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# Intrusion mechanisms in a turbidite sequence: the Voetspoor and Doros plutons in NW Namibia

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**Abstract**

Two syntectonic plutons of Cambrian age intruded Neoproterozoic metaturbidites in Namibia at the junction of the NS trending Kaoko and EW trending Damara belts. Sinistral transpression in the Kaoko Belt produced km-scale upright D1 folds overprinted by minor D2 folds. D3 is associated with N–S shortening in the Damara Belt. The plutons show two main pulses of intrusion: hornblende syenite intruded late during D1 or during D2 and biotite granite during D3. Each tectonic event produced a strain shadow defined by the shape of folds and the foliation trend around the plutons. The internal igneous fabric and the arrangement of wall rock xenoliths that locally make up 50% of the intrusion mass suggest that the plutons have a disk or wedge shape. A marginal shear zone indicates that one of the syenite intrusions descended during emplacement with respect to the wall rock. Emplacement is therefore inferred to be at least partly accommodated by descent of the intrusion floor. The biotite granite intruded into mega-strain shadows and tension gashes alongside and in the syenite during D3. The plutons show evidence of sinistral solid state rotation with respect to the wall rocks in response to D1–D2 transpression.


*Keywords:* Granite emplacement; Namibia; Lower Ugab domain; Neoproterozoic

## 1. Introduction

Emplacement mechanisms of igneous bodies and especially the related space problem are of considerable interest in crustal genesis and evolution. Plutons emplaced during regional deformation are important because they provide a means of dating deformation phases and to establish a link with magma generation elsewhere in the crust or mantle (Brun and Pons, 1981; Paterson and Vernon, 1995; Vigneresse, 1995; Cruden and McCaffrey, 2001). Models of magma ascent consider diapirism and related mechanisms, where a pluton rises through the crust, pushing aside the wall rock (Marsh, 1982; Bateman, 1984; Mahon et al., 1988; Paterson and Vernon, 1995) or ascent through dykes (Clemens et al., 1997; Petford et al., 2000), commonly associated with faults or shear zones in the host rock (D'Lemos et al., 1993; Petford et al., 1993; Clemens et al., 1997). Emplacement of large plutons is problematic since space has to be created in the crust for the intrusion (Petraske et al., 1978; Paterson and Fowler, 1993). Material transfer mechanisms such as stoping (Marsh, 1982; Paterson and Fowler, 1993) do not solve the large-scale space problem. Presently, most generally accepted and well-documented emplacement models are (1) ballooning or forceful emplacement through fluid pressure in the intrusion, displacing the wall rocks sideways (Bateman, 1985; Marre, 1986; Castro, 1987; Courrioux, 1987; Ramsay, 1989), (2) nested diapirism without lateral expansion (Paterson and Vernon, 1995), (3) uplift of the overlying sequence as in sills and

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laccoliths (Jackson and Pollard, 1988; Roman-Berdiel et al., 1995; Morgan et al., 1998) possibly including fracture systems that isolate part of the roof (Corry, 1988; Benn et al., 2000) and (4) footwall collapse, where space is created by sinking of the footwall along monoclines or a fault system (Cruden, 1998; Petford et al., 2000; Cruden and McCaffrey, 2001).

Although it may seem easy to reconstruct emplacement mechanisms in well-exposed areas, the problem has been a topic in the geological literature for over a hundred years, and is still only partly resolved. Reasons are that three dimensional exposure is usually limited, and that the geometry of an intrusion only represents the last stage of development (Paterson and Vernon, 1995; Grocott et al., 1999). Also, wall rocks are commonly hornfelsed and may have lost the delicate small scale structures that are the main tools of structural geology.

We analysed two plutons that invaded a belt of Neoproterozoic turbidites in Namibia, the Zerrissene Turbidite System, deformed during the Panafrican Orogeny of late Neoproterozoic and Cambrian age (Swart, 1992; Fig. 1). Both have a comparable composition and intruded the metaturbidite sequence with a nearly identical sequence of events. They are also similar in composition and tectonic age to other plutons in the area, which have not yet been studied in detail, including the large Omangambo intrusion (Fig. 1). The geological setting and excellent exposure in the Namibian desert constitute a favourable reference frame to study the structures associated with pluton emplacement. The area around the plutons was mapped on scale 1:10.000 and over 4000 structural measurements stored in a MapInfo Database.

## 2. Geological setting

The plutons and the surrounding metasedimentary successions of the Zerrissene Turbidite System (Swart, 1992) are localised in the southern Kaoko Zone (Miller, 1983; Miller et al., 1983; Hoffman et al., 1994; also called Ugab Zone, Goscombe et al., 2003a,b), a tectonic unit localised where the N–S trending Kaoko Belt merges into the NE–SW trending Damara Belt (Fig. 1). The limits of the southern Kaoko Zone are poorly defined because of its cover by late Palaeozoic and

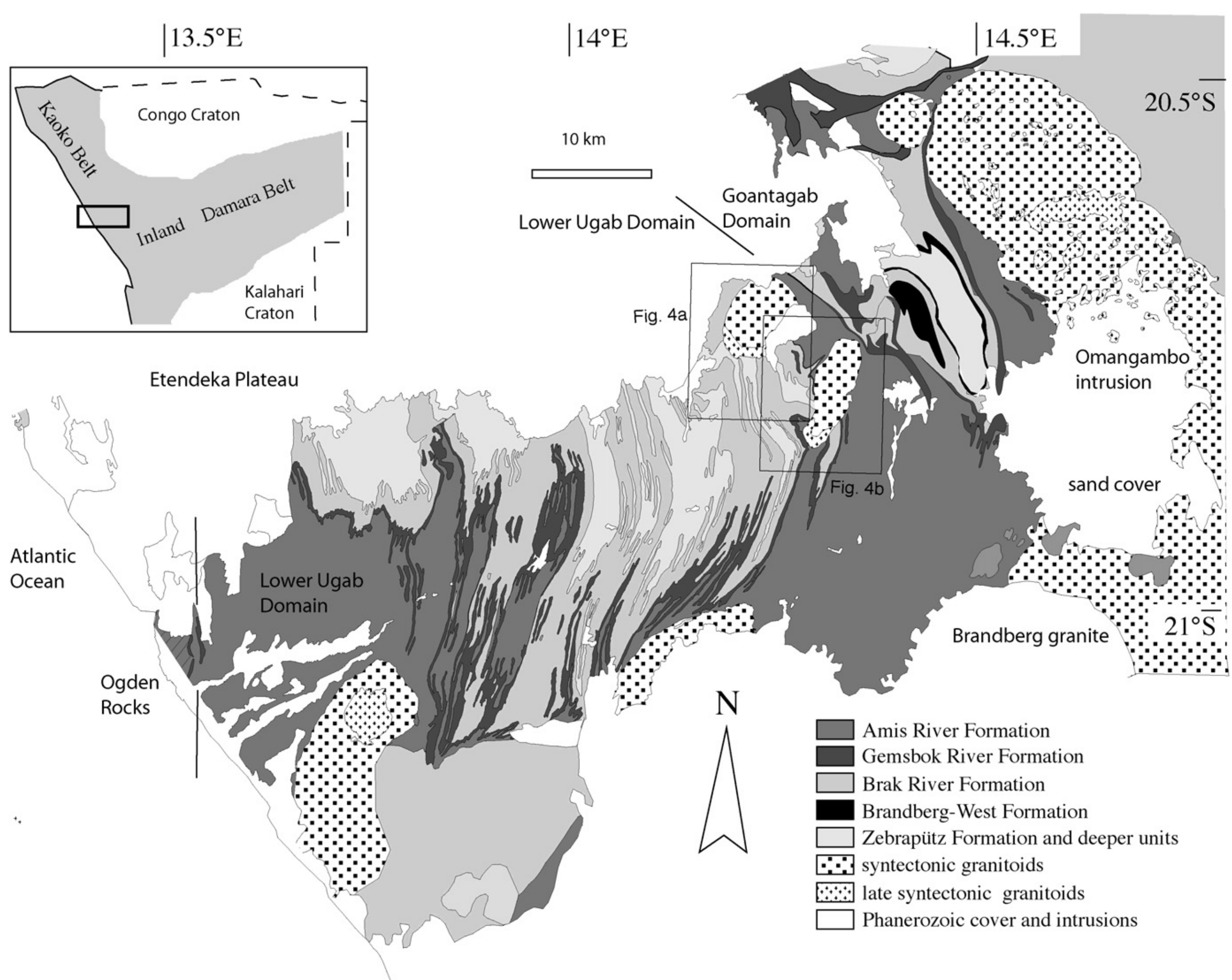


Fig. 1. Map of the Neoproterozoic rocks in the Lower Ugab and Goantagab Domains. Position of the Doros and Voetspoor plutons is indicated with squares.

Mesozoic sedimentary and volcanic rocks, both to the north and to the south (Fig. 1). This part of the Kaoko Belt was subdivided into three tectonic domains (Hoffman et al., 1994); the poorly exposed mylonitic Ogden Rocks Domain in the west, the Lower Ugab Domain occupying the central part of the belt and the Goantagab Domain in the northeast (Fig. 1; Hoffman et al., 1994; Passchier et al., 2002). The plutons are localised in the Lower Ugab Domain, close to the contact with the Goantagab Domain (Fig. 1). The Zerrissene Turbidite System in the Lower Ugab Domain is composed of a succession of metasedimentary rocks of about 1600 m minimum thickness. The basement and the top are not exposed. The succession consists of two turbiditic marble deposits separating three formations of pelitic and psammitic metaturbidite (Fig. 1). The tubiditic marble units consist of a lower, 15–20 m thick uniformly banded formation (Brandberg West Formation) and an upper ~200 m thick alternation of thinly banded marble and metapelite, including an upper massive marble unit (Gemsbok River Formation). To the east of the studied plutons, the stratigraphy changes; diamictite, coarse sandstone and quartzite lenses appear, and the marble units are more massive, indicating a sedimentary environment more proximal to the craton. The age of the metaturbidites has not been clearly established. Regional scale mapping, however, suggests that the Brandberg West and Gemsbok River marbles correspond to cap carbonates, which represent the end of the Sturtian and Marinoan glaciations, now dated at ca. 710 and 635 Ma (Kendall et al., 2004).

## 3. Metamorphism

Metamorphism in the Ugab Domain is of middle to upper greenschist facies, as indicated by the presence of abundant biotite in almost all rock types, besides albite/oligoclase, chlorite, carbonate and white mica. Garnet is rare, probably due to relatively low pressure and to rock composition. Amphiboles, mainly actinolite and possibly actinolitic hornblende, form poikiloblastic porphyroblasts in calcsilicates and impure marble. Superposed contact metamorphism around the granitic bodies produced cordierite and andalusite porphyroblasts, generally replaced by biotite muscovite chlorite aggregates. In calcsilicate layers hornblende, diopside and garnet grew, and wollastonite occurs locally along contacts between marble and the intrusions. Goscombe et al. (2004) give estimates of 540–570 °C and 2.5–3.2 kbar for the Ugab domain and this corresponds well with our estimates from the mineral parageneses. Goscombe et al. (2004) claim to recognise pervasive contact metamorphism in the Ugab Domain, attributed to extensive partly hidden granite intrusions similar to the plutons investigated here.

## 4. Deformation

The principal deformation phase (D1–D2) that affected the Turbidite System produced a sequence of upright to inclined tight to open large-scale folds, accompanied by the development of a penetrative slaty or spaced cleavage (Passchier et al., 2002; Goscombe et al., 2003a). The main folds, labelled D1, have a wavelength of 50–500 m depending on their tightness. The axes of D1 folds are subhorizontal and trend predominantly N–S (Fig. 1). D1 folds show a large-scale gradient in asymmetry associated with a cleavage fan; in the west of the area, axial planes dip gently east and folds verge westwards; close to the two plutons described in this paper, axial planes are subvertical and folds are upright and symmetric; east of the plutons, and especially in the Goantagab Domain, axial planes are west-dipping to subhorizontal and D1 folds become isoclinal and recumbent (Passchier et al., 2002; Fig. 2). The variable dip of the axial planes and the asymmetry of the folds is associated with a second phase of deformation D2, which also produced minor (D2) folds and S2 crenulation foliations throughout the area; in the west and centre, these D2-structures are weak and only locally developed, but east of the investigated plutons in the Goantagab Domain, upright D2 folds refolding recumbent isoclinal D1 folds become a dominant deformation feature (Fig. 2-profile; Passchier et al., 2002). The Goantagab Domain is an anticlinorium that structurally underlies the Lower Ugab Domain (Passchier et al., 2002).

Fold axes of D1 and D2 structures are normally parallel, and metamorphic conditions of formation are similar, indicating that D1 and D2 are related phases of one continuous tectonic event. The two phases may even be diachronous (Passchier et al., 2002).

D1 and D2 folds show evidence for stretching parallel to the fold axes, mostly through fibrous fringes around pyrite and boudinage of pelitic beds with E–W trending quartz veins filling the necks. Asymmetry of the boudin necks indicates a component of sinistral strike slip flow late during D1 and during D2. In fact, D1–D2 seems to have been a phase of transpressional deformation throughout the Lower Ugab Domain, which may have initiated, rotated and stretched the D1 folds (Passchier et al., 2002). In the Goantagab Domain, the stretching component of D1–D2 is stronger, and as a result N–S to NW–SE stretching lineations of D1–D2 age developed parallel to the fold axes (Fig. 3). Local peak metamorphic conditions prevailed during D1–D2 since biotite and rare garnet grew during these phases: included oblique foliation traces in garnet porphyroblasts and a slight outbowing of the external foliation S1 in the adjacent matrix show that garnet grew during D1.

D3 is a locally important refolding phase with upright folds and steep foliations that overprints D1–D2 structures at a large angle throughout the area (Fig. 2). The tightness and orientation of D3 folds and presence of an S3 foliation is variable over the area. Where D3 folds and foliations are ENE–WSW trending, local NNW–SSE shortening seems to be coaxial. NE–SW to NNE–SSW trending folds and foliations also occur, and these structures indicate a component of sinistral shear. This implies that D3 has been a phase of N–S bulk shortening. Metamorphic conditions during D3 were probably lower than during D1 and D2 since cordierite and andalusite porphyroblasts had broken down to biotite and chlorite, and the resulting aggregates were deformed. However, locally

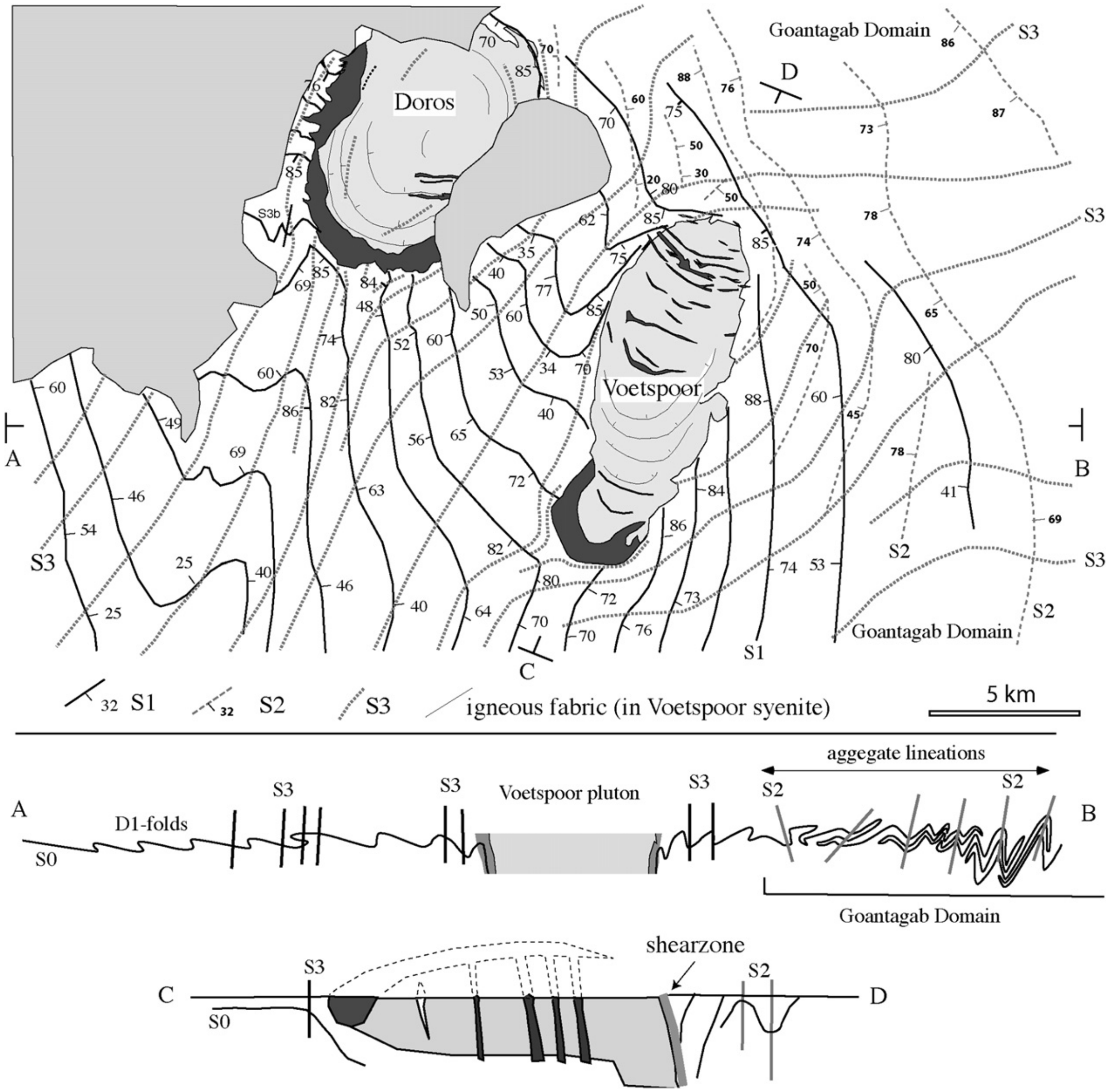


Fig. 2. Schematic presentation of the orientation of foliations in and surrounding the Voetspoor and Doros plutons. Dip values of S1 and S2 measurements are indicated with different font and size on lines presenting their strike trends. Where metasedimentary rock panels are present in the plutons, the orientation of S1-foliation within them is indicated.

recrystallisation of biotite has been observed. The intensity and orientation of D3 structures is mostly associated with the presence of granite plutons; the S3 foliations tend to concentrate between plutons, and to wrap around them. However, there are local high strain zones that are not associated with outcropping plutons, and these may be associated with buried or eroded plutons or to high-strain zones in the basement.

## 5. The Voetspoor and Doros plutons

The Lower Ugab Domain is intruded by a number of similar plutons, two of which, the Voetspoor and Doros plutons, are the subject of this study (Figs. 2–4). The Doros pluton is a roughly circular intrusion that is well exposed in the south, but less so in the NW where it is partly covered by drift, while in the SE it is partly covered by volcanic rocks of the Mesozoic Doros Mafic complex (Fig. 4a). The Voetspoor pluton is an elongate intrusive body, so named because of its resemblance to a giant footprint (‘‘voetspoor‘‘ in the Afrikaans language). It is well exposed except in its central southern part (Fig. 4b). The Voetspoor and Doros plutons are nested plutons both composed of two main components, hornblende syenite and biotite granite (Fig. 4). It was found that 60–70% of each pluton is composed of hornblende syenite with a range of compositions from quartz-syenite to gabbroic. The biotite granite occurs in the SW part of both plutons, along the western and southern rims, and in dykes in the centre. A minimum intrusive age of the hornblende syenite in the Voetspoor pluton

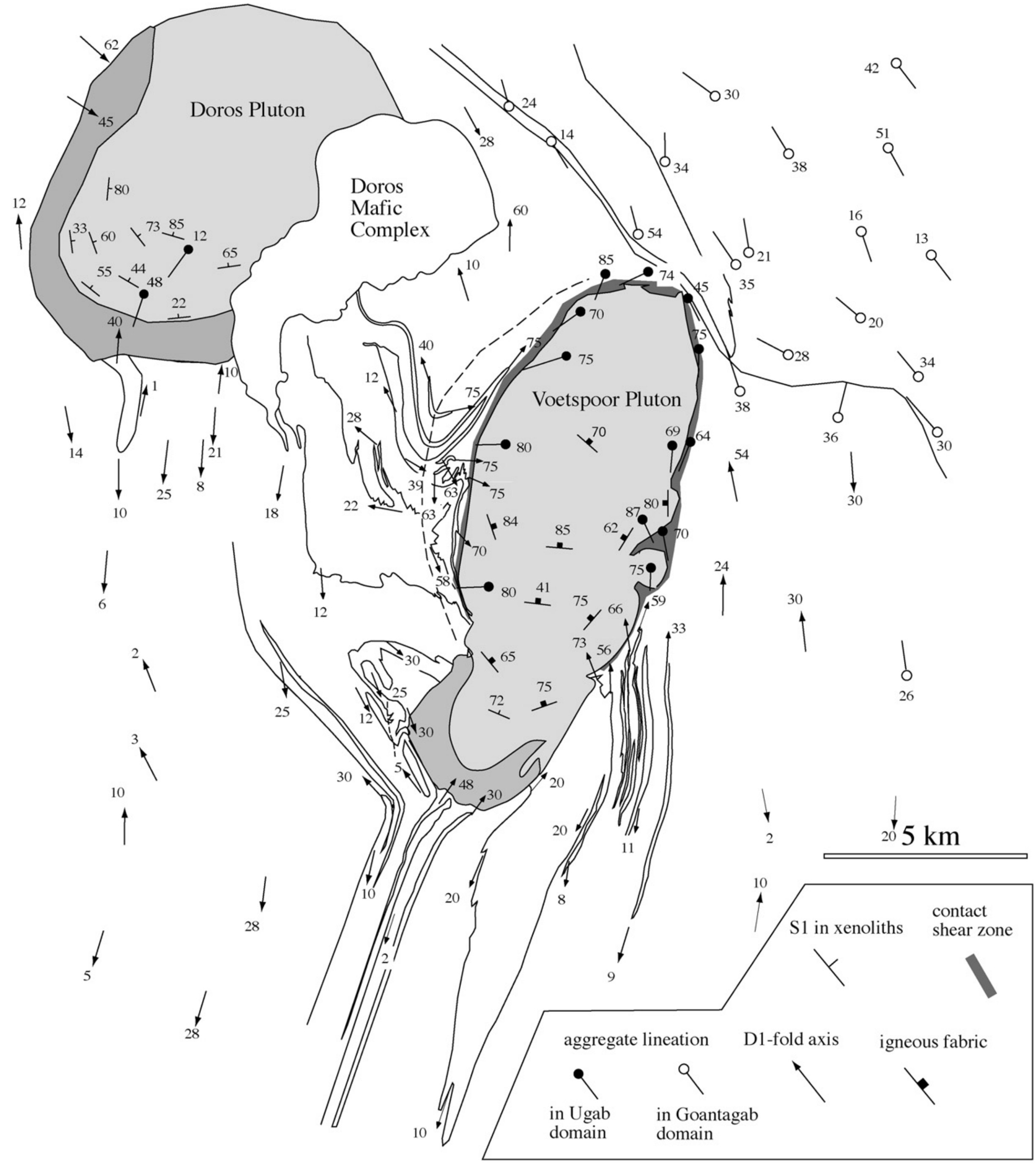


Fig. 3. Orientation of fold axes of D1 folds, and D1–D2 aggregate lineations around the Voetspoor and Doros plutons. Lithological contacts are traced for reference. A representative selection of measurements of S1 in metasedimentary rock panels and of igneous foliation in the syenite is shown.

was obtained at 530 ± 2 Ma by Pb evaporation on single zircons (Seth et al., 2000) and by conventional U-Pb dating of sphene at 541 ± 6 Ma (Jung et al., 2005). For the biotite granite, four idiomorphic zircon grains were evaporated individually and yielded comparable Pb isotopic ratios that provide a mean $^{207}Pb/^{206}Pb$ age of 513 ± 1 (Table 1, Fig. 5) and that we interpret to reflect the time of granite emplacement. For the Doros pluton there is a whole rock Rb-Sr age of 573 ± 33 Ma (Kröner, 1982).

### 5.1. Hornblende syenite

Hornblende syenite with a matrix grain size of 1–3 mm and a strong composition gradient is the main component of both plutons. In the NE of the Voetspoor pluton and the centre of the Doros pluton, dioritic composition dominates with 30–40% hornblende and up to 5% clinopyroxene. Towards the SW in the Voetspoor pluton and external parts of the Doros pluton, pyroxene disappears, the percentage of hornblende gradually

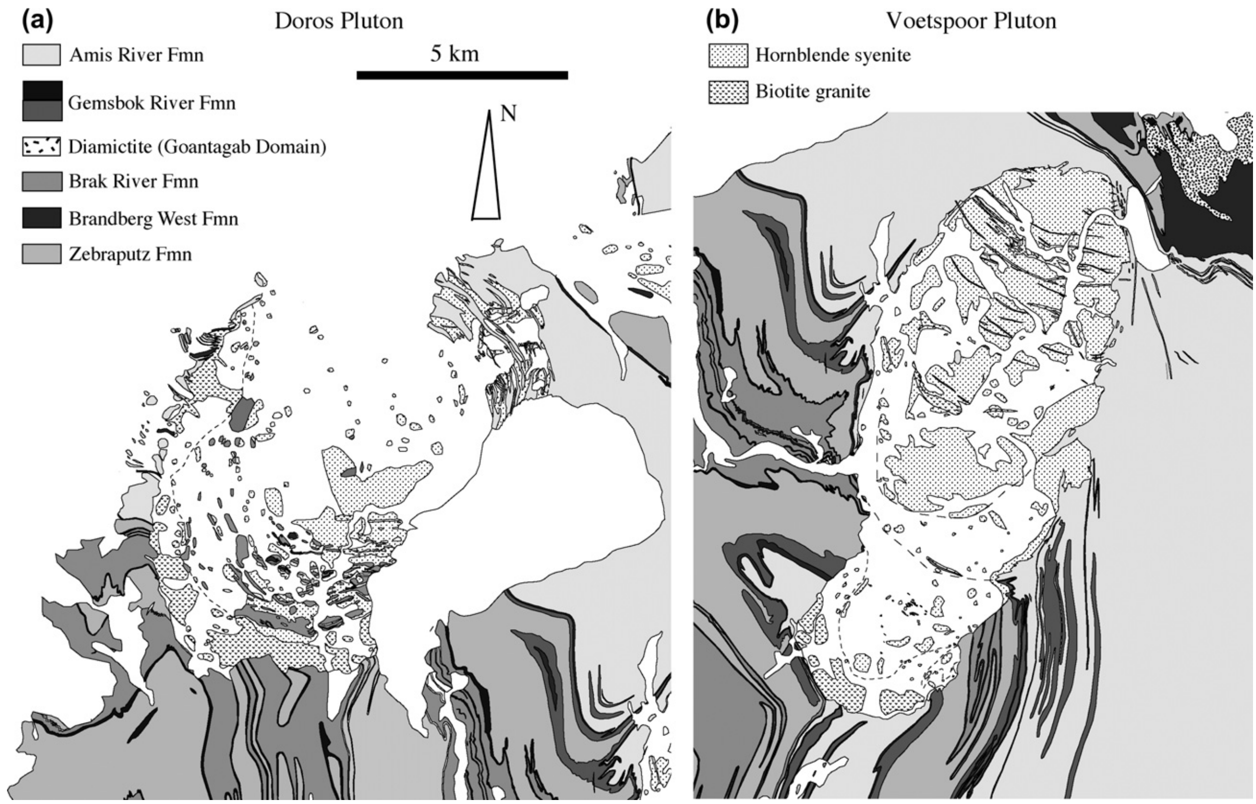


Fig. 4. Detailed maps of the plutons. (a) Geological map of the Doros pluton and surrounding units. (b) Geological map of the Voetspoor pluton and surrounding units. The maps are partly overlapping. Metasediment xenoliths in the intrusions cannot be attributed to individual formations. Location of both maps is indicated in Fig. 1. Mesozoic dolerite dykes have been omitted from the maps.

decreases to about 20% and idiomorphic K-feldspar phenocrysts gradually increase in size up to 3 cm in length. Composition changes gradually to a quartz-syenite or quartz-monzonite for the most felsic lithotype. All lithotypes plot as peraluminous, alkalic to subalkalic magmas with normative olivine and nepheline in the darkest phases and normative quartz in the lighter rocks (Jung et al., 2005). Enclaves of more hornblende and pyroxene-rich material, from cm to 100 m diameter are common, especially towards the northern part of the intrusions. These enclaves usually have sharp contacts with the more light-coloured syenite. The distribution of K-feldspar phenocrysts is variable; in darker parts of the intrusion they are fully absent, whereas in the lighter rocks they constitute up to 70 vol%. Locally, a Rapakivi structure with plagioclase rims around microcline was recognised. The range in composition observed for the hornblende syenites can be explained by simple crystal-liquid fractionation of K-feldspar, plagioclase, clinopyroxene and amphibole from a mantle derived alkaline magma (Jung et al., 2005). The contacts between the hornblende syenite and the wall rock are sharp and usually parallel to bedding or to a foliation in the wall rock. Jogs, sharp corners and thin apophyses invading the wall rock parallel to the fabric are common, suggesting that the viscosity of the magma was low during intrusion (Figs. 3, 4; Petford et al.,

Table 1
Isotopic data from single grain zircon evaporation[a]

| Sample number | Zircon colour and morphology | Grain # | Mass scans[b] | Evaporation temp. in °C | Mean $^{207}Pb/^{206}Pb$ ratio[c] and 2-σ error | $^{207}Pb/^{206}Pb$ age and 2-σ error |
|---|---|---|---|---|---|---|
| NA 20/9 | Stubby to long prismatic, light brown, idiomorphic | 1 | 84 | 1589 | 0.057559 ± 45 | 513.0 ± 1.7 |
| | | 2 | 127 | 1587 | 0.057562 ± 29 | 513.1 ± 1.1 |
| | | 3 | 149 | 1598 | 0.057558 ± 25 | 513.0 ± 1.0 |
| | | 4 | 108 | 1588 | 0.057580 ± 38 | 513.8 ± 1.4 |
| Mean of four grains | | 1–4 | 468 | | 0.057564 ± 16 | [d]**513.2 ± 1.0** |

[a] Single zircons were analyzed on a Finnigan-MAT261 mass spectrometer at the Max-Planck-Institut für Chemie in Mainz, Germany, using the evaporation technique developed by Kober (1987). Analytical procedures are detailed in Kröner and Hegner (1998). Data acquisition was by magnetic peak switching using the electron multiplier. The calculated ages and uncertainties are based on the means of all ratios evaluated and their 2-σ (mean) errors.

[b] Number of $^{207}Pb/^{206}Pb$ ratios evaluated for age assessment.

[c] Observed mean ratio corrected for non-radiogenic lead. Errors based on uncertainties in counting statistics.

[d] Error of combined mean age (in bold) is based on reproducibility of internal standard with error in $^{207}Pb/^{206}Pb$ ratio of 0.000026 ($2\sigma$).

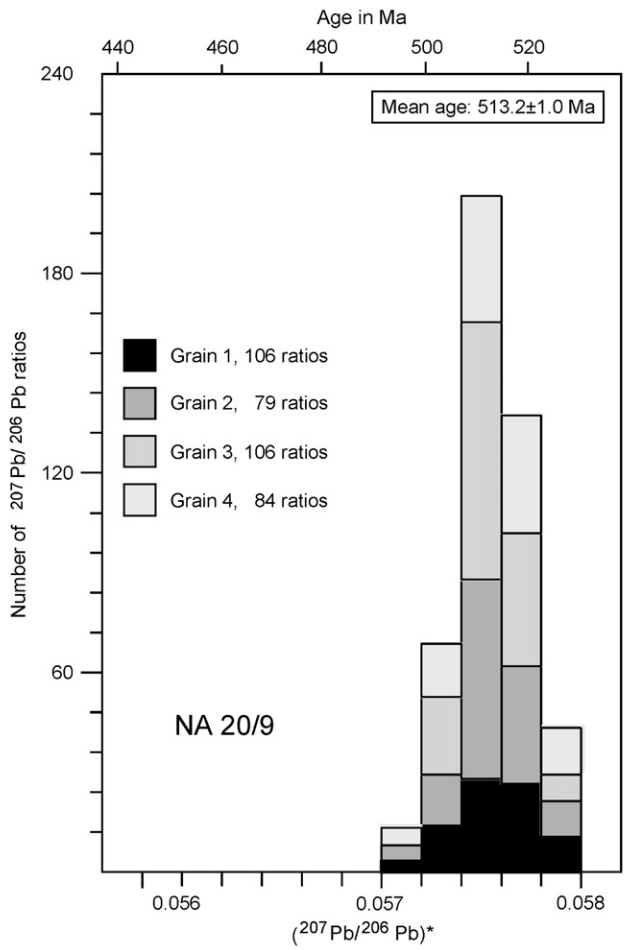


Fig. 5. Histogram showing distribution of radiogenic lead isotope ratios derived from evaporation of four zircon grains from biotite granite sample NA 20/9, Voetspoor Intrusion. The spectrum plotted has been integrated from 468 ratios. Mean age is given with 2-σ (mean) error.

1993, 2000). There are no chilled margins. The contacts are locally deformed, as discussed below.

K-feldspar phenocrysts have locally a strong preferred orientation interpreted to result from magmatic flow, because of the idiomorphic shape of the phenocrysts and lack of evidence for solid state deformation. The fabric is too weak to recognise a linear component although this may be present. The igneous fabric forms a more or less concentric pattern inside both syenite plutons and dips inwards (Fig. 2). In the more mafic part of the intrusion, hornblende can show a similar igneous fabric.

The hornblende syenite contains xenoliths and panels of metasediment up to 500 m in length and 100 m thick, mainly hornfelsed micaschist with some layers of marble (Fig. 4). These panels are oriented parallel to the igneous fabric, and have internal bedding and a foliation parallel to their elongate shape on the map. In most panels, the foliation is parallel or at a small angle to bedding. In the Doros pluton, tight to isoclinal folds were observed in bedding. Since there is no indication of strong ductile deformation in the panels after intrusion of the syenite, these folds are interpreted as predating intrusion. In the Doros pluton, the main foliation in the panels is locally overprinted by a steep second foliation, which could be an in-pluton foliation or S2. Because of these observations, and the different orientation of S2 and S3 around the plutons, the foliation in the panels is interpreted to be regional S1.

In the Voetspoor syenite the total volume of metasediment xenoliths is negligible, but in the central southern part of the Doros syenite up to 50% of the outcrops consist of metasediment panels (Fig. 4a). The syenite sheets in between the panels usually have a strong igneous fabric parallel to the contacts. Cathodoluminescence images of zircons from the hornblende syenite show that all contain inherited cores with small overgrowths (R. Schmitt, pers. comm.). This could imply that contamination with wall rock material has taken place.

At least three sets of dykes are associated with the hornblende syenite, invading both the main body of the pluton and the wall rock. In the pluton, their orientation is variable but in the wall rock the veins are normally parallel to bedding or S1, and at a small angle to the contacts of the pluton. The principal type of dykes are phenocryst-bearing light-coloured granite, pegmatite and bimodal dykes, consisting of a mafic core and a light rim.

### 5.2. Biotite granite

The south-western part of both plutons is made up of medium grained red light-coloured biotite granite that is intrusive in the hornblende syenite (Fig. 4). In the Voetspoor pluton, the granite is clearly distinct from the hornblende syenite by its higher quartz content (71–75 wt% against 47–60 wt% in the Voetspoor pluton) and lack of hornblende, although the difference in the Doros pluton is less pronounced. The biotite granite is more homogeneous and has fewer enclaves and xenoliths than the hornblende syenite, but xenoliths of both the metaturbidite wall rock and the hornblende syenite occur. The contact with the hornblende syenite and the metasediment enclaves is sharp. Angular contacts and thin dykes invading the wall rock indicate that the magma had a low viscosity (Petford et al., 1993, 2000). Chilled margins are lacking. According to Jung et al. (2005), the biotite granite may be fractionated from syenite, but is unlikely to have formed from the syenites in a closed system crystal fractionation scenario because of its distinct isotope composition.

The northern part of the Voetspoor hornblende syenite is cut by pink granite dykes of up to 40 m thick, mainly oriented NW–SE to E–W, approximately orthogonal to the longest dimension of the granite body on the map. Some of the veins are clearly arch-shaped (Fig. 4b). Intrusive contacts are sharp and vertical or dipping steeply inward, but since the vertical outcrop relief is less than 60 m, it is uncertain what the larger scale geometry of the contact is. These major veins are inferred to be of the same age as the biotite granite in the south-west because of their composition and relation with deformation in the wall rocks. Minor veins of biotite granite and associated pegmatite and aplite occur throughout the pluton and in the wall rocks, where they are mostly parallel to the foliation.

## 6. Deformation adjacent to the plutons

Field observations and thin section studies indicate that neither the hornblende syenite nor the biotite granite in both plutons underwent significant ductile deformation over most of the pluton volume. In the centre of both plutons, ductile strain does not exceed 10% shortening, manifested as minor undulose extinction in quartz (fabric Type 1; Hibbart, 1986; Hutton, 1988; Paterson et al., 1989; Cruden et al., 1999a). In the Doros pluton, refolding of some sediment panels occurs by open folds, some with a foliation, and this is attributed to D3. The wall rocks, however, were relatively strongly deformed during and after intrusion, as seen from the deformation of veins associated with the main intrusions. This contrast between a relatively rigid intrusion and ductile wall rocks may be due to the high percentage of feldspar and hornblende in the plutons, which would be relatively rigid at the deformation temperatures (<500 °C) manifest during D3 in the area.

The regional tectonic pattern changes considerably when approaching the two investigated plutons (Figs. 2, 3). S3 and vertical axial planes of D3 folds are deflected around the plutons, creating a D3-strain shadow at the SW corner of both bodies (Fig. 2). D3 deformation is strong between the plutons, as inferred by the tightness of D3 folds and the development level of S3. D3 folds between the two plutons are the result of coaxial NW–SE shortening, since the asymmetry of D1 quartz-filled boudin necks is symmetrically distributed around local D3 folds. However, D3 was a phase of non-coaxial flow on the west side of the Doros pluton; shear sense indicators of D3 age are all sinistral here. Moreover, SW of the Doros pluton S3 wraps around the pluton into a NW–SE trending orientation that is unique in the Lower Ugab Domain and is transected by a NS trending vertical foliation of the same style and metamorphic grade, which we labelled S3b (Fig. 2). This foliation is only present here and can be explained by rotation of S3 into the shortening field of bulk non-coaxial flow (Fig. 2). The strong D3 fabric on this side of the pluton grades to the SW into the D3-Bushman fold, also with evidence for sinistral non-coaxial flow (Passchier et al., 2002).

The effect of the plutons on D1–D2 structures is less clear because of the D3 overprint. D1 folds become E-vergent or vertical close to the plutons (Fig. 2). Along the south side of the Voetspoor pluton, a D1 syncline–anticline pair outlined in the rocks of the Gemsbok River Formation is cut by the pluton and folds have an unusual open geometry (Figs. 1, 4b). The interlimb angle of this structure decreases away from the pluton to the west and south, while S1 and the trend of minor folds are axial planar to the large fold, and run up to the contact without deflection (Figs. 2, 4b). On the NW side of the Voetspoor pluton, folds in the Gemsbok River Formation form a refolded pattern (referred to as the "heron-structure": Figs. 4b, 6) where isoclinal D1 structures with NE fold vergence, are refolded by upright NE–SW trending D3 folds (Figs. 2, 3, 6). The three boomerang-shaped structures in the position of the "wings" of the heron (Figs. 4b, 6) are D3-refolded tight NE-vergent D1 anticlines. The D1-fold that forms the neck of the heron structure spreads out towards

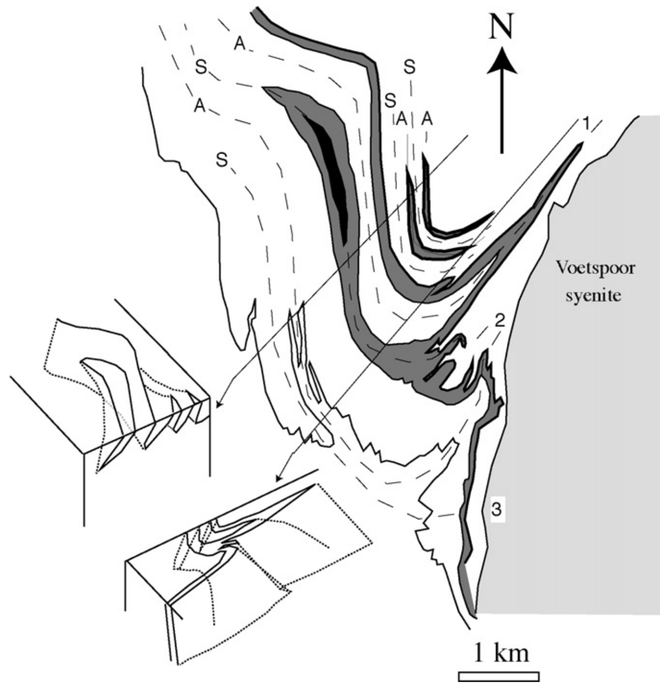


Fig. 6. Detailed map of the "heron-structure" NW of the Voetspoor pluton. The block diagrams show the structure of folded bedding in three dimensions. S – D1 syncline. A – D1 anticline. Numbers refer to parts of the fold structure discussed in the text.

the Voetspoor pluton into a star-shaped group of isoclinal anticlines and synclines (2 in Fig. 6) increasing the wavelength of the major fold structure from 1 km to more than 2 km. The plunge of D1 fold axes steepens from subhorizontal in the neck of the heron structure to vertical in this star-shaped cluster over a distance of a few hundred metres (Figs. 2, 3, 6). From the star-shaped cluster of isoclinal folds, bedding trends parallel to the pluton contact in a vertical sheet with isolated asymmetric isoclinal folds with vertical axes. In the north, one of these connects to a single isoclinal anticline with NE plunging axis (1 in Fig. 6), from which the bedding trends NW with shallowing dip. To the south (3 in Fig. 6), bedding runs parallel to the contact until it is transected by the pluton. This creates a fan-shape of D1 folds and S1 towards the pluton (Figs. 2, 3). If D3 shortening is removed from the "heron structure", its geometry is similar to that of the open D1-folds in the SW of the Voetspoor pluton.

All around the Voetspoor pluton and on the south side of the Doros pluton D1 folds, which on a regional scale have subhorizontal fold axes, show steepening of the fold axes close to the plutons in a narrow rim monocline (Fig. 3); over a distance of a few hundred metres, fold axes change from subhorizontal to subvertical. The sense of deflection indicates a relative downward motion of the pluton with respect to the wall rock. This deflection causes a map pattern with D1 fold closures close to the contact of the pluton (Fig. 4). Unfolding the D1 fold structures would place the surface of the pluton at least several hundred metres above its present level.

A subvertical mylonitic ductile shear zone with a width of 100–400 m lies along the entire NE rim of the hornblende syenite of the Voetspoor pluton (Fig. 3). This mylonite zone mostly affects the hornfelsed wall rocks and granite veins in it, but also a strip of hornblende syenite up to 100 m from the contact. Deformation is clearly solid-state, with strong deformation and recrystallisation of quartz, and ductile deformation of feldspar (Type 4 fabric; Hibbart, 1986; Hutton, 1988; Paterson et al., 1989; Cruden et al., 1999a). This mylonite zone may have even extended further south, since the younger biotite granite intruded along the contact, and may have erased traces of the mylonite zone. The foliation in the mylonite zone is parallel to the contact, while a strong stretching lineation is steeply SW plunging all around the pluton, both in the wall rock and in the syenite (Fig. 3). Boudinaged dykes form shear band boudins which, together with local shear band cleavage and sigma-type mantled porphyroclasts of feldspar, indicate relative downward motion of the intrusion with respect to the wall rock. The shear zone contains deformed dykes of phenocryst-bearing hornblende syenite, but is cut by bimodal dykes and by red granite, aplite and pegmatite dykes, which seem to be related to the biotite granite.

## 7. Relative age of intrusion and deformation

### *7.1. Hornblende syenite*

It is difficult to date the intrusion of syenite and granite in the plutons with respect to deformation in the wall rock. The main reason is that contact metamorphism destroyed delicate foliation overprint structures in the wall rock, while a syn-intrusive shear zone affected the Voetspoor pluton along most of the contact with the wall rock. Intrusive relations with veins can be established further away from the main body of the intrusions, but such veins are usually fine grained and of slightly different composition as the main lithotypes. It is therefore usually impossible to attribute a vein unambiguously to one of the intrusions. In all, the most reliable means of establishing relative age of intrusion and deformation in the wall rock is the large-scale geometry of deformation structures further away from the intrusions. The following relations have been observed in the field mainly around the well-exposed Voetspoor pluton:

1. A major D1 fold is cut at almost right angles by the main hornblende syenite in the SE.
2. Dykes of phenocryst-bearing hornblende syenite cut D1 folds.
3. A marginal shear zone affects the north side of the hornblende syenite in the Voetspoor pluton, with steep lineations and relative downward motion of the syenite against the wall rocks.
4. Undeformed bimodal dykes cut the contact shear zone along the edge of the syenite in the Voetspoor pluton. The bimodal dykes probably belong to the hornblende syenite; similar dark phases were never found in dykes associated with the biotite granite.
5. D1 structures form large strain shadow geometries on the W and SE side of the Voetspoor pluton; relatively open structures close to the pluton grade into closed folds further away.
6. Along the highly indented NW contact of the syenite a steep E–W trending S1 in the country rocks is cut approximately orthogonally by the equally steep N–S syenite contact, showing that D1 deformation largely precedes the syenite intrusion.
7. Along the NE rim of the Voetspoor pluton, many irregular D2 folds become tighter and more regular towards the pluton, with axial planes dipping at a shallow angle.
8. In metasedimentary rock panels in the Voetspoor and Doros plutons S1 is usually parallel to bedding that may show isoclinal D1 folds.
9. Minor folds with axial plane foliation deform the main foliation interpreted as S1 in metasedimentary rock panels of the Doros pluton.
10. Porphyroblasts associated with contact metamorphism of the intrusions are common in metapelitic hornfelses. These are mostly retrogressed to mica–quartz–feldspar aggregates, but from their shape and some remnants can be deduced to represent cordierite and andalusite. In a few cases, andalusite preserved deviation patterns of S1 that indicate porphyroblast growth late syn-D1 (Fig. 7).

From these observations we conclude that the hornblende syenite intruded late during D1 to early D2 in both plutons.

### *7.2. Biotite granite*

The biotite granite in the Voetspoor pluton is isotropic with few xenoliths. There is no shear zone or other deformation in the steep contact zone with the host rock. The biotite granite cuts through bedding and D1 structures and postdates the

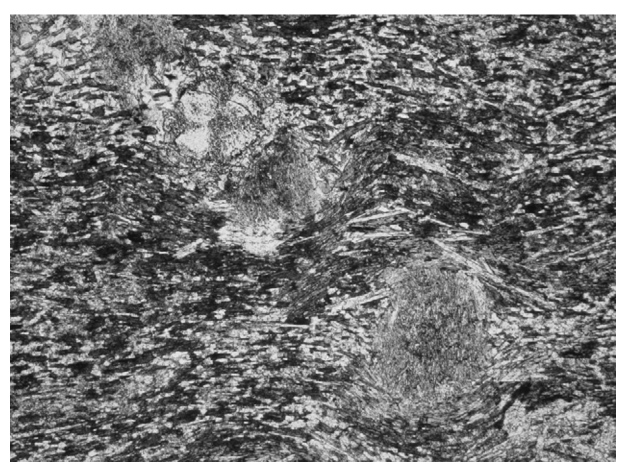

Fig. 7. Photomicrograph of porphyroblasts of andalusite in the Amis River Formation. Note andalusite crystal with chiastolitic inclusions in the upper left quarter; other porphyroblasts are substituted by quartz mica aggregates. Horizontal foliation S1 is included in some of the porphyroblasts and bends around them, indicating that the porphyroblasts grew late during D1. Hence, the Voetspoor Pluton must have intruded during D1 at this site. Width of view 5 mm. Plane polarised light. Location 100 m W of the central western contact of the Voetspoor pluton.

hornblende syenite and its marginal shear zone as indicated by crosscutting relations. The relationship with D3 structures is more complex. The biotite granite cuts through a major E–W trending D3 synform and apophyses of biotite granite and late light-coloured granite and pegmatite veins cut small D3 folds in the wall rock without being deformed. On the other hand, some biotite granite veins are folded by D3, and S3 can be seen to bend sharply around the pluton and the southern biotite granite of the Voetspoor pluton (Figs. 2, 4). This can be due to forceful intrusion of the biotite granite, or by ductile deformation of the wall rocks around the solidified pluton during D3.

## 8. Granite intrusion, general considerations

Commonly, elliptical to circular granite intrusions have been attributed to a diapiric origin and subsequent ballooning (Marsh, 1982; Bateman, 1984; Cruden, 1990; Paterson et al., 1996). This is mostly due to the fact that intrusion contacts are steep to vertical in many plutons (Cruden, 1998; Cruden et al., 1999a), with concentric patterns of magmatic foliation (Brown and McClelland, 2000) and the fact that the floor of plutons is rarely exposed, suggesting a dome shape. However, strain in the wall rocks of elliptical plutons is usually insufficient to allow an interpretation of ductile diapirism and ballooning, especially in high-level intrusions (Cruden, 1998; Benn et al., 1999; Brown and McClelland, 2000). Also, there is increasing evidence that the 3D shape of most plutons with circular and elliptic map-pattern is not spherical or balloon-shaped, but rather tabular or wedge-shaped. Some intrusions are exposed as tabular, subhorizontal sheets (Coleman et al., 1995; Scaillet et al., 1995; Sisson et al., 1996; Grocott et al., 1999) or such a shape can be inferred from magnetic susceptibility fabrics (Cruden and Launeau, 1994; Aranguren et al., 1997; Benn et al., 1999). Other plutons with a circular or elliptical shape on the map may have a bath-tub shape (Hamilton, 1988; Brown and McClelland, 2000) or more commonly a wedge-shape in 3D with floors dipping shallowly towards one or more conduits (McCaffrey and Petford, 1997; Cruden, 1998; Dehls et al., 1998; Benn et al., 1999; Grocott et al., 1999; Petford et al., 2000; Cruden and McCaffrey, 2001).

Characteristic for many plutons with inferred tabular or wedge-shape is down-folding of older wall-rock structures towards the contacts of the pluton in a rim monocline (Hamilton and Myers, 1967; Bridgewater et al., 1974; Brun et al., 1990; Grocott et al., 1999). Shear zones with down dip stretching lineations in the margin of these plutons, which have been active during emplacement, commonly indicate displacement of the granite downward rather than upward with respect to the wall rock (Tobisch and Cruden, 1995; Glazner and Miller, 1997; Dehls et al., 1998; Sylvester, 1998; Cruden et al., 1999b).

Based on a large number of described examples, the development of tabular and wedge-shaped granite bodies is presently imagined as follows (e.g. Grocott et al., 1999; Petford et al., 2000; Cruden and McCaffrey, 2001). Granitic magma may have a viscosity low enough to be transported to higher crustal levels along dykes (Dingwell et al., 1993; Pitcher, 1993; Clemens and Petford, 1999) and this is probably how tabular and wedge-shaped plutons are filled from below (Clemens and Mawer, 1992; Clemens et al., 1997; Petford et al., 2000). At a certain level, either depth related or due to some suitable lithology, intrusion of a sill-like body starts, which is then increasing in thickness and may grow laterally (Corry, 1988; Tobisch and Cruden, 1995; Cruden, 1998). This growth may occur by addition of km-scale tabular sheet-like intrusions by several magma pulses (Benn et al., 1999; Brown and McClelland, 2000; Petford et al., 2000). Commonly, the oldest sheets are at the bottom and the youngest at the top in such sequences (Cruden et al., 1999b; Brown and McClelland, 2000). Vertical inflation of individual intrusion sheets is common (Brown and McClelland, 2000; Petford et al., 2000). This leads to uplift of the wall rock to form a laccolith, depression of the floor to form a lopolith, or both. Laccolith formation is due to bending of the roof rocks (Dixon and Simpson, 1987; Jackson and Pollard, 1988; Roman-Berdiel et al., 1995; Morgan et al., 1998) or uplift along faults (Corry, 1988). Laccoliths dominantly form at shallow crustal conditions where surface uplift and erosion is easy, but can also form at greater depth if magma pressure is high (Petraske et al., 1978; Corry, 1988; Roman-Berdiel et al., 1995; Benn et al., 1999, 2000).

Most granites intruded at depth greater than 5 km have a lopolith rather than a laccolith form, testified by gravimetric data (Vigneresse, 1995), and in-bending of wall rocks (Myers, 1975; Paterson et al., 1996; Cruden, 1998). This may have happened by depression of the floor during intrusion: a magma reservoir lower in the crust may collapse when granite is transported upwards and induce sagging of the intermediate crust, creating space for the intruding granite (Brown and Walker, 1993; Cruden, 1998). Cruden (1998) showed that if sagging of the floor is by radial simple shear, larger plutons require slower ductile accommodating strain rates than smaller ones. Strain rates needed to accommodate granite intrusion at a speed sufficient to avoid solidification in the channel are not exceeding those proposed for ductile crustal deformation (Pfiffner and Ramsay, 1982). Accumulated crystalline material at the bottom of the magma chamber may have sunk together with the wall rock to form a layered basal part of the intrusion (Wiebe and Collins, 1998).

## 9. Intrusion mechanisms in the Voetspoor and Doros plutons

### 9.1. Ballooning as a possible intrusion mechanism

At first sight, diapiric intrusion and ballooning seems an attractive mechanism for intrusion of the hornblende syenite in the Voetspoor and Doros plutons. Metasedimentary rock panels and igneous foliation are arranged in circular patterns parallel to the edges of the hornblende syenite. At the southern rim of the Voetspoor pluton, D1 folds open towards the pluton and the shape of the “heron-structure” along the west side of the Voetspoor pluton is also suggestive of ballooning. One could consider the option that originally tight D1 folds further away from the pluton were opened by horizontal stretching of the wall rock around an inflating pluton to form the present geometry. Judging from the geometry of the folds around the Voetspoor pluton, the wall rock would have to be stretched by 200–300% parallel to the contact to create the present arrangement. However, intrusive relations indicate that the syenite intruded after S1 had developed. Ballooning of the pluton would have led to folding of this foliation and development of a foliation parallel to the contact, in addition to considerable vertical extension of the wall rock in the contact zone. It is difficult to conceive how this could have led to the gradually tapering geometry of the open folds in the SW of the pluton, to relative descent of the granite with respect to the wall rocks, and development of a narrow shear zone rather than a broad zone of intense deformation. The only foliation that runs parallel to the plutons and wraps around them to some extent is S3. This foliation postdates the intrusive phase of the hornblende syenite. We therefore favour a model whereby the pluton intruded into a relatively open fold, which then closed away from the pluton, while parts of this fold south of the pluton were protected in a large scale strain shadow. In the case of the Voetspoor pluton, D1 folds were still relatively open during intrusion, but the Doros pluton may have intruded later, after folds had tightened further; this is attributed to the lack of open D1 folds close to the Doros pluton and the parallel or subparallel orientation of S1 and bedding in metasedimentary rock panels in that pluton (Fig. 4a). D2 then further tightened and rotated D1 folds, especially around the Voetspoor pluton, and locally formed a new foliation and D2 folds where earlier structures had rotated into the shortening field. Therefore, some inflation of the pluton may have taken place during intrusion of the syenite, but this must have been of minor importance.

In the case of the biotite granite on the south side of the plutons, one could argue that the deviation of S3 around the pluton is the result of forceful intrusion postdating D3, but this would have led to associated stretching in the wall rocks. However, this does not explain the presence of similar S3 further away from the pluton. More likely, regional scale D3 deformation postdating intrusion created the D3 folds and their deflection around the biotite granite-syenite plutons.

The biotite granite occurs in an unusual arrangement, in a crescent-moon shaped main body along the south side, and as E–W to NW–SE trending dykes in the central and northern part of both plutons. This arrangement can be explained by syn-D3 intrusion. If D3 is indeed due to local NW–SE compression, as indicated by the geometry of S3, the arrangement of the biotite granite is that of a void-filling alongside a rigid object, in this case the older hornblende syenite. The curved dykes can be interpreted as tension gashes that fit the inferred orientation of the D3 stress field.

### *9.2. Intrusion mechanism of the hornblende syenite*

In both the Voetspoor and Doros plutons, the composition of the hornblende syenite shows a gradient, which can be attributed to crystal-liquid fractionation. Also, there are indications that the darker parts of the intrusion locally crosscut the more leucocratic parts, although the opposite was also observed. In the Voetspoor pluton sudden changes in dip of the igneous fabric (indicated as dotted lines in Fig. 4b), coincide with panels of metasedimentary rocks oriented parallel to the igneous fabric, and with asperities in the contact with the wall rock (Fig. 4b). Although no intrusive relationship has been observed on a small scale, these sudden breaks in the igneous fabric are interpreted as intrusive contacts between separate intrusive sheets of the hornblende syenite in the pluton. It is uncertain if the older sheets were completely solidified when the younger ones intruded. In the Voetspoor pluton, at least three such sheets can be recognised. Apparently, hornblende syenite in the Voetspoor pluton did not intrude in a single event or as a simple diapir-like mass of magma, but rather as a series of repeatedly intruding planar bodies. In the Doros pluton, the outcrop quality is insufficient to distinguish sudden changes in igneous fabric orientation, but the presence of numerous parallel concentric panels of metasediment suggests a similar situation (Fig. 4a).

Presently, most of the metasediment panels and included S1 and bedding in both the Voetspoor and Doros plutons are dipping towards the centre of the intrusion, parallel to the igneous fabric. In the Voetspoor pluton, most dips are over 60°, but in the Doros pluton, dips of the igneous fabric and metasedimentary rock panels are relatively gentle on the outside (20–30°), and increase in steepness inwards (Fig. 3). This orientation of the panels must have formed during intrusion, since the syenite itself is hardly deformed in the solid state. The orientation of S1 and bedding away from the intrusions is mostly steeply dipping in tight folds. This would suggest that the sediment panels were rotated from steep to gentle dips after they were incorporated in the intrusion.

Relative downward movement of an intruded granite with respect to its wall rock as observed for the Voetspoor pluton, with a rim monocline or shear zone has been observed adjacent to other plutons as well (Glazner and Miller, 1997; Dehls et al., 1998; Sylvester, 1998; Grocott et al., 1999; Cruden et al., 1999b). This phenomenon was explained by Glazner and Miller (1997) as an effect of sinking of a pluton after solidification because of a higher density than the wall rocks, and by other authors to sinking of the floor of the intrusion (Cruden et al., 1999b; Grocott et al., 1999; Petford et al., 2000). Sinking of the intrusions due to density contrast is unlikely, since the density contrast with the wall rock is small and metamorphic grade is low. Other possible mechanisms not mentioned by these authors are the effects of ballooning, or later deformation in the wall rock, which is folded around and over the existing rigid pluton. An often neglected problem of intrusion by floor-descent is that this can only be relevant if a magma chamber is present at limited depth below the pluton. Partial melting in the lower crust or upper mantle directly below the pluton may not be able to create enough volume for the floor of an intrusion to descend significantly. The strong composition gradient of the syenite intrusion, probably due to differentiation, favours the presence of such a magma chamber below the plutons. With the presently available data, it is difficult to decide if floor-descent is the only or dominant mechanism responsible for emplacement of the investigated plutons, or if other mechanisms have played a significant role as well. As discussed in Section 9.1, ballooning does not seem to be the main mechanism of emplacement. The rim monoclines of the plutons are very narrow, and dykes cutting the rim mylonite zone seem to indicate that intrusion was still in progress during the downward motion. Late ductile deformation in the wall rock is insufficient to explain the structures; a wider deformation aureole would be expected, and no dominant displacement direction, since the pluton contacts are vertical. Most likely, therefore, is that both the rim monoclines and the marginal shear zone are mainly due to sinking of the floor of the pluton during intrusion.

### 9.3. Development of the hornblende syenite intrusions

Based on the geometrical data presented above, the following scenario can be envisaged for intrusion of the hornblende syenite in the Doros pluton (Fig. 8).

A differentiated magma-chamber existed below the future site of the plutons, where pyroxene, hornblende and K-feldspar phenocrysts had segregated to the bottom and top of the chamber, respectively. First intrusion was sill-like into a gently folded sequence of metaturbidites transecting the level of the Gemsbok River formation (Fig. 8a). Unlike other models where a single planar sill is envisaged (Brown and McClelland, 2000; Cruden et al., 1999a,b; Grocott et al., 1999), the steep S1 foliation and bedding in D1 fold limbs in the metaturbidites may have caused sill-stepping and isolation of elongate wall rock panels close to where the feeder dyke joins the sill feeder dyke (Fig. 8a,b). Subsequent pulses of intrusion may have penetrated the initial sill and may have intruded along the roof, while the bottom of the pluton sagged downwards into the emptying magma chamber (Fig. 8a,b; cf. Cruden et al., 1999a,b; Grocott et al., 1999; Brown and McClelland, 2000). The isolated metasedimentary rock panels may have been pushed into the outer parts of the intrusion suffering boudinage, bending and rotation in the process to reach their present position and orientation. Subsequent intrusion pulses were in the upper reaches of the pluton, while the lower parts continued

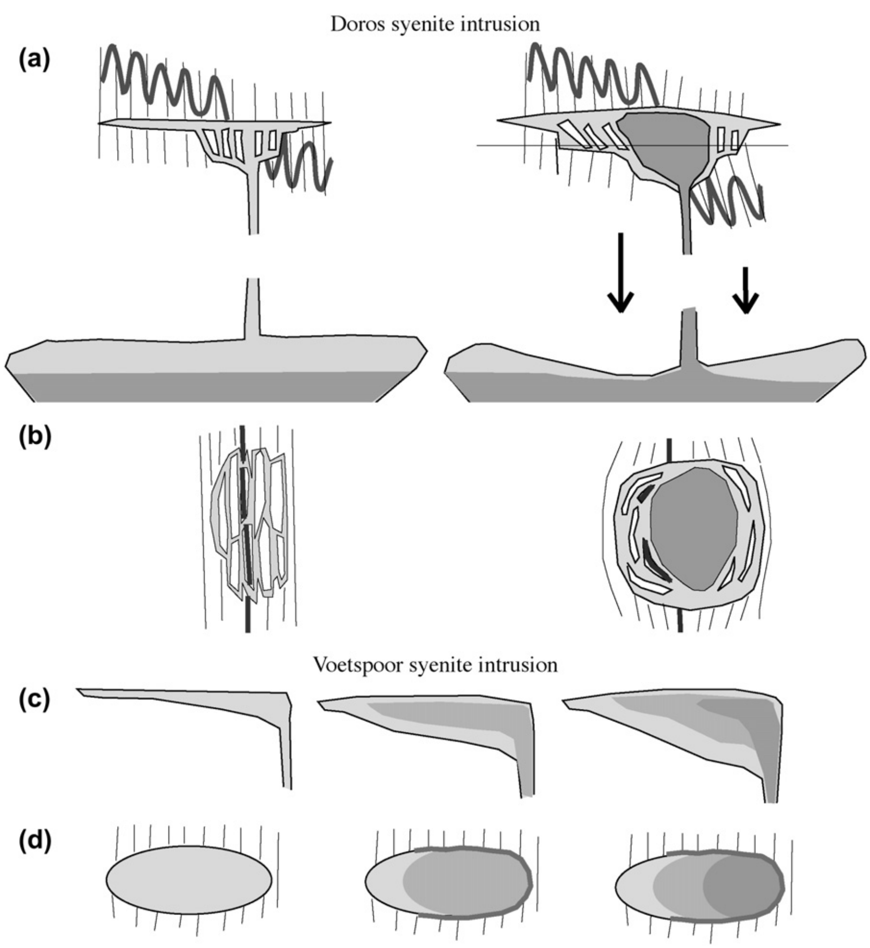


Fig. 8. Schematic presentation of the intrusion of hornblende syenite in the Doros (a, b) and Voetspoor (c, d) plutons. (a, c) SW-NE cross-sections at several stages during development of the plutons; (b, d) planar cross-sections. In the Doros pluton (a) intrusion detaches many panels of metasedimentary rock parallel to foliation, which are subsequently transported to the outer parts of the developing pluton. Space for the intruding magma is made mainly by descent of the floor into a composition-layered magma chamber at depth. In the Voetspoor pluton (c, d) development is similar but descent and intrusion are asymmetric. The rotation of the plutons in D1 is omitted for simplicity.

to sag and rotate downwards. It is uncertain, however, to what extent the roof of the intrusion was raised. From theoretical and experimental considerations, both elevation and sagging are possible at the depth at which the pluton intruded, depending mainly on magma pressure (Petraske et al., 1978; Benn et al., 2000). The final geometry is a wedge-shaped structure with increasing steepness of igneous fabric and metasedimentary rock panels inwards, and also increasing hornblende content and decreasing K-feldspar phenocryst content inwards (Fig. 8a,b). The floor of the pluton may have descended along faults oblique to bedding. The feeder dyke or vent may be located below the central part of the pluton.

For the Voetspoor pluton, the situation may have been slightly different. The intrusion may be slightly older than the Doros pluton, cutting more open D1 folds. The youngest and most mafic rocks are in the NE corner of the elliptical Voetspoor pluton, and not in the centre as in the Doros pluton. The marginal shear zone lies in the NE of the Voetspoor pluton, and this suggests the floor of the intrusion may have detached from the wall rocks along the marginal shear zone and sunk deepest below the NE part of the intrusion, where apparently the feeder vent was located (Fig. 8c,d).

### 9.4. Location of the hornblende syenite

Syenites with an inferred mantle source, like the Voetspoor and Doros plutons are commonly associated with steep shear zones that act as channels for the rising magma (Seth et al., 2000; Jung et al., 2005). However, such shear zones have not been recognised in the Lower Ugab and Goantagab domains despite detailed mapping (Figs. 1–4). More likely is that the intrusion is associated with D1 transpression, while the dominance of steep foliations may also have played a role. Tension faults could have opened oblique to D1 fold axial planes parallel to the instantaneous shortening direction and feeder dykes may have formed either along S1 in between such faults, or at the intersection of S1 and such faults (Fig. 8a,b). Plutons similar to the Voetspoor and Doros pluton are located along the southern and eastern margins of the Ugab Domain. A brief reconnaissance has shown that these have a similar bimodal nature and relation with regional deformation as the two plutons described in this paper. Domains with porphyroblasts similar to those in the contact aureoles of the hornblende syenite are found throughout the Lower Ugab Domain, and the local metamorphism is clearly at least in part a low pressure contact metamorphism with narrow anticlockwise P–T path (Goscombe et al., 2004). It is therefore likely that plutons similar to the Voetspoor and Doros plutons intruded throughout the area, but are now buried or eroded away in the central Lower Ugab Domain.

### 9.5. Intrusion of the biotite granite

The biotite granite is interpreted to intrude late syn-D3 on the SW side of the Doros and Voetspoor plutons, and in dykes inside the hornblende syenite. No evidence for ballooning or structures indicative of forced intrusion were encountered. Stoping is unlikely as a mechanism since xenoliths are rare in the biotite granite. The syn-D3 age of the intrusion makes it likely that space for the half-moon shaped biotite granite bodies along the SW side of both plutons was made by detachment of the wall rock from the hornblende syenite in the direction of D3 extension. This is remarkably similar to formation of small strain shadows around rigid objects that are common in many deformed rocks (Passchier and Trouw, 2005). The intrusion of dykes also fits a setting with NW–SE compression and NE–SW extension in the hornblende granite. Because of the subdued topography, we presently do not know the vertical arrangement of the syenite and the biotite granite. If the crescent-shaped dykes are feeder dykes, the biotite granite may have occupied the roof of the pluton. This would fit the observation that the Goantagab domain underlies the Ugab domain structurally, so that the exclusive presence of the biotite granite to the south side of both plutons can be attributed to tilting, enhanced uplift and erosion in the NE.

The similar distribution of biotite granite in the Doros and Voetspoor plutons could imply that they are somehow connected at depth; both plutons are separated by a major D3 syncline.

## 10. Rotation of the plutons

It is not unusual that granite plutons are affected by horizontal ductile shearing (McCaffrey, 1992; Vigneresse, 1995; Aranguren et al., 1997). Many plutons have an asymmetric shape that can be attributed to deformation of the entire pluton by intrusion into a major shear zone, deformation related to a later shear zone, or rotation of the entire pluton (Djouadi et al., 1997; Gleizes et al., 1997). The Voetspoor and Doros plutons show evidence for rotation.

Bedding and S1 in included pelitic metasedimentary rock fragments in the south of the hornblende-syenite of the Voetspoor pluton trend E–W (Figs. 2, 3, 4b), while the regional trend is NNE–SSW (Fig. 1). There are also hook-shaped folds of bedding on the south and SE side of the pluton, and a NW–SE trending major D1 syncline on the SW side (Figs. 1, 3, 4b). Since the aberrant D1 orientation is clearly associated with the presence of the pluton, we think that it may be due to *anticlockwise rotation* of the entire Voetspoor pluton with respect to the bulk S1 orientation over 45–60° (Fig. 9). The orientation of bedding and S1 in the ‘‘heron structure’’ in the NW, the open syncline in the SW, and the very tight anti- and synclines of the Gemsbok River Formation in the east can all be satisfactorily explained by such rotation (Figs. 4b, 9).

The situation in the Doros pluton is less clear because it is partly covered by younger strata. However, the pattern of inclusions in the pluton and deflection of bedding along the southern rim (Figs. 1, 4a) also support some anticlockwise rotation for this pluton. The tight D3 folds between the Voetspoor and the Doros plutons are interpreted as an effect of compression of material between the two plutons as they were brought closer together during D3 (Fig. 2).

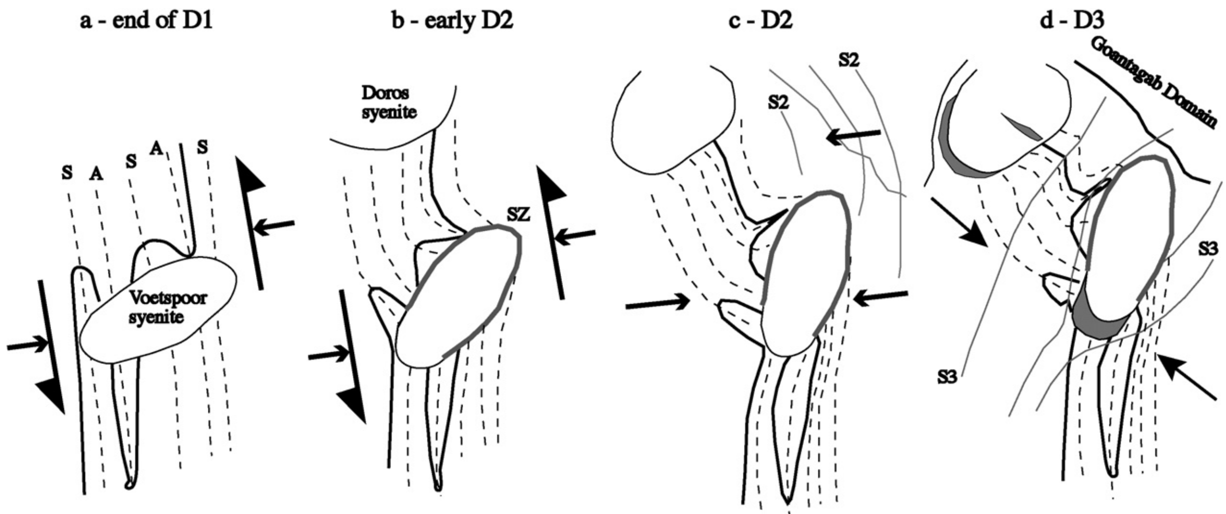


Fig. 9. Schematic presentation of the development of the Voetspoor and Doros plutons in map view. This series of images shows how after initial intrusion of the Voetspoor pluton in open D1 anticlines (A) and synclines (S), further D1 and D2 shortening associated with a sinistral shear component caused tightening of D1 folds and rotation of the pluton. The Doros pluton may have intruded at this stage. The shear zone (SZ) along the northern side of the Voetspoor pluton is due to intrusion and may be enhanced by E–W shortening. D1 folds to the west of the Voetspoor pluton are tightened into an asymmetric strain shadow due to ongoing D1 shortening and rotation of the plutons. During D3, the biotite granite intruded (dark grey) and D1 folds were slightly refolded.

The arrangement of S3 around the plutons is clearly asymmetric, and this could be used as an argument to attribute rotation to D3. However, on a regional scale D3 is a phase of coaxial N–S shortening. It is more likely that the rotation is of D1–D2 age. The arrangement of structures around the plutons mostly predates D3, notably the steepening of fold axes, and the marginal shear zone of the hornblende syenite. Moreover, N of the Voetspoor pluton D2 structures are arranged in W-vergent folds that correspond well with the rotational motion (Figs. 2, 9). Kinematically, rotation of the plutons is possible if they are relatively rigid bodies with a wedge or disk shape as indicated by our field data.

## 11. Conclusions

The Doros and Voetspoor plutons are composed of two compositional phases of intrusion, hornblende syenite and biotite granite, each of which intruded during a period of tectonic activity. The syenite apparently intruded in sill-like bodies, mainly by displacing the footwall downwards, disrupting existing D1 folds and foliation in the metaturbidites and including large panels of the metasedimentary wall rock, probably by several pulses of intrusion. Since the intrusion occurred during ductile transpression in the metaturbidites, strain shadows and deflection of structures developed similar to those around porphyroblasts on a microscopic scale. Another interesting effect is that the older syenite intrusion seems to have rotated considerably after consolidation in response to ductile transpression. A late phase of biotite granite intrusion probably used the same channels and intruded along tensional fractures in the syenite, and along the southern contacts in strain shadows related to D3. The geometry of intrusion of the biotite granite strongly resembles veins that open alongside rigid objects in microscopic fringe structures.

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