Fig 2.1).



**Fig 2.1:** Map of Gondwana showing the positions of the cratonic nuclei and the orogenic belts that weld the supercontinent together. SF = Sao Francisco craton, RP = Rio de la Plata craton. In the centre are the Kaoko, Gariep (coastal branch) and Damara (inland branch) belts (from Gray et al., 2008).

## 2.2 Architecture of the Damara belt



**Fig.2.2:** Simplified map of the Damara belt of Namibia divided into its discrete tectonostratigraphic zones (Miller, 1983).

The ENE-trending Damara belt is approximately 1000km along strike and has a width of about 400km (Miller, 1983). It is made up of pre-collisional, mainly Paleoproterozoic (ca.2.0-1.7Ga) basement inliers overlain by the Neoproterozoic (ca. 800-550 Ma) rocks of the Damara Supergroup, as well as a large number of intrusive granites related to convergent and collisional tectonics and the post-collisional equilibration of the belt (Miller, 1983, Porada, 1989; Jung et al., 1999; Gray et al. 2008).

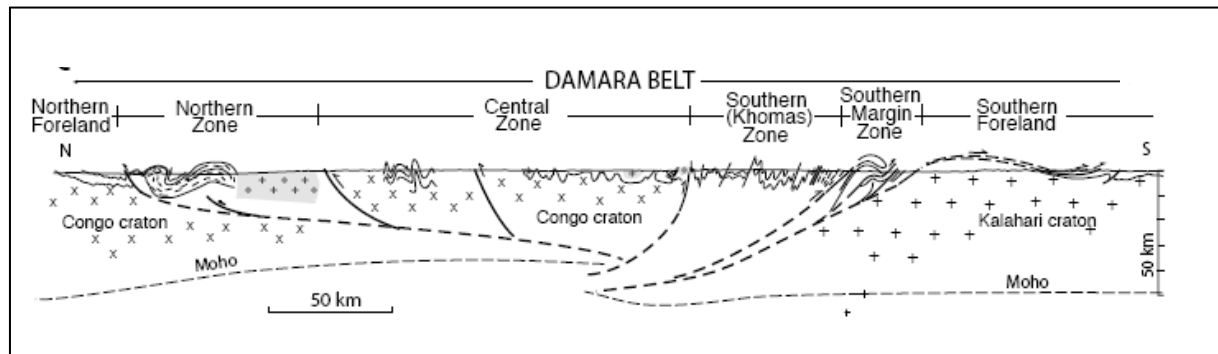
The belt has been subdivided into a number of discrete tectonostratigraphic zones (Fig.2.2 & Fig.2.3), based on the structure, stratigraphy, metamorphism, intrusions, geochronology and aeromagnetic

expression of rocks in the Damara belt (Martin, 1965; Martin & Porada, 1977; Barnes & Sawyer, 1980; Miller & Hoffman, 1981; Miller, 1983; Corner, 1983, 2000; Anderson & Nash, 1997).

These zones are, from N to S:

- Northern Platform (NP)
- Northern Zone (NZ)
- Central Zone (CZ - study area)
- Southern Zone (SZ)
- Southern Marginal Zone (SMZ)
- Southern Foreland (SF)

The Central Zone is further subdivided into the South Central Zone (SCZ), in which the study area is located, and the Northern Central Zone (NCZ) (Miller, 1983).



**Fig 2.3:** A schematic cross section through the different zones of the Damara belt (From Gray et al. 2008).

The Southern Foreland (SF) and Northern Platform (NP) represent the foreland basins of the Damara belt in the S and N, respectively. The Southern Marginal Zone (SMZ) and the Northern Zone (NZ) represent the fold-and-thrust belts of the Damara belt, verging SE and NW, respectively, resulting in the overall bivergent symmetry of the belt (Fig 2.3). The Southern Zone (SZ), made up of massive, several thousand meters thick meta-turbidite sequences, is thought to represent the accretionary prism of the Damara belt resting on the underthrust slab of the Kalahari Craton (Kukla & Stanistreet, 1991; Stanistreet et al. 1991; Kasch, 1983a). This notion is supported by the presence of the Matchless Amphibolite Belt, a narrow, but laterally extensive unit of mafic crust interpreted to represent an ophiolite sliver (Kukla & Stanistreet, 1991). The transition between the Central and Southern Zone is marked by the Okahandja Lineament Zone (OLZ), a zone interpreted to represent the suture between the Kalahari and Congo Cratons, representing the leading edge of the overriding Congo plate (Blaine, 1977; Miller, 1979; Sawyer, 1981; Downing & Coward, 1981). The OLZ is intruded by the supposedly post-tectonic Donkerhuk granite (Barnes & Sawyer, 1980; Sawyer, 1983;

Miller, 1983, De Kock & Walraven, 1995), for which available Rb-Sr and U-Pb age data suggest an emplacement at  $523 \pm 7$  Ma (De Kock & Walraven, 1995). The Central Zone (CZ) also referred to as the magmatic axis of the Damara Belt, intruded by a variety of syn-, late-, and post-tectonic granites between ca. 565 and 465 Ma (Hoffman, 1976; Blaxland et al. 1979; Miller, 1983, 2008; Hawkesworth & Marlow, 1983; Tack & Bowden, 1999; Jung & Mezger, 2003).

## **2.3 Evolution of the Damara belt**

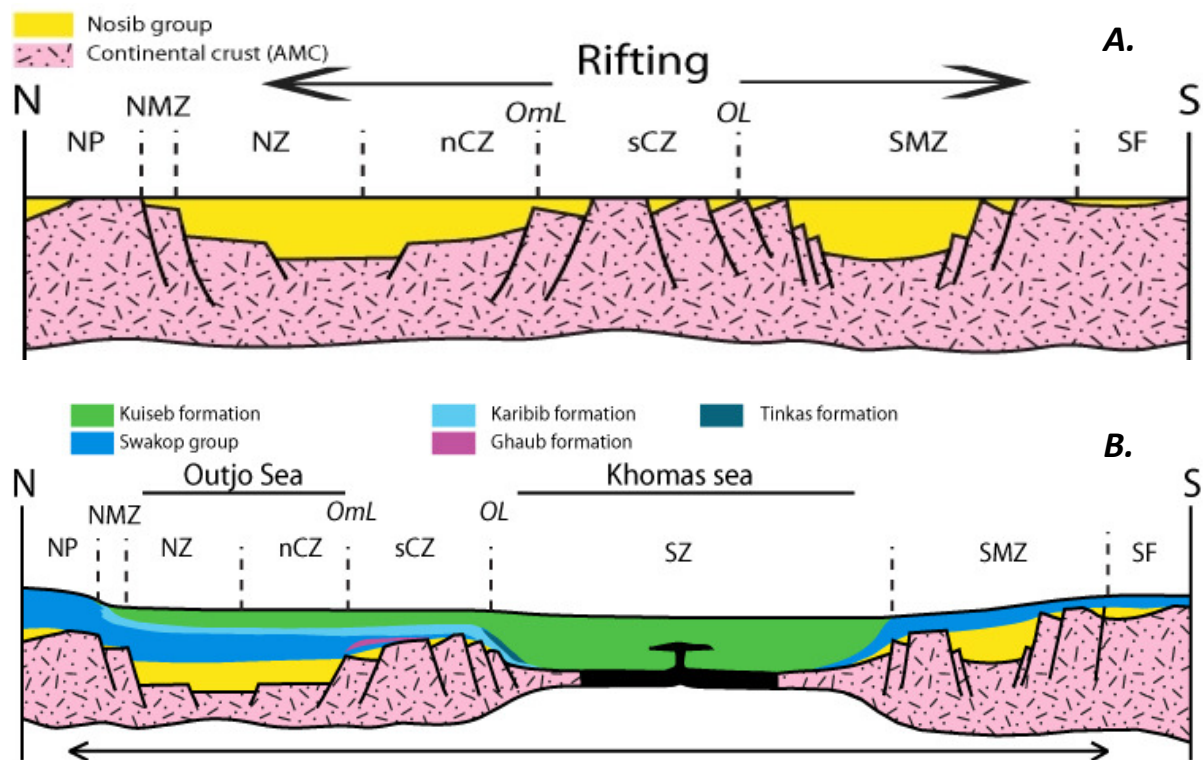
### **2.3.1 Rifting and basin evolution**

The onset of rifting between the Congo and Kalahari Cratons (Fig 2.4) and early sedimentation of the Damara Supergroup started with the deposition of continental sediments and volcanic rocks of the basal Nosib Group (Gevers, 1963; Smith, 1965; Martin & Porada, 1977; Porada, 1989; Kukla & Stanistreet, 1991; Stanistreet et al., 1991; Hoffman et al., 1996; Miller, 2008). The timing of initial rifting is indicated by the peralkaline volcanics (several thousand meter thick) of the ca. 760Ma Naauwpoort Formation in the NZ (Hoffman et al., 1996). Thickness and facies variations of the basal, coarse-clastic and volcanic Etusis Formation that unconformably overlies Mesoproterozoic basement rocks, indicates the segmentation of the continental rift into two subparallel, NE trending graben or half-graben structures (Henry et al., 1990; Stanistreet et al., 1991; Miller, 2008). A northern rift develops later into the so-called Outjo Sea (Miller, 2008). The several thousand meter thick sedimentary record of this basin is preserved in the NZ and NCZ. The southern rift developed on the Kalahari Craton side, in the area of the present-day SMZ. The northern part of this rift, the present-day SZ, progressively developed into the oceanic basin of the Khomas Sea.

Progressive rifting and basin subsidence led to first marine incursions indicated by the deposition of a mixed siliciclastic-carbonate sequence now represented by marble, schist and calc-silicate felsites of the Khan and Rössing Formations in the northern basin (NZ) and parts of the SCZ (Smith, 1965, Henry, 1992; Gray et al., 2006). This sedimentation was interrupted by a major Sturtian-age (ca. 730 Ma) glaciation event and the deposition of the diamictites of the Chuos Formation, the lowest unit of the Swakop Group (Badenhorst, 1992; Hoffmann et al., 2004). The diamictites are very widespread in large parts of the CZ and form a good marker horizon. Following the main glaciation event and associated sea-level rise, sedimentation is again characterized by a mixed siliciclastic-carbonate sequence indicating deposition along the transition from a shelf to a deep-sea environment (Miller, 2008). The shallow-water equivalent of the Swakop Group, the Otavi Group, occurs northwards on the Northern Platform of the Damara belt. Sedimentation occurred from ca. 730 Ma, the age of the Chuos Formation, to ca. 600 Ma, the youngest reliable age of  $635 \pm 1$  Ma being derived from thin volcanic horizons in the upper parts of the Swakop Group in the Ghaub Formation (Hoffmann et al.,

2004), which represents a second (Marinoan-age) glaciation event in the Damara belt. The Ghaub Formation is succeeded by a thick marble succession (up to 1000m thick) the Karibib Formation (Hoffmann, 1983 and Miller 1983a). The Karibib Formation grades into a mixed carbonate-clastic-sequence the Tinkas Formation along the OLZ thought to represent a deeper basin-slope facies variation of typically shallow marine carbonates of the Karibib Formation (Jacob, 1974; Downing & Coward, 1981).

The Kuiseb Formation forms the top of the Damara Supergroup. The schist-dominated turbidite sequence indicates the broadening of the oceanic basin and the connection of the northern Outjo Sea across the central high of the SCZ into the Khomas Sea (Porada & Wittig, 1983; Miller, 2008). There are no published ages for the deposition of the Kuiseb Formation, but sedimentation is thought to have occurred between ca. 600-580Ma (Kisters et al., 2007).



**Fig 2.4:** Continental rifting and the opening of the Khomas ocean, provides accommodation space for the deposition of clastic-sediments of the Damara sequence. Initiation of rifting leads to the depositon of the Nosib group (A) followed by further spreading and deposition of the Swakop group (B) (Oml = Omaruru lineament, OL = Okahandja lineament; Modified from Miller, 2008.).

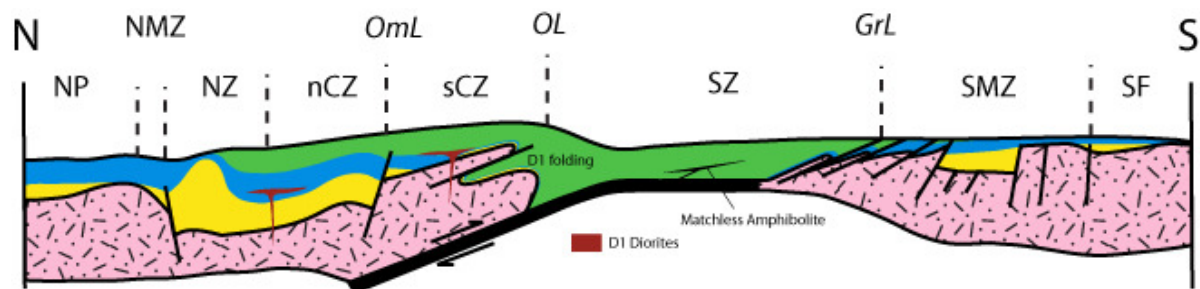
### 2.3.2 Crustal Convergence

Subsequent crustal convergence (Fig 2.5) occurred as a result of N- or NW-ward subduction of the Kalahari plate below the Congo Craton between 580-560Ma (Grey et al. 2006; 2008). Closure of the Khomas Sea resulted in the formation of SZ, (south of the Okahandja Lineament Zone)



thought to represent an accretionary prism formed during the time of final convergence (Kasch, 1983a; Kukla & Stanistreet, 1991).

The first, regionally widespread structures are formed during this initial stage of convergence (D1 deformation phase, ca. 580-575Ma, Miller, 2008). D1 is expressed in regional fabrics as a mainly bedding-parallel  $S_1$  foliation with associated folding and bedding transposition (F1). Early and relatively mafic plutons of gabbroic and dioritic composition intrude between 565 and 550 Ma and postdate the D1 deformation (De Kock et al., 2000; Jacob et al., 2000). D1 is followed by the formation of large-scale recumbent nappes and low-angle thrusts (F2) (Miller, 1983).



**Fig 2.5:** Onset of crustal convergence, with subduction towards the N or NW. Convergence leads to the formation of the first shallowly dipping D1 structures across the Damara belt. The intrusion of early granites into the CZ occurs during this time. (Colour coding as in Fig.2.4, GrL = The Gomab river line. Modified after Miller, 2008).

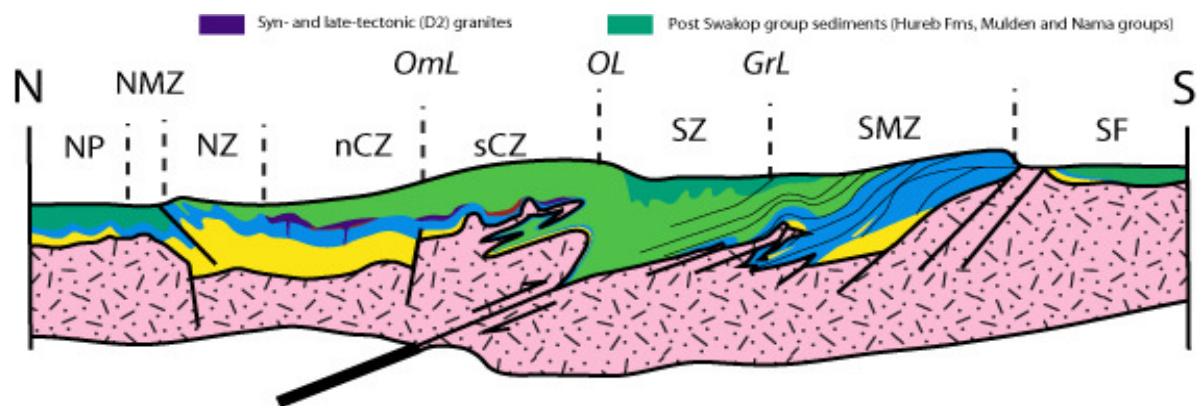
### 2.3.3 Progressive convergence and collision

Complete closure of the Khomas Ocean and, with it, collision of the Kalahari and Congo cratons is thought to have occurred and peaked by ca. 540 Ma in the CZ (Miller, 1983, 2008, Jacob et al., 2000, Kisters et al., 2004; Gray et al., 2006) (Fig 2.6).

Continental collision led to crustal thickening as well as the development of the bivergent symmetry of the orogen (Kröner, 1982; Barnes & Sawyer, 1980). The prominent ENE structural grain in the Damara belt was developed during the D3 event (note: D2 after Jacob, 1974; Poli & Oliver, 2001; Kisters et al., 2004), which led to the refolding of earlier low-angle fabrics and recumbent folds by regional-scale ENE-trending folds and the formation of pervasive ENE trending fabrics. These structures are discussed in further detail later in the chapter.

Crustal thickening related to continental collision is associated with the earliest M1 metamorphic event as well as the intrusion of most of the syn-tectonic granites in the SCZ and NCZ, the then magmatic arc of the orogen (Nex et al., 2001; Gray et al., 2006; Jacob et al., 2000). M1 is thought to have occurred at approximately 550-540 Ma (Nex et al., 2001). Though M1 was subsequently overprinted, it is estimated to have been a medium-P medium-T metamorphic

event, reaching pressures and temperatures of ca.  $4 \pm 1$  kbar and  $550\text{-}600^\circ\text{C}$  in the western parts of the sCZ (Nex et al., 2001). Granites intruding during this time include the regionally widespread Salem-type granites (Gevers, 1963) a term locally used to describe a suite of megacrystic biotite granodiorites and associated leucogranites (Smith, 1965; Jacob, 1974; Allsopp et al., 1983; Marlow, 1983; Miller, 1983; Hawkesworth et al., 1983; Jung & Mezger, 2003). Late- to post D2/3 leucogranites intruded at ca. 540 Ma (e.g. the Rote Kuppe Granite,  $539 \pm 6$  Ma Jacob et al., 2000) are interpreted to indicate the end of the collision in this part of the sCZ.

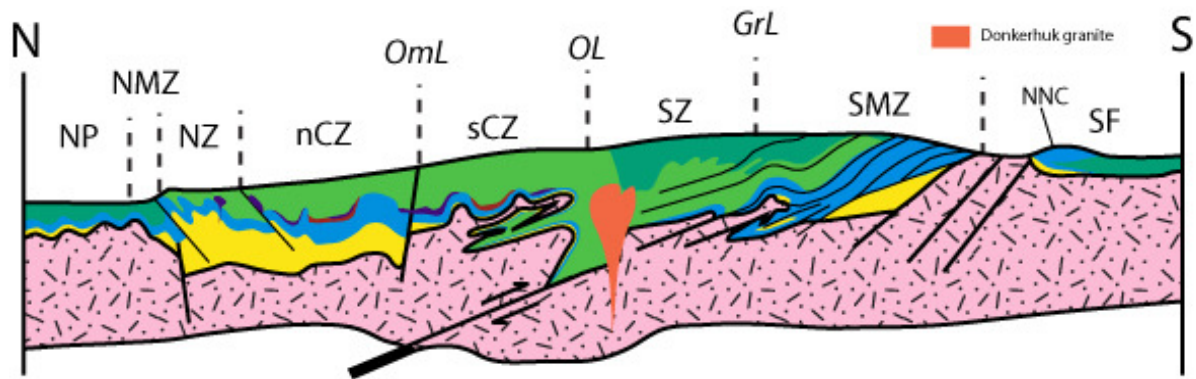


**Fig 2.6:** Collisional tectonics resulting in SE ward imbrication and folding in the SZ and SMZ on the Kalahari Plate and refolding of earlier D1 recumbent folds by more or less upright, ENE trending D2 folds and associated granite plutonism on the overriding plate of the Congo Craton in the N, preserved in the CZ (Modified after Miller, 2008).

#### 2.3.4 Post-Collisional evolution (Fig 2.7)

Collision was followed by the uplift and decompression of rocks, resulting in crustal thinning and extension (Tack and Bowden, 1999; Jung and Mezger, 2003; Johnson, 2006) (Fig 2.7). During this time the Damara belt experienced the intrusion of numerous post- and late-tectonic granites between 540-480 Ma (Hoffman, 1976; Jacob et al., 2000; Jung & Mezger, 2003). This final phase of granite plutonism is synchronous with a complex and seemingly episodic high-temperature, low-pressure M2 metamorphic event responsible for the current metamorphic assemblages found within the SCZ. This event is thought to have reached a number of discrete peaks between 505Ma and 478Ma (Bowden et al., 1999; Jacob et al., 2000; Jung & Mezger, 2003). Ar-Ar cooling ages of ca. 460 Ma indicate final uplift of the central parts of the Damara belt at that time (Gray et al., 2006).





**Fig 2.7:** After final collision a number of post tectonic granites, of which the Donkerhuk granite is the largest, were intruded. By this time a nappe complex, the Naukluft nappe complex (NNC) has been formed in the Southern Foreland. (Modified after Miller, 2008).

## 2.4 Regional Geology of the SCZ in the Karibib region

### 2.4.1 Lithostratigraphy (Table 2.1)

The present study investigates specifically the structural evolution of the SCZ between the Omaruru and Okahandja Lineaments (OmL and OL on Fig.2.7 above). Previous regional mapping of the SCZ in the Karibib region has been done by, inter alia, Gevers (1963), Smith (1965), Jacob (1974), Blaine (1977), Sawyer (1981), Brand (1985; 1987), De Kock (1989) and Badenhorst (1992). Sedimentation of the Damara Supergroup in the SCZ was strongly influenced by its paleogeography and its situation on the basement high between these two regional-scale structures separating the the Outjo Sea in the north from the Khomas Sea in the south (Fig 2.4.B). Consequently, many of the formations developed in the northern basin are not found or are only locally developed and with much smaller thicknesses in the SCZ compared to adjacent areas (Hoffmann, 1990; De Kock, 2001; Miller,2008). Miller (2008) suggested a modified stratigraphic subdivision of particularly the Swakop Group for different parts of the CZ, taking into account (1) facies and thickness variations in the sCZ compared to adjoining areas, and (2) the more recently identified second Marinoan-age glaciation event that is preserved in large parts of the Damara Belt (e.g. Hoffmann et al., 2004). Miller's (2008) subdivision (Table 2.1) combines earlier stratigraphies suggested for the NCZ and N parts of the SCZ (Jacob, 1974; Badenhorst, 1992) and the southern parts of the SCZ (Smith, 1965; De Kock, 2001).

Table 2.1: Lithostratigraphy of the SCZ. Formations and members in grey are not present in the study area. The stratigraphy has been subdivided and rearranged by Miller (2008) and is slightly different from that described by Badenhorst (1992), De Kock (2001) and Hoffman et al. (2004) (Table, adjusted from Miller (2008))

Group	Sub-groups	Formation	Members	Description	Thickness	Age	
Swakop	Navachab	Kuiseb		Quartz-biotite schist and meta-psammities as well as minor marbles and calc-silicate felses	40 to >3300 m (Smith, 1965; Badenhorst, 1987)		
		Tinkas	Quelle & Kuduleck	Banded schists, calc-silicate felses and marbles	Kuduleck member - up to 500m, Quelle member - 900m (De Kock, 2001; Miller, 2008)		
		Karibib		Recrystallised, brecciated and banded marbles	up to 1500m (Johnson, 2005)		
		Ghaub	Lievental, Daheim, Omusema & Kachab	Glacial diamictite and pelites, with basalts (Daheim member)	0-150m (Hoffmann et al., 2004)	635.5±1.2Ma (Hoffmann et al., 2004)	
	Usakos	Arandis	Oberwasser		Schists and minor calc-silicate felses	60-240m (Badenhorst, 1992; Kisters et al., 2004).	
			Okawayo		Marbles and micro calc-silicate felses	70-150m (Kisters et al., 2004)	
			Spes Bona		Meta-pelites, meta-psammities and minor calc-silicates	ca. 100m thick in the Usakos dome (Johnson, 2005)	
			Karub		Marble unit with interbedded calc-silicate felses and schist	<10m	
		Chuoss		Glaciogenic diamictite	up to 180m in the Karibib dome (Kisters et al., 2004)	ca. 750-720 Ma (Kaufmann et al., 1997)	
	Ugab	Rössing		Interbedded marbles, calc-silicate felses and siliciclastic rocks (highly variable).	up to 110m (Henry, 1992)		
	Nosib		Kahn		Calc-silicate felses, biotite- and graphite-schists and marble.	0 to 1500 m (Steven, 1993).	
Etusis				Quartzo-feldspathic arenites and minor grits	0-3000m (Smith, 1965)	756±2 Ma to 746±2 Ma (Hoffman et al., 1996)	

### The Abbabis Metamorphic Complex (AMC)

The AMC, locally also referred to as the Abbabis basement, is thought to represent parts of the leading edge of the Congo Craton onto which the Damara Supergroup was deposited. These rocks are named after the farm Abbabis, about 15 km SE of Karibib, on which one of the largest basement inliers is exposed (Gevers, 1931; Miller, 1983). These pre-Damara inliers are made up of a wide variety of rock types, dominated by red, homogeneous quartzo-feldspathic augen gneisses. Other rock types that occur within the AMC include amphibolites, quartzites, biotite schists and calc-silicate rocks. Around the Karibib/Usakos region, the AMC is further intruded by a number of granites and pegmatites (Brandt, 1987; Steven, 1993; Kisters et al., 2004). U-Pb Zircon dating of basement gneisses on the farm Abbabis has yielded ages of 1925Ma ± 330 Ma (Jacob et al., 1978). Zircon dating of rocks in the Kahn river near the Rössing uranium mine also gave evidence for Paleoproterozoic basement rocks with ages of 2014±39 Ma and 2093±51Ma as well as Namaqua-age overprint at ca. 1100 Ma (Kröner et al., 1991). Furthermore, the majority of Pan-African granites contain older, inherited zircons or zircon cores yielding ages between 1.7-2.0 Ga, indicating the presence of this Mesoproterozoic basement underlying large parts of the SCZ (Jung & Mezger, 2003; Johnson et al., 2006).

For more information on the basement stratigraphy the reader is referred to Brandt (1987).

## **Damara Supergroup (DSG)**

### **The Nosib Group**

On a regional scale, the Nosib Group in the SCZ is subdivided into the Etusis and Kahn Formations. The latter is not developed in the area around Usakos and Karibib nor further south towards the OLZ (Brandt, 1987; Badenhorst, 1992).

#### The Etusis Formation

This rift-type succession was deposited about 770-760Ma (Miller, 1983; Hoffman et al., 1996). The diagnostically pinkish-red rocks are made up of predominantly quartzo-felspathic meta-arkoses and quartzites, occasionally containing thin layers of calc-silicate felses, mica schists and grits as well as a few volcanic and pyroclastic horizons, e.g. E of the Navachab Mine (Kisters et al., 2004). The Etusis Formation shows well-preserved primary sedimentary structures, and cross-bedding is a very common feature. A basal conglomerate of the Etusis Formation often unconformably overlies the AMC (Brandt, 1987; Steven, 1993; De Kock, 2001). This formation is prominently developed within the SCZ, but varies in thickness from being absent to about 3300m, found in a thick wedge on the farm Abbabis, about 25km SSE of Usakos (Smith, 1965; Smith, 1966) The Etusis Formation is thought to have been deposited in half-graben structures and this accounts for considerable variation in this formations' thickness (Stanistreet et al., 1991; Henry, 1992; Miller, 2008).

### **The Swakop Group – Usakos and Navachab Subgroups**

The overlying Swakop Group represents a marine, shelf- to continental slope type depositional sequence overlying the original continental rift-type sediments of the Nosib Group (Smith, 1965; Jacob, 1974; Stanistreet et al., 1991; Henry, 1992; De Kock, 2001). Miller (2008) recently suggested a revised subdivision into a lower Usakos and an upper Navachab Subgroup. The base of each subgroup is marked by the occurrence of a diamictite unit, namely the lower, Sturtian-age (ca. 740Ma) Chuos Formation and the upper, Marinoan-age (ca. 630Ma) Ghaub Formation (Hoffmann et al., 2004). Rocks of the Swakop Group reach several thousand meters in thickness in the NZ and NCZ, but tend to be significantly thinner in the SCZ, outlining the paleohigh of the SCZ in the Neoproterozoic. A further consequence is that formations of the Swakop Group overstep each from NW to SE, so that progressively younger formations are resting unconformably on rocks of the AMC or the Nosib Group from N to S (summarized in Miller, 2008).

## **Usakos Subgroup – Chuos and Arandis Formations**

### **The Chuos Formation**

This formation was first described by Gevers (1931, 1963) as the Chuos tillite, and comprises pebbly schist, phyllites, pebbly marls and marbles. Now generally accepted to be a glaciogenic mixtite (Hoffman et al., 1998 and Badenhorst, 1992), this formation, although not always present, is a good marker horizon within the DSG. The clasts within the matrix show a wide compositional variety and can measure up to 80 cm in diameter. Fragments of granite, pegmatite, schist, vein quartz, calc-silicate felses and angular feldspars can all be observed within the Chuos Formation. Brandt (1985) suggested that a number of the granite boulders present within the Chuos Formation were derived from basement rocks. The Chuos Formation in the Karibib dome reaches a thickness of between 80-180m (Kisters et al. 2004).

The maximum age of the Chuos diamictite is  $746 \pm 2$  Ma, based on U-Pb zircon ages from the underlying volcanics of the Naauwpoort Formation in the NZ. This age correlates with the Sturtian glaciations in the late-Neoproterozoic (Hoffman et al., 1996).

### **The Arandis Formation - Spes Bona, Okawayo and Oberwasser Members**

The Arandis Formation consists of four members, but the lowermost marbles of the Karub Member are not developed in the study area. The remaining members had the rank of formations in previously devised stratigraphies for the northern parts of the SCZ and the NCZ, where these formations reach several thousand meters in thickness (e.g. Badenhorst, 1992). As mentioned above, the identification of the second and stratigraphically higher diamictite unit (Hoffmann et al, 2004) led Miller (2008) to suggest this revised stratigraphy for the SCZ.

### **The Spes Bona Member**

The Spes Bona Member is described as a meta-pelitic succession consisting of predominantly biotite-schists, as well as minor calc-silicate felses, meta-psammities and marbles layers. The name was first used by Badenhorst (1987) to describe a sequence of schist and calc-silicates rocks occurring below the Okawayo Formation on the farm Spes Bona (Badenhorst, 1987). The thickness of the Spes Bona Member is highly variable, but generally decreases from NW to SE. The largest thickness of > 2500m is found around Omaruru in the nCZ, to 600m in the Usakos dome (Steven, 1993; Johnson, 2005; Miller, 2008), to merely 15 m along the southern limb of the Karibib dome, although a structural excision of the formation is inferred (Kisters et al., 2004).

### The Okawayo Member

The Okawayo Member consists of a sequence of banded calcitic- and dolomitic-marbles as well as occasional bands of calc-silicate felses. Marbles are typically bluish-greyish or white in colour. They are either massive and/or finely laminated beds, and intraformational breccia horizons are common. This formation has only been identified in the NCZ and SCZ and reaches a thickness of 70 m in the Karibib dome and about 150m in the Usakos dome (Kisters et al., 2004; Steven, 1993).

### The Oberwasser Member

The Oberwasser Member is now defined as the lower part of the former Oberwasser Formation (after Badenhorst, 1992) occurring below the diamictite units of the Ghaub Formation (see below). The Oberwasser Member as described by Badenhorst (1992) is compositionally similar to the Spes Bona Formation. This formation consists predominantly of biotite- and biotite-cordierite-schists, with minor calc-silicate felses as well as marble breccias and marble horizons. As such it is easy to be mistaken with the Spes Bona Member in cases where the stratigraphy is not well known. The average thickness of the Oberwasser Formation in the Karibib dome is 60-80 m, but attains a maximum thickness of up to 150 m in the Usakos dome (Badenhorst, 1992; Kisters et al., 2004).

### **Navachab Subgroup – Ghaub, Karibib/Tinkas and Kuiseb Formations**

The newly defined Navachab Subgroup (Miller, 2008) consists of four formations. The lowermost Ghaub Formation is only developed in the northernmost parts of the study area, whereas the Karibib, Tinkas and Kuiseb Formations dominate this part of the SCZ.

### The Ghaub Formation – Kachab, Lievental, Omusema and Daheim Members

The Ghaub Formation encompasses the upper parts of the former Oberwasser Formation in the SCZ. The lowermost Kachab Member contains the diamictite (*sensu stricto*) horizons of the Ghaub Formation (Hoffman et al., 2004). It consists of quartzites and schists with dropstones, reaching a thickness of up to 150m outside of Karibib. U-Pb zircon ages from an ashbed in the Kachab Member indicate an age of deposition of  $635.5 \pm 1.2$ Ma (Hoffmann et al., 2004, Miller, 2008). In numerous places in the CZ, very contrasting diamictite types were previously mapped as the Chuos Formation, which also resulted in problematic stratigraphic correlations of units above and below the diamictite unit. It is likely that much of what has been mapped as Chuos Formation, especially near the Okahandja lineament, could now be classified as Ghaub

Formation. The present subdivision recognizes the second, Marinoan-age diamictite as a second and stratigraphically higher unit, separated by a time interval of ca. 110 Ma from the lower Chuos Formation. The Daheim Member is the second significant unit of the Ghaub Formation. The Daheim Member consist of alkaline mafic volcanics and volcanoclastics, now developed as amphibolites, amphibolitic scoria and breccias. Highly variable along-strike thickness variations are ascribed to the presence of volcanic centers (Badenhorst, 1992). The Daheim Member may reach a thickness of up to 120 m e.g. in the hinge of the Usakos dome, but laterally thins to narrow, only a few meter-wide amphibolite horizons that probably represent basaltic flows away from the central feeder structure (Badenhorst, 1992). The Daheim Member is also not developed in the study area.

#### The Karibib Formation

The Karibib Formation consists entirely of a thick marble succession with a few interlayered calc-silicate felses. It is a very characteristic and regionally developed unit throughout the SCZ. Around the town of Karibib, the Karibib Formation occurs as massive- to banded, grey and white calcitic and dolomitic marbles. These are often interspersed with marble breccia horizons, particularly in the lower parts of the formation. Finely laminated calc-silicate felses may form up to several meter wide units in the marbles. They commonly weather to a rusty-brown colour and are, thus, very obvious in the field. The competence contrast between marbles and calc-silicate felses commonly results in spectacularly developed folding and/or boudinage. In the Karbib and Usakos region, the marble unit reaches a thickness of 500-600m (Badenhorst, 1992; Kisters et al., 2004). Structural duplication may lead to thicknesses of well over 1.2km (Johnson et al., 2006, Kitt, 2008). In the northern parts of the study area around the Abbabis inlier, the Karibib Formation is dominated by white, commonly dolomitic marbles that are mined in a number of quarries. Here, and in most parts of the study area, the marbles are commonly pervasively recrystallized to coarse-grained, sugary beige-white marbles and primary sedimentary features are only rarely preserved. This is a distinct difference to marbles of the Karibib Formation further north in the Karibib and Usakos region, where primary sedimentary features are well preserved.

#### The Tinkas Formation

The Tinkas Formation only occurs along the NW margin of the Okahandja Lineament. It is thought to be an upward and lateral transition between the Karibib and Kuiseb Formation. Jacob (1974) considered it to be a lower member of the Kuiseb Formation (see also De Kock, 2001).

This formation is dominated by thick mica schists alternating with thin calc-silicate felses, marbles and minor amphibolites (Brandt, 1985). The rocks are interpreted as turbidites deposited on the upper continental slope facies and below the shelf-type Karibib Formation marbles that dominate further north. The interfingering of clastic and carbonate sediments is thought to represent the reworking of the carbonates of the Karibib Formation and deposition in deeper water environments so that large parts of the Karibib and Tinkas Formation are probably time-equivalents and merely facies variations. Numerous intraformational breccias mark the contact between the shelf- and continental-slope facies. Along the Okahandja Lineament Zone, De Kock (2001) and Miller (2008) distinguished a more proximal Kuduleck Member from a more distal Quelle Member. The Lievental Member (calc-silicates and marbles) of De Kock (2001) has subsequently been classified as part of the Ghaub Formation by Miller (2008).

#### The Kuiseb Formation

The regionally widespread Kuiseb Formation is the uppermost unit of the Damara Supergroup. The Kuiseb Formation is dominated by quartz-biotite schists as well as interlayered psammities and psammitic schists rich in feldspar and occasional calc-silicate layers. Cordierite, sillimanite and garnet are common porphyroblasts in the Kuiseb schists. The mineralogy of the Kuiseb Formation, however, varies with the composition and metamorphic grade across the CZ (Brand, 1985; Steven, 1993; De Kock, 2001).

Although few sedimentary structures (especially in the SCZ) are preserved in the Kuiseb schists, rock successions resemble Bouma-sequences that have led most authors to suggest that the Kuiseb Formation represents a turbidite succession (Porada & Wittig, 1983; Kukla et al. 1988). The rocks are developed from the NZ throughout the CZ and into the SZ, indicating the connection of the northern Outjo with the southern Khomas Sea (Miller, 2008). Although the top of the Kuiseb Formation is not preserved, Smith (1965) suggested a minimum thickness of 3300 m for the metaturbidites south of the Abbabis basement inlier in the SCZ. The Kuiseb Formation schists are the only formation developed in the SZ, located on the lower Kalahari plate, reaching a thickness of at least 10 km (Downing & Coward, 1981; Miller, 1983), although structural duplication must be assumed (Kukla & Stanistreet, 1991).

### **2.4.2 Intrusive units**

The CZ and parts of the NZ of the Damara belt are thought to represent the magmatic axis of the orogen, being intruded by numerous pre-, syn- and post-tectonic granitoids (Hawkesworth et al., 1983, Marlow, 1983; Miller, 1983; Jung et al., 1998, 2000, 2001).

Pan-African intrusive rock types are exposed across an area of ca. 70 000 km<sup>2</sup>. 96% of these intrusive units are granitic with the remaining 4% being gabbros, diorites and granodiorites (Miller, 1983; Jung, & Mezger, 2003). Miller (1983) has divided the CZ's intrusive units into 3 main groups. This subdivision has been used by numerous workers in the Damara belt.

These 3 groups are:

1. Fine- to medium-grained Red granites associated with rocks of the basement and the Nosib Group.
2. Coarsely porphyritic, biotite rich monzogranites and associated diorites. These include the broadly developed Salem-type granites and the Goas diorite suite.
3. Coarse leuco-granites, pegmatoidal-alaskites and pegmatites, all thought to be late- to post-tectonic.

In the region South of Karibib the two most common intrusive rock types are the diorites and granodiorites of the Goas diorite suite (Lehtonen et al., 1995; Jacob et al., 2000) as well as a number of porphyritic biotite-rich granites and diorites known as Salem-type granites (Gevers, 1963; Miller, 2008). At the OLZ lies the laterally extensive Donkerhuk granite. Associated with these extensive large intrusions are numerous smaller leuco-granites and pegmatites. Locally abundant, unique isolated granitoids (e.g. the Otjimbinge syenite) have also been intruded into the SCZ. Most of these smaller bodies however, fall into one of the three groups described above (Miller, 1983; Brandt, 1985; Jacob et al., 2000, Smith, 1965; Smith, 1966).

Granites are discussed in more detail in chapters 3 and 4.

### **2.4.3 Structure**

A polyphase deformation history has long been identified for the rocks of the SCZ (Miller, 1983). Structural work in the SCZ and the OLZ has been done by inter alia, Smith (1965), Jacob (1974); Sawyer (1981), Barnes & Sawyer (1980), Kasch (1983a), Coward, (1983) Kröner (1984), De Kock, (1989) Miller (1983), Oliver (1994) Poli and Oliver (2001), De Kock (2001), Kisters et al. (2004), Johnson (2005) and Miller (2008).



Although different authors describe a slightly different sequence of deformation phases for rocks across the SCZ, most workers in the Damara describe 3 main deformation phases (D1-D3) (Miller, 2008). Jacob (1974) Barnes & Sawyer (1980), Sawyer (1981), De Kock (1989) and Steven (1993) describe an additional D4 phase (Table 2.2).

Table 2.2: Deformation in the CZ. This table compares the deformation phases described by different authors, referring specifically to the structural features as being diagnostic of specific deformation stages. Many of these structural interpretations are based on studies at specific locations and probably does not refer to structural features across the entire CZ. Dates given for deformation phases were taken from Miller (2008).

Deformation phase	Jacob (1974)	Sawyer (1981)	Steven (1993)	Poli & Oliver (2001)	Kisters et al. (2004)	Miller, (1983; 2008)	Age of deformation
<b>D1</b>	<b>D1</b> - S1 is axial planar to mesoscopic isoclinal F1 folds. F1 folds refolded by later deformation.	<b>D1</b> - Tight to isoclinal folds with sub-horizontal axial planes.	<b>D1</b> - S1 is a bedding (S0) parallel fabric often seen wrapping around domes	<b>D1</b> - A strong S1-fabric is defined by biotite and quartz alignment parallel to S0. L1 fabrics are formed by sillimanite, cordierite and feldspar.	<b>D1</b> - A Bedding parallel/sub-parallel foliation and associated F1 intrafolial folds. Low angle thrusting and truncation faults	<b>D1</b> - Local D1 thrusts and large scale F1 recumbent folds	Prior to 560 Ma
<b>D2 (Early D2)</b>	<b>D2</b> - F2 defines the NE structural trend for many areas in the CZ, forming large scale antiforms and synforms. Folds can be isoclinal, but fold tightness is a function of variable rheology.	<b>D2</b> - A S2 fabric defined by biotite preferred orientation - orientated: 310-350/15-30	<b>D2</b> - Upright tight F2 folds trending 058°				
<b>D3 (Late D2)</b>	<b>D3</b> - Minor F3 open folds on steep F2 limbs (Scarce). A S3 fanning cleavage is occasionally developed.	<b>D3</b> - Subvertical F3 folds and associated schistosity trending 60°-240°	<b>D3</b> - Large scale upright open folds, forming F3 antiforms and synforms and a associated S3 schistosity.	<b>D2</b> - S2 planar-axial foliations are sporadically developed with no preferred orientation. Constrictional folding: sheath folds verging SW.	<b>D3</b> - NE trending doubly plunging domes defining the structure grain of the SCZ	< 542 Ma	
<b>D4 (Late D2)</b>	<b>D4</b> - F4 structures are superimposed on F2 folds. F4 seen as NNE trending crenulations (scarce).	<b>D4</b> - F4 folds dipping 327/42 and plunging 45/07	<b>D4</b> - NW-verging F4 folds (late-tectonic) as well as the development of a spaced axial planar S4 cleavage. F3 seems to be rotated by F4				

Recent studies by Oliver (1994), Poli & Oliver (2001), Kisters et al. (2004) and Johnson (2005) describes two main deformation phases (D1-D2), this implies that features that were previously described as a result of D3 and D4 are in fact the result of a extended progressive D2 phase.

An additional D5 phase was described by both Sawyer (1981) and De Kock (1989) for the strain aureole of the Donkerhuk granite intrusion along the OLZ.

This study follows the approach of Poli & Oliver (2001) and ascribes deformation to two progressive deformations events (D1 and D2).

D1 to D2 development in the SCZ:

Continental convergence D1, led to the formation of early recumbent folding and shallowly dipping thrusts as well as the development of an associated shallow, bedding ( $S_0$ ) parallel,  $S_1$  cleavage.

D2 led to refolding and reworking of D1 structures and resulted in the formation of numerous open to isoclinal NW-verging folds and a well developed (especially towards the margin of the OL) axial planar  $S_2$  cleavage. D2 structures often strongly overprint D1 structures, making them nearly impossible to identify. An example of fold overprinting is described for a structure called the Rooikuseb-anticline (near the OLZ) which clearly exposes a early regional scale F1 recumbent fold overprinted by a later F2 folds. F2 folding at the OLZ verges SE (Downing & Coward, 1981).

A constrictional linear fabric ( $L_2$ ) is often found associated with  $S_2$ . A  $L_2$  fabric has been described by Smith (1965) as the result of F1 and F2 fold interference.  $L_2$  is best observed as mineral-preferred orientations and mineral-stretching lineations plunging NE throughout the CZ (Smith, 1965; Downing & Coward 1981; Miller, 1983; Poli & Oliver, 2001).

A subsequent late-D2 (D3 of Miller 1983) event is interpreted to have formed km-scale NE-SW trending oval dome shaped structures which are very characteristic of the CZ. The formation of these domes remain controversial and a variety of mechanisms have been proposed, e.g. Smith (1965), Barnes & Sawyer (1980) Jacob (1983), Miller (1983), De Kock (1989) and Sawyer (1981), Steven (1993) and Johnson (2005). Poli and Oliver (2001) suggest dome formation to be the result of “mid-crustal, constrictional folding”, the result of progressive late-D2 deformation and not necessarily as the result of a distinct additional deformation episode producing interference structures. For more information on the nature of these dome structures the reader is referred to Johnson (2005).

Across the SCZ a change in structural style is seen from SW to NE. The granulite-facies SW portion of the SCZ is dominated by SW extruded sheath fold structures, whereas more brittle deformation features such as thrust imbrication dominates in the NE part of the SCZ where lower grade metamorphic rocks lie exposed. Kisters et al. (2004) explains this feature to be the result of different rheologic responses between lower crustal granulite facies rocks in the SW and lower-grade amphibolite and greenschist facies rocks in the NE of the SCZ.

## Structure of the OLZ and the SZ

The Okahandja lineament zone lies at the boundary between the SCZ and the SZ and is believed to be a late-tectonic (F3) downfold of the crust (Miller, 1979). The OLZ is not only a structural transition, but also a clear stratigraphic one representing the passive margin of the Congo craton (De Kock, 1989). Structurally this area is characterised by tight and isoclinal upright folding. Folds along this margin are upright often verge slightly towards the SE (Blaine, 1977; Sawyer 1981). Towards the Donkerhuk granite along the OL a  $S_5$  foliation thought to be associated with the intrusion of the Donkerhuk body is described by De Kock (1989) and Sawyer (1981). Sawyer (1981) describes this  $S_5$  fabric as a “crenulations cleavage with associated pressure solution giving rise to a dark biotite-rich striping.”

In the SZ NE-trending fabrics are more linear than basement-dome geometries of the CZ. Only a single foliation ( $S_3$ , Miller, 2008) is present in schist just to the south of the OLZ as a well developed spaced cleavage. A lack of compositional banding in homogenous schist of the SZ makes folding hard to identify (Downing & Coward, 1981).

### 2.4.4 Metamorphism

The CZ (NCZ & SCZ) of the Damara belt is a classic high-T low-P metamorphic terrain. Peak metamorphism increases from NE to SW along the central zone changing from amphibolite facies conditions around Karibib at  $3\pm 1$  kbar and 550-600 °C to lower-granulite facies towards Swakopmund where peak metamorphic conditions of  $7\pm 0.5$  kbar and  $760\pm 50$  °C were reached (Puhan, 1983; Masberg, 2000, Jung & Mezger, 2003; Ward et al., 2008). The increase in metamorphic grade along strike towards the SW indicates that the current erosional surface in the CZ exposes an oblique crustal section, with deeper crustal levels being exposed along the Atlantic seaboard (Barnes & Downing, 1979).

Masberg (2000) explains metamorphism in terms of a single protracted complex clockwise P-T metamorphism for the Damara orogen, explained by burial-heating followed by exhumation, isothermal decompression and subsequent cooling (Masberg, 2000). Nex et al. (2001) suggest the development of 2 main metamorphic events, M1 and M2 in the CZ. Indeed an earlier M1 metamorphic episode has been identified by early authors such as Nash (1971), Blaine (1977) and Sawyer (1981) who found early metamorphic assemblages not related to M2 metamorphism. M1 was a high-T, med-P event thought to have occurred during the main collisional episode (Miller, 2008). M1 is not well defined and has been estimated to have occurred at about 540-530Ma

reaching pressures of up to  $4\pm 1$  kbar and 550-600 °C in the SCZ (Nex et al., 2001; Jung & Mezger, 2003).

Peak metamorphic conditions in the SCZ were reached by 535-470 Ma (or 535 Ma according to Miller, 2008) during the M2 event (Jung et al., 2001; Jung & Mezger, 2003). Jung & Mezger (2003) describe M2 as being a more complex event, consisting of a number of discrete heat pulses, and M2 is thought to be the result of voluminous post-tectonic granite intrusion across the CZ.

Metamorphism in the SZ, SMZ and the OLZ.

The SZ is a high-P, med-T metamorphic terrains. M1 and M2 metamorphism occurred in both the SZ and SMZ, with a drop in temperature occurring between these 2 events. Peak M2 metamorphism in the SZ occurred during the same time as in the SCZ and reached peak metamorphic conditions of about 7.6 Kbar and 550 °C (in the Gamsberg-Kuiseb river) and 8.4 kbar and 570 °C (around Omitara) (Kasch, 1983b; Miller, 2008).

Symmetric metamorphic isograds along the margin of the Donkerhuk granite at the OLZ imply that this large batholith has a metamorphic aureole (Sawyer, 1981; Kasch, 1983b). No detailed metamorphic work has been done on this suggested aureole, but a number of unique metamorphic assemblages adjacent to the Donkerhuk granite are attributed to its thermal effects (Hoffer, 1977, Kasch, 1983b).

## **Chapter 3 Lithostratigraphy and Petrography**

### **3.1 Introduction**

The regional traverse studied here transects a number of distinct lithological facies changes across the SCZ so that different stratigraphic subdivisions have to be adopted for different areas. For the northernmost areas around Karibib, the stratigraphic subdivision follows that of Miller (2008) (Table.2.1). A significant facies change occurs southwards and towards the OLZ. Previous works have explained lithological variations and the occurrence and/or disappearance of units in this part of the SCZ mainly as the result of a morphologically structured basement high in the southern parts of the SCZ during sedimentation. Although clear facies variations are found across the traverse, the present study also suggests that significant aspects of the lithological variability encountered in the SCZ may be structurally controlled. This will be discussed in chapter 4. This chapter presents the lithological inventory and stratigraphic make up across the traverse. It also describes plutonic, mainly granitic rocks, along this traverse.

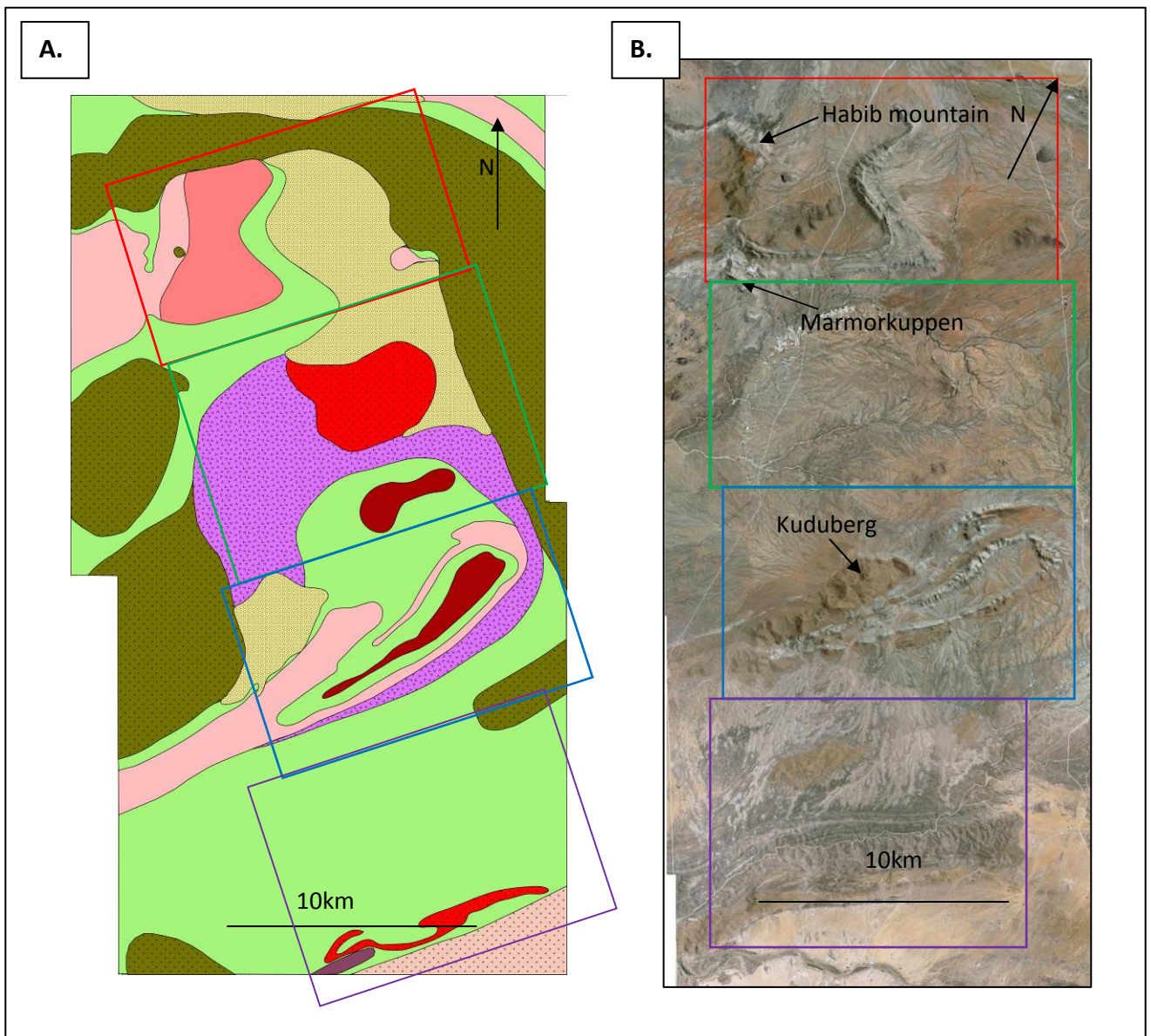
### **3.2 Tectonostratigraphic domains**

In order to account for the lithological and also structural variability (chapter 4) across the traverse, the field area has been subdivided into a number of discrete tectonostratigraphic domains (Fig 3.1).

The division of the field area is based on (1) different stratigraphic relationships, (2) the abundance and geometry of granite plutons, and (3) different structural styles, fabric development and fabric intensities across the traverse. This chapter only considers the stratigraphic changes across these domains. The mapping domains are named after the farms on which they occur and the resultant maps are included as Appendices I-IV.

The tectonostratigraphic mapping domains, from NW to SE, are (Fig 3.1):

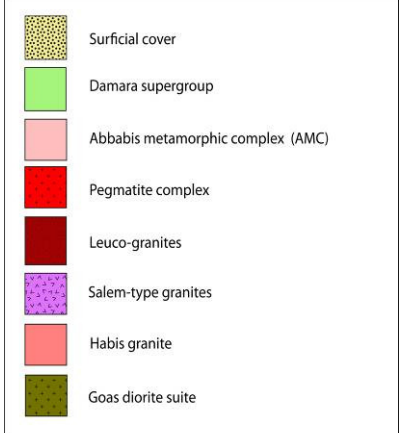
1. Etusis
2. Dorneb
3. Audawib
4. Otjimbingwe



Subdivision of mapping domains:

1. Etusis
2. Dorneb
3. Audawib
4. Otjimbingwe

**A.)** Distribution of granites and metasediments along the ca. 40km long and 10-12 km wide structural transect across the SCZ between Karibib and Otjimbingwe. (modified from Smith, 1966) **B.)** Aerial photo (Google Earth image) of the traverse, with outlines of the four mapping domains.



**Fig 3.1:** The four tectonostratigraphic mapping domains

### 3.2.1 Etusis domain (Appendix I)

The Etusis domain is the northernmost domain of the structural transect, forming the transition between the fully developed sequence of the Damara Supergroup (DSG) in the Karibib/Usakos region north of the AMC (Badenhorst, 1992) with a greatly reduced and changed stratigraphy to the south of the AMC (De Kock, 1989, 2001, Miller 2008). The Etusis domain is dominated by a large anticlinal inlier of the AMC, surrounded by Damaran metasedimentary rocks. The eastern extent of this AMC inlier is intruded and completely obscured by the intrusion of the Pan African Habis granite, for which, to date, no geochronological data is available. Except for a diamictic calc-silicate unit directly below marbles of the Karibib Formation, the stratigraphy conforms to that described by Badenhorst (1992). All of the lower formations of the Damara sequence (Etusis Formation to Ghaub Formation) pinch out and disappear towards the eastern part of this domain. Marbles of the Karibib Formation directly overlie the Habis granite (Smith, 1966). Further east, only surficial cover is present. At the northern margin of the domain, massive marbles of the Karibib Formation are underlain by the Mon Repos diorite body (Goas diorite suite).

### 3.2.2 Dorneb domain (Appendix II)

The Dorneb domain is dominated by extensive platforms underlain by rocks of Salem-type granites that are intruded into the Kuiseb Formation and the NW and SE boundaries of this domain are drawn roughly along this intrusive contact. The Salem granites are in contact with rocks of the Gamikaub diorite and the Okongava diorite (Chapter 3.4) in the SW and NE boundaries of this domain, respectively.

### 3.2.3 Audawib domain (Appendix III)

The Audawib domain is dominated by a central, complexly folded, regional-scale fold structure, henceforth referred as the Audawib fold complex. The fold complex can be mapped as several prominent NE trending ridges, exposing rocks of the AMC, Etusis, Karibib, and Kuiseb Formations. It is surrounded by mainly biotite-quartz schists of the Kuiseb Formation in low-lying areas. The Salem granites outcrop towards the NW, NE and SE of the Audawib fold complex and completely encloses it. Much of the Kuiseb Formation in this domain has been intruded by fine-grained distinct leucogranites.

### 3.2.4 Otjimbingwe domain (Appendix IV)

The Otjimbingwe domain follows to the immediate SE of the Audawib fold complex. The NW-portion of this domain forms a large peneplain dominated by open- to isoclinally folded rocks of the Kuiseb Formation. The metaturbidites grade into rocks of the Tinkas Formation towards the SE, characterized by upright folding and pervasive bedding transposition. This zone corresponds to the OLZ defined and described by Smith (1965), Downing & Coward (1981), Downing (1983) and Henry et al. (1990). In the SE, the upright rocks of the Kuiseb Formation have been pervasively intruded by sheeted leucogranites, forming part of the intrusive aureole of the large Donkerhuk granite that forms the SE boundary of this mapping domain. The contact aureole of this pluton reaches about 3 km into the Tinkas Formation, where numerous granite and pegmatite sills related to the Donkerhuk granite have been intruded. Two large plutons, the Palmental diorite and the Otjimbingwe syenite, have also intruded into metasediments in the SE of this mapping domain.

Correlations of the geological formations across these mapping domains are presented hereafter in Fig 3.2.

## 3.3 Lithological inventory

### 3.3.1 The Abbabis Metamorphic Complex (AMC)

#### Description

The AMC contains a variety of rock types, which include crenulated schists, calc-silicate felses, and quartzo-feldspathic augen gneisses, which are by far the most common type of basement rock found in the field. The AMC is intruded by numerous, pinkish-orange quartzo-feldspathic pegmatites.

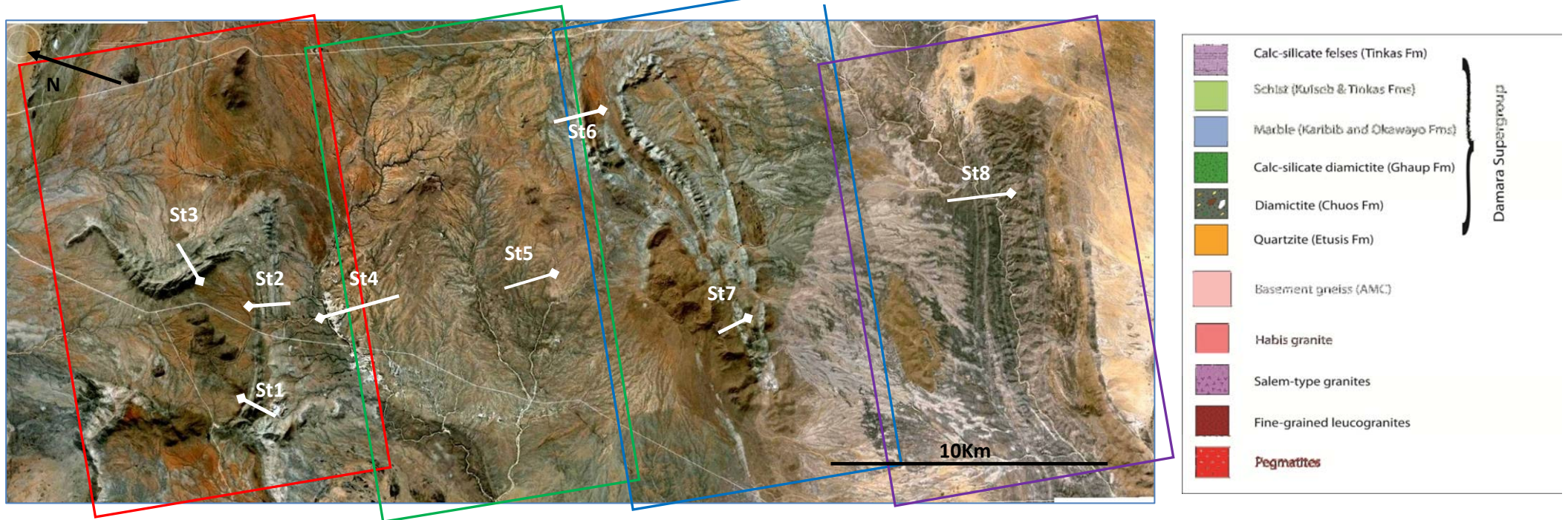
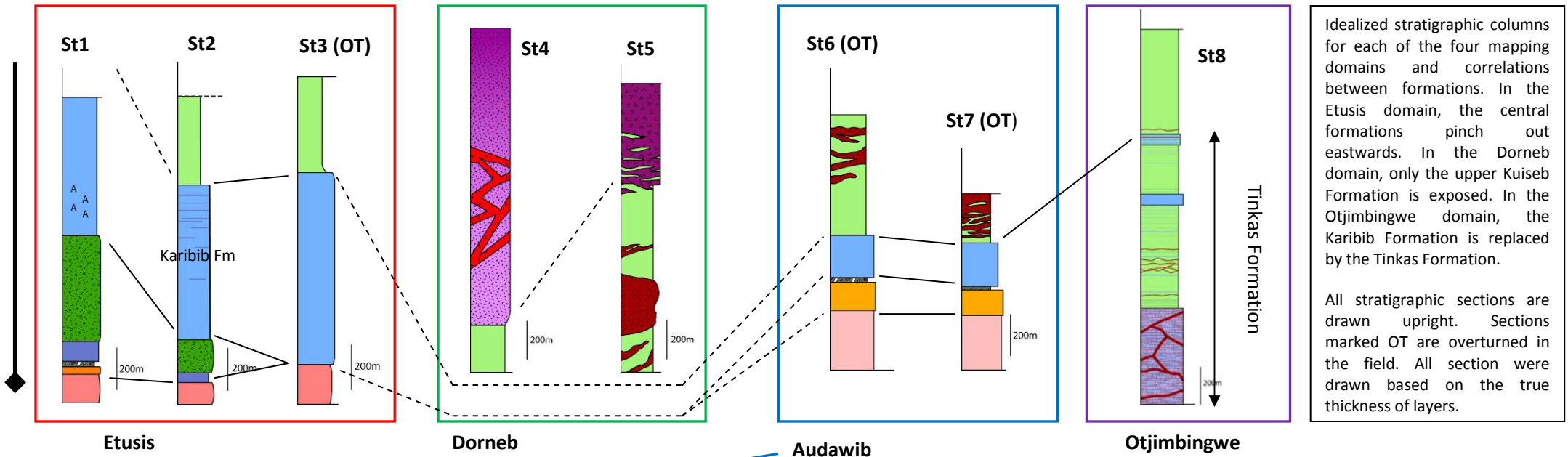
#### Distribution

Basement rocks of the AMC occur in the Etusis, and Audawib mapping domains. In the Etusis mapping domain, the rocks of the AMC are exposed in a basement window in the middle of this domain, surrounded by rocks of the DSG. These are intruded by the Habis granite body that occurs towards the E of the main AMC exposure. Contact relationships in the field appear rather gradational and a clear distinction between basement gneisses and foliated megacrystic granites is not always clear. Augen gneisses occur at the westernmost part of this domain, where outcrop is scarce.

In the Audawib domain, orange-coloured augen gneisses of the AMC are found on the farms Gamikaub and Audawib. The prominent Kuduberg mountain (22°12'21"S; 15°54'13"E) on the farm Audawib consists entirely of these gneisses.



**Fig 3.2: Stratigraphic correlation of the four domains**



Basement rocks show very few preserved primary sedimentary features and, in most cases, their origin cannot be determined. Pervasively developed fabrics and recrystallization of the rocks (at least in the northern Etusis domain) and unconformable contacts with the overlying rocks of DSG indicates that the gneisses have been through an earlier tectonometamorphic cycle (e.g. Brandt, 1987; Steven, 1993).

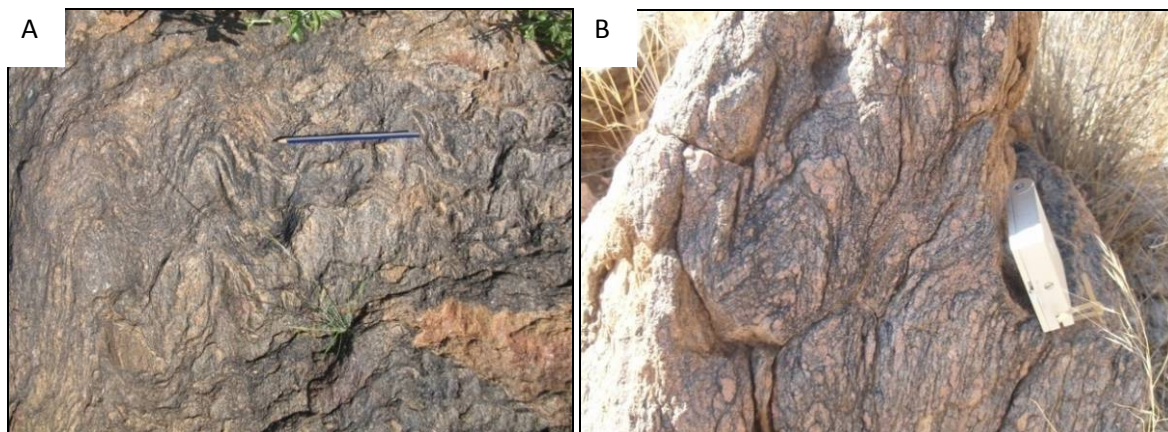
## Petrography

### Muscovite schist

Fine-grained, crenulated quartz-muscovite schists are locally developed in the Etusis mapping domain. Other minerals include feldspar, biotite, chlorite and opaques. These accessory minerals (other than quartz and muscovite) make up about 20% of the modal mineralogy of the rocks and are distributed in approximately equal proportions (Fig 3.3.A & Fig 3.4.A).

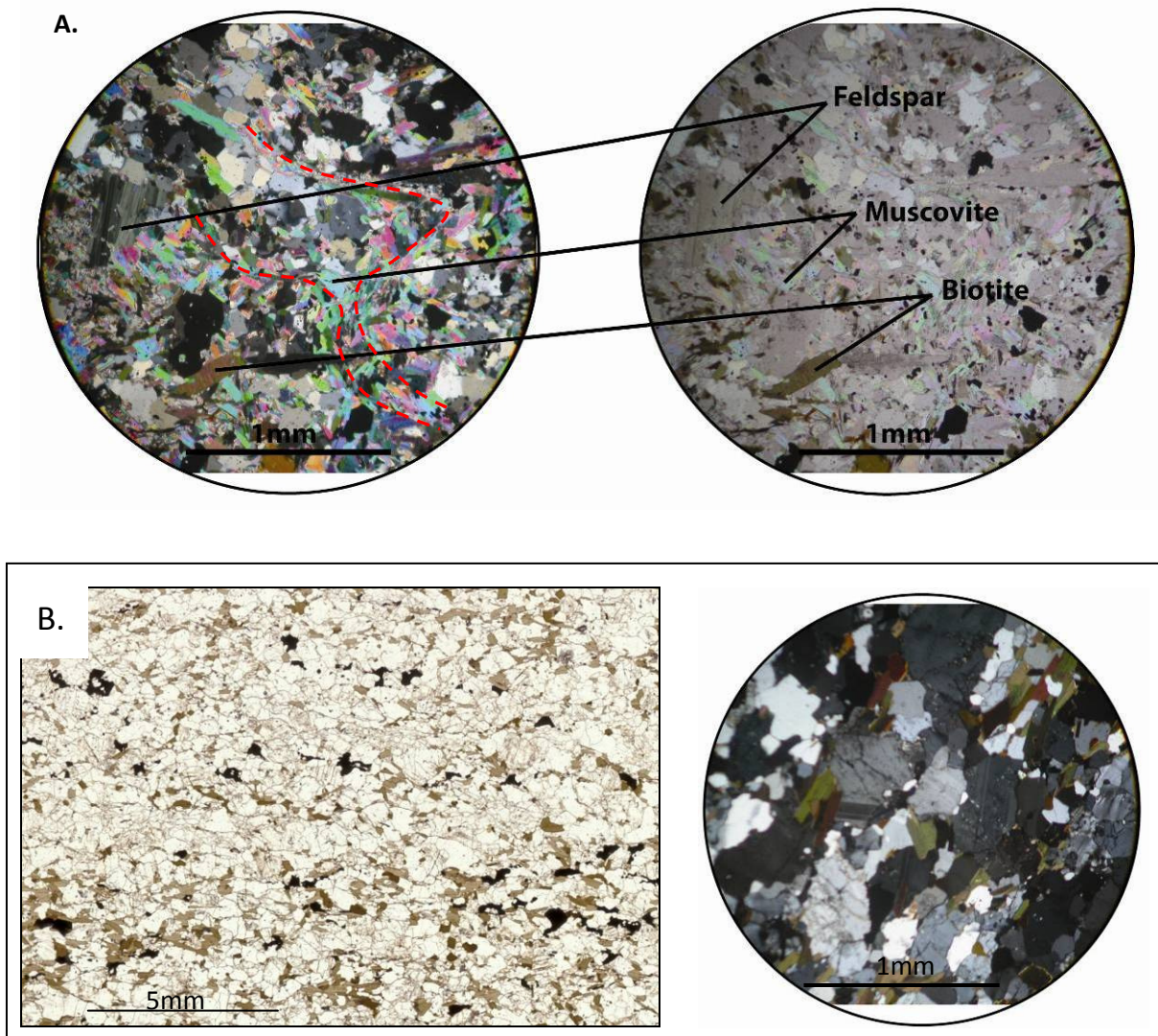
### Banded gneisses and augen gneisses

Compositional banding in these gneisses is defined by finely-banded streaks of K-feldspar, quartz and biotite (Fig 3.3.B & Fig 3.4.B). The rocks are commonly pinkish in colour. Biotite makes up about 15 vol % of the mineralogy, shows a preferred orientation aligned to the dominant foliation defined by ovoid, 1-3cm large K-feldspar augen and the grain-shape preferred orientation of quartz-feldspar aggregates (Fig.3.3.B). The crystal matrix is medium- to fine-grained (ca. 1mm grain size) and, for the most part, statically recrystallized (Fig.3.4.B).



**Fig 3.3: A.)** Crenulated AMC schists on the farm Etusis at 22°05'02"S;15°48'09"E. Photograph looking onto foliation plane. **B.)** Red augen gneisses that make up the majority of the AMC basement rocks, showing pink K-feldspar augen set in an anastomosing quartz-biotite-feldspar matrix, that defines the gneissosity. (Photograph taken on the farm Audawib at 22°13'05"S;15°53'49"E).





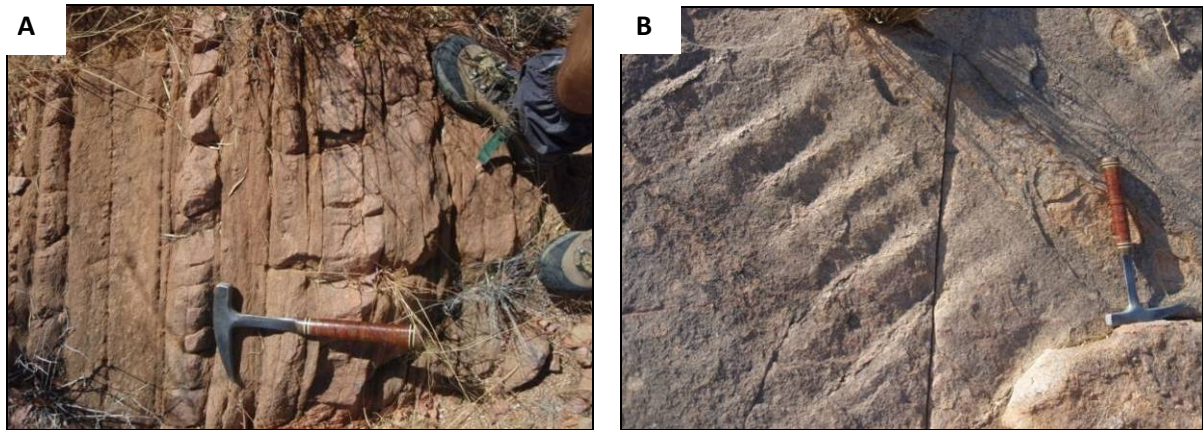
**Fig 3.4: A.)** The AMC schist. The highly birefringent mineral is muscovite, with larger qtz and feldspars appearing greyish. Red dashed lines indicate a weak crenulations formed by the alignment of muscovite crystals (left: plane light; Right: crossed polars). **B.)** AMC gneiss of predominantly feldspar and brown biotite (greenish under crossed polars) crystals. AMC gneisses show a clear foliation, in this case a Damaran overprint. The left-hand square was photographed in plane light and the right-hand photo was taken under crossed polars.

## The Damara Supergroup (DSG)

### 3.3.2 The Etusis Formation

#### Description

The Etusis Formation is always developed as the basal formation of the DSG along the traverse and consists of fine-grained pink/orange feldspathic-quartzites. Conglomeratic layers are particularly common in its basal parts where the rocks unconformably overlie basement rocks of the AMC. The weathering resistant Etusis Formation often forms topographic highs. Primary sedimentary features, such as primary bedding, cross-bedding and ripple marks are often preserved (Fig 3.5 & 3.6.A)



**Fig 3.5:** **A.)** Typical Etusis quartzite, showing primary bedding still preserved in cross-section. **B.)** Ripple marks preserved on the erosion surface of the Etusis quartzites. Both photos were taken on the farm Audawib.

### Distribution

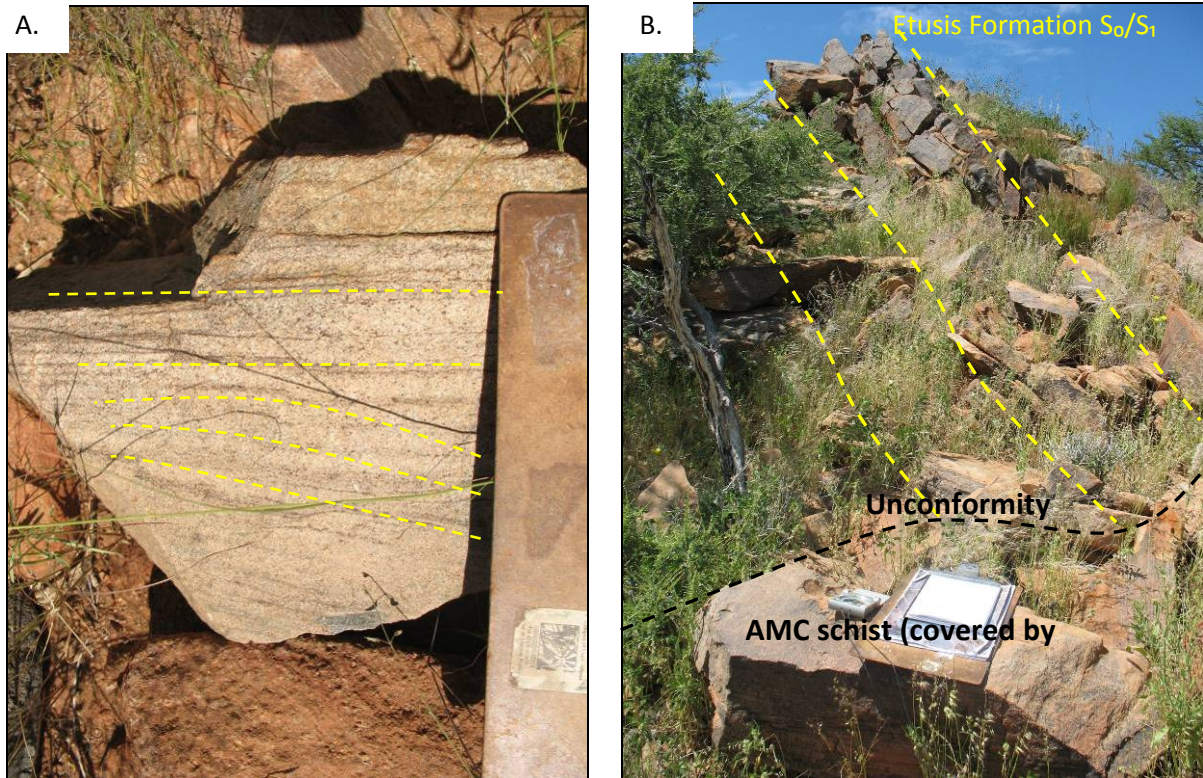
Rocks of the Etusis Formation occur in the Etusis, Audawib and Otjimbingwe domains.

In the Etusis domain, the rocks are developed along the southern boundary the AMC. The Etusis Formation is at its thickest in the W of the domain, reaching a true thickness of ca. 90 m. This formation thins and then completely pinches out eastwards. It is not found to the west of the C32 road between Otjimbingwe and Karibib. Well-preserved unconformable contacts between the Etusis Formation and the AMC were recognised at co-ordinates, 22°5'36"S; 15°48'29"E (Fig 3.6.B). Conglomerates are often present in the Etusis Formation along this contact (Fig 3.7.A)

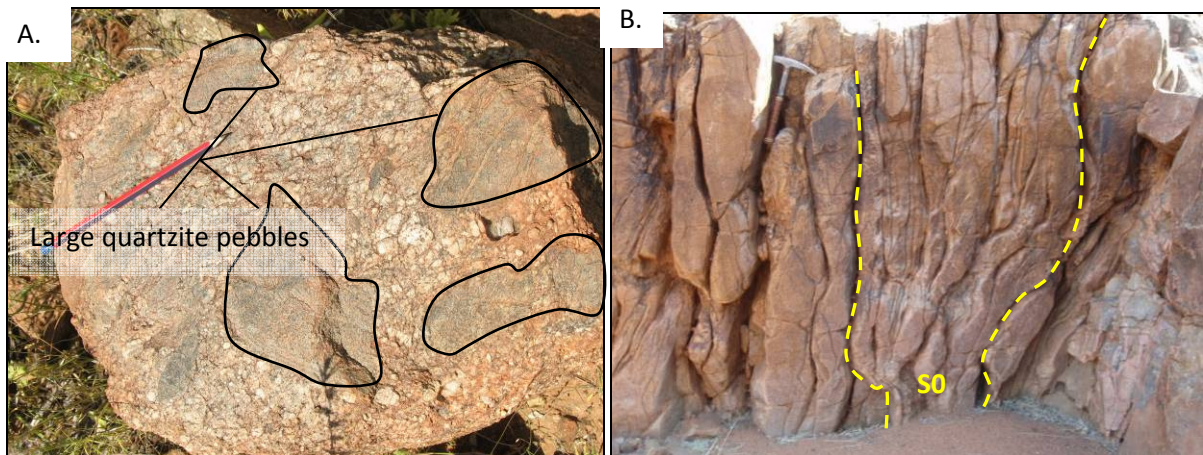
Further SE in the Audawib domain the Etusis Formation is well developed reaching a maximum thickness of 120 m and an average thickness of about 30 m. Although no basal conglomerate is developed here, cross-bedding can occasionally be seen and, in one locality (22°12'50.01"S; 15°53'29.22"E), ripple marks have been preserved (Fig 3.5.B). In the Audawib domain, unconformable contacts between the AMC and the Etusis Formation cannot be identified. Instead, both basement gneisses and quartzites of the Etusis Formation contain a pervasive, contact-parallel foliation. In this domain, the Etusis Formation commonly underlies basement gneisses, indicating a higher degree of tectonism and the overturned nature of stratigraphy.

An isolated hill of highly-strained Etusis Formation quartzite occurs in the northern portion of the Otjimbingwe domain (see Appendix IV). Here, primary sedimentary features have largely been obliterated and the  $S_0$  surfaces are crenulated and overprinted by a tectonic ( $S_1$ ) foliation (Fig 3.7.B), defined by the preferred orientation of biotite. Bedding is only preserved in lithon-like low-strain domains on a decimetre scale. The lack of continuous bedding features hampers any determination of the true thickness of this km-scale outcrop of Etusis Formation.





**Fig 3.6: A)** Cross-bedding preserved in a boulder (yellow dashed lines) of the Etuis Formation quartzite (sectional view). **B.)** Inclined strata of the Etuis Formation unconformably overly shallowly-dipping, foliated schists of the AMC (looking west). Both photos were taken at 22°05'35"S; 15°48'29"E with a A4 clipboard for scale.



**Fig 3.7: A.)** A Basal conglomerate of the Etuis Formations. Quartzo-feldspathic fragments lie in gritty medium-grained matrix. Photo was taken at 22°05'37"S; 15°48'39"E on the farm Etuis. **B.)** A sectional view of sheared Etuis Formation, characterized by a pervasively developed, anastomosing, subvertical  $S_1$  (Photos was taken at 22°16'55"S; 15°55'32"E, looking W).

### Petrography

These medium- to fine-grained (< 1-2 mm), equigranular rocks consist predominantly of quartz (60-80%) and K-feldspar (30-20%) as well as muscovite, biotite and iron-oxides, the latter often forms darker bands delineating the foresets of cross-beds (Fig 3.6.A). Pyroclastic horizons described from

the Etusis Formation in the Karibib dome to the immediate N (Kisters et al., 2004) were not identified.

### **3.3.3 The Chuos Formation**

#### Description

A grey diamictite conformably overlying the Etusis Formation has been classified as the Chuos Formation. This singular unit contains angular feldspar and granite clasts set in a fine-grained, grey, feldspar and quartz matrix (Fig 3.8.A,B).

#### Distribution

The Chuos Formation forms a mappable unit only in the Etusis domain when it is up to 15m thick. The diamictite overlies quartzofeldspathic rocks of the Etusis formation and is overlain by a marble unit interpreted to represent the Okawayo Formation. This section of the Chuos Formation gradually thins and eventually pinches out over a distance of ca. 3 km towards the eastern boundary of the farm.

Further south on the farm Audawib, a thin, up to 3 m thick band of diamictite is intermittently developed between the Etusis Formation and the marbles of the Karibib Formation. This unit can most likely be correlated to the Chuos Formation, but its actual stratigraphic position must remain tentative.

#### Petrography

The grey diamictic rocks of the Chuos Formation have a mortar texture and show very little grain sorting (Fig 3.8.B). A wide range of clast sizes co-exists in a fine- to medium-grained wacke-like matrix. The clasts range in size from granite boulders up to 40 cm in diameter to angular feldspar fragments of 0.25 mm in size. The matrix is made up of fine-grained, recrystallized, equigranular quartz and feldspar of ca. 40-120  $\mu\text{m}$  in diameter. Biotite makes up about only about 5-10 % of the modal mineralogy. Green fine (<0.05 mm) calc-silicate minerals, most commonly diopside are occasionally observed, but make up < 1% of the modal mineralogy. On average the fine-grained matrix makes up between 50-40 % of the total modal composition. Hardly any depositional banding is visible, but the preferred orientation of biotite, as well as the re-orientation and flattening of many clasts (primarily feldspar), define a weak tectonic fabric (Fig 3.8.B).



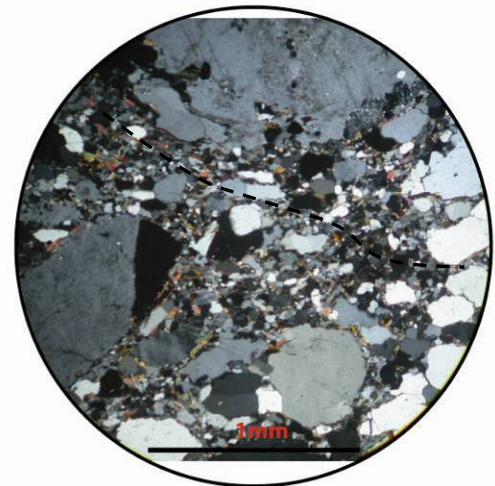
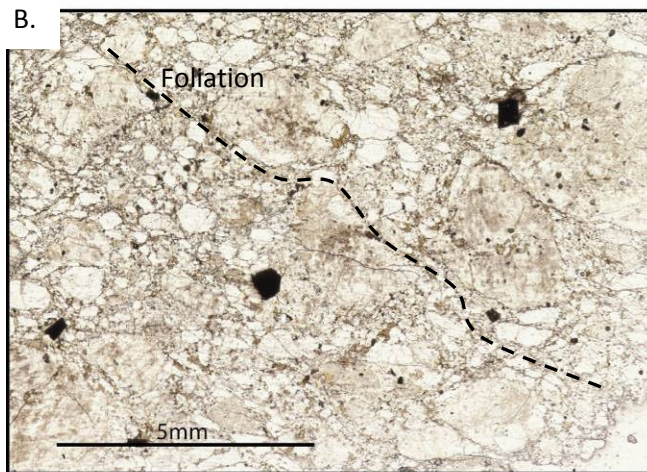


**Fig. 3.8: The Chuos diamictite**

A.) In the field, large K-feldspars and granite fragments are clearly visible in a fine-grained grey matrix.

B.) Thin sections show the very poor sorting and the variety of fragment sizes within a fine grained matrix. Large clasts are mostly feldspars, while the fine matrix consists of quartz and feldspar with occasional opaques and brown biotites. A black dashed line outlines a weak foliation.

Photo (A) was taken at 22°05'39"S; 15°49'08"E (plan view) and thin section (B) samples are from this locality slides show a section in plane light (left) and under crossed-polars (right).



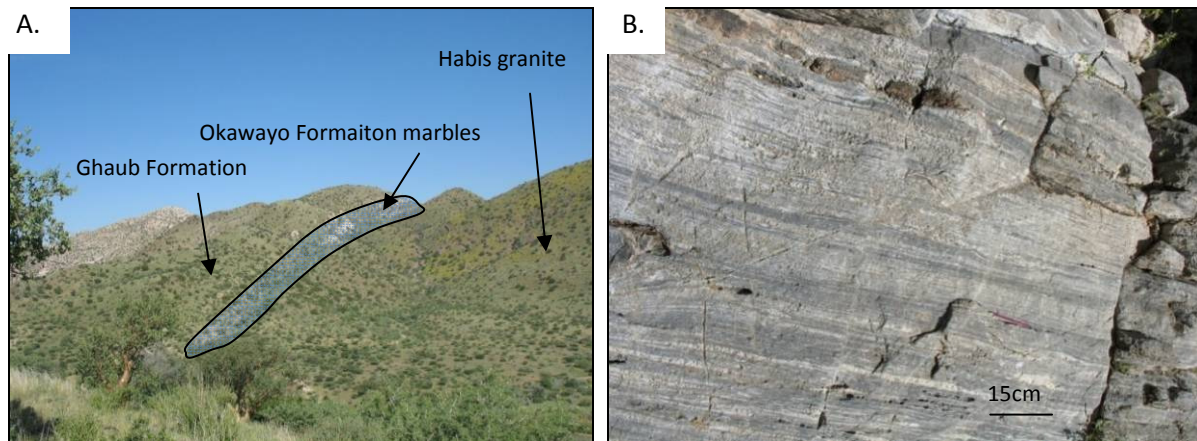
### 3.3.4 The Okawayo Formation

#### Description

The Okawayo Formation occurs as a marble unit directly above the Chuos Formation (Fig.3.9.A). This marble occurs as either a greyish banded marbles (Fig 3.9.B), or as white, homogenous, coarsely recrystallized and sugary-textured marble.

#### Distribution

The Okawayo Formation is only present in the Etusis domain where it is located stratigraphically above the Chuos Formation, along the southern margin of the AMC basement inlier. The Okawayo Formation reaches a maximum thickness of 85 m in the centre of the domain, but thins to a about 35 m further to the west. Marbles of the Okawayo Formation gradually lose their compositional banding and become more and more recrystallized and homogenous, before thinning and eventually pinching out completely in the eastern portion of the domain.



**Fig 3.9: A.)** The Okawayo Formation marbles occur as a thin strip between the Chuos Formation and the Ghaub Formation. This photo was taken from 22°05'33"S; 15°39'41"E towards the SWW and shows the Okawayo Formation at its' thickest. **B.)** Compositional banding in marbles of the Okawayo Formation. This photo was taken in plan view at 22°05'35"S; 15°49'41"E.

### Petrography

The marbles consist of medium-grained (up to 3mm in diameter), colourless calcite of equal grain size. Westwards, bands of creamish-white dolomitic and greyish-blue calcitic marbles can be distinguished (Fig 3.9.B).

### 3.3.5 The Ghaub Formation

#### Description

Rocks mapped originally by Smith (1966) as the Oberwasser Formation are now interpreted as part of the Ghaub Formation (Hoffmann et al., 2004; Miller, 2008) on the farm Etusis. The rocks consists of predominantly clast-bearing calc-silicate felses alternating with layers of marble- and clast-bearing biotite schists. This formation lies conformably below massive marbles of the Karibib Formation and above marbles of the Okawayo Formation. Rocks of the Ghaub Formation contain numerous and often massive (up to 70 cm in diameter) granite, marble and calc-silicate clasts (Fig 3.10).

#### Distribution

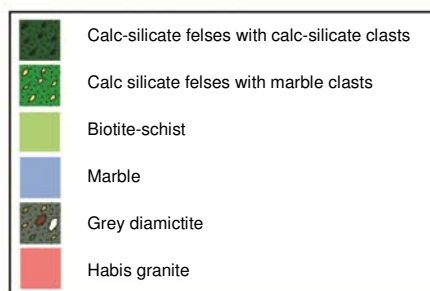
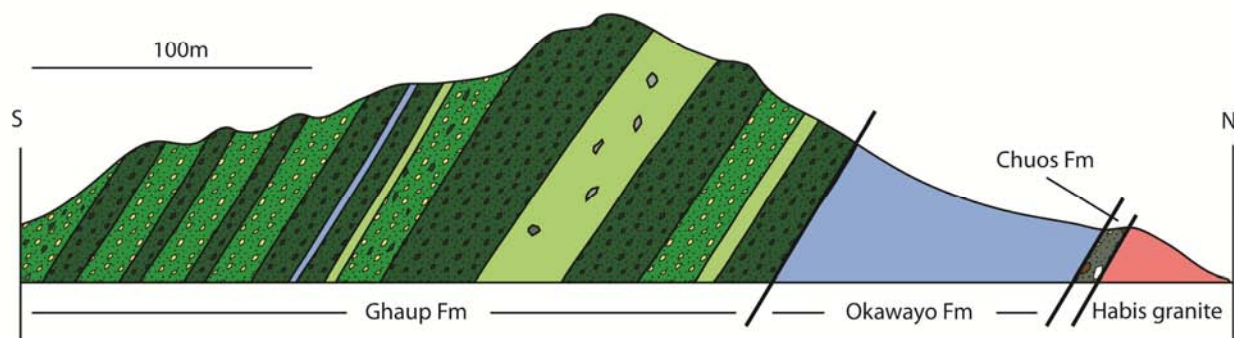
This formation is only present in the Etusis domain. As with the underlying formations, in this domain the Ghaub Formation also pinches out towards the E of the farm Etusis and attains a maximum thickness of about 510m on the farm Etusis 75 at 22°5'48"S, 15°48'41"E, emphasizing the overall wedge-shaped geometry of this package as a whole. The typically clast-bearing biotite schists and calc-silicate felses suggest a correlation with the Ghaub Formation (Hoffmann et al., 2004), rather than the schist-dominated Oberwasser Formation (Smith, 1966). Calc-silicate dominated diamictites of the Ghaub Formation also occur as a few isolated slivers (ca. 100 meters long) below sub-horizontal marbles of the Karibib Formation just north of the AMC in the Habib mountains (Fig 3.1, Appendix I).



## Petrography

This formation consists of alternating layers of biotite schist, marble and calc-silicate felses with clasts of mainly granite and marble. Green, calc-silicate dominated felses make up the bulk of this formation (80%) (Fig 3.10).

The dominant calc-silicates layers in this formation are fine-grained and seemingly homogenous, but on closer inspection, the rocks consist of an extremely fine-grained calc-silicate matrix (diopside, tremolite, calcite, plagioclase) containing large calc-silicate or marble clasts. A strong tectonic fabric is defined by the alignment and stretching of clasts and crystals aggregates. Smaller clasts of quartz or angular feldspar are also common. The matrix of the rocks is microcrystalline, often containing fine streaks of very fine quartz aggregates occurring as lenses along the mineral foliation. Biotite-quartz schists in this formation are extremely-fine grained (grains <<1mm in diameter).



**Fig.3.10: The Ghaub Formation**

Schematic section through the Ghaub Formation, exposed on the farm Etusis (22° 5'43.79"S; 15°49'8.73"E) – A succession of diamictic calc-silicate felses and schists, underlain by a massive recrystallized marble unit mapped as the Okawayo Fm. The Etusis Fm is not developed in this section. This figure looks westwards through the succession. Colour-coded photos illustrate typical rock types.

### 3.3.6 The Karibib Formation

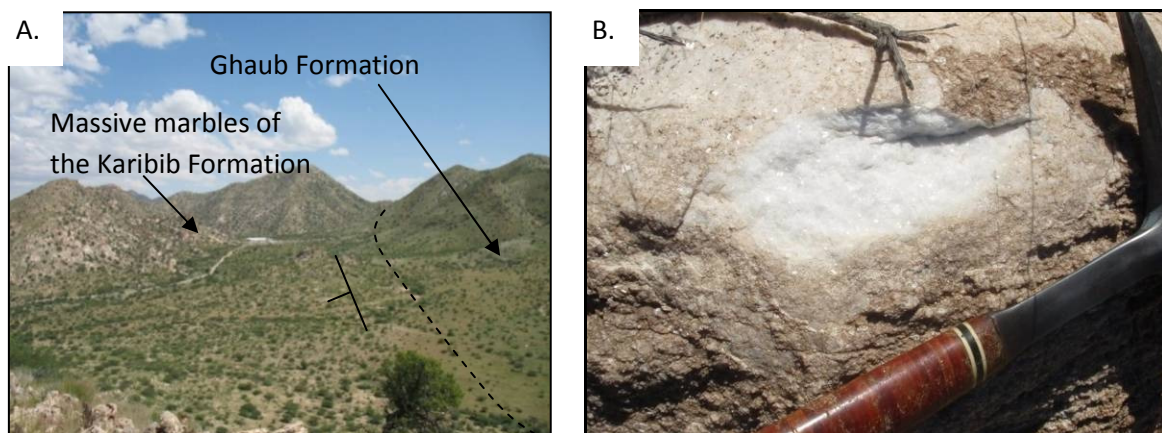
#### Description

The Karibib Formation is a marble-dominated unit and one of the most prominent formations occurring within this part of the SCZ. Karibib Formation marbles along the traverse occur predominantly as largely recrystallized dolomitic and, to a lesser extent, calcitic marbles. As such, marble units appear, for the most along the traverse, as sugary-white, medium- to coarse-grained units, with compositional banding only occurring sporadically. The marbles have an average thickness of about 600 m in the Etusis domain, decreasing to only a few tens of m thickness in the southern Audawib and Otjimbingwe domains.

#### Distribution

Marbles of the Karibib Formation occur in the Etusis, Audawib and Otjimbingwe domains. In the Etusis domain, Karibib Formation marbles can be found all around the AMC inlier where the carbonates reach a maximum thickness of about 1000m (Fig 3.11.A). Towards the N and E of the AMC, the Karibib Formation directly overlies either basement rocks or Habis granite. In this domain, the basal 200 to 400 m of the marble of the Karibib Formation is pervasively recrystallized (Fig 3.11.B,C), and primary compositional banding is only developed in some distance above the marble-granite and/or –basement contact. In the northernmost parts of the Etusis domain, the Karibib Formation marbles overlie the Mon Repos diorites.

Marbles of the Karibib Formation are widely distributed in the Audawib domain. The marbles occur between rocks of the Etusis Formation and biotite schists of the Kuiseb Formation. The marbles in this domain have a relatively consistent thickness of approximately 80 m, but may reach a thickness



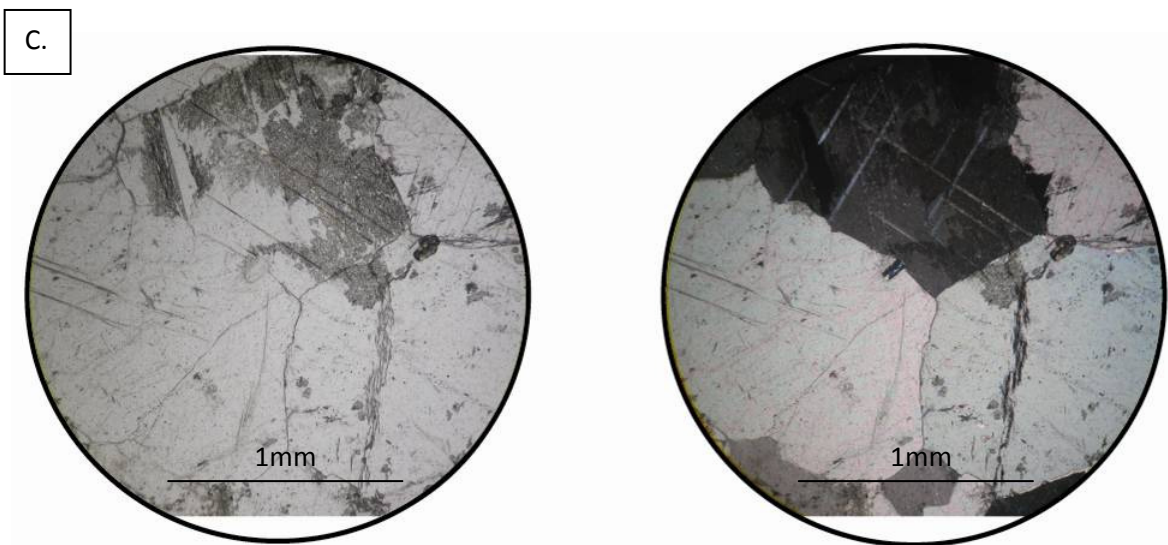
**Fig 3.11: A.)** The Karibib formation appear as large white marble hills and can be identified from afar in the Etusis/Habis domain, the photo was taken looking westwards from 22° 5'55.75"S; 15°49'37.88"E. Here Karibib marbles reach their maximum thickness of about 1000m . **B.)** On closer inspection Karibib Formation marbles have a white sugary texture.

of as much as 400 m in the central parts of this domain, possibly the result of tectonic thickening. For the most part, marble units in the Audawib mapping domain appear coarsely recrystallized and foliated, although primary banding is, locally, indicated by the alternation between calcitic and dolomitic marble bands.

In the SE parts of the Otjimbingwe domain, thick (>100m) marbles of the Karibib Formation are replaced by the lateral downbasin turbiditic facies equivalent the Tinkas Formation (Porada & Wittig, 1983).

#### Petrography

Recrystallized marbles of the Karibib Formation are relatively coarse-grained (up to 1cm large grains) and consist of mainly calcite and/or dolomite. In the field, the calcitic or dolomitic nature of the marbles can be tested by their reaction to diluted HCl. In thin section, carbonates show straight grain boundaries, very distinctive 120° triple junctions and mechanical twins are common (Fig 3.11.C). Most marbles show accessory amounts of quartz, muscovite and diopside. The colouration of dark-grey and black marbles seems to be the result of ultrafine graphite dispersed throughout the marble matrix.



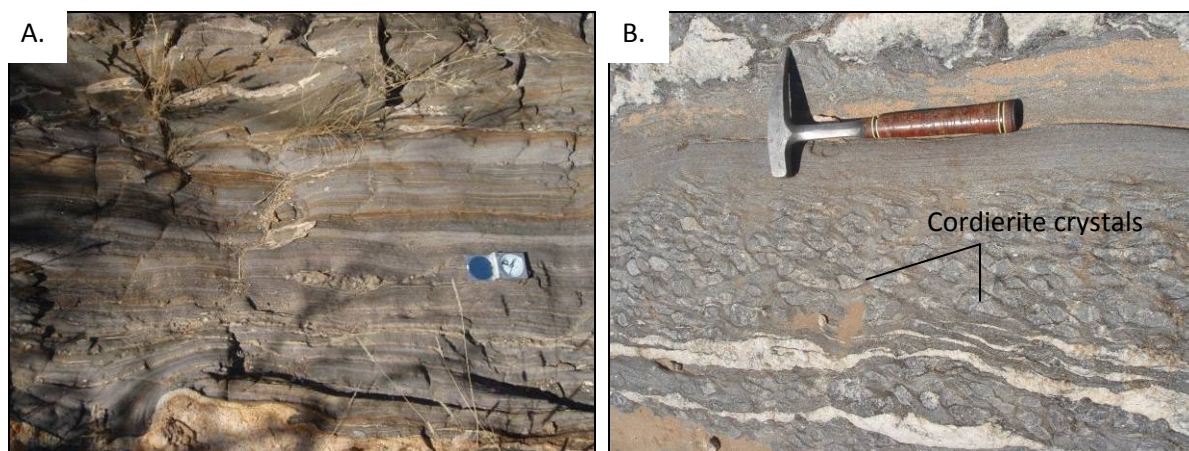
**C.)** In thin section, large (1-2mm) carbonates show clear 3 point junctions implying dynamic recrystallization and statically annealed textures.(Left image: plainlight; Right image: polarized light)



### 3.3.7 The Kuiseb Formation

#### Description

The Kuiseb Formation is a siliciclastic sequence, consisting entirely of interlayered banded quartz-biotite schist and psammites, commonly interpreted to indicate an original turbidite sequence (Kukla et al., 1990). Rock types include common porphyroblastic cordierite- and sillimanite-schist (Fig 3.12 & Fig 3.13). Since this is the uppermost formation of the DSG, a true stratigraphic thickness cannot be established. The biotite-schists commonly contain a strong, bedding-parallel foliation (mostly  $S_1$ , locally also  $S_2$ , see chapter 5) defined by the preferred orientation of biotite and the grain-shape preferred orientation of quartz and quartz-feldspar aggregates (Fig 3.14 & Fig 3.15).



**Fig 3.12: A.)** The typical banded appearance seen in schists of the Kuiseb Formation is the result of discrete compositional differences. **B.)** Though all rocks have a roughly schistose composition, a number of porphyroblast containing units are interspersed in the banded sequence. In this case, large (1-2cm) ovoid cordierite porphyroblasts can be clearly distinguished. Both photos were taken in the Audawib river in plan view.

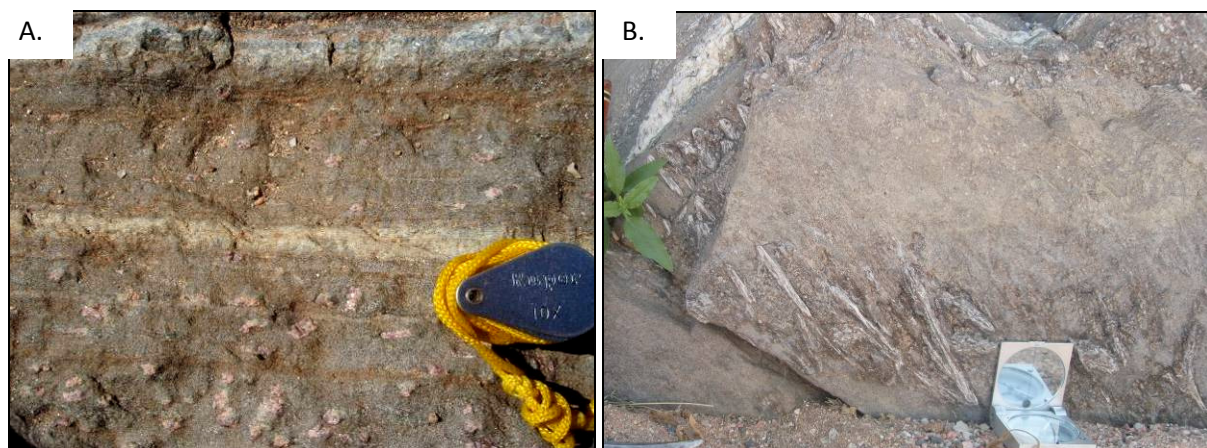
#### Distribution

This formation is well developed throughout the traverse, although thicknesses and abundance increase dramatically in the southern domains (Dorneb, Audawib and Otjimbingwe mapping domains). Kuiseb Formation schists are best exposed in the Otjimbingwe mapping domain, where the compositional banding is well preserved despite the strong structural overprint. At the south east boundary of the study area the Kuiseb Formation grades into a succession of schists, calc-silicate felses and marbles, mapped as the Tinkas Formation.

In the central part of the traverse, in the Dorneb and Audawib domains, the Kuiseb Formation is commonly intruded by fine-grained leucogranites as well as Salem-type granites (see 3.4 and chapter 4).

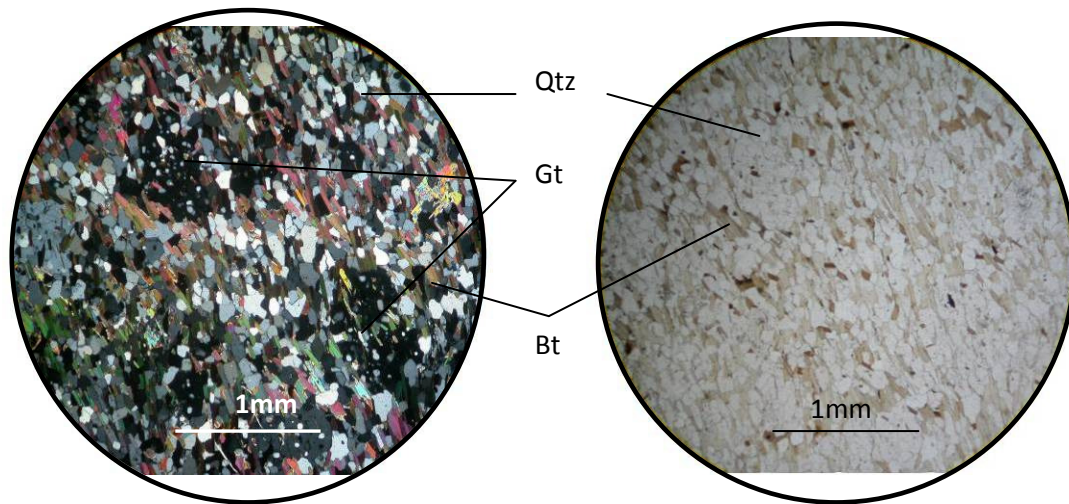
## Petrography

Schists of the Kuiseb Formation commonly consists of a thick succession of interlayered, cm- to dm-thick metapelites and metapsammites. Depending on the biotite and feldspar content, colours range from dark-brown and blackish to a light grey (Fig 3.12). In the metapelites are made up of biotite, quartz, plagioclase and K-feldspar, comprise up to 90% of the modal composition. In many places, cordierite is common, constituting up to 20% of the rocks. Cordierite porphyroblasts range in size from merely millimetres, then giving the rocks a spotted appearance, to up to 5cm in diameter. Sillimanite porphyroblasts are more common in schists in the Otjimbingwe mapping domain. Garnet is a constituent of the biotite-schist units only in the Otjimbingwe domain and close to and within the OLZ (Fig 3.13 & Fig 3.14). Individual sillimanite crystals may reach lengths of up to 20cm and both acicular sillimanite and stretched cordierite may define a lineation (Fig 3.13.B & Fig 3.15). Sillimanite and cordierite schists are characteristic for the Kuiseb Formation, but only constitute approximately 10% of all schist units of the metaturbidites, the bulk of which is made up of foliated, equigranular quartz-biotite schist (Fig 3.14). Grey metapsammites and dark greenish to reddish calc-silicate felses are present but subordinate (<5-10%).

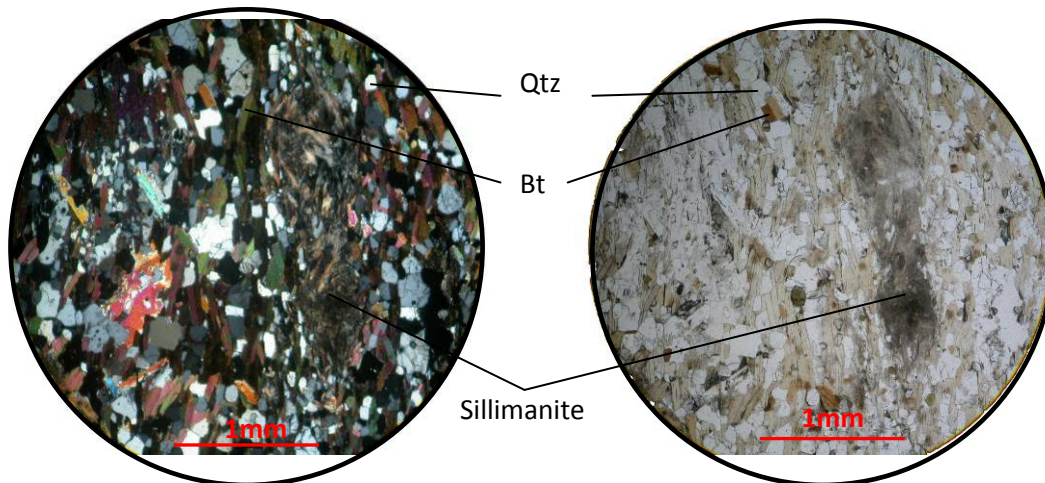


**Fig 3.13** – Metamorphic porphyroblasts are common in the Kuiseb Formation. **A.)** Garnet porphyroblasts are largely restricted to the southern Otjimbingwe domain. Here they occur as up to 5mm large, flattened garnets in transposed schists of the Kuiseb Formation. **B.)** Sillimanite porphyroblasts are also confined to the southern domains. Sillimanite grows preferentially on foliation planes and may reach lengths of up to 20cm, commonly defining a well-developed mineral lineation. Both photos were taken in the Audawib river, photo A is a plan view, whereas photo B is oblique, looking onto the foliation plane.





**Fig 3.14:** Thin section of psammitic schist of the Kuiseb Formation containing small poikiloblastic garnet. Brown/green biotite defines a clear foliation. Garnets as small as these are really only visible under cross-polarized light (left). Clear minerals are quartz, K-feldspar and plagioclase, defining a grain-shape preferred orientation parallel to the biotite foliation



**Fig 3.15:** Thin section of a sillimanite schist of the Kuiseb Formation. The proportion of sillimanite increases within ca. 1km from the Donkerhuk granite pluton. Sillimanite is large and obvious in hand samples as well as being clearly visible in thin-section as fibrous grey-colourless highly birefringent minerals. Sillimanites are orientated with the long-axis parallel to a  $S_2$  foliation defined most clearly by biotite crystals.

### 3.3.8 The Tinkas Formation

#### Description

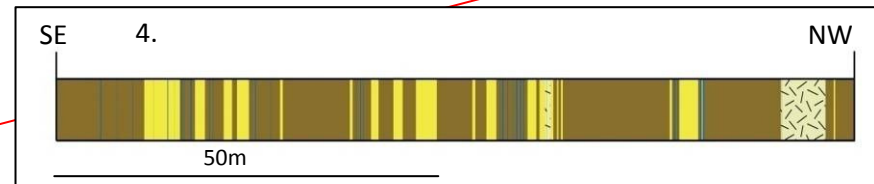
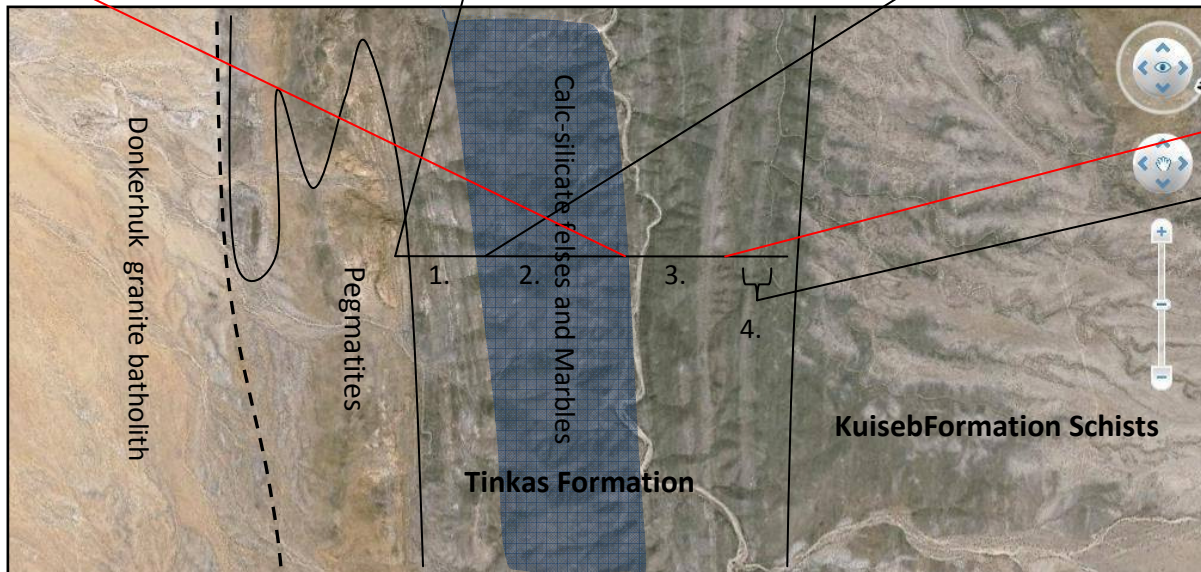
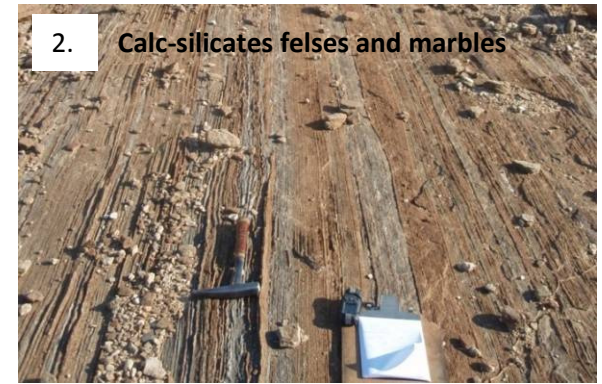
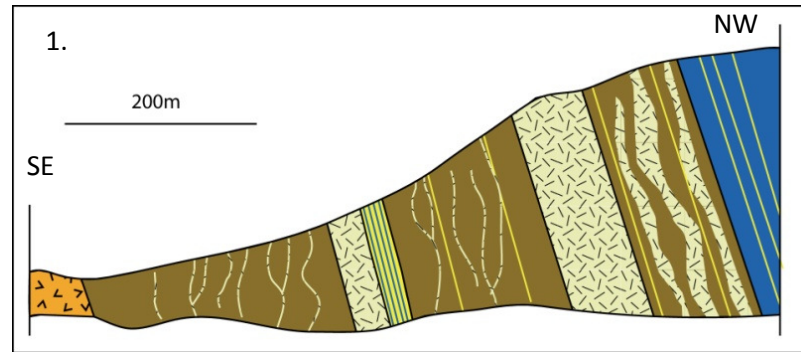
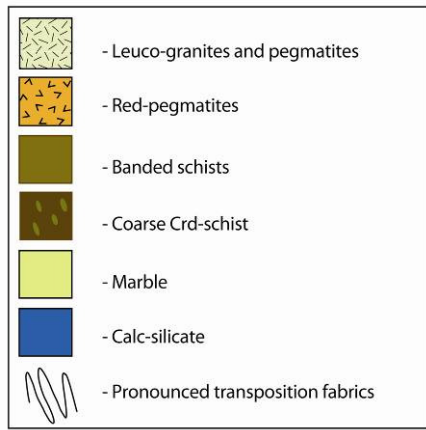
This formation consists of a variety of biotite schists, isolated marble bands, thickly banded marble units and calc-silicate layers (up to 50 cm thick) (Fig 3.16). The Tinkas Formation is commonly regarded as a lateral facies equivalent of the Karibib Formation (Porada & Wittig, 1983; Miller 1983; De Kock, 2001). Porada & Wittig (1983) estimate this formation to be 4-5 km thick, intense folding of this formation along the OLZ, however makes a true thickness very hard to determine. The Tinkas Formation shows a gradational contact with schist of the Kuiseb Formation. The banded sequence is intruded by granite and pegmatite sills in the OLZ.

## **Distribution**

The Tinkas Formation only occurs at the SE margin of the study area in the Otjimbingwe Townlands. The contact between the Kuiseb and Tinkas Formations is not a sharp one and was taken as the first occurrence of marble bands and layers of calc-silicate felses that mark the lower contact of the Tinkas Formation within the otherwise biotite- and biotite-quartz schist dominated domain. The central parts of the Tinkas Formation show a ca. 1km thick horizon consisting of interlayered calc-silicate felses, marble bands and minor biotite schist. This unit is cross cut by leucogranite and pegmatite dykes, forming apophyses from the large Donkerhuk granite to the south. This rock unit is similar in composition to ones that De Kock (2001) referred to as the Lievental member of the Tinkas Formation, though it has not been mapped as such. An up to 90 m thick (apparent) marble unit within the Tinkas Formation might be what De Kock (2001) referred to as the Karibib Formation marble unit within this area (Fig 3.16).

Although De Kock (2001) developed a unique stratigraphic classification for this part of the DSG, in keeping with Miller's (2008) stratigraphic classification, all the rocks that consists of banded calc-silicates felses, schist and marble are here classified as the Tinkas Formation. Tight upright folding of lithologies within this formation make it difficult to determine the succession of rock units.

# The Tinkas Formation



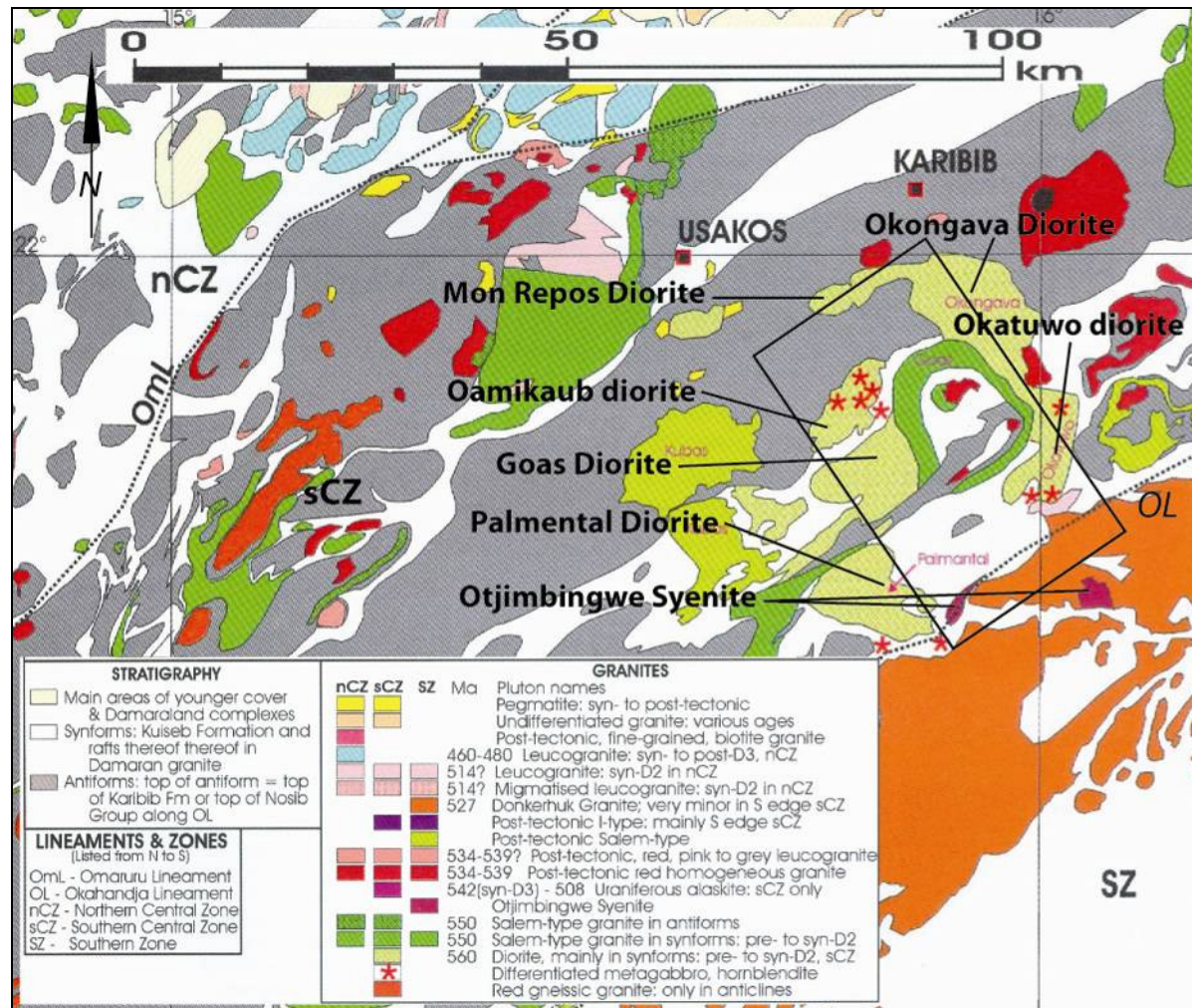
**Fig.3.16: Tinkas Formation**

This formation consists predominantly of biotite schists, with numerous layers of marble and calc silicate felses. Layering persists on a macroscopic, mesoscopic as well a microscopic scale. Towards the margin of the Donkerhuk batholith, this formation is intruded by pegmatites and granites, making the identification of marker layers difficult. The central section of the formation consists of calc-silicate felses and marbles.



### 3.4 Granites and pegmatites

Although most of the granites in the field area fall into the subdivisions of Miller (1983), there is enough variation in intrusive rocks south of Karibib to justify a further sub-division into eight main granite types (Gevers, 1963; Smith, 1965; Jacob, 1974; De Kock, 1989; Miller, 1983, 2008; Lehtonen et al., 1995).



**Fig 3.17:** Simplified geologic map showing the relative distribution of different granite types as well as metasedimentary rocks across the SCZ. The outlined square in this image represents the study area. Large plutons in this area have been annotated. Smaller granite intrusion such as the Habis granite are not shown in this image. Modified after Miller, 2008.

These are:

- 3.4.1 - Salem granite suite (e.g. Jacob, 1974; Miller, 1983)
- 3.4.2 - Goas diorite suite (e.g. Lehtonen et al, 1995; Jacob et al., 2000)
- 3.4.3 - Donkerhuk granite batholith (e.g. Gevers, 1963; Smith, 1965)
- 3.4.4 - Habis granite suite (e.g. Gevers, 1963; Brandt, 1985)
- 3.4.5 - Otjimbingwe syenite (e.g. Jung et al., 2004)
- 3.4.6 - Red granites (e.g. Smith, 1965)

3.4.7 - Fine-grained leucogranites (e.g. Blaine, 1977)

3.4.8 - Pegmatites (e.g. Roering, 1966)

The following section focuses on the distribution and composition of the granites in the study area. Structural features such as mineral fabrics and contact relationships for individual granites are presented in chapter 4.

### **3.4.1 Salem-type granites**

This group of characteristic granites (named after the farm Salem on which it was first identified) shows a widespread distribution throughout the CZ. Salem-type granites consist of a variety of characteristic biotite-rich granites and diorites. These spatially closely associated granitoids are thought to be early to syn-tectonic (D2) and it has long been recognized that they occur almost exclusively in synforms of schists of the Kuiseb Formation (Gevers, 1931; Gevers, 1963; Smith, 1965 and Miller, 2008).

This suite consists of a number of mineralogically and texturally distinct, spatially closely associated granitoid phases (e.g. Jacob, 1974). The two most widespread and common phases include a medium- to coarse-grained, biotite- and/or hornblende-rich, K-feldspar megacryst-bearing granodioritic phase and a much more leucocratic, medium- grained, commonly garnetiferous leucogranite that may or may not contain K-feldspar megacrysts.

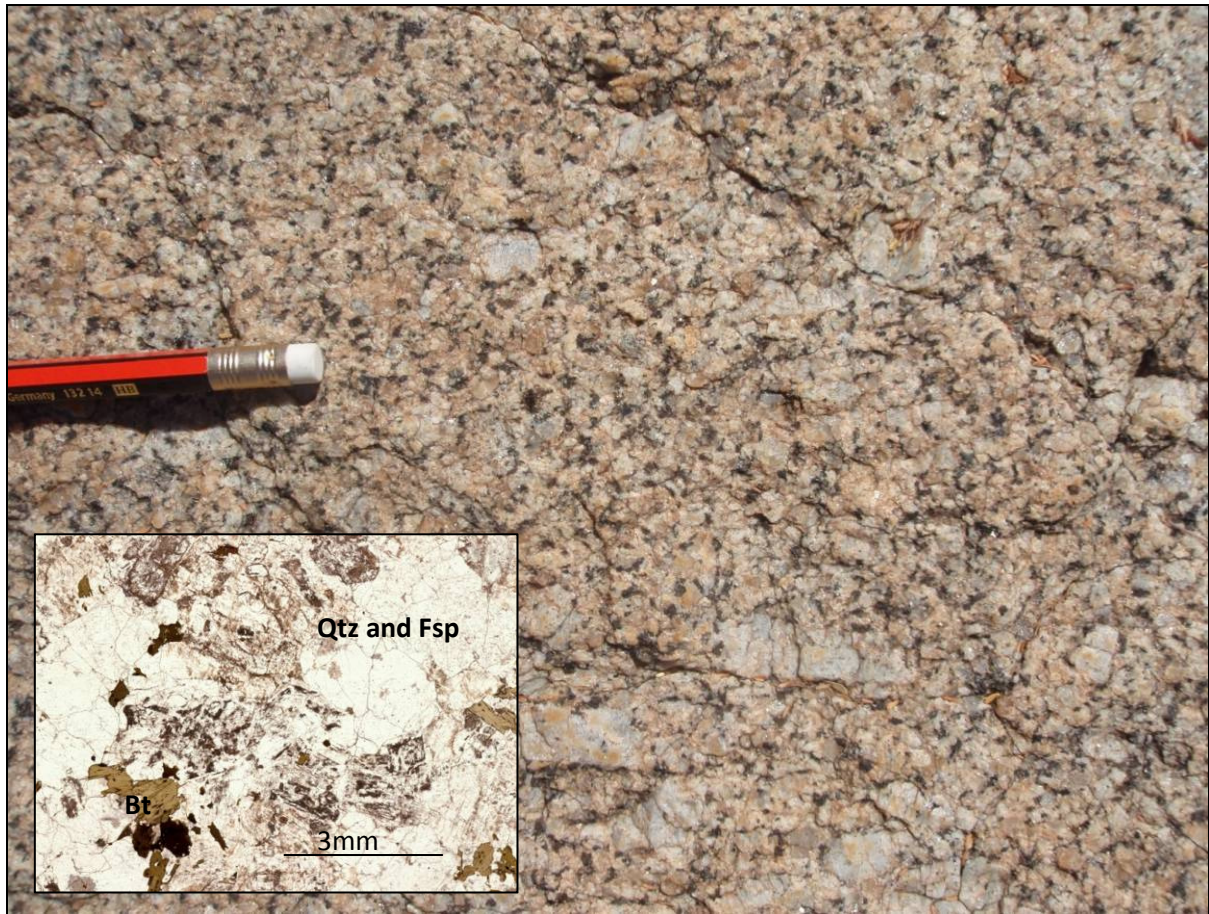
Recent U-Pb zircon ages from the Stinkbank granite SW of Usakos suggests intrusion of the Salem granites around Karibib to have occurred at  $549\pm 11$ Ma (Johnson et al., 2006). Previous ages for the Salem granite were not very robust, showing errors of as much as 20-60Ma, yet these dates suggests similar emplacement ages for the granites at around  $\pm 570-530$  (Kröner, 1982; Hawkesworth et al., 1983; Haack & Martin, 1983; Johnson et al., 2006; Miller, 2008)

In the study area, Salem-type granites are exposed in laterally extensive granite platforms in the central portions of the traverse, making up most of the Dorneb and large parts of the Audawib domain. These granites cover an area of more than 110 km<sup>2</sup>.

The leucogranite variety (Fig 3.18) consists, on average, of K-feldspar and plagioclase (45-65%), quartz (25-45%), biotite and hornblende (<15%), muscovite (<5%), and garnet (<2%) with accessory zircon and monazite. Chlorite, sericite and epidote are secondary minerals. K-feldspar phenocrysts can reach up to 3 cm in length, but are far less abundant compared to the biotite-rich Salem variety.



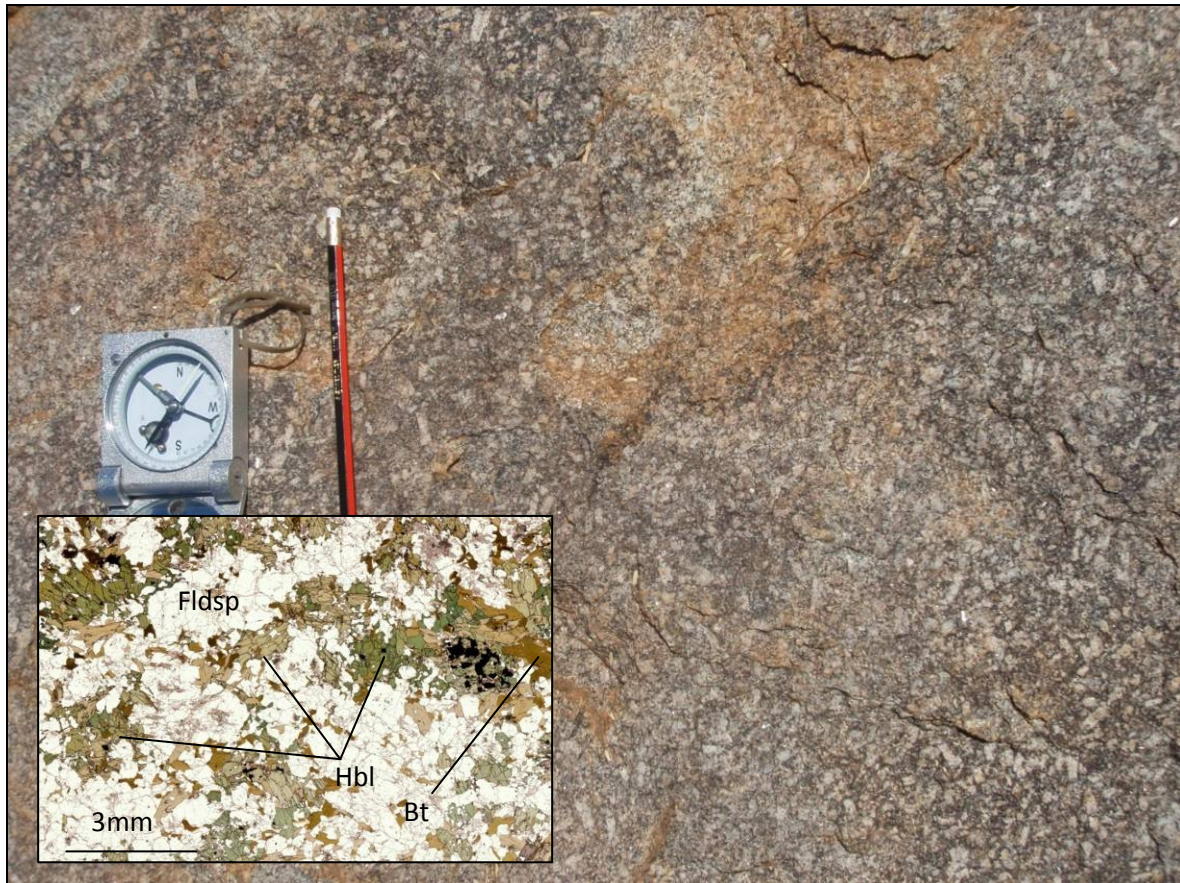
The leucogranite phase is most common in the NW parts of the large Salem-type granite in the Dorneb domain (Fig 4.3.2 in the following chapter).



**Fig 3.18:** “Leuco”-Salem-type granite has a light pink or white colour, with sporadic K-feldspars phenocrysts. The rock consists of primarily quartz and feldspar, with only minor amounts of biotite and even less hornblende. This photo was taken at 22°8'4"S; 15°50'0"E in the Gamikaub river.

The dark-coloured megacrystic Salem-type granite (Fig 3.19) has an equigranular med- to fine-grained (minerals are ca. 0.2-3mm in diameter) matrix containing between 35 and 55% hornblende and biotite. In many places hornblende is far more abundant than biotite. The granite typically contains less than 20% quartz. The granite is porphyritic containing large (between 0.5 and 4cm in length) euhedral K-feldspars aligned to define a foliation in the matrix. In many parts of the Audawib fold complex these rocks can be observed to have intruded as sheets of slightly different compositions (expressed as discrete colour and phenocryst size differences).





**Fig 3.19:** Typical biotite-hornblende megacrystic Salem-type granite. This variety contains a larger modal amount of mafic minerals compared to the leucogranite variety, giving it a much darker colour. The most notable feature is the abundance of euhedral K-feldspar phenocrysts. These are often aligned to define a prominent magmatic foliation in the rock. A mineral foliation can be hard to identify in thin section. This photo was taken on the farm Audawib at 22° 9'8.97"S; 15°58'8.75"E.

### 3.4.2 The Goas diorite suite

The Goas diorite suite was described and subdivided by Lehtonen et al. (1995) into 6 intrusive bodies consisting of metagabbros and diorites in the region south of Karibib. The individual diorite plutons are the Okatuwo, Mon Repos, Oamibkaub, Gamikaub, Okongava and Plamental diorites (Fig 3.17). Intrusive plutons of the dioritic suite defines a clear circular structure on aeromagnetic maps, and Miller (2008) sees this as a implication for a genetic and intrusive relationships between these diorites (Lehtonen et al. 1995; Miller, 2008).

S1 foliations observed by Mcdermott (1986) and De Kock (1989) within schist xenoliths in the diorites have led them to believe that the Goas diorites were intruded at an early stage during deformation (prior to Miller's (2008) D2). Recent studies indicate that the Goas diorite suite, at least in the Mon Repos area, form sheet like bodies (Ameglio et al., 2000).

U-Pb zircon ages for rocks of the Mon Repos, and Okangava diorites indicate an emplacement age at  $564 \pm 5$  and  $558 \pm 5$  Ma, and these ages give a indication of the maximum age of D2 deformation, since the diorites were intruded post-D1 and pre-D2 (De Kock et al., 2000; Jacob et al., 2000).

The following plutons of the larger Goas diorite suite occur within the field area:

- Mon Repos diorite (Etusis)
- Okongava diorite (Dorneb)
- Oamikaub diorite (Etusis)
- Gamikaub diorite (Dorneb)
- Palmental diorite (Otjimbingwe).

The Mon Repos diorite is exposed along the northern margin of the Etusis domain where it forms a roughly oval body of about 19 km<sup>2</sup>. The Oamikaub diorite intrudes the Damara Supergroup along the SW margin of the Etusis domain. It is a ca. 38 km<sup>2</sup> intrusion found between marbles of the Karibib Formation and schist of the Kuiseb Formation. The NE parts of both the Etusis and Dorneb domains are underlain by parts of the extensive ca. 90 km<sup>2</sup> Okangava diorite. The SW margin of the Dorneb domain is intruded by a narrow tongue of the Gamikaub diorite situated between schists of the Kuiseb Formation and Salem-type granites in the W. A part of the Palmental diorite occurs only in the SW corner of the Otjimbingwe domain.

These plutons have similar compositions, although e.g. Ameglio et al. (2000) and Jacob et al. (2000) point out compositional variations within individual plutons. For example, the Mon Repos diorite is commonly associated with a lighter, granodioritic phase. Most notably, the Oamikaub diorite has an associated ring-dyke complex of coarse meta-gabbro (the Neikoes gabbro), consisting almost exclusively of hornblendite (Lehtonen et al., 1995; Miller, 2008). This ring complex is situated in the centre of the Oamikaub intrusion.

Diorites of this suite are medium- to fine-grained (0.5 - 2mm grain size) and consist, for the most part, of hornblende, plagioclase, biotite, K-feldspar and quartz (Fig 3.20). Most of the dioritic bodies in the SCZ are foliated. This is the case for e.g. the Mon Repos and Palmental diorites that contain strong solid-state fabrics, particularly along their margins. The exception to this is the Oamikaub diorite that only very occasionally contains a weak magmatic fabric. The fabric development and relative timing of emplacement of the diorites will be presented and discussed in the following structural chapter (chapter 4).





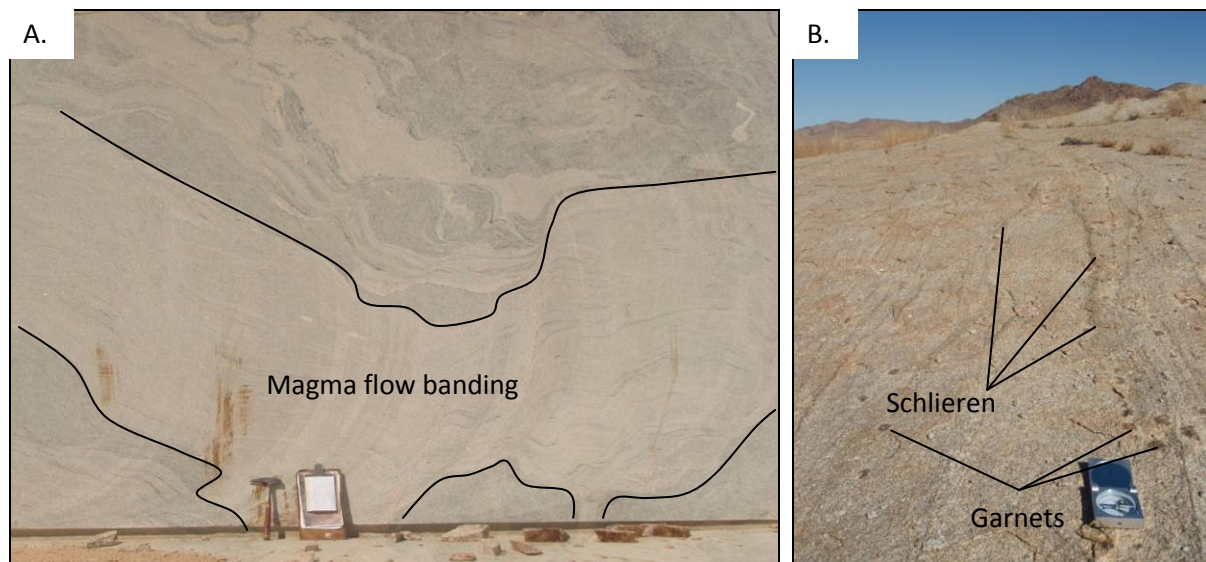
**Fig 3.20:** Compositionally and texturally heterogeneous diorite of the Mon-Repos diorite suite (plan view). The diorites are fine- and/or medium-grained and consist of predominantly black hornblende (greenish-brown in thin-section) (35%), plagioclase and minor K-feldspar (55%) as well as quartz (10%). This photo was taken on the farm Mon Repos at 22° 1'52.00"S; 15°48'43.92"E and shows a heterogeneous foliation particularly well-developed in the central section of the photograph.

### 3.4.3 The Donkerhuk batholith

The massive Donkerhuk batholith is the largest single granite intrusion in Namibia and occurs at the northern margin of the SZ along the OLZ and is exposed for about 330 km along strike, being up to 45 km wide. It was first described and named by Gevers (1963). This is a medium-grained, light coloured, two-mica, syeno- to monzo-granite. This granite has been suggested to be a late- to post-tectonic granite and is believed to have an extensive structural and metamorphic aureole (Barnes & Sawyer, 1980; Sawyer, 1983; De Kock, 1989). This large composite granite is associated with the intrusion of numerous pegmatites and leucocratic-granite apophyses into schists along its contacts (Jacob, 1974, Sawyer, 1983).

Rb-Sr whole rock ages for the Donkerhuk pluton are between  $532 \pm 8$  Ma and  $527 \pm 3$  Ma. (Blaxland et al., 1979; Haack & Gohn, 1988; Miller, 2008). Miller (2008) considers an age of  $505 \pm 4$  Ma obtained by Kukla (1993) too young for the intrusion, although the Donkerhuk granite is undoubtedly made up of a number of discrete phases.

The NW margin of the batholith forms the SE boundary of the Otjimbingwe domain. On a regional scale, the composition of the Donkerhuk granite is relatively homogeneous and the batholith is typically developed as a pinkish to white leucogranite. This granite is commonly a medium-grained (1-3mm) and a two mica-granite (biotite and muscovite), with quartz, K-feldspar, and plagioclase as the main constituents. Garnet is common (Fig 3.21) whereas, zircon, monazite, apatite and oxides are accessory. Although this rock is relatively homogenous, magmatic layering and/or sheeting is common. The complex and composite internal structure and compositional heterogeneity of the Donkerhuk granite is best observed in abandoned dimension stone quarries (e.g. 22°21'15.61"S; 15°57'9.85"E, where internal flow banding and schlieren are clearly visible (Fig 3.21) or in the very large, whaleback-type granite outcrops next to the C32 (road) on the way to Otjimbingwe. Here, the preferred alignment of cm-scale K-feldspar phenocrysts defines a pronounced and pervasive, steep NW dipping magmatic foliation parallel to the granite country-rock contact in the NW. The granite has a lit-par-lit intrusive contact with the schist units of the Kuiseb Formation, and this has led to a minor degree of partial melting at the contact zone. Leucocratic pegmatite and granites sills of similar composition that occur within Kuiseb Formation as much as 4 km NE of the contact seems to be a continuation of the lit-par-lit injection of granites in the marginal zone of the batholith (Gevers, 1931; Sawyer, 1983; De Kock, 1989; Miller, 1983 2008).



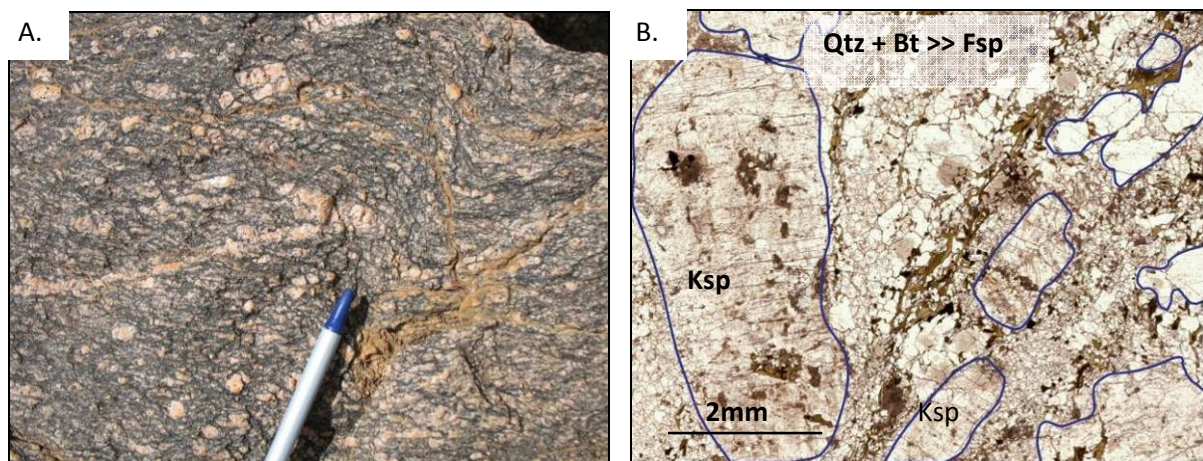
**Fig 3.21:** The Donkerhuk granite. **A.)** Subtle colour and compositional changes may indicate flow banding. Flow banding is clearly exposed within a abandoned quarry in the middle of the pluton. This illustrates the internal heterogeneity of the granite. **B.)** Localised melanosomes with peritectic garnet occur close to the contact with the Kuiseb Formation schist and could be a indication of localised melting along this contact. Photo **A** was taken at 22°21'15.61"S; 15°57'9.85"E looking SE. Photo **B** was taken at 22°20'33"S; 15°56'17"E looking SW. The contact with the Kuiseb Formation lies about 50m directly to the right (NW) of this photo.



### 3.4.4 The Habis granite

This strongly foliated porphyry granite occurs exclusively at its type locality in the central part of the Etusis domain. This pluton covers an area of approximately 16 km<sup>2</sup> and is well exposed in a number of prominent hills at 22° 4'17.15"S; 15°49'59.30"E next to the C23 (road). The Habis granite contains large (1-5 cm) K-feldspar phenocrysts in a fine-grained matrix of predominantly quartz with some biotite and K-feldspar and plagioclase. Gevers (1931) described it as, "a coarse-crystalline biotite granite with K-feldspar phenocrysts". This granite has a very strong magmatic fabric, defined by the preferred orientation of K-feldspar megacrysts. The magmatic fabric is overprinted by a pervasive solid-state fabric, defined by quartz ribbons, marginally recrystallized and stretched, ovoid feldspar augen and the preferred orientation of recrystallized biotite (discussed in chapter 5) (Fig 3.22). The exact mineral proportions in the matrix can vary, but biotite makes up between 10% and 25% of matrix minerals, the rest being formed by quartz, K-feldspar, plagioclase, minor muscovite and accessory zircon. Chlorite, epidote and sericite are secondary minerals.

This porphyritic granite has been suggested by Brandt (1985) to be part of the AMC. In the Etusis domain it occurs adjacent to AMC rocks to the immediate W. A gradational contact exists between rocks of the AMC and the Habis granite, muscovite schists of the AMC schist grading into more quartzo-feldspathic rocks in the E, accompanied by the gradual appearance of K-feldspar phenocrysts. This results in eventually igneous textures characteristic for the Habis granite. The contact between the DSG and the Habis granite is sharp and is discussed in more detail in the following chapter (chapter 4).



**Fig 3.22: A.)** Typical Habis granite consists of orange/pink ovoid and euhedral feldspar phenocrysts in a fine-grained, dark-coloured biotite-quartz-feldspar matrix. This photo was taken on the farm Etusis at 22°03'49"S; 15°50'24"E (plan view). **B.)** Thin section (plain polarised light) of the Habis granite, showing large K-feldspar megacrysts a finer-grained, partly recrystallized and foliated matrix of equi-granular quartz, feldspar (larger, transparent and cloudy) ,and biotite (green/brown).

### 3.4.5 Otjimbingwe syenite

The Otjimbingwe syenite was intruded as a number of small plutons within and outside the OLZ near the town of Otjimbingwe. One of these intrusive plugs (6km long, 1km wide) is found within the study area in the OLZ adjacent to the Donkerhuk granite. On aerial photos, this large syenite body stands out as a prominent red-coloured hill in the Otjimbingwe domain. Rocks of the Otjimbingwe syenite have been described as a “moderately foliated, dark-grey, porphyritic hornblende and pyroxene-bearing massive syenite” (Jung et al., 2004). The Otjimbingwe syenite is thought to post-date D1 (Jung, et al., 2004; Miller, 2008). Jung & Hellebrand (1998) obtained a U-Pb titanite age for the Otjimbingwe syenite of  $520\pm 9$  Ma, this is much more accurate than a earlier whole rock Rb-Sr age of  $570\pm 41$  Ma, determined by Haack et al. (1988).

The reddish-orange syenite is texturally and compositionally relatively homogenous and consists of almost exclusively black hornblende and red K-feldspar. Quartz is locally present in small amounts so that the rocks is, strictly speaking, a quartz syenite (Fig 4.6). The medium-grained (0.5-3mm) matrix of the rock contains rodded, up to 6-8cm long K-feldspar phenocrysts, resulting in a very pronounced linear fabric in the rock (Fig 3.23). These minerals also define a somewhat less-prominent planar foliation.



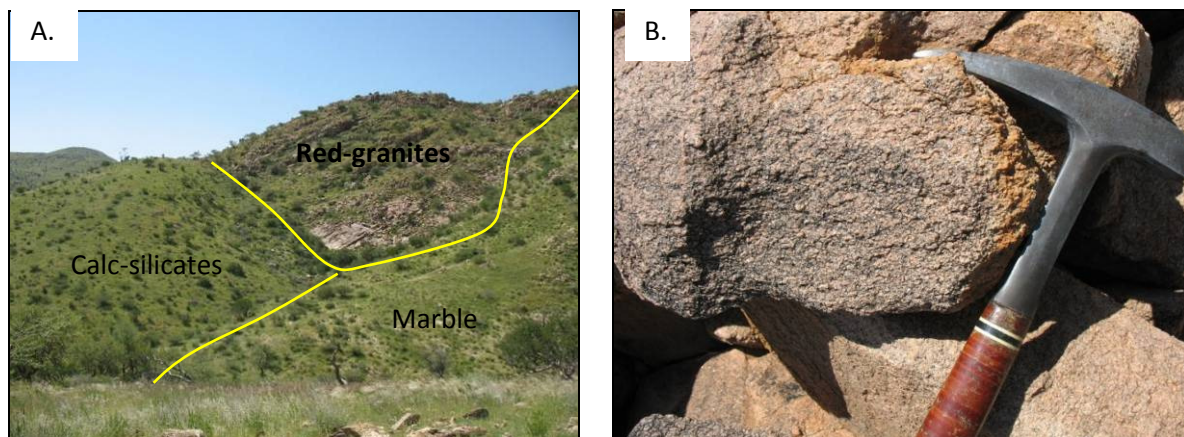
**Fig 3.23:** The Otjimbingwe syenite occurs in the field as a texturally and compositionally distinct reddish-orange and streaky rock. The streaky appearance is the result of aligned and stretched K-feldspar phenocrysts and black hornblende in the matrix, resulting in the rodded textures typical for the Otjimbingwe syenite. Rodding of minerals form a very distinct mineral lineation. Photo was taken in the Audawib river at  $22^{\circ}22'26''S$ ;  $15^{\circ}53'17''E$  and shows typical appearance of the Otjimbingwe syenite. Outcrop here is not in-situ and the thin section is rotated at right angles to the photo



### 3.4.6 Red-granites

These granites are relatively uncommon in the study area. The Red granite rocks are often found in regional anticlinal structures and typically intruded into stratigraphic levels below the Karibib Formation marble. The rocks are thought to have formed due to the partial melting of AMC gneisses and arkoses of the Etusis Formation (Smith, 1965; Hoffmann, 1976; Miller, 2008). Ghost structures that resemble augen or compositional banding may, indeed be inherited from the protolith.

The only significant exposures of these granites are in the Etusis domain. The granites are commonly associated with pegmatite intrusions. The most notable exposure is a wedge of red-granite some 200 m in diameter at 22° 5'37.45"S; 15°48'9.76"E (Fig 3.24.A) occurring between rocks of the AMC and rocks of the Nosib and Swakop Groups. The orange to pink-coloured granites are equigranular, medium- to fine-grained (0.2-2mm), containing quartz, K-feldspar, muscovite and biotite as the main rock-forming minerals (Fig 3.24.B).



**Fig 3.24:** A.) One of the largest exposures of Red granites is in the study area, in between units of the Swakop group. This photo was taken at 22°05'47"S; 15°48'04"E looking NE. This intrusion is adjacent to a large NW-SE trending fault on the opposite side of the hill B.) Red-granites are often found close to the AMC as homogeneous, equigranular white or pinkish, feldspathic granites. Pegmatites are commonly associated with Red Granites

### 3.4.7 Fine-grained leucogranites

Fine-grained leucogranites (Fig 3.25.A) are commonly found in schists of the Kuiseb Formation in the Dorneb and Audawib mapping domains and might share some kind of relationship to the red granites described by Miller (1983). Smith (1966) for example mapped many exposures of leucogranites as red granites. These granites similar to Salem-type granites intrude preferentially into schists of the Kuiseb Formation. Leucogranites, however, occur at higher structural levels in the Kuiseb Formations and are finer, more homogenous and poorer in mafic phases than Salem-type granite varieties. Many areas on the farm Audawib that have been mapped as schist in regional maps (e.g. Smith, 1966), consist entirely of these leucogranites, with schist only occasionally appearing as xenoliths within the larger intrusive bodies. Schists of the Kuiseb Formation within the



centre of the Audawib domain have been almost completely swamped by leucogranites (e.g. around 22°11'42.27"S; 15°56'55.86"E). The contacts with the Kuiseb Formation schists are commonly developed as very thin (10-500 cm wide) lit-par-lit type contacts (Fig 3.25.B).

These rocks are typically fine-grained, friable on surface and leucocratic. The main constituents of these granites are K-feldspar, plagioclase and quartz and minor muscovite and biotite (<5%). These intrusive bodies show no structural fabrics.



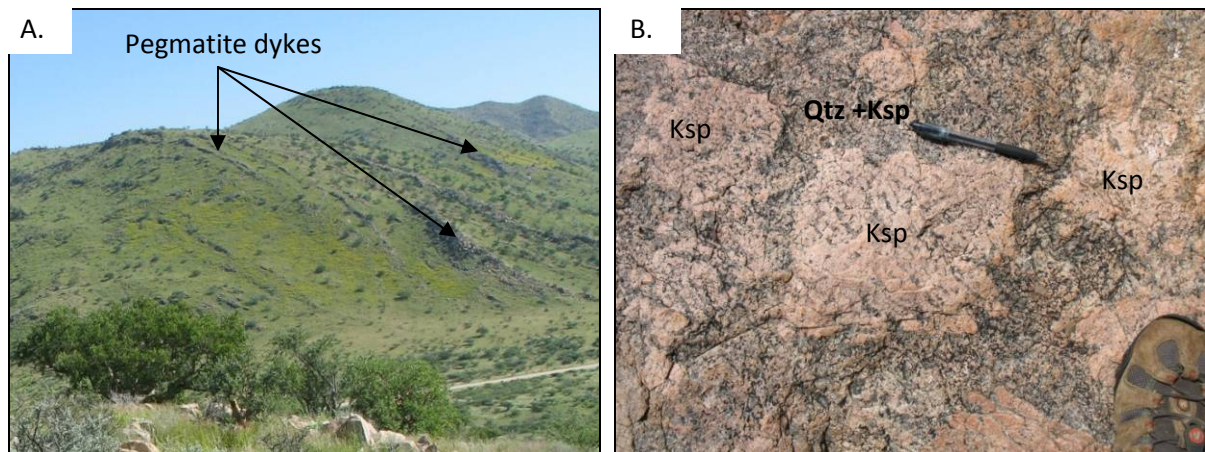
**Fig 3.25:** **A.)** Leucogranites have a weathered white/orange appearance and consists of equi-granular fine quartz and feldspar crystals. **B.)** These granites have been intruded into schist and this intrusive relationship is often seen on a fine scale where only small schist slivers are preserved in white granites. These photos were taken in the Audawib mapping domain at, 22°8'42"S; 15°56'2"E.

### 3.4.8 Pegmatites

Abundant late stage pegmatites intrude throughout the study area, and these are thought to have been emplaced during the end of phases of granite intrusion during embrittlement of the crust (Roering, 1966; Miller, 2008). Although some pegmatites are and have been mined for tin, lithium and gem quality crystals (e.g. Tourmaline and beryl), the vast majority of these are of little economic value (Steven, 1993, Miller, 2008).

The late- to post-tectonic pegmatites cross cut all rock types. Although these pegmatites are found in all lithologies, they are commonly found in gneisses of the AMC and are scarcely seen cross-cutting marbles of the Karibib Formation. Pegmatites typically occur as 1-5m wide dykes reaching strike length of 10 to 200 m. They commonly occur as NW-SE trending dyke swarms that may amalgamate to form composite pegmatite plugs and stockworks. Such composite pegmatites may form hills and ridges (Fig 3.26.A). particularly in the NE of the Dorneb mapping domain and along the contact between the Donkerhuk granite and the Kuiseb Formation in the Otjimbingwe domain.

Pegmatites consist almost exclusively of large (2-15 cm) K-feldspar. Quartz occurs as an interstitial phase between intergrown K-feldspar crystals (Fig 3.26.B). Large biotite or tourmaline crystals are commonly also found inter-grown between feldspar. Phaneritic proportions of K-feldspar and quartz are common.



**Fig 3.26:.A.)** Pegmatites typically occur as long, positively weathering dykes, making them easy to identify from afar. This photo was taken at 22°05'26"S; 15°48'12"E looking North. Here pegmatites have intruded in schist of the AMC. **B.)** Pegmatites are typically orange or pink due to the colour of the large K-feldspar crystals which make up most of pegmatites. K-feldspar crystals show graphitic intergrowths with quartz.

## Chapter 4: Structural Geology

### 4.1.1 Introduction

The CZ of the Damara belt exposes polyphase deformed mid-crustal rocks, exhibiting pervasively developed ductile fabrics and, in many places, high-strain fabrics. The variable rheology of rock types in the SCZ, as well as the intrusion of numerous syn- to late-tectonic granite phases has had a profound influence on the structural evolution and architecture of the SCZ (e.g. Kröner, 1984; Coward, 1983; Kasch, 1983a)

This study takes the structural subdivision of Poli and Oliver (2001) and Kisters et al. (2004) and considers deformation as a progressive event in which a variety of structures were developed during, in particular, the D2 deformation. The structural evolution is, thus, not sub-divided into a multitude of discrete deformation increments, but rather two main progressive deformation events have been identified, namely D1 and D2 (Table 4.1).

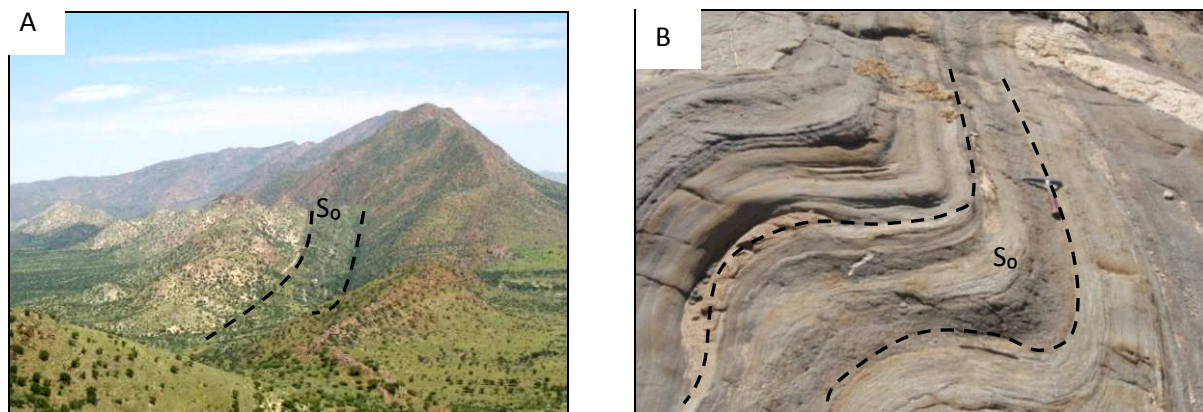
**Table 4.1:** Structural notations of different types of structural elements in the field area, the deformation phases they are related to and their presence in the different domains.

Deformation stage	Fabric manifestation	Fabric Description	Domain
-	S0	Bedding	All domains
D1	S1	Bedding-parallel mineral foliation, flattening of diamictite clasts, intrafolial folds	All domains, but most pronounced in the Etusis and Audawib Domains
	F1	Meter- to km- scale recumbent folds deforming S0. Centimeter- to decimeter-scale intrafolial folds within marble units	All domains
D2	S2	Upright, axial-planar foliation developed primarily in schists	All domains
	F2	Upright to NW-verging, open- to isoclinal folds. Developed on a cm- to km- scale	All domains
	L2m	NE-plunging mineral lineation defined by sillimanite and cordierite on the S2 surface	Otjimbingwe Domain
	L2c	NE-plunging crenulations lineation on the S2 surface of schists	Otjimbingwe Domain
	L2i	NE-plunging mineral lineation in intrusive granite	Otjimbingwe Domain



## Bedding ( $S_0$ )

Bedding ( $S_0$ ) banding defined by subtle compositional variations (banding between dolomitic- and calcitic-marble) or differences in rock types such as layered marbles and schists or calc-silicate felses.  $S_0$  is most prominent between formations where major lithological changes occur (Fig 4.1.1.A). In the Kuiseb Formation,  $S_0$  is often overprinted by later foliations and as a result is difficult to identify within the homogenous schists. Here,  $S_0$  can be distinguished by subtle lithological variations between biotite schists, psammitic schists, cordierite and sillimanite schists (Fig 4.1.1.B).



**Fig 4.1.1: A.)** On a regional scale,  $S_0$  can be distinguished between formations. This picture looks SW from Marmorkuppen, the large orange hill consists of quartzites of the Etusis Formation overlain by the marbles of the Karibib Formation of the Swakop Group, Etusis domain. **B.)** Folded bedding ( $S_0$ ) in schists of the Kuiseb Formation in the Otjimbingwe mapping domain in the Hases river (plan view), hammer for scale. Light layers are psammitic units, darker units are biotite-schists.

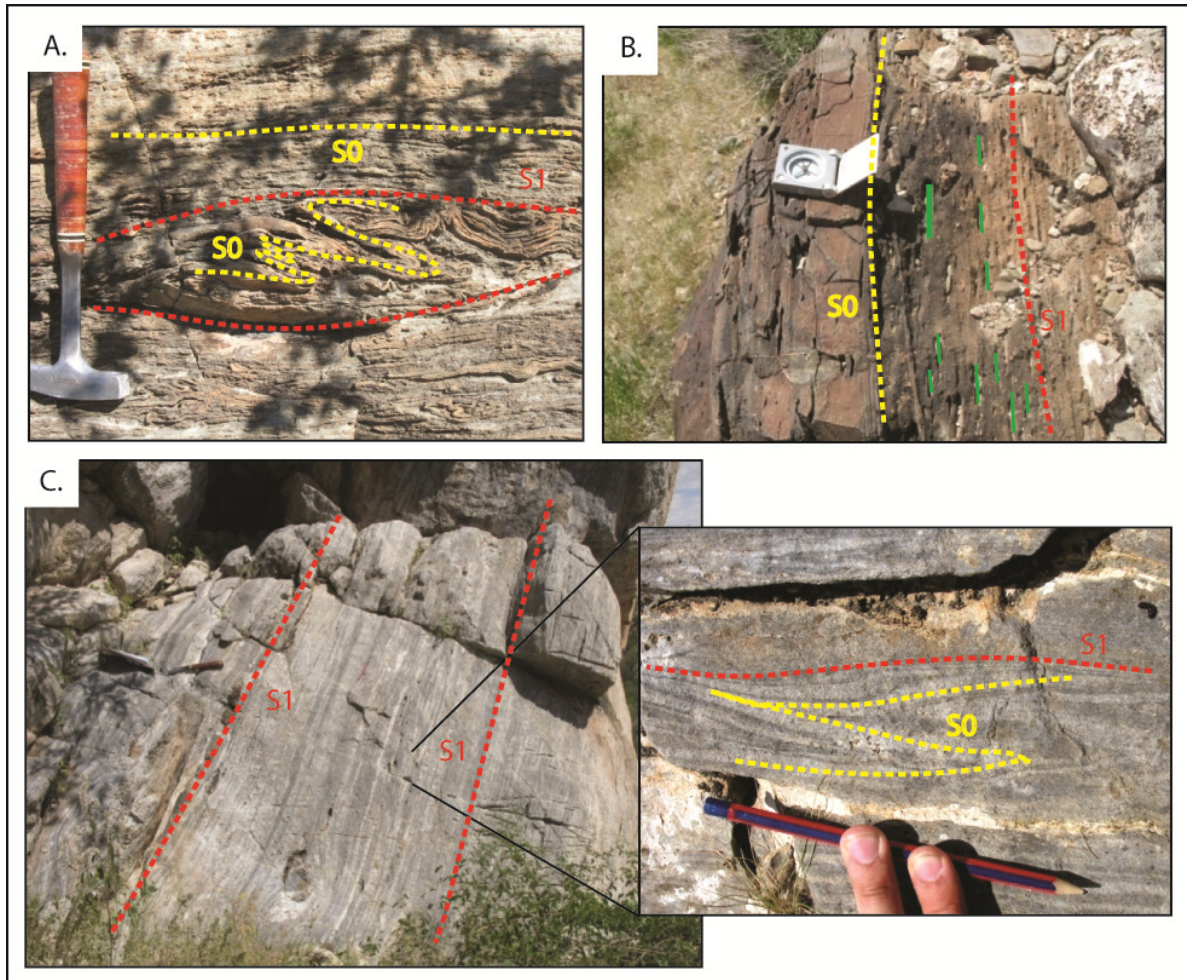
### 4.1.2 Structural Fabrics ( $S_1$ , $S_2$ , $L_2$ )

#### $S_1$ – Bedding parallel foliation

The oldest tectonic fabric is a bedding-parallel foliation,  $S_1$ . This fabric is not always easy to distinguish from the later  $S_2$  foliation, as both fabrics are defined by identical minerals and mineral assemblages and the fabrics tend to be sub-parallel, particularly on the limbs of upright F2 folds.  $S_1$  is also axial planar to intrafolial folds (F1) within individual units (Fig 4.1.2.A,C). It is seen in marbles, but is most pronounced in biotite schists of the Kuiseb Formation where metamorphic minerals such as hornblende and biotite define a  $S_1$  foliation parallel to  $S_0$ . Marbles of the Karibib Formation, in contrast, are often pervasively recrystallized, leaving little trace of  $S_1$ .  $S_1$  is also expressed as a shape fabric foliation by the preferred orientation of flattened clasts in diamictites of the Ghaub Formations (Fig 4.1.2.B.).

The  $S_1$  foliation is much better preserved and more pronounced in the Etusis Domain compared to the southern domains, where the steep  $S_2$  foliation has largely transposed and overprinted earlier bedding and the  $S_1$  foliation.



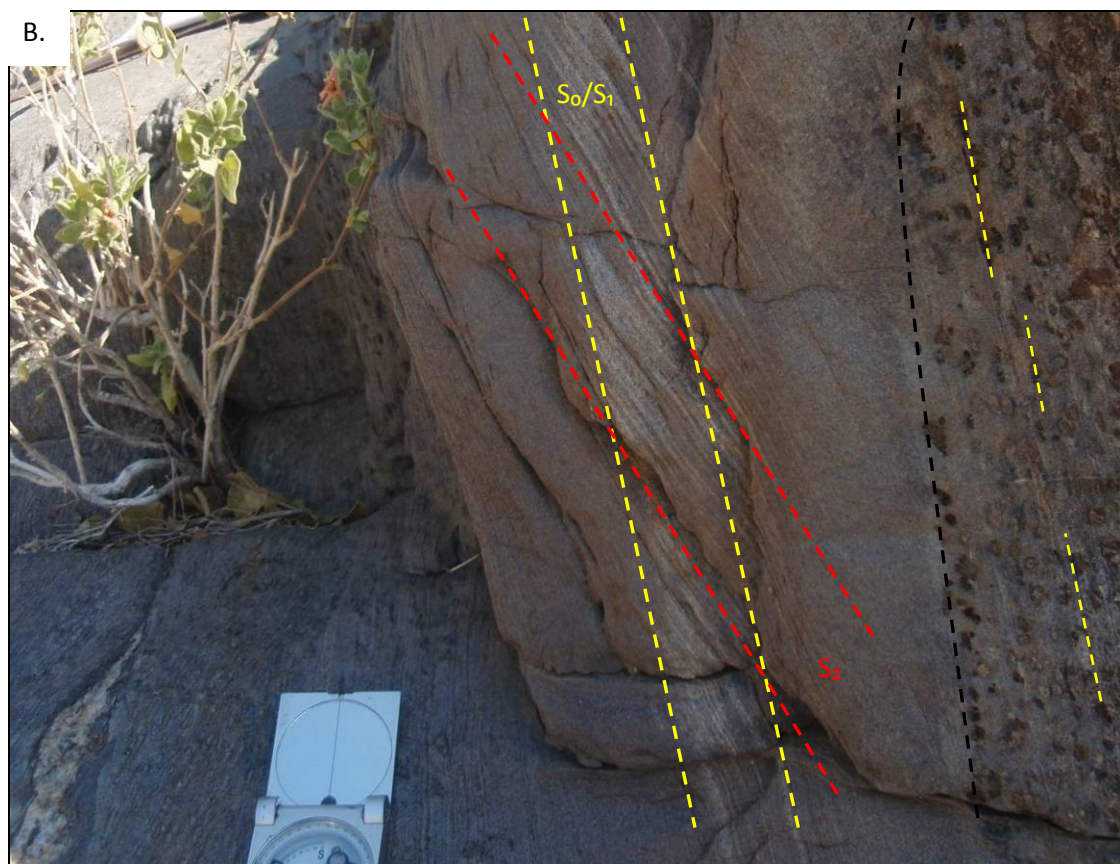
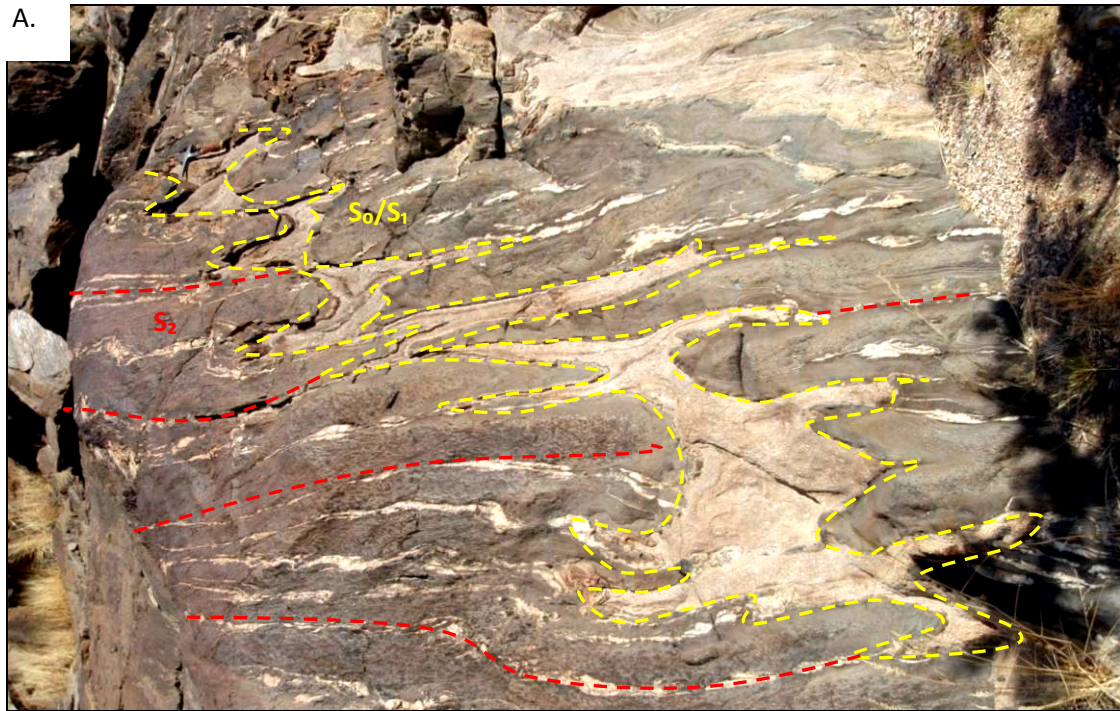


**Fig 4.1.2** A.) F1 folding of  $S_0$ , in this photograph  $S_0$  forms the enveloping surface to  $S_1$  within marbles (Karibib Formation). B.)  $S_1$  is defined by elongated calc-silicate clasts in a diamictites of the Ghaub Formation parallel to bedding ( $S_0$ ) C.) On closer inspection banding parallel to the  $S_0$  foliation shows intra-folial folds in the Okawayo Formation. These photos were taken around the area of the Marmorkuppen in the Ghaub (B) and Okawayo (A,C) Formations

**$S_2$**  – Steep, axial-planar foliation (Fig 4.1.2 & Fig 4.1.3)

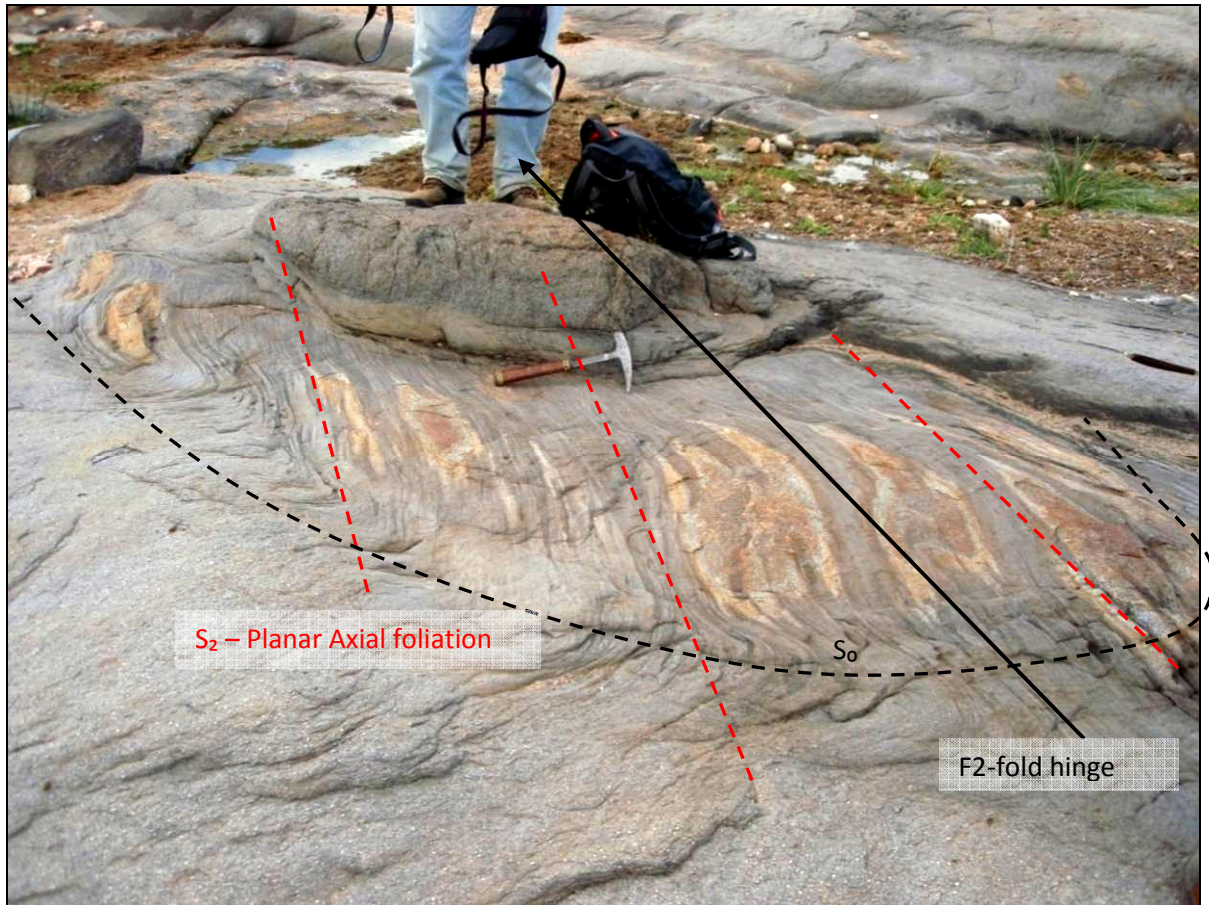
This is a predominantly steeply dipping (but shallow in D2 thrusts) NE trending fabric developed parallel to the regional grain of the Damara belt. The foliation is developed axial-planar to both open as well as tight- to isoclinal F2 folds (Fig 4.1.4). Variable dips of  $S_2$  also delineate different fold and thrust vergences of large-scale structures. The vergence of D2 structures changes systematically along the NW-SE traverse. These foliations can be hard to distinguish from  $S_1$  in the Etusis mapping domain, but forms a pervasive NW-dipping upright foliation in rocks of the Otjimbingwe mapping domain. In zones of pervasive bedding transposition,  $S_2$  can easily be mistaken for  $S_1$ . A  $S_2$  mineral foliation can easily be observed (Fig 4.1.3.B).





**Fig 4.1.3.A.)**  $S_2$  foliation (annotated as red dashed lines) cross-cutting  $S_0/S_1$  (yellow lines). The light-coloured marble forms mullion-type, flame-like structures protruding parallel to  $S_2$  into the more competent schistose rocks. This photo was taken at  $22^{\circ}21'23''S$ ;  $15^{\circ}53'39''E$  in the Audawib river. Where a marble is sharply cross-cut by the  $S_2$  foliation, the relatively plastic marble layer is completely buckled. **B.)** The  $S_2$  foliation is very pronounced in schists where it cross-cuts  $S_0/S_1$ . Here  $S_2$  foliation planes are defined by pressure solution seams, resulting in the somewhat streaky appearance and fuzzy bedding planes of the steeply-dipping sandy unit ( $S_0$  annotated).





**Fig 4.1.4:** Strong  $S_2$  axial-planar foliation development in F2 fold within the calc-silicate-schist sequence of the Tinkas Formation. Pressure solution of calc-silicates create flame like structures along a F2 fold. This photograph was taken in the Hases river about 1.5 km north of the Audawib river.

#### $L_2m$ – Mineral lineation

A mineral lineation is often developed in rocks of the DSG. This lineation is developed on the  $S_2$  foliation surface and is most pronounced in schist units of the Kuiseb Formation where it is often defined by the preferred orientation of predominantly sillimanite. A mineral stretching lineation is defined by stretched cordierite and or stretched quartz-feldspar aggregates. These linear features are best developed in metasedimentary rocks in the Otjimbingwe domain and gneisses in the Audawib domain (chapter 4.5).

#### $L_2c$ - Crenulation lineation

$S_2$  is locally crenulated and the hinges of these minor folds define  $L_2c$ . This crenulation is often found adjacent to with the mineral lineation ( $L_2m$ ), though  $L_2m$  has not been observed developed on the crenulation surface. These are primarily found in the Otjimbingwe terrain. Nearing the margin of the Palmental granite suite, prolate fabrics defined by the stretching of magmatic enclaves, rodding of mineral aggregates and rotation of fold hinges are

pervasive and this crenulation axis becomes even more pronounced. These fabrics are further discussed later in the chapter (chapter 4.5).

$L_{2i}$  – A strong mineral preferred orientation lineation is displayed in some of the intrusive units in the SCZ. This lineation is best developed in the Palmental granite and seems to be consistently parallel to crenulations and mineral lineations in the surrounding rocks of the DSG (Chapter 4.5). The lineation is similar in orientation to  $L_{2m}$ , but distinguished from it, since it occurs in intrusive rocks.

#### **4.1.2 Folding (F1-F2)**

Folds are a common manifestation of deformation in the SCZ ranging in wavelength from a few cm ( $4^{\text{th}}/5^{\text{th}}$  order) to many kilometres ( $1^{\text{st}}$  order folds). Based on overprinting relationships, at least two distinct fold generations can be distinguished. Coaxial refolding of folds is common and this is interpreted to reflect the progressive nature of convergent and collisional tectonism recorded in the SCZ (Poli & Oliver, 2001).

##### **F1 – Recumbent folds and bedding-parallel intrafolial folds**

Earlier authors have described F1 folds as recumbent or near-recumbent folds, that are related to the early convergent history of the Damara belt (Downing & Coward, 1981; Kasch, 1983a; De Kock, 1989, Miller, 2008). F1 folds occur most commonly as cm- to dm-scale, intrafolial, tight to isoclinal folds restricted to individual beds (Fig.4.1.2.A). This suggests that, at least, parts of the succession have undergone bedding transposition and that stratigraphic thickness have to be viewed with caution. Intrafolial F1 folds are common in rocks of the Kuiseb and Karibib Formations, but are relatively rare in the Etusis Formation that tends to preserve original sedimentary features, showing a less strain, owing to its competence.

##### **F2 – Upright-to overturned, open to isoclinal folds**

F2 folds commonly show NE trends, defining the regional structural grain of the Damara belt. Most folds in the study area are interpreted to be F2 structures. These folds range from open to isoclinal and are typically upright, but also overturned. Folding is pervasive in the Otjimbingwe domain, where isoclinal upright folds transpose early fabric elements (Fig.4.1.4). F2 is associated with a very pronounced NE trending axial-planar  $S_2$  foliation.

The F2 folds verge SE along the OLZ in the southern parts of the study area and NW in the northernmost part of the traverse in the Etusis domain. This transition from SE to NW vergent folding can be seen in the Audawib mapping terrain where doubly-plunging folds are bi-vergent to



both the SE and NW. The detailed presentation of data and analysis of fold structures follows in chapters 4.2 to 4.5.

#### **4.1.3 Shear zones and high strain fabrics.**

Two prominent ductile shear zones can be followed for several kilometres along strike in the Etusis and Audawib domains (Appendix I and III). A number of discrete structural features such as mylonites and/or exotic xenoliths along contact margins and the excision of lithologies also indicate the existence of two large-scale shear zones. Details are discussed in the respective mapping domains.

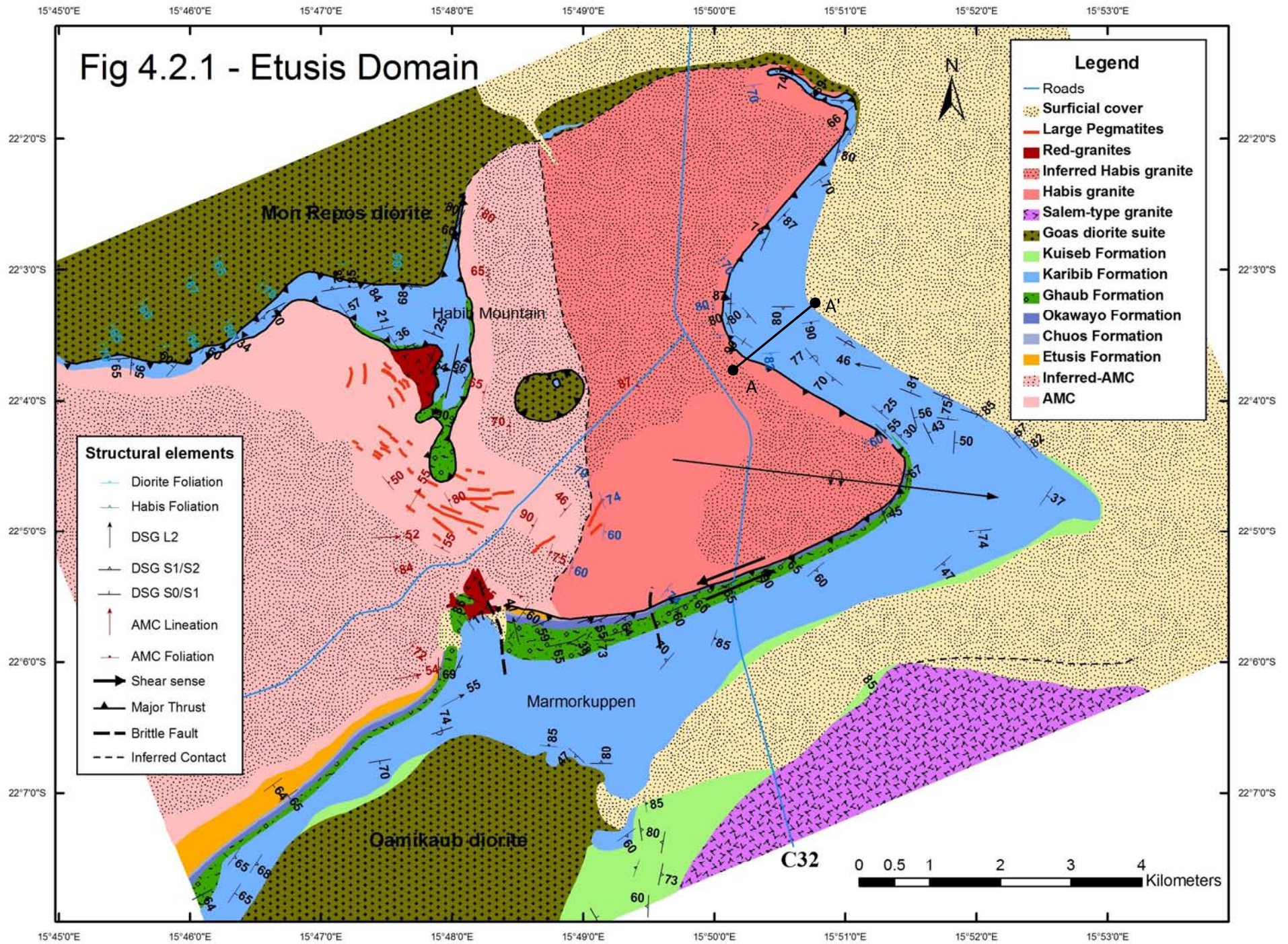
### **4.2 Etusis domain**

#### **Domain description**

The NW part of the domain on the farm Mon Repos is almost entirely underlain by rocks of the Mon-Repos diorite and granodiorite, sandwiched between marbles of the Karibib Formation in the hangingwall and either basement of the AMC or the Habis granite in the footwall of the shear zone (Fig 4.2.1, Appendix I). The central part of this domain is cored by biotite schist, pink feldspathic gneisses and calc-silicate felses of the AMC and foliated rocks of the Habis granite suite (Fig 4.2.1 and 4.2.2). Thin (< 30-40 m) sheets of Mon Repos diorite and/or rocks of the Ghaub and Karibib Formations occur as isolated inselbergs overlying the basement and the Habis granite. To the south, the AMC and the Habis granite are overlain and enveloped by a more complete section of rocks of the DSG. The Karibib Formation dominates in the E parts of this domain, whereas lower formations, from the Etusis to the Ghaub Formations are developed in the W (Fig 4.2.1.). Most of the rocks at the basement-cover contacts and particularly those of the Karibib Formation are strongly recrystallized with, in places, mylonitic fabrics developed.

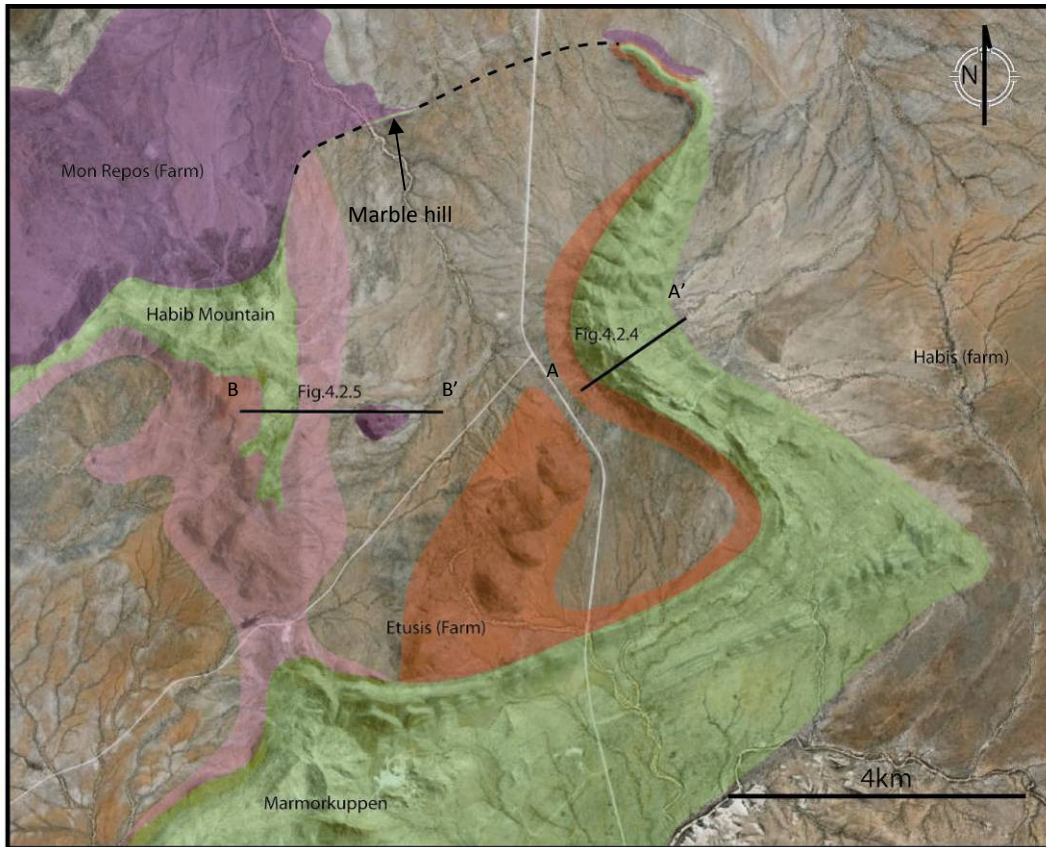
#### **AMC**

Rocks of the AMC occur in the central part of the domain, underlying rocks of the DSG towards the NW, E and SE. The biotite-muscovite schists and feldspathic gneisses contain a penetrative, often crenulated foliation (Fig 4.2.3.A and B). The steeply dipping planar fabric undulates, but shows overall N-S trends (Fig 4.2.3.C). As such, fabrics in the AMC are commonly at an angle to those developed in rocks of the overlying DSG (Fig 4.2.3.C & Fig 4.2.6).



For a detailed version of this map see Appendix I





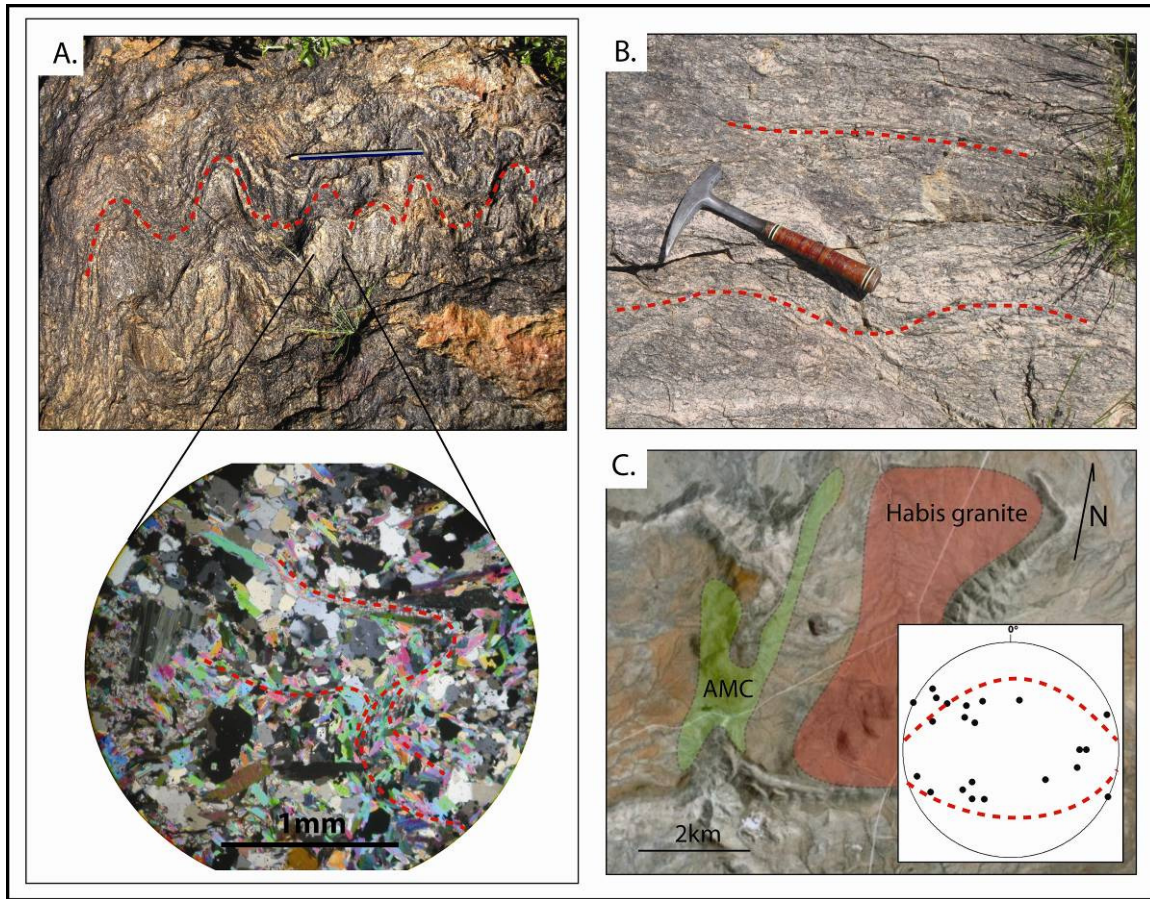
**Fig 4.2.2:** Google earth image of the Etusis domain. The DSG is shown in green, AMC in pink, Habis granite in red and Mon Repos diorite in purple. This illustration gives a approximate outline of features in the Etusis domain and the locations of the cross-sections in Fig 4.2.4 and Fig 4.2.5.

### Damara Supergroup

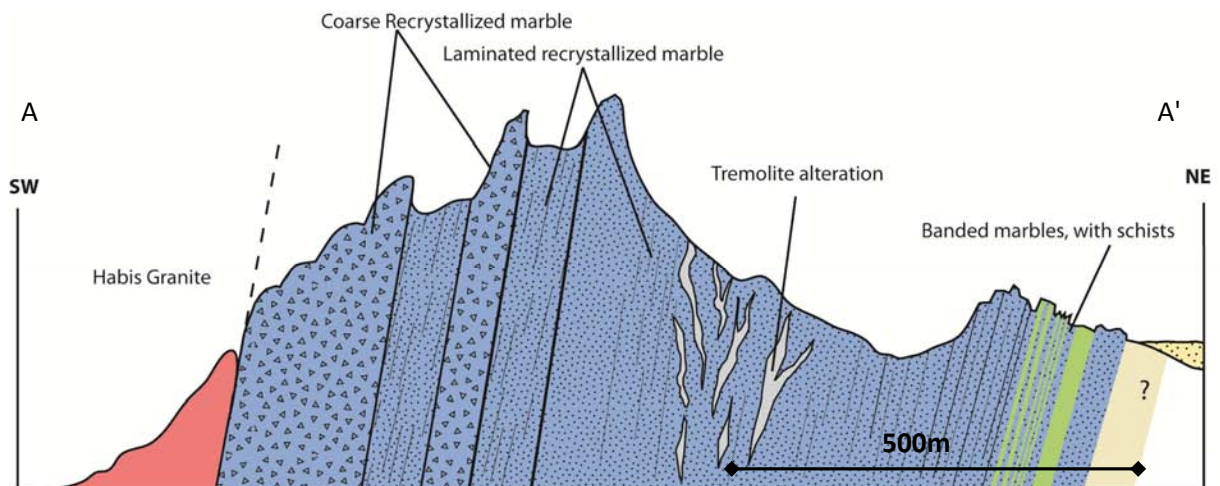
The marbles of the Karibib Formation show a gradual decrease in the amount of matrix-recrystallization away from the contact with the underlying AMC and Habis granite (outwards from the AMC inlier e.g. from A to A' in Fig.4.2.2) (Fig 4.2.4 & chapter 3.3). A decrease in the matrix grain size is accompanied by an increase in recognizable bedding features and compositional banding in the Karibib Formation (Fig 4.2.4).

Following the Karibib Formation SW and around the Habis granite-marble contact along the Marmorkuppen hills (Fig.4.2.2), the degree of recrystallization decreases and evidence of primary bedding is recognized, all of which is thought to be an indication of a decrease in strain intensity. The decrease in fabric and strain intensity is accompanied by the re-appearance of formations of the lower DSG underlying the Karibib Formation (Appendix I), including, from bottom to top, the Etusis, Chuos, Okawayo and Ghaub Formations. These formations increase in thickness further W, situated below the marbles of the Karibib Formation.

Although the lower formations of the DSG are only developed towards the SW of the domain, slivers of diorite as well as remnants of layered calc-silicate facies, (probably of the lower Ghaub Formation) occasionally occur along the basement-marble contact just south of the Habib mountain (Fig 4.2.5).



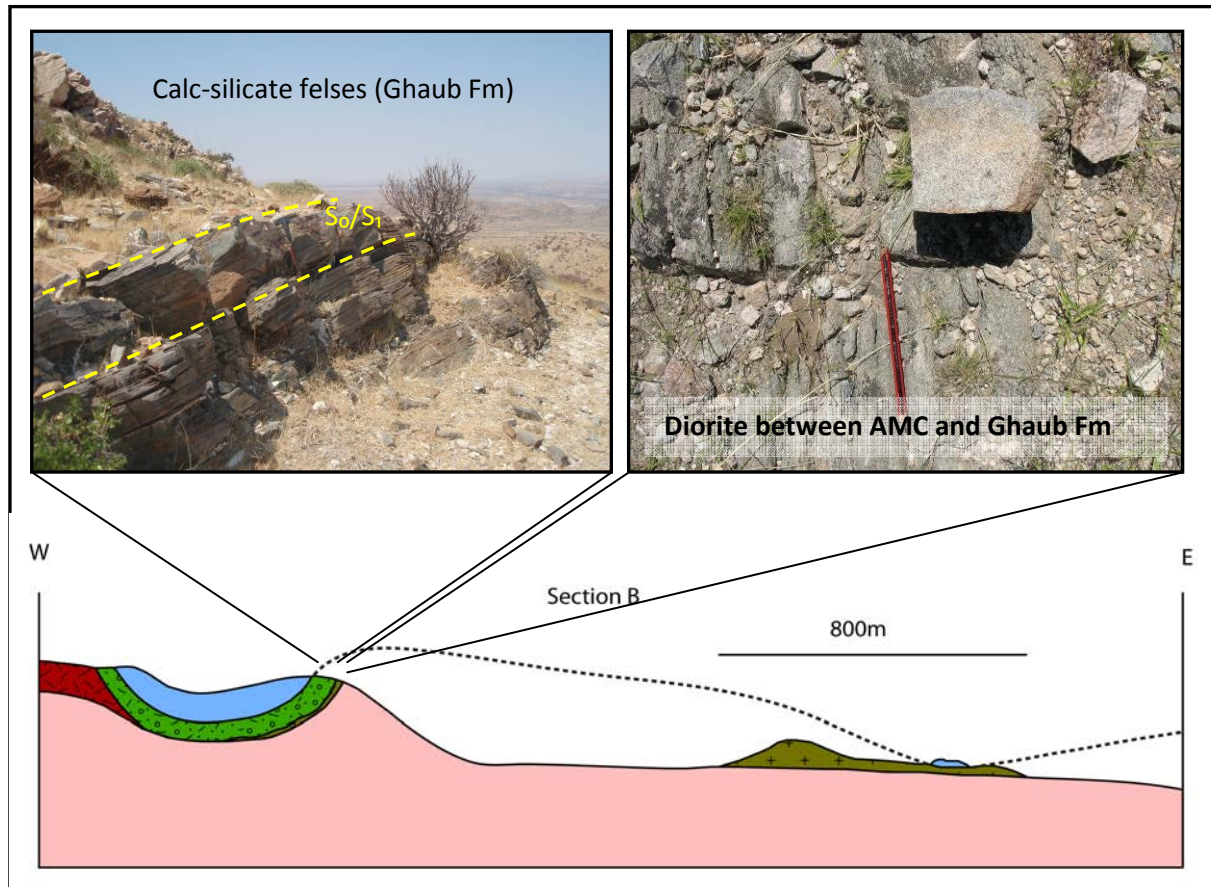
**Fig 4.2.3:** **A.)** AMC schists show a crenulation cleavage defined by alignment and folding of muscovite on macroscopic and mesoscopic scale. **(B)** Augen gneisses have a foliation defined by the preferred orientation of masses of quartz and K-feldspar, resulting in a planar fabric on a mesoscopic scale. **(C)** Both schist and gneisses show a rough trend N-S, as foliation poles roughly lie between 270° and 90°.



**Fig 4.2.4:** Traverse of the Marble hills on the Farm Habis, showing the progressive decrease of fabric intensities and strain (recrystallization) in marbles of the Karibib Formation. Away from the contact, coarsely recrystallized, homogeneous marbles give way to finer-grained marbles with visible compositional banding and primary bedding features. Rocks such as BIF's and biotite schist occur eastwards. Blue in the NE corner of the traverse indicates surficial cover. Vertical scale is exaggerated. This traverse looks NW and is shown as A-A' in Fig 4.2.2.



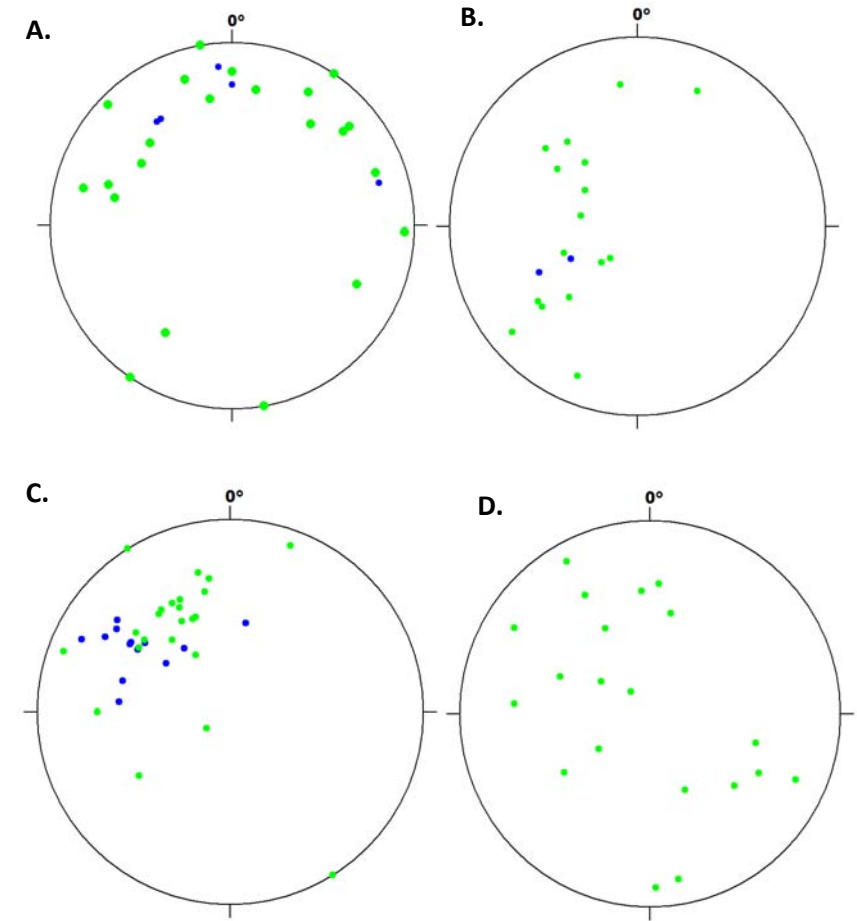
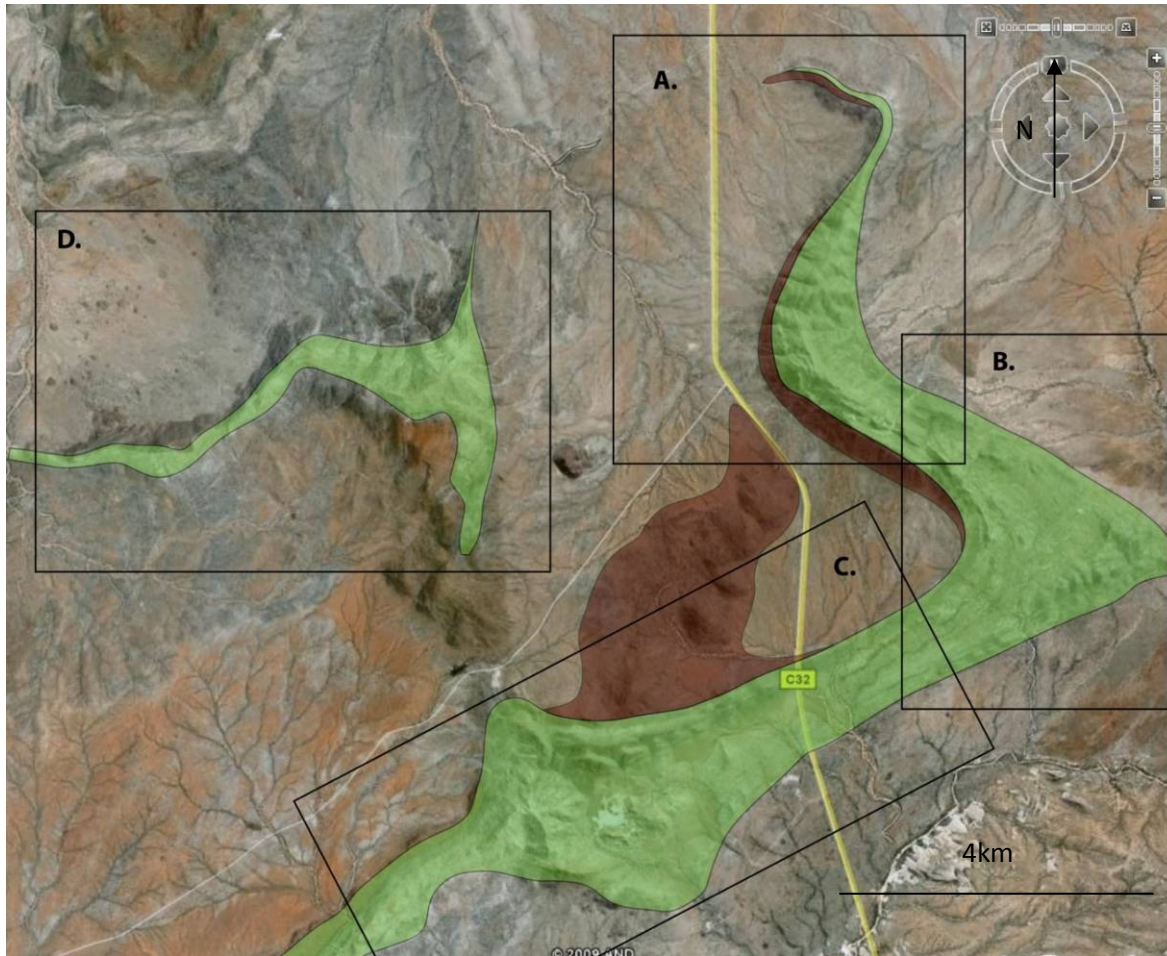
Slivers of diorite up to 1 m thick occasionally occur along this AMC/DSG contact (Fig 4.2.5). Slivers of banded calc-silicate felses can be up to 30 m thick, showing well preserved  $S_0$  layering parallel to and conformable with the overlying Karibib Formation marbles (Fig 4.2.5).



**Fig 4.2.5** – A sliver of layered calc-silicate felses about 30 m thick (top left) and a patch of weathered diorite about 30cm thick (top right) lies between gneisses of the AMC (orange in cross section) and marbles of the Karibib Formation. Eastwards an erosional klippe is preserved where a small relic of marble is found on a sheet of Mon-Repos diorite overlying AMC basement rocks. The cross-section is taken as B-B' in Fig 4.2.2. In the above cross-section, red-striped rocks are red-granites, pink shows the AMC, brown-green indicates diorites and green the calc-silicate felses, with light-blue marbles on top.

Re-crystallized marbles of the Karibib Formation often preserve a macroscopic lamination. This lamination is parallel to  $S_1$  developed in the lower formations of the DSG. This foliation is also common in calc-silicates of the Ghaub Formation, being defined by a preferred orientation of matrix minerals as well as the long axis of deformed marble clasts within calc-silicate diamictites (Fig 4.1.2.B). This  $S_1$  foliation is consistently parallel to  $S_0$  banding and the contacts between formations.

The aforementioned main foliation in the DSG is moderately dipping ( $30^\circ$ - $65^\circ$ ) and follows the contact around the AMC/Habis inlier. A pronounced SE-dipping foliations persists S-SE of the AMC inlier (trending NE parallel to the AMC/DSG contact.), in all of the DSG formations. This NE



**Fig.4.2.6:** Comparison between foliations of the Habis granite (poles to foliation: blue) and those of the DSG (green) in four different sub-domains (A-D) of the Etusis domain. In the field, a  $S_1$  and  $S_2$  foliation is indistinguishable. **(A)** The spread of poles to foliations indicates that the foliation in the Karibib Formation is parallel to and follows the contact with the Habis granite. The presence and clustering of SE dipping foliations in the Karibib Formation and the Habis granite indicates the existence of the  $S_2$  foliation. **(B.)** Foliations define a large 2<sup>nd</sup> order fold with a crude great circle distribution of poles pointing to a moderately E plunge of the fold. **(C.)** In sub-domain C, the foliation in both the DSG and the Habis granite suite dips consistently towards the SE.  $S_0$  bedding also dips SE in this area and  $S_1$  and  $S_2$  are likely to be parallel. **(D.)** In the Habib mountains, a shallowly-dipping, but widely scattered  $S_1$  foliation persists indicating the shallow sheeted nature of marbles overlying the basement. Where rocks of the Habis granite suite are exposed, these show a foliation approximately parallel to that of the DSG. These rocks are however exposed mainly in sub-domain C and in the centre of the domain, showing a moderate SE-dipping foliation parallel to  $S_2$ .