

STRUCTURAL ANALYSIS OF THE LOWER WITWATERSRAND SUPERGROUP IN THE VREDEFORT DOME, SOUTH AFRICA

A Thesis submitted to the Faculty of Science, University of the Witwatersrand,
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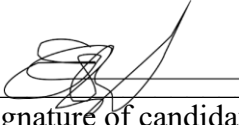


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DECLARATION

I declare that this thesis is my own, unaided work. It is being submitted for the Degree of Doctor of Philosophy at the University of the Witwatersrand, Johannesburg. It has not been submitted before for any degree or examination at any other University.


(Signature of candidate)

06 day of May, 2025 in Johannesburg

ABSTRACT

Hypervelocity impact craters are sites of exceptionally large, focused, energy release that results in extreme stresses, finite strains and strain rates in target crust. In particular, large impact structures are characterised by complex internal morphologies involving second- and third-order structures. The formation of second-order features, such as central uplifts are successfully reproduced by numerical models such as iSALE, which assume gross hydrodynamic behaviour of the crust, even if the reason for such behaviour is not fully understood. However, within central uplifts, field and petrographic studies show that strain is highly heterogeneous at all scales. This study examines brittle and ductile features in the 2.02 Ga Vredefort Dome, the largest and most deeply eroded central uplift on Earth. It focuses particularly on three areas close to the contact between the massive crystalline core and the basal parts of layered supracrustal sequence located between 20 km and 22 km from the centre of the Dome, where the lithologically-heterogeneous lower West Rand Group of the Witwatersrand Supergroup is well exposed. Geometric analysis of shatter cones, branching (horsetail) fractures and a near-orthogonal pair of cm-spaced, so-called shatter cleavages shows some preferred centrosymmetry linking them to the shock wave within the first few seconds of impact. A 3D analysis of two large faults with individual slips of 400 - 500 m provides the first evidence of a conjugate pair of originally tangentially-striking normal faults attributed to collapse of the transient crater wall approximately 1 minute after impact. Structural analysis of large pseudotachylyte dykes and vein networks establishes that they formed by frictional melting related to these faults. Brittle-ductile folding with steep tangentially-striking axial planes and shallow-plunging hinges and wavelength of ~ 800 m and amplitude of ~ 100 m is linked to central uplift collapse between 3 and 4 minutes after impact, and is distinct from previously recognised radially-oriented constrictional folds that formed ~ 70 -90 s after impact. This was followed by late-stage concentric and radial-conjugate normal faults with centre-down slip attributed to central uplift collapse, which also contain significant pseudotachylyte volumes. The final stage involves the intrusion of the granophyre impact melt intrusion into tensional fractures. The evidence of multiple stages of faulting emphasises the need for more careful analysis of the fault patterns in the collar of the Dome. Although no bilateral symmetry was detected in fault and bedding strike patterns shown in published maps, future work establishing the 3D geometry and relative chronology of the faults is recommended before testing this aspect again.

The cumulative field evidence indicates that second-order structures in the collar of the Vredefort Dome preserve a complex, multistage record of evolving strain that can be resolved into initial transient crater wall collapse, intermediate convergent and upward flow (constriction) related to central uplift rise and later divergent and downward flow (flattening) linked to its collapse, and that can be matched to the general stress-strain predictions of iSALE numerical modelling. This suggests that applying conventional structural mapping techniques to central uplifts in multi-ring impact structures can assist in understanding their highly dynamic evolution.

DEDICATION

In memory of

Frontline workers who lost their lives due to COVID-19

and

Constable Ryan Slater

23 January 1985 - 5 December 2020

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LIST OF SYMBOLS

$S_1 > S_2 > S_3$	Deviatoric principal stresses
HEL	Hugoniot elastic limit
WRG	West Rand Group
CRG	Central Rand Group
TC	Transient Crater
P	Pressure
K	Temperature (in degrees Kelvin)
ρ	Density
$\sigma_1 > \sigma_2 > \sigma_3$	Principal compressive stress (positive in compression)
L	Impactor diameter
V_i	Impactor velocity
r	Radial distance from the estimated point source
UTM	Universal Transverse Mercator
EPMA	Electron probe microanalyser
DF	Dark Field
XPL	Cross-polarised
PPL	Plane-polarised light
XRF	X-ray fluorescence
EDX	Energy-dispersive spectrometry
WDX	Wavelength-dispersive spectrometry
EDS	Energy-dispersive X-ray spectroscopy
WGS 84	World Geodetic System 1984
GPS	Global Positioning System
ANEOS	ANalytical Equation Of State
EoS	Equations of State
CPPR	Computational cell per projectile radius
CSQ	Chart-D Squared
iSALE	Impact Simplified Arbitrary Lagrangian Eulerian
t	Time
TIMA	Tescan Integrated Mineral Analyser
WAMLAB	Wits Automated Mineralogy Lab
CPPR	Cell Per Projectile Radius
SALEB	Simplified Arbitrary Lagrangian Eulerian code, version B
FSI	Fracture Saturation Index
MSJS	Multiple-Striated Joint Surfaces
PDF	Planar Deformation Features
n.d.	Not detected
CLA	Shatter Cleavage A
CLB	Shatter Cleavage B

Chapter 1

INTRODUCTION

The formation of central uplifts in complex impact craters remains one of the most poorly understood aspects of cratering mechanics ([Lieger, 2011](#); [Rae et al., 2019](#); [Ebert et al., 2020](#)). Central uplifts are proposed to form at strain rates as high as $10^4 - 10^6 \text{ s}^{-1}$ involving displacements of metres to tens of kilometres within tens to hundreds of seconds ([Melosh, 1977](#)). Central uplifts vary morphologically: small complex craters show a central peak, whereas larger complex craters show peak rings. Peak rings form when the rising central peak over-heightens and collapses outwards over material sliding inwards from the crater walls, resulting in extensive rotation and overextension of its upper levels ([Morgan et al., 2000](#); [Lana et al., 2003](#); [Jahn and Riller, 2009](#); [Kenkmann et al., 2014](#)).

Numerical modelling can explain the varying gross morphologies of central uplifts, but not the internal complexities that would depend on such factors as whether the target rock is crystalline and/or layered, or how gross hydrodynamic flow behaviour is affected. In particular, the limitation of numerical models to incorporate the inherent heterogeneity of layered rocks and the observed internal second- and third-order structures found in exposed central uplifts inhibits their ability to model internal complexities. However, detailed structural analysis of impact structures is providing new insights into the internal complexities of central uplifts ([Kenkmann and Von Dalwigk, 2000](#); [Kenkmann et al., 2010, 2012, 2014](#)). This new information can be used to improve numerical models and enhance our understanding of cratering mechanics and central uplift formation. Where the target is layered, pre-existing markers provide an excellent opportunity to study internal small-scale strain patterns and identify associated faulting and folding.

This study focuses on the deeply exhumed central uplift of the world's largest known impact structure, the Vredefort Dome ([McCarthy, 2006](#)). Through field-based structural mapping and microstructural analysis of layered metasedimentary rocks of the lower West Rand Group (WRG), Witwatersrand Supergroup, in the inner collar of the dome, it aims to reconstruct the sequence of impact-induced structural changes following the impact. It sets out to provide a clearer understanding of the deformation mechanisms, kinematic history and rheological behaviour of these rocks throughout the various phases of central uplift formation. This research has implications for planetary sciences, where central uplifts can be used to probe the

subsurface and the crustal evolution of planets (structure, mineralogy and chemistry) during the intense bombardment, following the period of accretion and differentiation.

The study is subdivided into five fundamental thematics: 1) a field-based structural and kinematic analysis of large-scale folds within the collar of the dome to understand their origin and timing within the evolution of the impact structure; 2) a field and microscopic investigation of a closely-spaced fracture cleavage phenomenon in quartzites in the inner collar and its possible link to shatter cones; 3) a structural, petrographic and geochemical analysis of pseudotachylyte dykes and their associated faults, aimed at investigating and understanding their origin and timing; 4) a desktop study of the (a)symmetry of structural elements in the dome; and 5) numerical modelling of the stress and strain evolution of a rock volume equivalent to the inner collar locations investigated in themes 1 - 3 as an aid to constraining the timing of the documented structures.

1.1 Location of the Study Areas

The study areas are located in the WRG of the Witwatersrand Supergroup in the collar of the Vredefort Dome (Figure 0.1). Although the desktop studies encompass the entirety of the inner collar, four detailed study areas should be noted. The largest of the detailed study areas is located approximately 12 km northwest of Vredefort on Nauwpoort, Deelfontein, Steenkampsberg and Thwane farms (Figure 0.2). It is $\sim 12 \text{ km}^2$ in extent and covers the most well-layered, lithologically heterogeneous part of the Vredefort Dome, rocks of the Parktown, Brixton and Bonanza formations of the Hospital Hill Subgroup, of the WRG of the Witwatersrand Supergroup (Figure 0.3). The second study area is located in the northeastern collar on the farm Geluksoord and is $\sim 600 \text{ m}^2$ in extent (Figure 0.2), covering the rocks of the Bonanza Formation (Figure 0.3). It is located 12 km Northeast of Parys. The third study area is 5 km^2 in extent and covers part of the Archaean Basement Complex, Dominion Group and WRG of the Witwatersrand Supergroup in the innermost collar at the core-collar contact (Figure 0.2). It covers the Kopjeskraal and Donkervliet farms 8 km northwest of Parys. The fourth and smallest study area is a shatter cone outcrop in the Randfontein Formation of the Johannesburg Subgroup of the Central Rand Group (CRG) at the Schoemansdrift Bridge, which is 300 m^2 in extent. It is 17 km northwest of Vredefort.

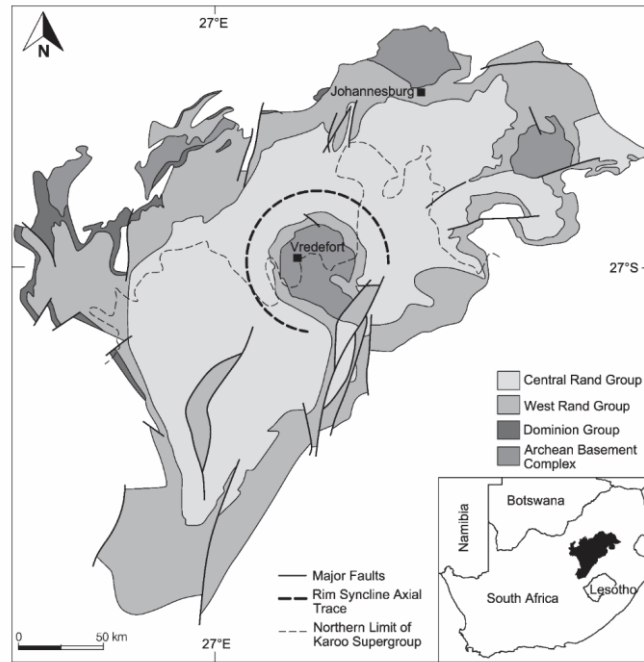


Figure 0.1: Geological map of the Witwatersrand Basin, stripped of all cover sequences, showing the location of the Vredefort Dome ~ 120 km southwest of Johannesburg. The axial trace of the rim syncline (Potchefstroom Synclinorium) and the northern limit of the Karoo Supergroup are also shown. Modified after [McCarthy \(2006\)](#).

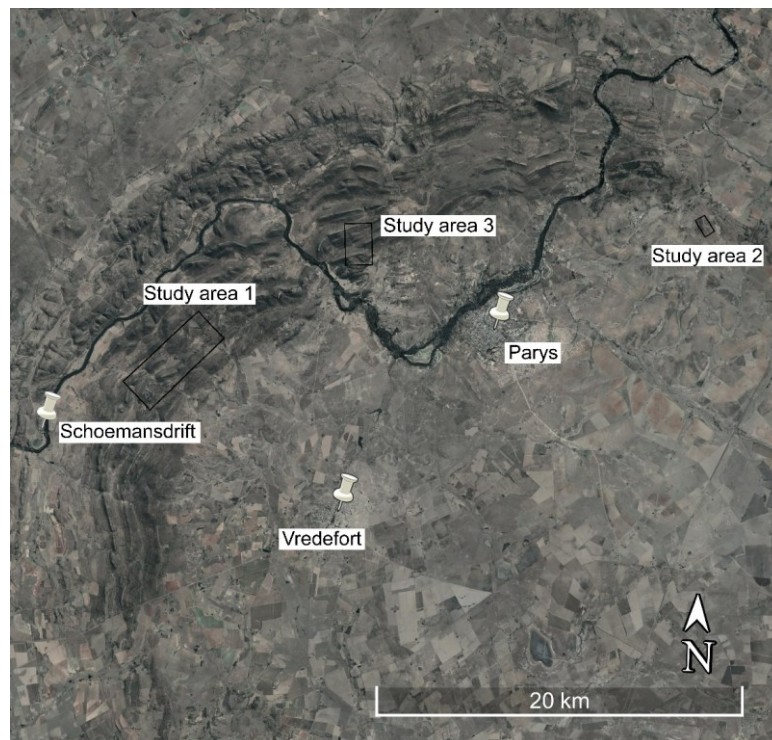


Figure 0.2: Google Earth Pro® image showing the various study areas covered in this thesis. Study Area 1 of Nauwpoort, Deelfontein, Steenkampsberg and Thwane farms; Study Area 2 on Geluksoord farm; and Study Area 3 at the Donkervliet and Kopjeskraal farms. The Schoemansdrift locality is also shown (Landsat imagery date: 30/01/2020).

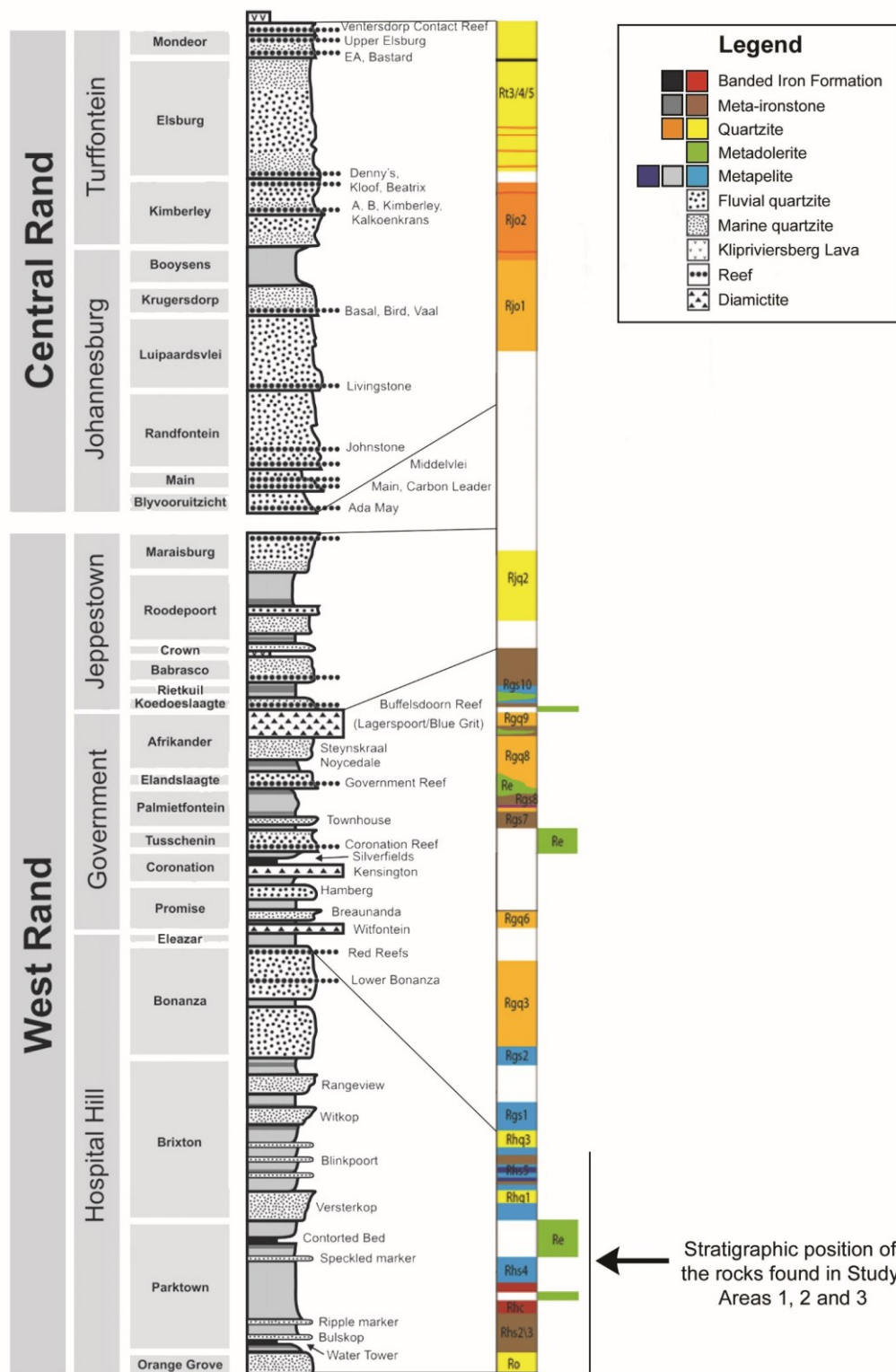


Figure 0.3: Stratigraphic column of the Witwatersrand Supergroup showing the stratigraphic position of the rocks and associated unit codes found within the four study areas. Study areas 1, 2 and 3 cover the Parktown, Brixton and Bonanza Formations within the Hospital Hill Subgroup of the WRG. The Schoemansdrift locality is found in the Turffontein Subgroup of the Central Rand Group unit code Rjo1 (Bisschoff, 2000). Modified from Guy et al. (2012) and Bisschoff (2000).

1.2 Aims and Objectives

This study aimed to document all macroscopic structural features (faults, folds, fractures, cleavages, shatter cones and pseudotachylytes) observed within the study areas to establish their relative geometric, spatial and temporal relationships. Kinematic analysis of structures of impact origin was then used to refine further their timing relative to the 3 phases of cratering and substages of each phase.

This study aimed to compare structures recorded in the field with the stresses and strains provided by numerical modelling to constrain the exact conditions of formation. This also validates or nullifies the numerical model's findings.

This study aimed to identify whether an axis of symmetry exists within the collar of the Vredefort Dome at depths in the central uplift, through desktop studies of previously published bedding and fault orientation data.

This study aimed to assess the mechanism by which the central uplift is formed and the structures created to accommodate its rise and collapse through field-based structural and kinematic analysis of large-scale folds and faults.

This study aimed to investigate and understand the timing, genesis and emplacement of pseudotachylyte dykes through field-based structural, petrographic and geochemical analysis.

This study aimed to test whether a locally intense-spaced fracture cleavage could be linked to the impact and shatter cones through detailed mapping, back rotations, and microscopic investigation.

1.3 Literature Review

This literature review covers broad topics of the Vredefort impact structure and impact-related topics that form the basis for understanding the bulk of this thesis. More detailed literature reviews on chapter-specific topics can be found at the beginning of individual chapters.

1.3.1 Geological Setting

The Vredefort Dome, located ~ 120 km southwest of Johannesburg, South Africa (Figure 0.1), formed in response to a 2.023 ± 0.004 Ga (2σ ; concordant U-Pb single zircon age; [Kamo et al., 1996](#)) impact event. The 80 - 90 km dome denotes the remaining root of the central uplift of a complex crater after ~ 10 km of overburden has been removed ([Gibson et al., 1998](#)). As a result, the Vredefort Dome allows one to study the intensely deformed levels beneath the original

impact crater. The original crater diameter is thought to have been 250 - 300 km, making it one of the largest impact structures in the world ([Gibson and Reimold, 2008](#)). The dome is well exposed in the northern and western sectors but largely covered by the Phanerozoic Karoo Supergroup in the southern and eastern sectors ([Gibson and Reimold, 2008](#)). Its ~ 40 km wide core comprises > 3.08 Ga Archaean granitoid gneisses and minor metamafic and metasedimentary gneisses ([Armstrong et al., 2006](#)). The core is surrounded by a 20 - 25 km wide collar of unconformity-bounded, 2.96 - 2.1 Ga, supracrustal strata containing a variety of intrusions that were deposited in a series of basins located on the Kaapvaal Craton ([McCarthy, 2006](#); [Paprika et al., 2021](#)). The collar rocks are generally overturned and are arranged as a broadly concentric sequence of layers younging away from the core of the dome ([Gibson and Reimold, 2008](#)). The collar is flanked by the Potchefstroom Synclinorium, an ~ 50 km wide structural trough (Figure 0.1) ([Simpson, 1978](#)).

The supracrustal collar rocks comprise the 2.96 Ga Dominion Group ([Paprika et al., 2021](#)), the 2.99 - 2.85 Ga Witwatersrand Supergroup ([Kositcin and Krapež, 2004](#)), the 2.791-2.683 Ga Ventersdorp Supergroup ([Gumsley et al., 2019](#)) and the 2.67 - 2.05 Ga Transvaal Supergroup ([Hofmann, 2011](#)). The intrusive rocks comprise metadolerite of the Klipriversberg Group 2.791 - 2.779 Ga ([Gumsley et al., 2019](#)), dolerite and alkali granite ± ultramafic complexes of the 2.055 Ga ([Mungall et al., 2016](#); [Scoates et al., 2021](#)) Bushveld Complex, and the Umkhondo, 1.05 Ga, Anna's Rust dolerite ([Reimold et al., 2000](#)).

This study focuses on rocks of the Witwatersrand Supergroup that are subdivided into the lower WRG and upper CRG, but also include parts of the Archaean Basement Complex and Dominion Group in places. The WRG comprises interbedded quartzites, metapelites and meta-ironstones. Although strongly metamorphosed, the lithologies retain sedimentary structures ([Gibson and Reimold, 2008](#)). In the quartzites, which are highly recrystallised, structures such as asymmetric ripples and planar and trough cross-bedding are preserved. The latter has been extensively used as a way up indicator, showing that the collar rocks are largely overturned ([Bisschoff, 1988](#); [Wieland et al., 2005](#); [Gibson and Reimold, 2008](#); [Jahn and Riller, 2009](#)). The quartzites are exposed as relatively straight ridges that individually stretch several kilometres and are cross-cut by faults mostly with small strike-separations of less than 10 m but up to 100 s of metres in places.

The rocks in the collar of the dome show evidence of a regional metamorphic overprint ([Bisschoff, 1982](#)). The metamorphic grade during this event increased with depth, from lower greenschist facies in the CRG to amphibolite facies in the WRG ([Bisschoff, 1982](#); [Gibson and](#)

[Wallmach, 1995](#)). Although [Bisschoff \(1982\)](#) and [Gibson and Wallmach \(1995\)](#) proposed a Bushveld age for the pre-impact event, [Beaton et al. \(2022\)](#) obtained Ventersdorp (2.796 ± 0.0015 Ga) Sm-Nd garnet ages in the northwestern collar of the dome, with a relatively minor overprint possibly linked to the Bushveld Event. The rocks within the Vredefort Dome experienced renewed metamorphism immediately after the impact, induced by shock heating and impact-induced decompression ([Gibson et al., 1998](#)). Temperatures attained during this event decreased centrifugally from the core, which experienced temperatures of ≥ 700 °C (granulite facies) to ~ 525 °C (lower amphibolite facies) in the lower WRG to ~ 300 °C (lower greenschist facies) in the CRG ([Gibson et al., 1998](#)). The lithologies that are currently exposed at the surface were buried after the impact at depths of 8 - 10 km ([Gibson et al., 1998](#); [Gibson and Reimold, 2008](#)) and thus accommodated extensive recrystallisation and/or metamorphic reaction, including comprehensive recrystallisation and metamorphic mineral growth in the pseudotachylyte matrices ([Gibson et al., 1998](#); [Deseta et al., 2017](#)).

Layering in the collar rocks, particularly in the WRG where strong competence contrasts exist on scales of tens to hundreds of metres, provides markers that help constrain the structural response of the target rock on impact and, thus, the strain evolution in the root of a central uplift. Given the exceptional age of the rocks (and, therefore, the strong possibility that they have experienced polyphase tectonism), it is necessary to consider the pre-impact tectonics of the Witwatersrand Basin first.

1.3.2. Structural Geology of the Witwatersrand Supergroup Prior to Impact

The known structural information on the Witwatersrand Supergroup has chiefly been obtained from 2D and 3D seismic data, underground mapping and borehole data from gold mines located on the outskirts of the basin (Figure 0.1). The information was obtained to identify structural controls associated with gold mineralisation and document faults that displace gold-bearing horizons that affect mine productivity (c.f., [Coward et al., 1995](#); [Jolley et al., 2004, 2007](#); [Beach and Smith, 2007](#); [Dankert and Hein, 2010](#); [Mambane et al., 2011](#); [Mashabela, 2011](#); [Manzi et al., 2013](#)).

Currently, four deformational events are considered to have affected the Witwatersrand Supergroup, based on data collected from the goldfields (c.f., [Bickle and Eriksson, 1982](#); [Burke et al., 1985, 1986](#); [Killick, 1992](#); [Jolley et al., 2004](#); [Frimmel, 2005](#); [Beach and Smith, 2007](#); [Dankert and Hein, 2010](#); [Manzi et al., 2013](#)):

D1 regional deformation, termed the Asazi Event by [Manzi et al. \(2013\)](#), describes early compressional tectonics during the deposition of the CRG at ~ 2.90 Ga. The deformation is visible in 3D seismic sections on a regional scale as an unconformity ([Manzi et al., 2013](#)). Deformation was associated with uplift and erosion, and localised fault formation and block tilting of the WRG prior to and during the early stages of deposition of the CRG ([McCarthy et al., 1982](#); [Roering et al., 1991](#); [Barnicoat et al., 1997](#); [Brink et al., 2000](#); [McCarthy, 2006](#)).

D2 deformation, termed the Umzawami Event ([Dankert and Hein, 2010](#)), is attributed to a single compressive event at 2.73 Ga ([Beach and Smith, 2007](#); [Jolley et al., 2007](#); [Manzi et al., 2013](#)). This age association may need revision with the latest ages for the Klipriviersberg Group of 2.779 - 2.791 Ga ([Ashwal, 2021](#)). The deformation occurred along the western and northern basin margins, forming an easterly-verging open fold-thrust belt ([Brink et al., 2000](#)). A well-documented example of such a D2 feature, described by [Manzi et al. \(2013\)](#), is the Libanon Anticline in the West Rand Goldfield. The Anticline is an upright, northerly-trending, tight-to-open fold with a wavelength of 8 km and an amplitude of 2 km that is associated with second-order parasitic folds, as well as reverse faults that verge to the east.

D3 craton-wide deformation, termed by [Manzi et al. \(2013\)](#) the Hlukana-Platberg Event, describes an extensional tectonic event that occurred at 2.70 - 2.64 Ga. The extension was accommodated by collapse along normal listric faults of variable orientation, typically with displacements of 2 - 3 km in magnitude, forming horst and graben structures or half grabens (c.f., [Stanistreet et al., 1986](#); [Tinker et al., 2002](#); [Beach and Smith, 2007](#); [De Kock et al., 2012](#)). The listric faults described by [Bickle and Eriksson \(1982\)](#), [Burke et al. \(1985\)](#), [Stanistreet et al. \(1986\)](#), [Brink et al. \(2000\)](#), [Frimmel and Minter \(2002\)](#) and [Tinker et al. \(2002\)](#) displaced the Witwatersrand Supergroup and the Ventersdorp Supergroup. The hanging wall is characterised by rollover anticlines and drag folds. The decollement zone is near the base of the WRG ([Manzi et al., 2013](#)). The faults are truncated by the unconformity at the base of the Transvaal Supergroup ([Gibson et al., 2000](#); [Mambane et al., 2011](#)). [Clendenin et al. \(1988\)](#), [Crow and Condie \(1988\)](#), and [Eriksson et al. \(2001, 2009\)](#) associated the extension and lithospheric thinning with mantle plume heating. [Brink et al. \(2000\)](#), [Jolley et al. \(2004, 2007\)](#), and [Beach and Smith \(2007\)](#) proposed that the eruption of the Klipriviersberg Group lavas marked the start of the event. This mechanism for the initiation needs revision given the latest ages for the Klipriviersberg Group of 2.779 - 2.791 Ga ([Cornell et al., 2024](#)).

[Mashabela \(2016\)](#) mapped two large arcuate strike-slip faults in the northern collar of the Vredefort Dome. The larger of the two has an offset of ~ 2.5 km, and the smaller 1.2 km. (Figure 0.4). He interpreted these as pre-impact conjugate listric faults that displayed an original NNE strike with a west-side-down slip, which were passively rotated during doming. In this original orientation they constitute an asymmetric half-graben. These faults are similar to those described by [Tinker et al. \(2002\)](#), [Jolley et al. \(2004, 2007\)](#), [Beach and Smith \(2007\)](#), [Dankert and Hein \(2010\)](#), and [Manzi et al. \(2013\)](#) in the Carletonville and West Rand goldfields and are linked with the D3 Hlukana-Platberg Event.

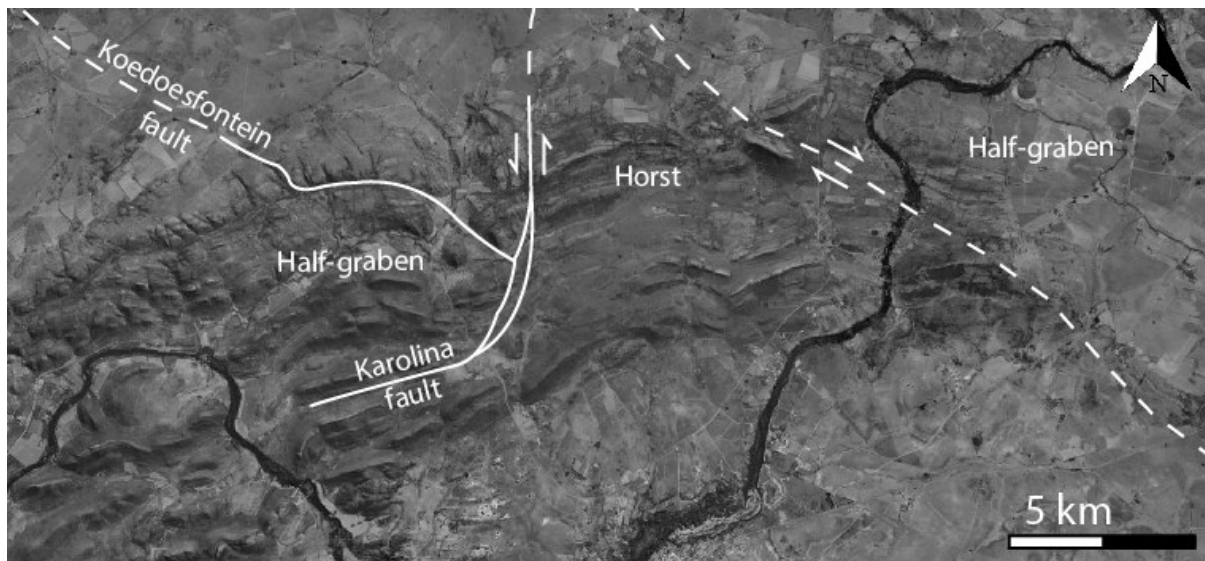


Figure 0.4: Google Earth Pro® image showing pre-impact structural features. The arcuate strike-slip, eastern fault described by [Bush \(2014\)](#) and Karolina fault that forms a horst and half-graben structure described by [Mashabela \(2016\)](#) (Landsat imagery date: 30/01/2020).

D4 regional deformation, termed by [Dankert and Hein \(2010\)](#) the Ukubambana event, describes orogenic compressional deformation that occurred after the deposition of the upper Pretoria Group at 2.20 Ga ([Bumby et al., 1998](#); [Killick and Scheepers, 2005](#); [Alexandre et al., 2006](#)). The deformation formed regional cleavages, a thrust-fold belt, NE and NW trending strike-slip faults ([Dankert and Hein, 2010](#)) and mylonitic shear zones ([McCarthy et al., 1986](#); [Gibson et al., 1999](#); [Alexandre et al., 2006](#)). [Gibson et al. \(1999\)](#) and [Alexandre et al. \(2006\)](#) showed that the cleavage post-dated contact metamorphic assemblages of the Bushveld Complex and Bushveld intrusions but predated the Vredefort impact at 2.02 Ga. The deformation had a N to NNE vergence ([Gibson et al., 1999](#); [Alexandre et al., 2006](#)); however, [Alexandre et al. \(2006\)](#) reported varying vergence directions and attributed them to polyphase deformation. [Alexandre et al. \(2006\)](#) previously attributed these features to the Transvaalide

Event. [Molezzi et al. \(2019\)](#) proposed that the Transvaalide and Ukubambana events are synonymous.

1.3.3 Mechanics of Complex Impact Crater Formation

The formation of complex impact craters can be divided into 3 phases that may partially overlap or vary spatially according to the location within the impact structure ([Gault and Wedekind, 1978](#); [Melosh, 1989](#); [Melosh and Ivanov, 1999](#)): 1) The Contact and Compression Phase; 2) The Excavation Phase; and 3) The Modification Phase. The first two phases last seconds to, at most, a few minutes and involve the generation and passage of the shock wave through the target, forming a transient cavity (TC). The first phase is characterised by exceptional shock pressures and temperatures, leading to a variety of mostly microscopic shock metamorphic phenomena, including melting ([Melosh, 1989](#); [Melosh and Ivanov, 1999](#)). The third phase lasts at most 10 minutes, even in the largest craters ([Melosh, 1989](#); [Melosh and Ivanov, 1999](#)).

I. Contact and Compression

The Contact and Compression Phase begins when the meteorite (impactor), travelling with a cosmic velocity of 17 - 25 km/s ([Le Feuvre and Wieczorek, 2011](#)), comes into contact with the target rock. Compression of the target rock and the impactor results in pressures ≥ 100 GPa at their point of contact ([Stöffler, 1971](#)). These high pressures are generated by the internal partitioning of the impactor's kinetic energy into the target rock and the impactor. As a result, a high-pressure stress wave is formed, called a shock wave ([Hugoniot, 1887](#)). A shock wave is a propagating disturbance that moves faster than the speed of sound in the given medium ([Wang et al., 2021](#)). As a result, the medium experiences a sudden, nearly discontinuous change in pressure, temperature, and density ([Courant and Friedrichs, 1999](#)). The shock wave moves outward from the point of contact, hemispherically down into the target rock and up into the impactor, causing compression ([Gault and Wedekind, 1978](#)). The associated deformation occurs as the rock experiences stress above its Hugoniot Elastic Limit (HEL) ([Winkler et al., 2018](#)). The HEL, as described by [Ahrens et al. \(1973\)](#), is the maximum stress a rock can resist under shock compression without experiencing plastic deformation and internal rearrangement. Above the HEL, a rock behaves in a strengthless manner, like liquids, and will undergo a shock-metamorphic overprint ([Winkler et al., 2018](#)). The generalised peak pressures required to elevate rocks above their HEL: felsic rocks 3 GPa; intermediate rocks 4 - 5 GPa; sediments <

1 GPa; mantle rocks ~ 9 GPa and mafic rocks 6 GPa ([Melosh, 1989](#)). The HEL for quartz is between 5 - 8 GPa ([Stöffler and Lagenhorst, 1994](#)). When the shock wave reaches the back end of the impactor, it is reflected backwards as a rarefaction wave ([Ahrens and O'Keefe, 1972](#); [Melosh, 1989](#)). Rarefaction waves decrease the density of the medium through which they travel ([Riemann, 1860](#)). The contact and compression phase ends when the rarefaction wave enters the target rock, releasing the shock pressure ([Melosh and Ivanov, 1999](#)). The rarefaction wave increasingly modifies the orientation of stress and instantaneous strain within the shock wave with time ([Rae et al., 2021](#)). The rarefaction waves result in partial vaporisation and melting of both the impactor and target rock in the direct vicinity.

At peak shock pressure, the maximum principal compressive stress (σ_1) is radial in both steep and shallower transects at distances of ~ 2.5 r/L (Figure 0.5), where r is the radial distance from the point source, and L is the impactor diameter. Near the point of impact (<1.5 r/L), σ_1 can be steeper than radial by $\leq 30^\circ$ (Figure 1.5) ([Rae et al., 2021](#)). At shock wave release, the maximum principal compressive stresses are almost radial, near and far from the point of impact (Figure 1.5), whereas at intermediate distances (~ 1.5 and 3.5 r/L), they are $30^\circ - 60^\circ$ shallower than radial (Figure 1.5) ([Rae et al., 2021](#)). During shock deformation, the maximum principal compressive stresses rotate between 40° and 60° . The radial variation of σ_1 orientations is constant given an impactor size of 0.1 - 10 km, at an impactor velocity of 15 km/s, and an incident angle of 90° for impacts on Earth ([Rae et al., 2021](#)). However, impactor velocity does affect stress orientations. Near-coaxial deformation is limited to directly beneath the impactor ([Rae et al., 2021](#)). The principal stresses are compressional throughout shock deformation, with σ_2 always being in the hoop direction. During the rise phase, i.e., the time between when the pressure first exceeds the HEL and the peak pressure, the stress ellipse is a prolate spheroid with $S_2 \approx S_3$ and $S_2 < 0$. However, once the peak pressure is reached, which is coincident with the arrival of the rarefaction wave, the stress ellipse close to the point source changes to an oblate ellipsoid ($S_2 > 0$) through a prolate ellipsoid ([Rae et al., 2021](#)). Further away, where pressures are < 10 GPa, the stress ellipsoid changes directly from a prolate spheroid to a prolate ellipsoid ([Rae et al., 2021](#)). The contact and compression phase lasts no longer than a few seconds, even in impacts as large as Vredefort ([Melosh and Ivanov, 1999](#)).

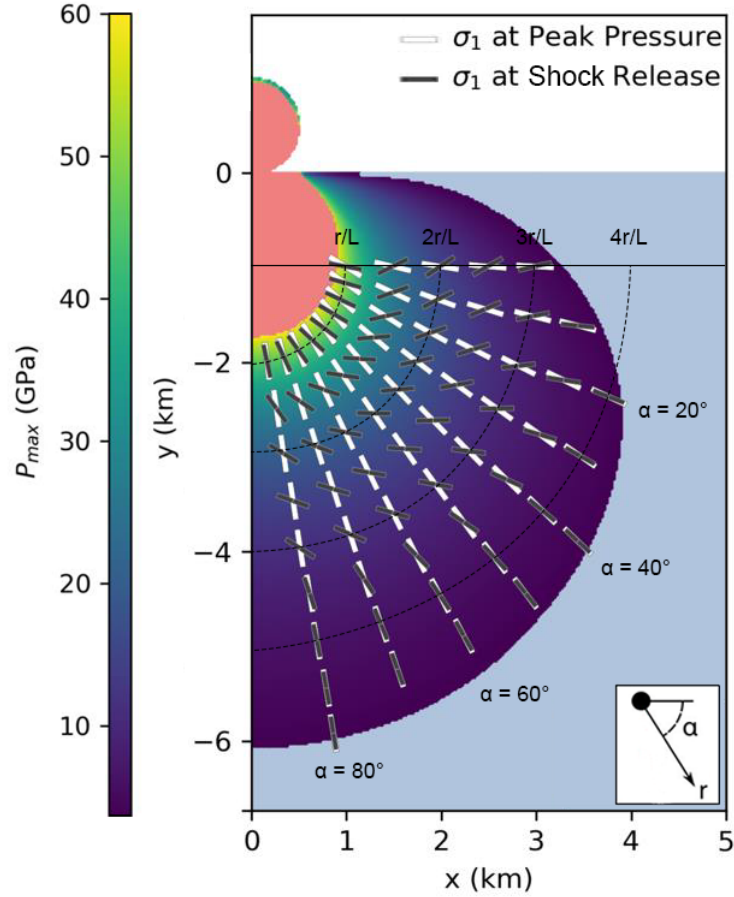


Figure 0.5: Image showing the orientation of the maximum principal compressive stress (σ_1) at peak pressure and shock release. The white and black lines indicate the orientation at peak pressure and shock release, respectively. Various radial transects of radial distance (r) are described as a function of impactor diameter (L). The transects' plunges are defined by (α) the angle of inclination from the horizontal. Dashed lines indicate various radial distances from the point source in terms of r/L . Peak shock pressures (white lines) are radial for steep transects and distances greater than $\sim 2.5 r/L$. σ_1 at peak shock pressure can be steeper than radial by as much as 30° at distances $< 1.5 r/L$. σ_1 at shock wave release (black lines) are almost radial, near and far from the point of impact. However, between 1.5 and $3.5 r/L$, they are $30^\circ - 60^\circ$ shallower than radial. Modified from [Rae et al. \(2021\)](#).

II. Excavation

The Excavation Phase describes the opening of the TC. Vaporisation and melting occur near the point of contact. The expanding hemispherical shock wave's energy attenuates with distance from the impact ([Holsapple, 1987](#)). The shock pressure and strain rate attenuate faster horizontally than vertically ([Rae et al., 2021](#)). This difference occurs owing to the counteraction of the mechanical strength of the target rock, the confining pressure, proximity to the free surface, and the angle of the shock and rarefaction wave fronts. When pressures are attenuated to $1 - 2$ GPa, the shock wave is turned into a seismic (elastic) wave ([Kieffer and Simonds,](#)

[1980](#)). When the tensional stress exerted by the shock and rarefaction waves exceeds the mechanical strength of the target rock, the material undergoes a downward and outward acceleration (≥ 10 km/s), forming a hemispherical TC ([Melosh, 1989](#); [French, 1998](#)). The deformation in this regime occurs as the rarefaction wave encroaches on the shock wave, and pressures fall below the HEL, creating a concentric hoop stress ([Rae et al., 2019](#); [Winkler et al., 2018](#)). The upper third of the cavity undergoes excavation and ejection (an upward and outwards movement), ejecting significant amounts of material into the atmosphere ([Gault and Wedekind, 1978](#); [Melosh, 1989](#)). The lower two-thirds are deflected downwards and outwards, forming a displaced zone within the cavity ([Maxwell, 1977](#)). This phase lasts several minutes. It ends when the cavity stops growing and the shock and rarefaction waves are not sufficiently energetic to eject material outside the TC rim ([Melosh and Ivanov, 1999](#)). This can be a product of the target rock weight or its cohesive strength, also known as gravity and strength-dominated cratering, respectively ([Kenkmann et al., 2012](#)).

III. Modification

The Modification Phase is characterised by the broadening and shallowing of the TC and a rapidly changing high-stress environment. This phase can be further broken down into four stages. Although this is a continuous process and stages may overlap, the subdivision into stages helps distinguish different processes in the evolving structure ([Gault and Wedekind, 1978](#); [Melosh, 1989](#); [Melosh and Ivanov, 1999](#); [Jahn and Riller, 2009, 2015](#)) (Figure 0.6).

- 1) Inward movement of the TC wall: The inward movement occurs via concentric normal faulting of the TC wall, which vertically thickens the centre of the TC. This produces one or more crater-rim normal faults and/or a crater-rim monocline. (Stage 1: Figure 0.6)
- 2) Upward movement of the central part of the TC floor: If the size of the TC exceeds a given threshold, which is controlled by target rock type and surface gravity ([Pike, 1988](#)), the force of buoyancy in the bottom of the TC will exceed the target rock strength ([Kenkmann et al., 2014](#)). This starts in the centre of the TC floor, where the force of buoyancy is strongest owing to centripetal forces and constrictional rock flow, causing the target rock to rise in a central peak. Central peak formation is accommodated by sub-vertical radial faulting and radial folding. (Stage 2: Figure 0.6)
- 3) Outward and downward movement during the collapse of the central uplift ([Osinski and Spray, 2005](#)): At a specific threshold diameter, > 2 km for sedimentary targets and

> 4 km for crystalline targets on Earth ([Grieve and Robertson, 1979](#); [Grieve, 1987](#)), central uplifts start to collapse under their weight, due to gravitational instability. This results in the radial spreading of the central peak ([Kenkmann et al., 2014](#)). If the central peak overheightens, the strata will collapse outwards and overturn, forming a peak ring ([Morgan et al., 2000](#); [Lana et al., 2003](#); [Jahn and Riller, 2009](#)). The strata may also collapse outwards when the crater reaches a critical size of $\geq 13 - 23$ km, depending on the lithologies involved ([Pike, 1985](#)). Outward collapse can be accommodated by overturning strata in the periphery of the central uplift. The resultant peak ring may separate an inner basin from the annular trough. Thus, structures formed during Stages 1 and 2 are overprinted by those formed during Stage 3; for instance, Stage 1 normal faults are rotated into reverse fault geometries. (Stage 3: Figure 0.6)

- 4) Inward movement of crater-rim material: Gravitational instability in the elevated crater-rim results in inward slumping that will also trigger tangential shortening, forming slump terraces (Figure 0.7). The overturning, outward-moving strata of the collapsing central uplift collide with the inward-slumping terraces, forming a topographically elevated complex interference zone called the peak ring ([Morgan et al., 2000](#)).

This phase lasts for no more than 5 - 10 minutes, even in the largest craters ([Melosh, 1989](#)). However, impact-related deformation, such as normal faulting, may continue for some time afterwards, owing to gravitational adjustment and cooling of the target rock and impact crater fill ([Kenkmann et al., 2014](#)).

Stages of the modification phase

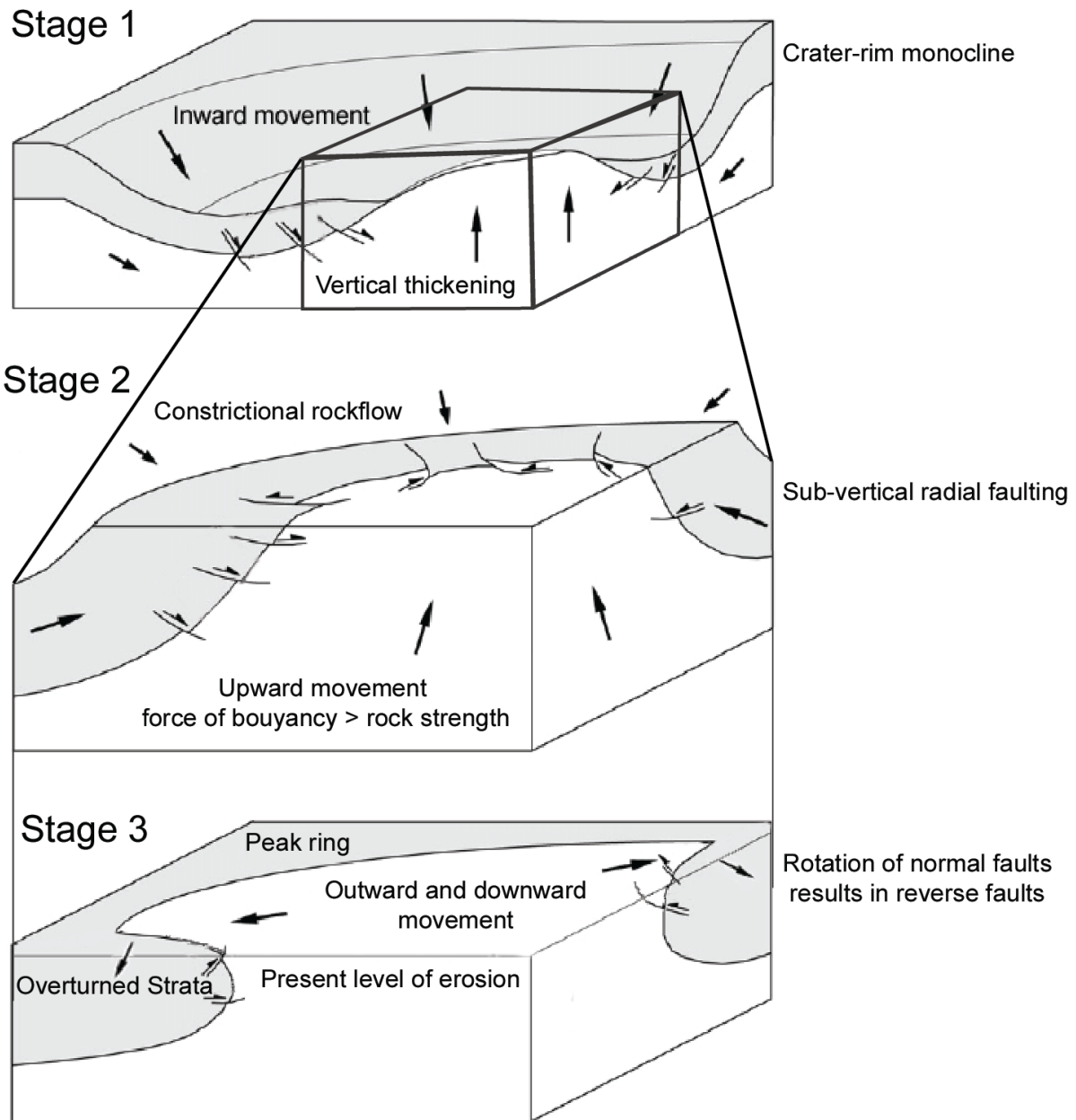


Figure 0.6: Simplified stages of the modification phase (Modified from [Jahn and Riller, 2009](#)): Stage 1: Inward movement of the TC wall produces a crater-rim monocline/normal fault and vertically thickens the centre of the TC. Stage 2: Constrictional rock flow increases the force of buoyancy in the centre of the TC floor, causing the target rock to rise in a central uplift. Sub-vertical radial faulting and radial folding accommodate this. Stage 3: Gravitational instability results in the radial spreading of the central uplift, which is accommodated by the overturning of strata and the rotation of normal faults into reverse faults. Extensive rotation overturns the strata and enables outward collapse, forming a peak ring.

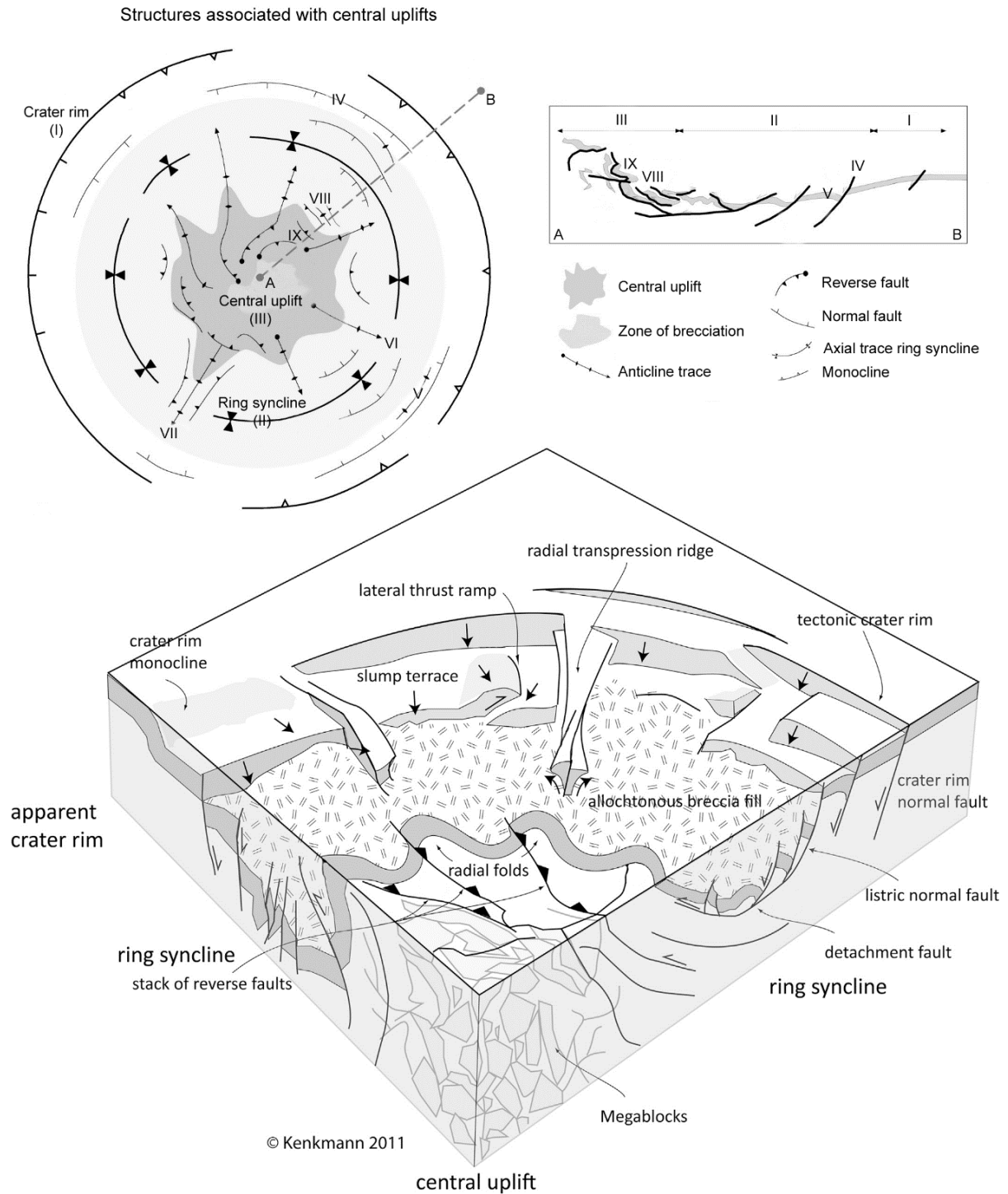


Figure 0.7: Schematic block diagram showing the structural complexity and spatial relationships between structures in complex impact craters. First-order elements consist of a crater-rim normal fault (I), a ring syncline (II), and a central uplift (III). Second-order structures that occur within or cut across first-order structures include normal listric faults (IV), rollover antiforms (V), radial anticlines (VI), radial transpression zones (VII), and recumbent folds (IX). Modified from [Kenkmann \(2002\)](#) and [Kenkmann et al. \(2011\)](#).

1.3.4 Structural Elements of Complex Impact Craters

[Kenkmann \(2002\)](#) separated the structural features within complex impact craters into first-and second-order elements (Figure 0.7). First-order elements comprise the crater-rim monocline/normal fault, ring syncline and central uplift. Second-order structures occur within or cut across first-order structures. These include but are not limited to, listric normal faults, rollover antiforms, radial anticlines, radial transpression zones, and recumbent folds. The structural complexity and intensity of impact deformation increase from the crater-rim, with an accumulated strain of ~ 0.25 , centripetally towards the central uplift, with an accumulated strain of ~ 1 ([Collins et al., 2004](#)). The increase of strain is related to increasing constrictional rock flow towards the centre of the crater during the modification phase ([Kenkmann, 2003](#)). It is important to note that there is a significant contrast between the finite and accumulated strain. The finite strain and associated features can vary dramatically with location and time ([Collins et al., 2004](#)).

In the outer zone, comprising the crater-rim and the ring syncline, low-angle concentric listric normal faults dip toward the crater centre, with associated rollover antiforms in the hanging walls ([Kenkmann, 2002](#)). These faults commonly merge into detachments at depth in stratified rocks where the bedding planes are used as glide planes where the target is horizontally layered. In the crater-rim region, significant displacements are concentrated along fault planes ([Kenkmann, 2002](#)).

The inner zone of a layered target, comprising the central uplift, shows kink and buckle folding of the steeply uplifted strata of the peak ring ([Kenkmann et al., 2014](#)). Spatial incompatibilities increase radially inwards, leading to increasing fold tightness. These spatial incompatibilities may initiate reverse faults along steeply plunging fold hinge zones. In the periphery of the central uplift, recumbent folds with concentrically oriented hinge lines and/or thrusts occur (Figure 0.7), where the hanging wall moves outward ([Kenkmann, 2002](#)). Reverse/thrust faults increase in frequency and displacement toward the centre of the central uplift. The faults propagate into fold limbs, detaching limbs from one another, forming sheet-like blocks bounded by fault breccia along thrust planes. These blocks stack up onto the brecciated core of the central uplift in an imbricate manner ([Kenkmann, 2002](#)).

1.3.5 Acoustic Fluidisation

The gross fluid-like response to form central uplifts is not yet fully understood; however, acoustic fluidisation is the favoured model for explaining this time-dependent process.

Acoustic fluidisation, as described by [Melosh \(1979, 1996\)](#), entails target rock blocks fragmented by the passage of the shock wave that undergoes pressure fluctuations around lithostatic stresses. When pressures are high, compression results in local increases in the frictional resistance between blocks. Pressure lows facilitate blocks to have momentary reduced normal stress between them. Thus, the frictional resistance is decreased sufficiently to enable slip events to occur at the boundaries. Therefore, the bulk material behaviour becomes that of a viscous fluid ([Melosh, 1979, 1996](#)).

The critical parameters that control acoustic fluidisation are the dominant wavelength and period of the pressure vibrations. The wavelength is directly proportional to the block size ([Ivanov and Artemieva, 2002](#)). It controls the acoustically fluidised rock mass's viscosity and the timescale for the decay of vibrations ([Dörfler and Kenkmann, 2020](#)). As the acoustic wavefield decays, small displacements along closely spaced faults, which occur earlier in the cratering process, give way to fewer large-scale displacements that occur along widely spaced slip surfaces. The target rock blocks cease oscillating as they regain internal friction and shear strength. This behaviour change is reflected by a change in deformation style from distributed cataclastic flow to localised shear-faulting, marking the end of acoustic fluidisation. Field examples created by such pressure fluctuations include (ultra-)cataclasite zones, narrow-spaced fault networks, blocks and mega-blocks ([Kenkmann et al., 2014](#); [Riller et al., 2018](#)).

[Ivanov and Kostuchenko \(1997\)](#) further applied acoustic fluidisation to develop the block oscillation model. The oscillation of individual blocks within a mass of breccia intermittently reduces the friction at the boundary, enabling slip. This mechanism results in a bulk material behaviour like that of a Bingham fluid ([Melosh and Ivanov, 1999](#)). These two models give rise to the block model of acoustic fluidisation used in most impact simulations ([Dörfler and Kenkmann, 2020](#)). A constant effective viscosity and a time-dependent yield strength are applied, resulting in a time-varying Bingham fluid rheology (e.g., [Collins et al., 2002](#); [Ivanov and Artemieva, 2002](#)).

Observations from the Chicxulub core ([Riller et al., 2018](#)) found that the blocks were smaller than expected and thus concluded that high-frequency vibrations had to be sustained for the duration of crater collapse. They explained this by the regeneration of acoustic energy during the cratering process, similarly to [Johnson et al. \(2016\)](#), who described this process in landslides. The acoustic wavefield may also regenerate by an increase in the dominant vibrational wavelength and block size during crater formation. [Rae et al. \(2017\)](#) had to use a long decay time to match their simulations and the observed morphology from the West

Clearwater Lake impact structure. This led them to suggest that acoustic waves regenerate during central uplift collapse.

1.3.6 Pseudotachylyte

The type locality for pseudotachylyte is the Vredefort Dome. It was named by [Shand \(1916\)](#) after its similar appearance to that of volcanic glass, tachylyte. Two spellings are presently in use: pseudotachylyte and pseudotachylite. To avoid confusion, pseudotachylyte is used as a petrographic and field name, consistent with Shand's [\(1916\)](#) initial definition. It took its name before the Vredefort structure was accepted to be of impact origin. Since 1916, it has been recognised along fault planes, landslide surfaces, and impact structures ([Kenkmann et al., 2014](#)).

Pseudotachylyte occur as irregular veins, massive vein networks or pods along faults, and fractures. These occurrences consist of brown-black, aphanitic, matrix-supported, dyke-like rocks of unspecified origin. Sometimes, they preserve evidence of a melt phase presented by any combination of the following features: glass; chill margins; flow banding; vesicles and amygdales; microlitic and spherulitic textures; melting and/or growth of minerals; and dominance of angular quartz clasts over rounded and embayed feldspar clasts ([Maddock, 1983](#); [Maddock et al., 1987](#)). These melt features have all been successfully replicated in frictional melting experiments ([Spray, 1987](#)).

Pseudotachylyte was initially thought to consist of a glass or a devitrified groundmass, based on optical microscopy, and was proposed to have formed along slip surfaces by frictional melting ([Holland, 1900](#); [Clough et al., 1909](#)). However, when [Willemse \(1937\)](#) analysed the Vredefort pseudotachylyte using powder X-ray diffraction, he found that it comprised ultra-fine-grained crystalline minerals. Thus, the groundmass consists of a mass of crystalline fragments comminuted to a rock flour and can only infrequently be termed melt-rock. Melt-rock does, however, occur in the centre of pseudotachylyte bodies, parallel to its walls. The discovery of [Willemse \(1937\)](#) began the debate as to whether pseudotachylyte was the product of melting or comminution.

[Sibson \(1975, 1977\)](#) determined that frictional melting upon a fault plane formed pseudotachylyte. This is supported theoretically within a fault zone when slip exceeds more than a few centimetres, even for shear resistances as low as 10 MPa ([McKenzie and Brune,](#)

[1972](#); [Cardwell et al., 1978](#)). Seismic melts form within a few milliseconds of slip initiation ([Violay et al., 2014](#)). [Wenk \(1978\)](#) contended, based on microstructural analysis of pseudotachylyte veins, that numerous brittle deformation features indicate formation through shock deformation. This was supported by shock experiments by [Johnson and Violet \(1958\)](#), showing that rock fuses at shock pressures ≥ 10.5 MPa. Wenk's argument was further strengthened by the emerging consensus in 1978 that the Vredefort structure was of impact origin and would have experienced associated shock pressures ([Martini, 1978](#)).

[Spray \(1995\)](#) showed that comminution is a precursor to friction melting. Thus, it is now accepted that a gradation between comminution and melting exists to form pseudotachylytes. The precursor minerals' fracture strength controls the comminution degree ([Reimold and Gibson, 2006](#)). Mechanical abrasion within an unconsolidated matrix and against the wall rock within the crush zone results in randomly oriented sub-rounded to well-rounded clasts ([Clarke et al., 1998](#)). The melted portion often contains microscopic crystallites and flow banding that can locally be seen in the adjacent, fine-grained material ([French, 1998](#)). Microlites and crystallites indicate quenching, suggesting that melting only occurs for a short duration. Pseudotachylyte melts form under non-equilibrium conditions ([Maddock, 1992](#); [Spray, 1987](#)). Most experiments are conducted under equilibrium conditions, and the individual temperatures at which minerals melt are not well known ([Papa et al., 2021](#)). Phyllosilicates have the lowest melting points, $\sim 700 - 900$ °C ([Spray, 2010](#); [Hung et al., 2019](#); [Dobson et al., 2021](#)). Plagioclase, under equilibrium conditions, melts at $1200 - 1400$ °C ([Bowen, 1913](#)). Quartz melts at 1715 °C ([Tuttle and Bowen, 1958](#)); however, [Lee et al. \(2017\)](#) report that slip melting could occur at temperatures as low as $1350 - 1500$ °C. Garnet melts at ~ 1200 °C ([Papa et al., 2021](#)).

However, in the context of impact craters, the genesis of pseudotachylyte is ambiguous, as one must distinguish between millimetre-sized veinlets and occurrences extending numerous kilometres, which are up to several tens of metres wide ([Kenkmann et al., 2014](#)). Given the high strain rates ($\geq 10^4$ s⁻¹; [Spray, 1997](#)) accompanying impact cratering, impact structures are considered favourable locations for forming friction melts. However, several potential mechanisms may be responsible for melt generation: frictional melting ([Spray and Thompson, 1995](#)), shock melting ([Fiske et al., 1995](#)), decompression melting ([Reimold and Gibson, 2005](#)), drainage from a superheated impact melt pool into tensional fractures ([Lieber et al., 2009](#); [Riller et al., 2010](#)), a combination of shock and decompression melting ([Mohr-Westheide and](#)

[Reimold, 2011](#)), and/or seismic shaking ([Garde and Klausen, 2016](#)). It is controversial how one distinguishes between these processes and their significance concerning the mechanisms of cratering. [Reimold and Colliston \(1992\)](#) introduced the term pseudotachylytic breccia to be used within impact structures for rocks that meet pseudotachylyte criteria but have an unknown genesis. However, the pseudotachylyte originally defined by [Shand \(1918\)](#) did not have a genetic implication but was instead a descriptive term used for identification, thus making the term “pseudotachylytic breccia” redundant. In addition, the term pseudotachylytic breccia is a poor description as the fragments can be anything from sub-rounded to well-rounded, regardless of clast size.

The pseudotachylyte-forming process was long thought to be self-limiting ([Bjørnerud and Magloughlin, 2004](#)) as the formation of pseudotachylyte reduces friction along the slip surface, but has recently been seen to only apply in experimental work as opposed to in natural systems where the fracture surface is constantly evolving, and melt is removed from the fault surface ([Spray and Biren, 2021](#)).

A and B-type pseudotachylyte

At least two generations of pseudotachylyte have been proposed to form during complex impact cratering. The first is referred to as A-type ([Martini, 1978, 1991](#)) or S-type (shock-related; [Spray, 1998](#)) pseudotachylyte, which forms during shock compression (hereafter referred to as A-type). A-type pseudotachylyte veins are generally < 1 mm wide, are randomly distributed throughout crater floors, have a shock origin and contain high-pressure polymorphs (coesite ± stishovite) in quartz-bearing rocks ([Martini, 1978, 1991](#); [Spray, 1998](#)). Melt is formed by localised high temperatures (800 °C - 1350 °C), pressures (> 45 - 60 GPa) and stresses during the contact and compression phase underneath the TC. This is due to shock wave propagation and rarefaction. The pseudotachylyte may concentrate along the lithological boundaries where shock wave-induced micro-shearing causes friction and ultracataclasis, decompression melting and vaporisation ([Kenkmann et al., 2000](#); [Heider and Kenkmann, 2003](#)). High temperatures are induced by frictional heating of the wall rock, where differential acceleration occurs along collapsing pre-existing joints, fractures, pores and heterogeneities ([Kenkmann et al., 2000](#); [Heider and Kenkmann, 2003](#)). A-type pseudotachylyte was successfully produced in shock compression experiments along heterogeneities where shock impedance contrasts existed ([Lin and Shimamoto, 1998](#)). A-type pseudotachylytes thus have no dependency on fault systems ([Martini, 1978, 1991](#); [Fiske et al., 1995](#); [Kenkmann et al.,](#)

2000; [Langenhorst et al., 2002](#); [Reimold and Gibson, 2006](#)). The bulk chemical composition of A-type pseudotachylyte tends to correlate to that of the wall rock, supporting in-situ formation ([Martini, 1991](#)).

The second-generation pseudotachylyte is thought to form during the excavation and modification phases, as decompression and/or frictional melts, and is termed B-type ([Martini, 1991](#)) or E-type (endogenic-related; [Spray, 1998](#)) pseudotachylyte (hereafter referred to as B-type). B-type pseudotachylyte is more voluminous, comprising the majority of occurrences forming along faults ([Martini, 1991](#); [Spray, 1998](#)). These reach considerable thicknesses from 1 mm to a few hundred metres ([Grieve and Theriault, 2000](#)). They lack diagnostic shock features and are proposed to be produced along faults by frictional melting during the crater modification phase. The bulk chemical composition of B-type pseudotachylyte tends to correlate to that of the wall rock, suggesting in-situ formation ([Martini, 1991](#)). They lack high-pressure polymorphs as the quenching time to shock pulse duration ratio is unfavourable due to their thickness ([Shaw and Walton, 2013](#)).

Geometric classification and implications of pseudotachylyte

[Sibson \(1975\)](#) proposed a simple pseudotachylyte classification based on field occurrence and vein geometry. In this classification, the pseudotachylyte vein along the fault plane (generation zone/plane) is termed a fault vein. Individual veins and vein networks branch obliquely off the fault vein, filling minor tensional fractures (emplacement zones) and are termed offshoot/injection veins ([Grocott, 1981](#); [Magloughlin and Spray, 1992](#)). Fault veins generally have a planar morphology, being several millimetres to several centimetres thick. Offshoot veins generally form complex network morphologies that record no offsets with wedges that thin away from the fault vein. Generation zones are well-developed and relatively straight, confined to a fault plane (single-generation zone) or shear zone (paired-generation zone).

Offshoots form on the extensional side of dynamic faults and grow towards the maximum compressive far-field (remote) stress ([Alneasan et al., 2020](#)). A new offshoot's orientation (branching angle) changes based on the loading conditions and is directly proportional to the fault tip speed, fault inclination angle, friction between fault surfaces, confining stress, and Poisson's ratio. The branching angle of the offshoot varies between approximately 77° and 115° ([Alneasan et al., 2020](#)). This is why field observations report tensile cracks to appear to be perpendicular to fault surfaces. Tensile cracks, emplacement zones and offshoot veins are mode-I fractures, which obstruct the propagation of dynamic faults in the direction of the

maximum shear stress ([Gangi et al., 1977](#); [Anders and Wiltshko, 1994](#); [Griffith et al., 2009](#); [Ngo et al., 2009, 2012](#); [Ferrill et al., 2017](#); [Collettini et al., 2019](#)). [Grasemann et al. \(2011\)](#) and [Ngo et al. \(2012\)](#) concluded that the density of tensile cracks is inversely proportional to the distance from the main fault because they prohibit growth. It is noteworthy that for tensile and hybrid fractures, the ductility index (DI) of the host should be $DI < 2$ as opposed to the case of shear fractures, where $DI > 2$ ([Ishii, 2016](#)). DI is defined as the far-field effective mean stress normalised to the tensile strength of the intact rock ([Ishii, 2016](#)).

Generally, tensile cracks occur around faults at shallow depths, as the confining stress resists their formation at greater depths ([Reches and Dewers, 2005](#)). The tensile branch direction is directly proportional to the stress angle ([Alneasan et al., 2020](#)). This causes the tensile branch on normal faults to form a positive angle (anticlockwise) with the fault plane, whilst for reverse faults, the branching angle is negative (clockwise) ([Alneasan et al., 2020](#)). An increase in the stress ratio causes the branching angle in normal faults to increase, whilst, in reverse faults, it decreases ([Alneasan et al., 2020](#)).

[Lin \(2008\)](#) reported that most generation zones are paired varying in thickness from several centimetres to several metres and are fragment-dominated (Figure 0.8). Less common, single-generation zones are < 5 cm thick and are melt-dominated ([Lin, 2008](#)). The pseudotachylyte melt and matrix material in generation zones are autochthonous ([Sibson, 1975](#)).

Emplacement zones form simultaneously with melt generation. Pseudotachylyte emplacement involves rapidly injecting a thermally pressurised frictional melt into tensional fractures ([Lin, 2008](#)). Pseudotachylyte geometries in emplacement zones range from orthogonal sets of veins, oblique offshoots veins, sporadic pods and dykes, en echelon arrays to complex networks that lack any preferred vein orientation. Based on granitoid samples from the Sudbury impact structure, [Al Barazi et al. \(2009\)](#) reported that the compositions of offshoot veins and their wall rocks were dissimilar overall. In addition, jigsaw-like margins indicate that hardly any material was removed from the wall rock during fracturing ([Pollard and Segall, 1987](#)). Thus, structural and geochemical data suggest that pseudotachylyte melt and clastic matrix material in emplacement zones are allochthonous or parautochthonous ([Reimold and Gibson, 2006](#)).

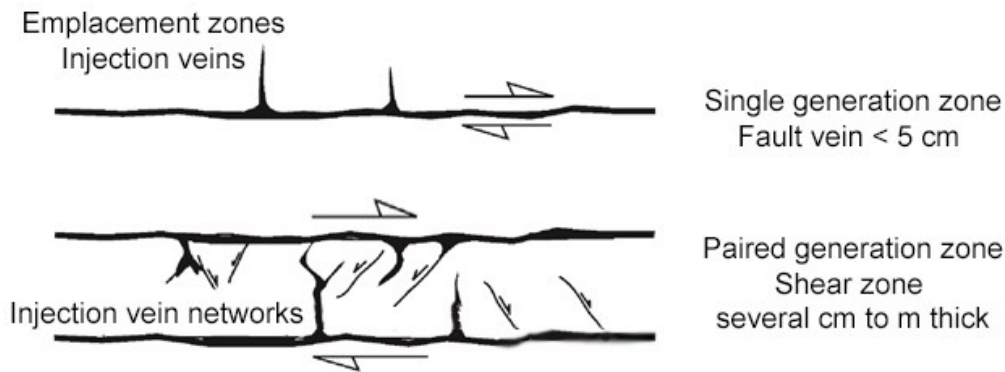


Figure 0.8: Common pseudotachylyte geometries from tectonic settings. Adapted from [Lin \(2008\)](#). Pseudotachylyte veins in generation zones are well-developed and relatively straight. They can be restricted to a fault plane (single-generation zone) or a shear zone bound by two fault planes (paired-generation zones).

[Sibson \(1975\)](#) demonstrated a linear relationship between the thickness of pseudotachylyte generation veins and their host fault displacement magnitude, whilst [Di Toro and Pennacchioni \(2005\)](#) reported a non-linear relationship. In most cases, confirming this relationship is complicated as the melt is removed from the generation zones into the emplacement zones, making it difficult to quantify the amount of melt generated along a single fault as the pseudotachylyte thickness in the generation zone would represent a minimum melt volume. The geometric variability of emplacement zones makes quantifying the volume of pseudotachylyte difficult, which is at least partly dependent on precursor lithologies ([Lin, 2008](#)). According to [Riller et al. \(2010\)](#), no correlation exists between pseudotachylyte width and wall rock displacement within impact structures. However, minimum and maximum target rock rotation zones positively correlated with dyke thickness. This relationship may signify that target rock rotation was achieved by tensile fracturing and the localisation of bending strains, which occur orthogonally to the maximum dilation directions ([Riller et al., 2010](#)). Pseudotachylyte occurrences increase in number and volume within large-scale fault zones that cross-cut central uplifts and in the hinge zones of folds ([Riller et al., 2010](#)). This contrasts with [Bisschoff \(1988\)](#), who stated that pseudotachylyte is not associated with major faults or offsets within the Vredefort structure. This suggests that pseudotachylyte distribution is kinematically related to the total accumulated strain during central uplift formation ([Lieber et al., 2009](#)).

In their discussion of large-volume pseudotachylyte occurrences in impact structures, [Thompson and Spray \(1996\)](#) and [Melosh \(2005\)](#) suggested that the friction melt is fluidised and continuously removed from the sliding surface during large-magnitude slip events,

accumulating in low-pressure reservoirs in the adjacent rocks. This model allows frictional sliding along the generation plane to remain active by keeping it narrow. However, this would require very large displacements to produce voluminous pseudotachylyte. Accordingly, [Spray \(2010\)](#) suggested that extremely large (up to kilometre-scale) displacements of unpinned blocks occur during the crater modification stage, causing the formation of large melt volumes. However, this idea is inconsistent with the absence of such large displacements within the Vredefort structure (e.g., [Wilshire, 1971](#); [Reimold et al., 2017](#)) and with the generally locally-derived composition of pseudotachylyte melt (e.g., [Mohr-Westheide et al., 2009](#); [Harris et al., 2013](#); [Reimold et al., 2017](#)). However, as shown in [Chapter 2](#), the total displacement experienced by the volume of rock is not always visible, as the net displacement can obscure the actual displacement. For example, a fault plane can undergo 100 m of normal slip followed by 90 m of reverse slip, preserving 10 m of slip in the rock record. This can obscure the relationships between pseudotachylyte thickness and fault displacement.

Autochthonous vs allochthonous origin

Within generation zones, melt transport can be geochemically and petrographically traced for metres ([Reimold, 1991](#)), from more prominent pseudotachylyte veins toward smaller ones and into offshoots ([Lieger, 2011](#)). Compositional differences can be explained by the movement of melt and entrained clasts or admixture of an exotic melt component ([Lieger et al., 2009](#); [2011](#)). Where melt transport cannot be proved, the difference in geochemistry between the wall rock and pseudotachylyte should still be considered as autochthonous by invoking partial melting, caused either by dissimilar fracturing behaviour of various mineral phases ([Schwarzman et al., 1983](#)) or by preferential melting of hydrous ferromagnesian minerals ([Shand, 1916](#); [Reimold, 1991](#)), wall rock heterogeneity ([Killick et al., 1988](#)), metasomatism within fault zones, and/or assimilation of fragments and crystals into the pseudotachylyte melt. The mineral fracture toughness hierarchy (from more susceptible to less susceptible to melting) is the following: phyllosilicates < amphiboles < pyroxenes < orthoclase < plagioclase < quartz < orthosilicates ([Spray, 1992](#)). This, in conjunction with melting point, would promote selective melting of ferromagnesian minerals ([Gibson et al., 1997](#); [Spray, 2010](#)) and, consequently, a slightly more mafic composition of the melt compared to the wall rock (e.g., [Reimold et al., 1985](#); [Maddock, 1992](#); [Magloughlin and Spray, 1992](#); [Killick, 1994](#); [Thompson and Spray, 1996](#)). Mafic domains, rich in biotite and amphibole, are thus observed to be more favourable to pseudotachylyte formation than felsic domains in granitic gneiss (e.g., [Reimold et al., 1985](#),

[1994](#); [Spray, 2010](#)). This systematic compositional difference between the matrices and wall rocks of pseudotachylyte has been documented for pairs from the Sudbury impact structure and the Vredefort Dome ([Speers, 1957](#); [Reimold, 1991](#); [Al Barazi et al., 2009](#)). For instance, the matrices of pseudotachylyte bodies within the felsic granitoids of the Vredefort Dome are relatively depleted in SiO₂ but enriched in Fe₂O₃, MgO, Al₂O₃, Na₂O, CaO and K₂O ([Schwarzman et al., 1983](#); [Reimold et al., 1985](#); [Reimold and Gibson, 2006](#)). In contrast, the pseudotachylyte matrix within the mafic wall rock is enriched in SiO₂ but depleted in MgO and CaO ([Reimold and Gibson, 2006](#)). Thus, based on geochemical analyses, pseudotachylyte matrices have largely been interpreted to form by in situ melting without introducing an exotic melt. The numerous wall rock clasts within the pseudotachylyte matrix make it unlikely that the bulk pseudotachylyte composition and the wall rock will differ much; however, this is not conclusive evidence for an autochthonous melt ([Lin, 2008](#)).

Pseudotachylyte veins usually contain more water than the wall rock as hydrous phases melt first preferentially to form the matrix of the pseudotachylytes ([Toyoshima, 1990](#); [Spray, 1993](#); [Lin, 1994](#)), although [Camacho et al. \(1995\)](#) report similar water contents. The melting point for quartz in the SiO₂ - H₂O system is ~ 1100 °C at pressures greater than 200 MPa ([Kennedy et al., 1962](#)).

Pseudotachylyte in the Vredefort Dome

Most pseudotachylyte samples analysed within the Vredefort Dome come from the basement granitoid rocks. This is due to the large amount of exposed basement granite and the contrast between the dark pseudotachylyte and the light felsic rocks. The majority of analyses support the in situ formation of the pseudotachylyte with the same stoichiometry as the wall rock minerals ([Schwarzman et al., 1983](#); [Killick, 1994](#); [Dressler and Reimold, 2004](#); [Reimold and Gibson, 2006](#)). Sulphide and silicate mineralogy indicate pseudotachylytic melt temperatures of at least 1200°C - 1700°C ([Lieger, 2011](#)). In the metapelites of the WRG, [Gibson et al. \(1997\)](#) linked pseudotachylyte to the formation of shear fracture cleavage. [Gibson and Reimold \(2005\)](#) explain pseudotachylytes as being formed by shock melting, which later accumulates in extensional sites during central uplift formation. [Reimold and Gibson \(2006\)](#) linked the pseudotachylytes to cataclasite zones created during the crater's inward collapse along bedding surfaces and lithological contacts.

[Lieger et al. \(2009\)](#) proposed that an allochthonous melt from an impact melt pool was emplaced into a tensional fracture system in the granitic crater floor during the collapse of the

Vredefort central uplift to produce the large-volume pseudotachylyte dykes. In a low-volume dyke-vein system also from the granitic core of the dome, [Mohr-Westheide and Reimold \(2011\)](#) concluded an autochthonous origin for pseudotachylyte formed by decompression melting in the central uplift of the Vredefort impact structure without a fault. [Mashabela \(2016\)](#) and [Wieland \(2006\)](#) linked the pseudotachylytes to radial faults that cross-cut the collar of the dome. [Mashabela \(2016\)](#) showed that the volume of pseudotachylyte mapped is larger in the CRG than in the WRG, with a general decrease in the volume of pseudotachylyte in the collar with respect to the core.

1.4 Methodology

1.4.1 Field-based Data Collection

In addition to basic geological mapping involving lithological description, recording of rock distribution and analysis of field relationships between mappable units, the techniques for structural mapping involved the measurement of the azimuth and plunge of lineations and the strike, dip and dip directions of planes with a Brunton Tru Arc compass-clinometer at UTM (WGS 84) georeferenced station points using a Garmin etrex[®]20 Global Positioning System (GPS). Planar features that were measured include bedding, shatter cleavage, pseudotachylyte dykes, joints and fractures. Oriented linear features that were measured include ripple crests on bedding surfaces and slickenlines along faults.

Representative, oriented samples were taken from selected sites and cut into thin sections for microstructural analysis. This was done to investigate the microscopic nature of the various lithologies and pseudotachylyte, shatter cleavage, grain-scale shock effects and post-shock annealing. As the primary field mapping area forms part of an ecotourism hiking trail within the Vredefort Dome World Heritage Site, samples were removed in the most discrete locations wherever possible.

The UTM (WGS 84) GPS coordinates of key outcrops, contacts, samples and photograph locations were recorded and captured into a database (Electronic appendix sheet labelled GPS Station Points). The stratigraphic unit codes used in this thesis are in accordance with the code system used by [Bisschoff \(2000\)](#).

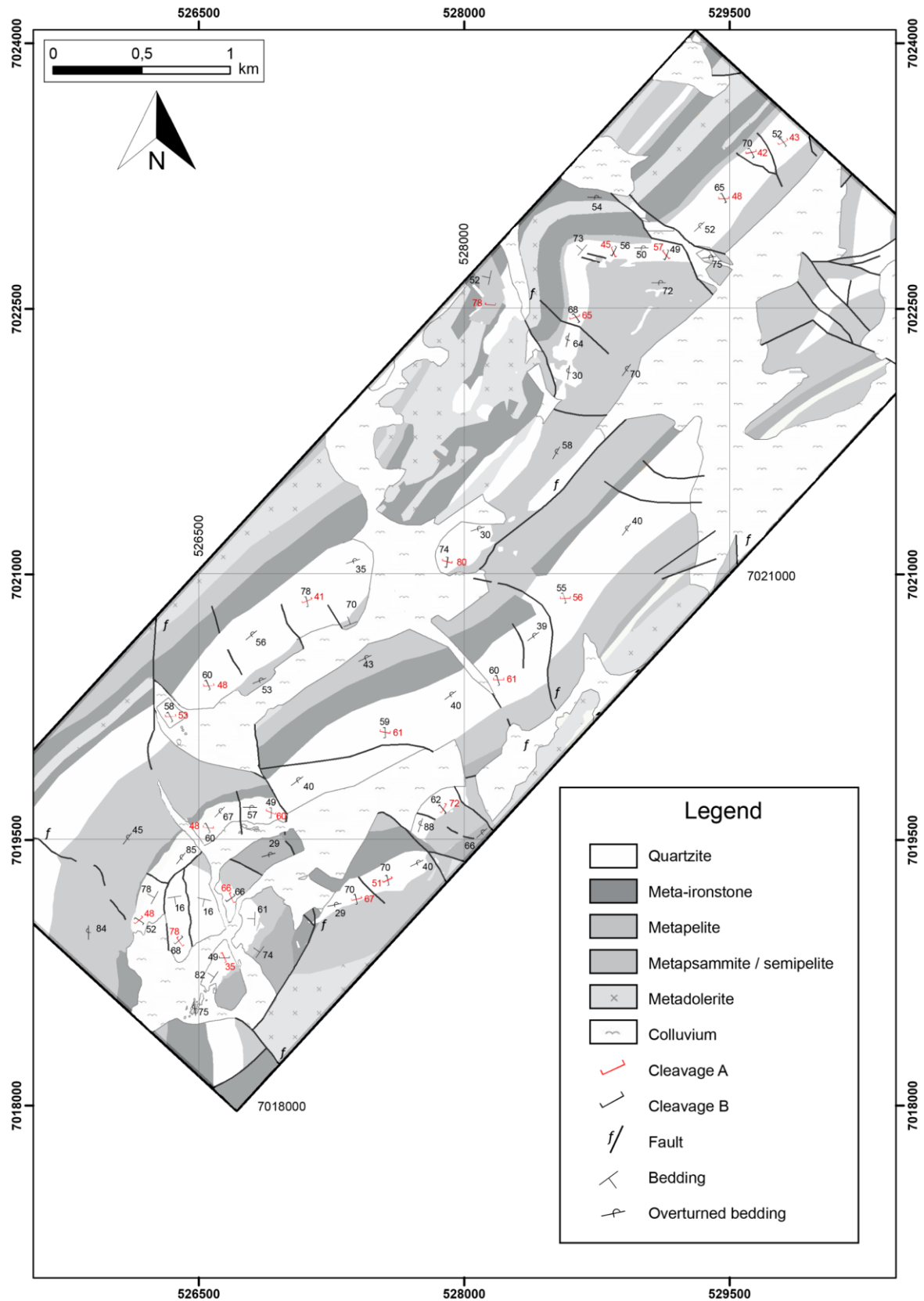


Figure 0.9: Geological map of Study Area 1 showing major lithologies, faults, cleavage and bedding measurements. Cleavage A and B refer to impact-induced shatter cleavage, where the tick indicates a dip direction (see [Chapter 5](#)).

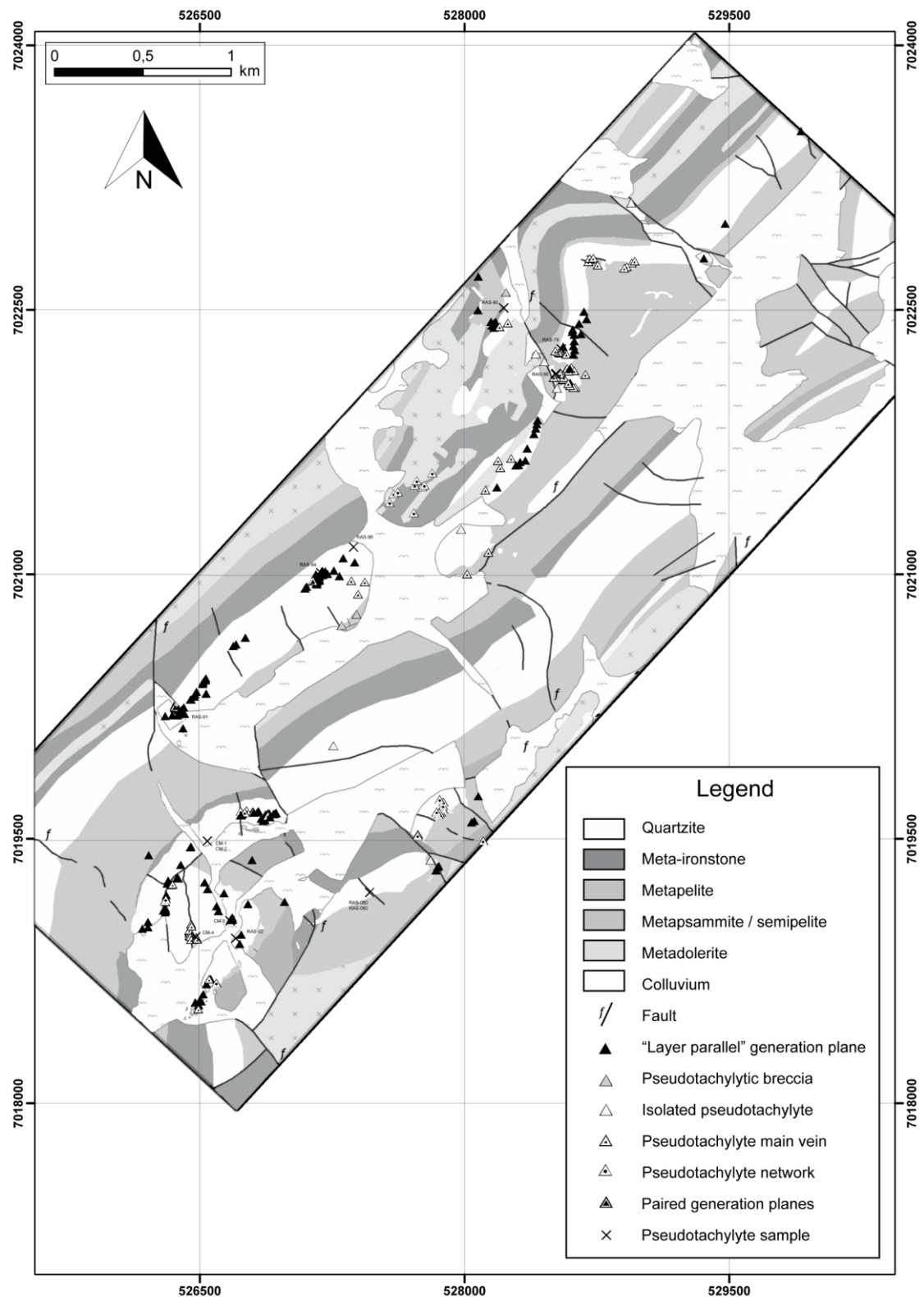


Figure 0.9: Geological map showing 420 pseudotachylyte occurrences by type within Study Area 1. Study Area 1 covers the Nauwpoort, Deelfontein, Steenkampsberg and Thwane farms. Pseudotachylytes along faults tend to display network morphologies. The majority of pseudotachylyte occur within layers without visibly associated faults. The majority of faults cross-cut layers that contain pseudotachylyte. See Figure 0.5 for a detailed view of pseudotachylytes at the Nauwpoort locality.

1.4.2 Laboratory-based Data Collection

Polished thin sections of selected samples were prepared at the thin section laboratory in the School of Geosciences at the University of the Witwatersrand, Johannesburg. The thin sections were analysed using transmitted light microscopy and photographed at the Microscopy and Microanalysis Unit, University of the Witwatersrand, Johannesburg, in Dark Field (DF), cross-polarised (XPL) and plane-polarised light (PPL) using an Olympus BX-63 microscope. The CAMECA SX5-FE electron probe microanalyser (EPMA) from the Microscopy and Microanalysis Unit of the University of the Witwatersrand, Johannesburg, was used to further analyse the mineralogy of the pseudotachylyte and its relation to its wall rock mineralogy at 15 kV and 15 nA. Analyses were calibrated using Smithsonian and Astimex international mineral standards. The calibrated elements and respective average detection limits [in ppm] are: Si [230], Na [79], Fe[205], K [31], Ti [124], Al [158], Mg [66], Ca [38], Mn [97], P [125] and S [75]. Counting times were 30 seconds on peak and 15 seconds on background, with a beam diameter of 35 nm. Energy-dispersive (EDX) and wavelength-dispersive spectrometry (WDX) were used to determine the mineralogy of the wall rock and adjacent pseudotachylyte matrices. The results of the EPMA work were complemented by x-ray fluorescence (XRF) analyses of the pseudotachylyte and their associated wall rocks. Crushed samples were analysed by M. Patchappa at the Earth Laboratory, School of Geosciences, at the University of the Witwatersrand. The Axios Max from Panalytical was used to analyse the samples at 50 mA and 50 kV. Major element analyses were performed using SuperQ and the fused disk method with a tetrabate flux. Selected thin sections of very fine-grained pseudotachylyte were characterised using the VEGA3 X64 TESCAN TIMA 1.7.0. Energy-dispersive X-ray spectroscopy (EDS) generated phase maps and elemental maps using the VEGA3 SEM instrument and the VEGA3 X64 Tescan Integrated Mineral Analyser (TIMA) 1.7.0. software, housed at the School of Geosciences Wits Automated Mineralogy Lab (WAMLAB) at the University of the Witwatersrand, Johannesburg. The beam conditions during quantitative analysis on the VEGA3 X64 TESCAN SEM were 25 kV accelerating voltage, 43 μ A emission current, with a working distance of 21.33 mm, a specimen beam current of 17.68 nA and a spot size of 590.00 nm with an absorption current of < 1 pA. The scanning mode was set to resolution mode, the heating was at 47 %, the filament live time was at 431 hours, and the gun pressure was at 1.6×10^{-3} Pa while the column pressure was at 5.5×10^{-3} Pa.

1.4.3 Data Processing

The areas' representative geological maps and cross-sections were made by overlaying UTM (WGS 84) GPS-referenced field station points onto a series of georeferenced Google Earth Pro[®] images using ArcGIS 10.3.1[®]. All the measurements from a single location were then plotted onto an equal area stereonet (Schmidt net) using Stereonet 9.9.4[©] to determine structural trends, fold geometries, and relationships between bedding and shatter cleavages. The summarised data were then transferred onto the geological map. Rotations of bedding and associated features were performed to determine the relative timing with respect to folding and the various phases of complex crater formation, as described in [Chapter 1.3.3](#). The structural data and the geological maps were then imported into Leapfrog Geo[®] 5.1 to create 3D models of selected study areas to enable 3D visualisation of the various fold structures.

1.4.4 Limitations

The geological map produced has a ± 5 m error on all GPS locations used to collect structural data, plot contacts between lithologies and faults, and record the location of photographs and samples.

Accuracies for the major element XRF analyses (all in wt. %) are: SiO₂ [0.5]; Al₂O₃ [0.1]; Fe₂O₃ [0.05], MgO [0.05], CaO [0.05], Na₂O [0.05] and K₂O [0.05]; TiO₂ [0.01], MnO [0.01] and P₂O₅ [0.01].

1.5 Thesis Organisation

The thesis is divided into eight chapters:

➤ Chapter 1: Introduction.

This chapter is a basic introduction to the pertinent literature required to understand the core chapters of the thesis.

➤ Chapter 2: Numerical Modelling

This chapter uses a numerical simulation of the Vredefort impact to evaluate bulk stresses and strains in a predetermined volume of the model corresponding to the location of field areas 1,2, and 3.

➤ Chapter 3: Large-Scale Symmetry Aspects in the Dome

This chapter assesses the symmetry in the orientations of faults and bedding across the dome in order to identify whether the dome is asymmetric.

- Chapter 4: Paper I: Dynamics of Collapse of an Impact Central Uplift: Evidence From Folds and Faults in the Collar of the Vredefort Dome, South Africa
This chapter contains my published paper, which documents the impact-induced structural evolution of study areas 1 and 2. It identifies a second generation of impact folding and extensional faulting related to late-stage central uplift collapse.
- Chapter 5: Impact-Related Fracture Phenomena
This chapter describes the geometry and timing of two orthogonal, closely spaced, impact-related fracture cleavages in Study Area 1 and their potential relationship to shatter cones.
- Chapter 6: Paper II: Pseudotachylyte Genesis Linked to Friction Melting in Large Impact-related Fault Zones in the Vredefort Dome, South Africa
This chapter contains my second paper, which is in preparation for submission in March 2025. It describes a pair of large-scale faults in Study Area 3 that are interpreted as a conjugate set and explores their timing and the implications for pseudotachylyte development.
- Chapter 7: Discussion and Conclusions
This chapter synthesises the various findings presented in Chapters 2 - 6 and their implications for the structural evolution of a large central uplift. It highlights the findings of this research and recommends future work..

Chapter 2

NUMERICAL MODELLING

2.1 Introduction

In [1961, Bjorke](#) published the first numerical model of a meteorite impact but could not account for gravity or the target rock strength ([Bjork, 1961](#)). Early models before 1999 could not account for non-vertical impacts, i.e. oblique impacts. Oblique impact models were first published by [Pierazzo and Melosh \(1999\)](#). The advance from 2D to 3D modelling has revealed that 2D simulations are accurate enough to recreate the gross crater morphology (as the length and width are sufficient to determine the size and overall shape of the crater) except in highly oblique impacts ([Collins et al., 2017](#)).

One of the greatest shortcomings of numerical models is the ability to describe the material models which govern how the target and the projectile materials behave under given impact conditions (stress) ([Collins et al., 2013](#)). Material models combine Equations of State (EoS) and the deviatoric stress/strength model. The EoS describe how the material behaves in relation to pressure, temperature, and density changes. These are particularly important in the contact and compression phase. ANalytical Equation Of State (ANEOS) is currently the most widely used EoS for solids; however, it is limited to a single solid-solid or melt transition. This significantly limits modellings as complex materials may require several solid-solid phase transitions ([Melosh, 2007](#)). Sophisticated EoS are needed when the impactor velocity is greater than 7 km/s, or particular interest is given to the melt or vapour production during impact ([Collins et al., 2004, 2013](#)). EoS for mixtures of materials (i.e., porous rocks containing water/air) and mixtures of minerals are beyond current modelling capabilities. At each timestep, EoS need to be calculated for each cell. This significantly increases computational time as computers have finite memory allocations. The deviatoric stress model defines the non-hydrostatic part of the stress tensor as a function of strain, strain rate, pressure, and internal energy/temperature ([Collins et al., 2013](#)). The deviatoric stress model is used to elucidate the material's resistance to a change in shape, which is important in the modification phase. Deviatoric stress models are based on laboratory experiments that overestimate rock strength and fail to produce gravitational collapse when extrapolated to scale. Thus, weakening processes must be considered([Collins et al., 2013](#)).

Additional problems are: 1) Modelling shear bulking/dilatancy and compaction, which affects porous materials. These result in volume changes that occur during the model's course,

affecting the EoS and the deviatoric model, which are modelled separately. [O’Keefe et al. \(2001\)](#) made an initial attempt to incorporate shear bulking/dilatancy in the impact code. The latest adjustment is implementing a porous compaction model by [Wünnemann et al. \(2006\)](#). 2) Attempting to portray a continuum with a finite number of points. A sufficient number of cells must be used to avoid errors due to lack of resolution. The measure of the resolution used is computational cell per projectile radius (CPPR). The greater the cell-to-impactor radius, the more detailed the model is, and the smaller the individual cell size ([Collins et al., 2013](#)). The modeller has to consider the feature or problem of interest and decide on the best resolution. Limited computational memory and calculation time must be weighed against accuracy. 3) Numerical models still use a static point source ([Rae et al., 2021](#)) as opposed to a more accurate point source that migrates with time to describe the excavation flow ([Austin et al., 1980](#); [Anderson and Schultz, 2006](#)). 4) The target rock strength is still greatly simplified as damage within the modelled cell causes a loss of cohesion such that the single cell behaves more like a granular material ([Collins et al., 2008](#)). Thus, the size of the cells modelled is dramatically larger than the volumes of coherent rock. 5) Computer models are still unable to model the nuanced layering of sedimentary/metasedimentary units, which act as planes of weakness for deformation. Thus, targets are often greatly simplified.

Computational and unit cell limitations continue to leave a disconnect between modelling and structural geology. Recent Chicxulub models ([Rae et al., 2019](#)) and subsequent papers ([Rae et al., 2021](#)) have advanced the frontiers of numerical modelling by analysing the instantaneous and finite stress and strain parameters throughout the impact event. One of the consequences is a much more refined modification phase, which can be broken up into several sub-stages.

2.1.1 Vredefort Models

Four models have been made of the Vredefort crater. The first was by [Turtle and Pierazzo \(1998\)](#), which was updated by [Turtle et al. \(2003\)](#), then later by [Ivanov \(2005\)](#) and most recently by [Allen et al. \(2022\)](#). The models are summarised below in Table 2.1. All three earlier models obtained crater sizes between 150 and 200 km. These results are significantly smaller than 270 - 300 km ([Therriault et al., 1997](#)) estimated by field data and 250 km ([Henkel and Reimold, 1998, 2002](#)) estimated by geophysics. The latter model by [Allen et al. \(2022\)](#) addressed this discrepancy, producing a crater size of 252 - 264 km. The simulation was used to generate an inventory of the stress and strain states of a unit cell volume corresponding to the inner collar locations studied in the field. The timing and geometry of the observed

structures could then be interpreted within the complex stress, strain, displacement and rotation history that was necessary to form the Vredefort Dome.

Table 0.1: Table summarising the details of the four previous numerical models of the Vredefort crater alongside the parameters used in this study.

	<u>Turtle and Pierazzo (1998)</u>	<u>Turtle et al. (2003)</u>	<u>Ivanov (2005)</u>	<u>Allen et al. (2022)</u>	Manzi (2025)
Crater diameter	150 - 200 km	150 - 200	172 km	252 - 264 km $\pm$ 10 km	160 km
Projectile	14 km	14 km	14 km	20 - 25 km	14 km
Projectile composition	Dunite	Dunite	Granite	Granite	Asteroid (2.670 kg/m ³)
Projectile Velocity	20 km/s	20 km/s	15 km/s	20 - 25 km/s	15 km/s
Impact Angle	90°	90°	90°	90°	90°
Target	Quartzite 14 km Granite 31.5 km Dunite	Quartzite 14 km Granite 31.5 km Dunite	Quartzite 14 km Granite 31 km Dunite	Quartzite 15 km Granite 25 km, Dunite 88 km	Quartzite 17 km Granite 28 km Dunite
TC	100 km	100 km	100 km	120 km	
Thermal Gradient			15 - 20 K/km	15 K/km	
Melt volume			13 000 km ³	130 000 – 160 000 km ³	
Melt thickness			400 - 800 m		
Concentric shortening			~ 25 %		
Run time			400 s	1500 s	600 s
CPPR		20	20	19	35
Model Details	Hydrocode CSQ - Contact and Compression Phase Z model - Excavation Phase Finite element modelling - Collapse Modification Phase	Hydrocode CSQ - Contact and Compression Phase Z model - Excavation Phase Quasi-static finite element model - Collapse Modification Phase	SALEB Continuous model	iSALE2D	iSALE
Major Limit	No inertia	No inertia			
Focus	Shock features and size	Temperatures	Comparison to other large craters	Constraining size	Model stress and strain of a volume of rock that ends up at a radial distance of 21 km and 10 km depth.

2.1.2 Model Parameters

The formation of the Vredefort impact structure was modelled to determine the stress and strain history in the inner collar at a radius of ~ 21 km in the lower WRG and compare it to detailed mapping results, which can be found in [Chapters 4, 5 and 6](#). This position was chosen because the lower WRG contains well-defined sedimentary layering at scales from mm to 100's m, providing the best markers for measuring ductile and brittle strain. The numerical modelling was done to compare the observed structural sequence, constrain duration and determine the relative sequence of features mapped in the study area.

The formation of the Vredefort impact structure was modelled using iSALE (impact Simplified Arbitrary Lagrangian Eulerian). iSALE is based on the SALE hydrocode by [Amsden et al. \(1980\)](#), a multi-rheology, multi-material shock physics code that has included several modifications: 1) porous compaction model ([Wünnemann et al., 2006](#)); 2) fragmentation models ([Melosh et al., 1992](#)); 3) several equations of state ([Ivanov et al., 1997](#)); 4) elastoplastic constitutive model; and 5) dilatancy model ([Collins, 2014](#)).

The impactor parameters are that of a 14 km diameter spherical projectile, with a velocity of 15 km/s and a density of 2.670 kg/m^3 , representing asteroid densities ([Britt et al., 2002](#)). The vertical velocity component of 15 km/s was chosen due to the statistical probability that impact occurs at a 45° angle ([Shoemaker, 1962](#)) and that the impactor would be travelling at the average asteroid impactor velocity of 20.5 km/s ([Le Feuvre and Wieczorek, 2011](#)). After a literature review, a 3-layer target was carefully chosen to represent the cover, crust and upper mantle. From the surface downwards: 1) 17 km of crystalline metavolcanics and metasedimentary rocks are represented by quartzite. Quartzite was chosen to be comparable to previous models; 2) 28 km thick granitic lower crust represented by granite; 3) mantle that begins at a depth of 45 km represented by dunite. Tables generated by the EoS program, ANEOS ([Thompson and Lauson, 1974](#)), were used to constrain the thermodynamic behaviour of the materials in the model. The resolution cell width is 200 m, 35 computational cells per impactor radius (CPPR), with a final rim-to-rim diameter of 160 km and a core-rim syncline boundary at a 40 km radial distance at a post-impact crustal depth of 10 km. The model ran from 0 - 600 s, in 2 s time steps, which is enough time for the final crater shape to form. The various parameters were presented to Auriol S.P. Rae (Department of Earth Sciences, University of Cambridge, UK), who provided the best-fit iterations and ran the simulation, which the author then interpreted as part of this thesis.

A numerical model was run for the volume of rock, hereafter referred to as the unit cell, 200 m x 200 m, at the location corresponding to that of the Nauwpoort locality (Study Area 1) which is representative of the Lower WRG. The unit cell was selected based on its final position at a radial distance of 21.4 km and 10 km depth with respect to the surface following the impact. The model indicates that this cell would have originated from a depth of 15.5 km and a radial distance of 26.3 km and had an ambient pressure of 0.52 GPa immediately prior to the impact. The model predicts cratering would have been largely completed within 600 s. The 600 s over which the model runs are divided into phases and stages that describe the gross changes in crater morphology. The absolute timing and duration of these phases/stages are relevant to the gross crater morphology and applied to the relevant unit cell. For instance, a volume of rock at a greater radial distance will experience the shock wave later than a volume closer to the impactor. Modelling has become so detailed that specifying the volume of rock in question is of greater importance as different portions of rock may experience markedly different strain regimes during the same gross morphological phase. See [Chapter 1.3.3](#) for more information on how these regimes are distinguished.

2.2 Results

The model is divided into 3 phases with substages that may partially overlap or vary spatially according to the location within the impact structure. For more details, see [Chapter 1.3.3](#). The whole process takes approximately 10 minutes.

- 1) The Contact and Compression phase begins when the meteorite impacts the ground and ends when the rarefaction wave enters the target rock, releasing the shock pressure.
- 2) The Excavation Phase begins with the rarefaction wave and opening of the TC.
- 3) Modification Phase Stage 1 is marked by the onset of inward movement of the TC wall
Modification Phase Stage 2 is marked by the onset of upward movement of the central part of the TC.

Modification Phase Stage 3 is marked by the onset of outward and downward movement during the collapse of the central uplift.

In this model, at the specific unit cell chosen, the 3 phases last from 0 to 2 s, 3 to 30 s, and 31 to 420 s, respectively (Figure 0.2). Phases and stages are overlain in the form of grey zones onto graphs plotted using the raw data, which can be found in the electronic appendix sheet: Numerical Modelling. The contact and compression stage is too brief to be indicated by a zone on the graphs as it is represented by the first few seconds upon contact, depending on the

volume of rock in question. Thus, the contact and compression phase is merged with the grey zone of the excavation stage.

For the cell corresponding to the inner most collar representing the WRG, the model predicts that the unit cell rises from 15.5 km to 12.7 km at 50 s before dropping again to 15.9 km below the surface at 103 s. It is then exhumed to its minimum depth of 10 km below the surface at ~ 193 s (Figure 0.2).

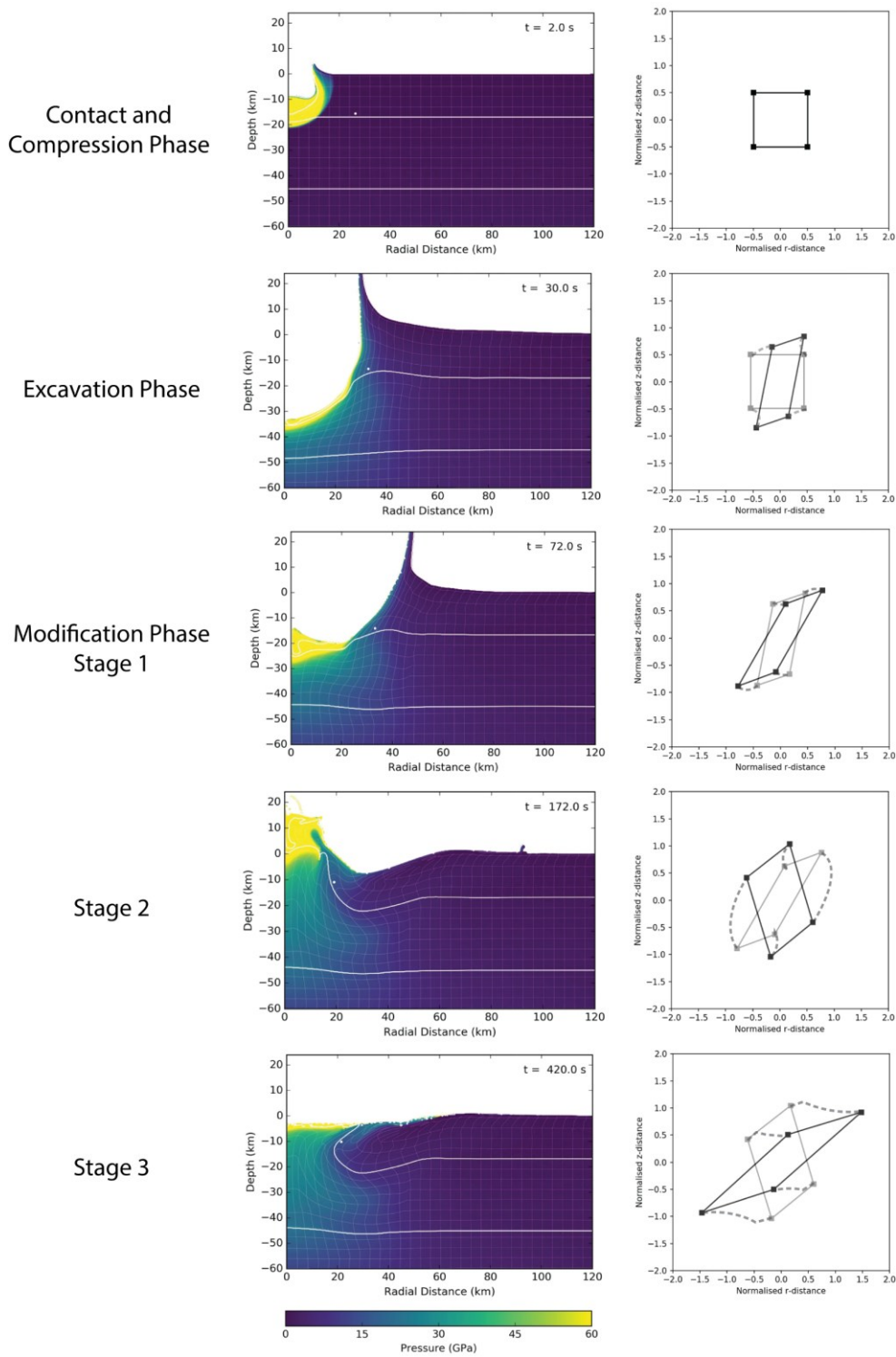


Figure 0.1: A 5-stage model that shows the crater morphology, shock pressure distribution and position of the lower West Rand Group rock volume at the end of each significant step/stage in the Vredefort cratering process is shown on the left. The white tracer particle depicts the location of the 200m x 200m lagrangian pseudo-cell. The thick white lines mark the boundaries between the granitic basement and the mantle at 45 km depth (lower) and the 17 km deep crystalline metavolcanic-granite contact (upper). On the right is the shape of the lagrangian pseudo-cell arrangement (shown in black) in which r and z are dimensionless distances normalised to 0 and 1. The grey dashed lines show the pseudo-cell path from the previous step/stage (grey solid lines).

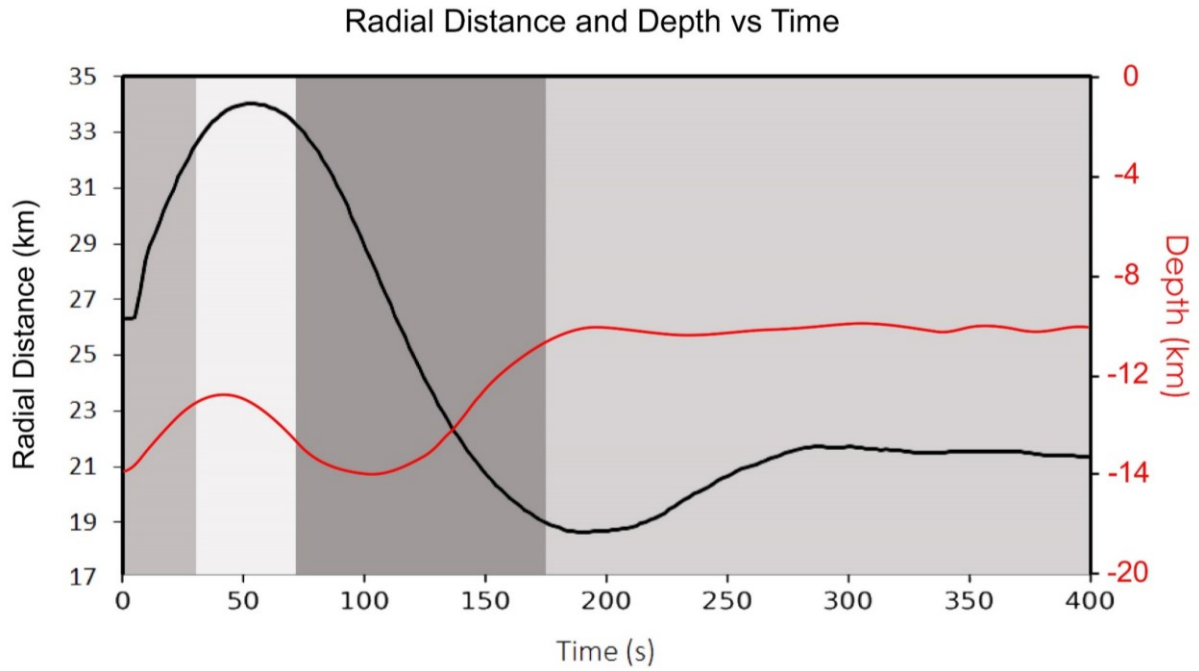


Figure 0.2: Graph showing the radial distance (black curve, left side axis) and depth (red curve, right side axis) of the lower West Rand Group lagrangian unit cell with respect to the centre of the dome or pre-impact surface versus time. From left to right, grey zones are Contact and Compression and Excavation Phases, Modification Stage 1, Modification Stage 2, and Modification Stage 3. The unit cell originates at a radial distance of 26.3 km at $t = 0$ s. It reaches its maximum radial distance of 33.9 km at $t = 59$ s and a minimum radial distance of 19 km at $t = 193$ s before settling into its final radial distance of 21.4 km at 269 s. The unit cell rises from its original location of 15.5 km depth to 12.7 km at $t = 50$ s before dropping again to 15.9 km below the surface at $t = 103$ s. It is then exhumed to its minimum depth of 10 km below the surface at $t = 193$ s.

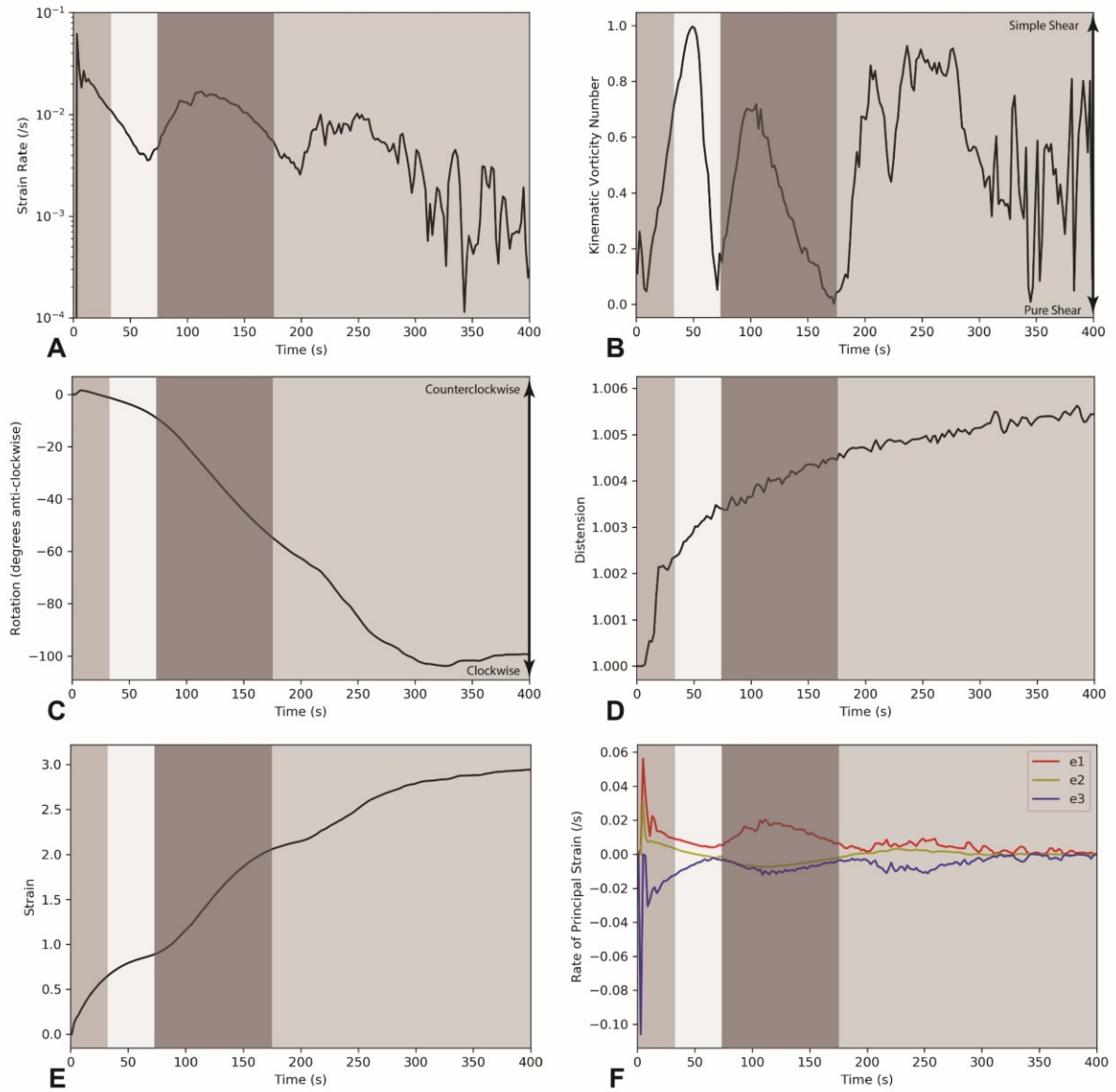


Figure 0.3: Graphs quantifying strain during impact crater formation for the lower West Rand Group lagrangian unit cell. Overlain grey zones, from left to right: Contact and Compression and Excavation Phases, Modification Stage 1, Modification Stage 2, Modification Stage 3. Graphs of time vs. (A) strain rate, (B) kinematic vorticity number where 0 corresponds to coaxial deformation, (pure shear) and 1 to rotational deformations (simple shear) and values in between represent sub-simple shear, (C) cumulative material rotation, (D) distention, (E) total deviatoric strain, and (F) rate of principal strain.

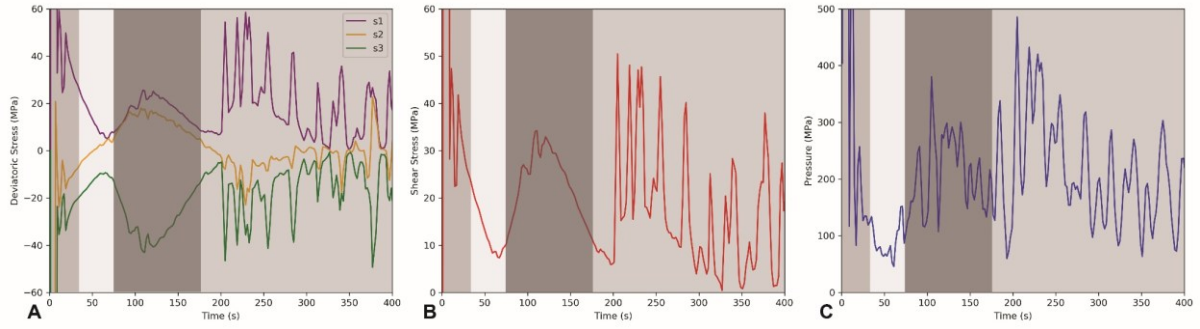


Figure 0.4: Graphs for the lower West Rand Group lagrangian unit cell of time vs. (A) principal deviatoric stress magnitudes, (B) maximum resolvable shear stress, and (C) pressure. Overlain grey zones from left to right Contact and Compression and Excavation Phases, Modification Stage 1, Modification Stage 2, Modification Stage 3.

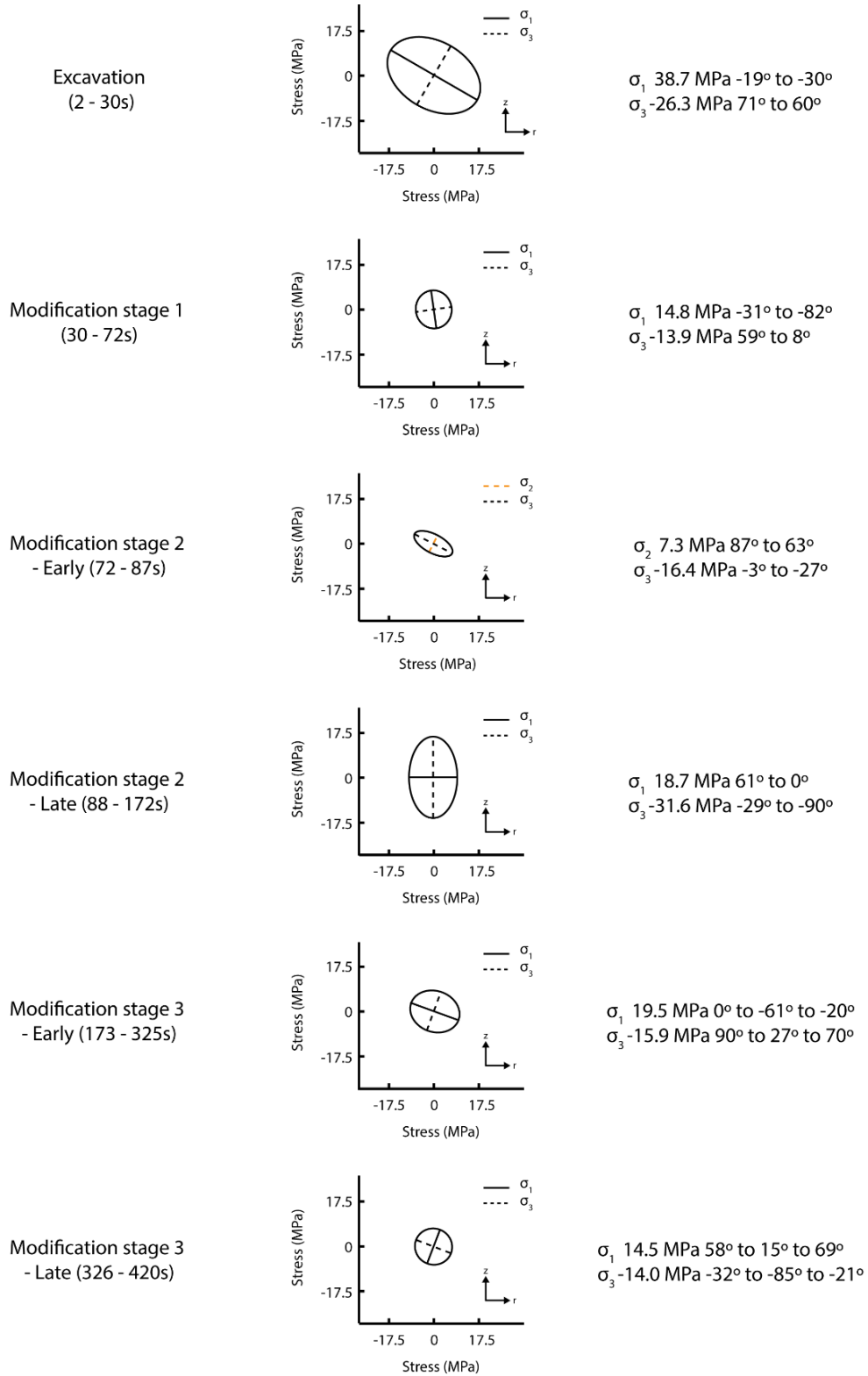


Figure 0.5: Sectional deviatoric Lamé's stress ellipses through the phases of cratering. The image displays the orientation at the end of each time step. On the right are the stress values for the two stress axes oriented within the 2D space and the angles through which they rotate in the given time step.

It is important to remember that the gross crater behaviour constrains the phases and not when the particular unit cell experiences the consequences of the phase. i.e. the contact and compression phase ends when the rarefaction wave enters the target rock and not when it passes the unit cell in question.

2.2.1 Contact and Compression Phase (0 s - 2 s)

The contact and compression phase begins when the meteorite impacts the ground and ends when the rarefaction wave enters the target rock, releasing the shock pressure. It is important to note that, due to the resolution of the modelling, inaccuracies in such a rapid process will exist as the time step is limited to 2 s. However, the information acts as a good guideline for understanding these processes. Due to the scale of the graphs provided in this chapter, values for the Contact and Compression Phase are taken directly from the modelled data available in the electronic appendix sheet labelled Numerical Modelling, as they are not visible from the graphs.

The volume of rock in question experiences a confining pressure of approximately 0.52 GPa immediately prior to impact. The shock wave arrived 3 s after first contact (Figure 0.4). However, the maximum pressure of 9.97 GPa experienced by the volume occurred at 5 s. The shock wave exceeds the HEL of the quartzite, which is ~ 4.5 GPa ([Melosh, 1989](#)), causing pressures to increase in the form of a plastic wave ([Ahrens et al., 1973](#)). After 7 seconds, pressure is once again less than the HEL. Shock pressure conditions are sustained between 3 s and 5 s after impact in the given volume of rock. It is then rapidly released by the rarefaction wave between 5 - 9 s, which reduces the pressure to 0.116 GPa —i.e., below the original confining pressure of the rocks. After that, the pressure experienced is generally less than the original confining pressure of 0.52 GPa except on three occasions at 219, 229 and 233 s after impact (Electronic Appendix Sheet: Numerical Modelling). The stress field thus inverts at approximately 5 s without the volume having been displaced more than 50 m (Figure 0.2, Figure 0.3).

During the contact and compression phase, the kinematic vorticity number is always less than 0.3 (Figure 0.3). This indicates that deformation is dominated by pure shear. The shock deformation is characterised by horizontal compression with negligible rotation and volume reduction (Figure 0.3). The cell's temperature rises from 520 K to 712 K (256 °C to 438 °C) (Electronic Appendix Sheet: Numerical Modelling).

2.2.2 Excavation Phase (2 s - 30 s)

The excavation phase begins with the rarefaction wave and opening of the TC. During the excavation phase, the volume of rock moves upward by 2.1 km to a depth of 13.4 km (Figure 0.2) and outward by 6.4 km to a radial distance of 32.7 km (Figure 0.2). During excavation, the kinematic vorticity number is always less than 0.6 (Figure 0.3b). This indicates that deformation is dominated by pure shear. The pressure falls from 0.403 GPa to 0.134 GPa (Figure 0.4c). The strain rate increases rapidly during this stage (Figure 0.3a). The volume of the rock rapidly increases (distends) (Figure 0.3d). The material rotates clockwise by 1° in the hoop direction. This rotation is largely insignificant. The temperature rises slightly, from 712 K to 720 K (438 °C to 556 °C). During this phase, σ_1 averages ~ 38.7 Mpa and is oriented between 19° to 30° inwards; σ_2 is in the hoop direction and not quantifiable; the average σ_3 is ~ -26.3 Mpa (extensional) and oriented between 84° to 60° outwards (Figure 0.5). The stress ellipse is a prolate spheroid with $S_1 \gg S_2 \approx S_3$ in this phase (Figure 0.5).

2.2.3 Modification Stage 1 (30 s - 72 s)

The inward movement of the TC wall marks the modification phase stage 1 (Figure 0.6). In the selected unit cell, this stage is characterised by a decrease in strain rate (Figure 0.3A) and material deceleration as the crater loses energy to expel material from within the cavity. The crater wall moves inwards with minimal rotation (Figure 0.3C). The strain remains relatively constant during this stage (Figure 0.3E). Kinematic vorticity peaks at this stage, being close to 1, indicating that this stage is dominated by simple shear (Figure 0.3B). Shear stress decreases rapidly during this phase to below 10 MPa, and the pressure is at its minimum at 0.046 GPa at $t = 61$ s but ranges from 0.119 GPa at the beginning of the stage to 0.87 GPa at the end (Figure 0.4B). The pressure at $t = 61$ s is significantly lower than the original confining pressure of 0.52 GPa. By the end of this stage, the principal deviatoric stresses S_1 and S_2 are approximately equal (Figure 2.5A). During this stage, the unit cell moves 1 km outwards before returning radially inwards to its original radial distance, from where it continues to move inward in the subsequent stage (Figure 0.2). The unit cell rises 0.5 km between $t = 30$ s – 49 s and then falls 1.5 km between $t = 50$ s - 72 s with a net displacement in the Z direction of 1 km to a depth of 14.2 km by the end of the first stage (Figure 0.2). The temperature remains relatively constant from 719 K to 725 K (445 °C to 451 °C). The stress ellipse begins this phase as a prolate spheroid $S_1 \gg S_2 > S_3$ and progressively changes towards an oblate ellipsoid $S_1 \geq S_2 > S_3$. During this phase σ_1 averages ~ 14.8 MPa and is oriented between -31° to -82° inwards; σ_2 is in

the hoop direction and not quantifiable; the average σ_3 is ~ -13.9 MPa (extensional) and oriented between 59° to 8° outwards (Figure 0.5).

2.2.4 Modification Stage 2 (72 s - 172 s)

Modification Stage 2 marks the start of significant rotation and the rise of the central peak (Figure 0.6). The upper portion of the TC moves outwards while the material in the lower regions moves inwards and upwards as the force of buoyancy is overcome ([Jahn and Riller, 2009](#); Figure 0.6). This stage shows the most dramatic change in the lateral and vertical position of the unit cell as it moves inward and downward to a depth of 16 km during the start of this stage and then changes to move inward and upwards to a depth of 10.9 km by the end of this stage (Figure 0.2). The unit cell ceases to move vertically at the end of this stage whilst it continues to move radially in the 3rd stage of modification. In the 100 s covered by the stage, the net displacement in the z direction is 5.1 km, and the net displacement horizontally is 14 km radially inwards (Figure 0.2).

The strain rate and shear stress increase during this stage, but the net changes are approximately zero by the end of the stage as the central uplift reaches its peak and becomes gravitationally unstable (Figure 0.3A and Figure 0.4B). The strain also rapidly rises during this stage (Figure 0.3E). The principal deviatoric stresses S_1 and S_2 are approximately equal throughout this stage (Figure 0.4A). The kinematic vorticity number begins the stage at approximately 0.2; it peaks at 0.7 in the middle of this stage before decreasing to approximately 0.2 at the end (Figure 0.3B). The pressure rises from 0.116 GPa to 0.134 GPa (Figure 0.4C). During this stage, the temperature increases slightly from 725 K to 758 K (451°C to 484°C). The stress ellipse is an oblate ellipsoid $S_1 \geq S_2 \gg S_3$ in this stage. This stage is divided into early (72 – 86 s) and late (87 – 172 s) sub-stages based on the orientation of the hoop stress. In the early part of Modification Stage 2 σ_1 is in the hoop direction, the average σ_2 is ~ 7.3 MPa oriented between 87° and 63° outwards and the average σ_3 is ~ -16.4 MPa oriented between -3° and -27° inwards (Figure 0.5). In the late part of Modification Stage 2, σ_2 is in the hoop direction, the average σ_1 is ~ 18.7 MPa oriented between 61° and 0° outwards and the average σ_3 is ~ -31.6 MPa oriented between -20° and -90° inwards (Figure 0.5).

2.2.5 Modification Stage 3 (172 s - 420 s)

The unit cell, although no longer moving vertically, continues to move radially inward before reversing motion to outward at $t = 201$ s and finally coming to a static position at 21.4 km from the centre of the dome at approximately $t = 269$ s (Figure 0.2). The unit cell becomes overturned

at $t = 200$ s. During central peak collapse, the kinematic vorticity number increases from approximately 0 to 0.8 (Figure 0.3B), indicating a simple shear dominated process. The shear stress increases rapidly at the beginning of this stage as the central peak collapses but then gradually declines with time, similar to the pressure (Figure 0.4 B, C). The principal deviatoric stresses S_2 and S_3 are almost equal during this stage (Figure 0.4a). The strain rate decreases over this stage as the crater material comes to rest and rotation halts (Figure 0.3A). The temperature increases slightly from 758 K to 776 K (484 °C to 502 °C). The stress ellipse is a prolate spheroid in this phase with $S_1 \gg S_2 > S_3$. During the early part of this phase σ_1 averages ~ 19.5 MPa and is oriented between 0° to -61° to -20° inwards; σ_2 is in the hoop direction and not quantifiable; the average σ_3 is ~ -15.9 MPa (extensional) and oriented between 90° to 27° to 70° outwards (Figure 0.5). During the late part of this phase σ_1 averages ~ 14.5 MPa and is oriented between 58° to 15° to 69° outwards; σ_2 is in the hoop direction and not quantifiable; the average σ_3 is ~ -14 MPa (extensional) and oriented between -32° to -85° to -21° inwards (Figure 0.5).

2.3 Discussion

Shock compression occurs at 3 - 5 s after impact in the given volume of rock, reaching a peak of 9.97 GPa at 3 s. It is then rapidly released by the rarefaction wave between 5 - 7 s, reducing the pressure to 0.116 GPa. After that, the pressure experienced is generally less than the pressure prior to impact of 0.403 GPa except for three discrete seconds at 219, 229 and 233 s after impact. This explains why the majority of features identified in this study and described in [Chapters 3 - 6](#) are extensional. [Lilly \(1981\)](#) estimated the shock pressures from planar deformation features (PDF's) in the collar quartzites to be 148 -59 kbar (14.8 – 5.9 GPa). These values are significantly higher than those calculated from the model here of 9.97 - 0.107 GPa.

One can see that the decompression associated with the excavation phase is experienced later in the majority of the crater at 50 s vs 30 s at the centre (Figure 0.1). The radial displacement of the unit cell continues long after the vertical displacement has settled (Figure 0.2). Thus, the maximum change in vertical displacement is only 5.5 km versus the radial maximum change of 14.9 km.

It is evident from the plots, particularly in Modification Stage 3, that there is dramatic fluctuation in the principal deviatoric stresses, shear stress, and pressure experienced by the unit cell. It is instructive that a given volume of rock is undergoing dramatic changes in a short period with little space or time to respond to these changes.

Compared to [Ivanov \(2005\)](#), who modelled the temperature of the given volume of rock at 773 K (500 °C), the temperatures experienced by the rock volume are similar at 776 K (502 °C). The pressures, however, are significantly different from those given by [Ivanov \(2005\)](#) which are between 20 and 30 GPa and our model's pressure at a maximum of 9.97 GPa. Although [Allen et al. \(2022\)](#) model a crater-rim that matches more closely that proposed by field geologists, the probability of having an impactor greater than 20 km striking the earth is very low, 1 in 300 million or 1 in 3 billion years ([Le Feuvre and Wicczorek, 2011](#); [Marchi et al., 2009](#)). Additionally, the simplest constraint of the collar core contact is modelled as occurring at 40 km when it occurs at 20 km, which is a large discrepancy. Numerical modelling still makes many assumptions about rock strength and the strength degradation process which affects crater size. The field estimates may be wrong, especially since the structure is deeply eroded. In addition, the maximum diameter of a multiring basin is very difficult to constrain as the rim is a series of fault scarps that get smaller and more indistinct outwards.

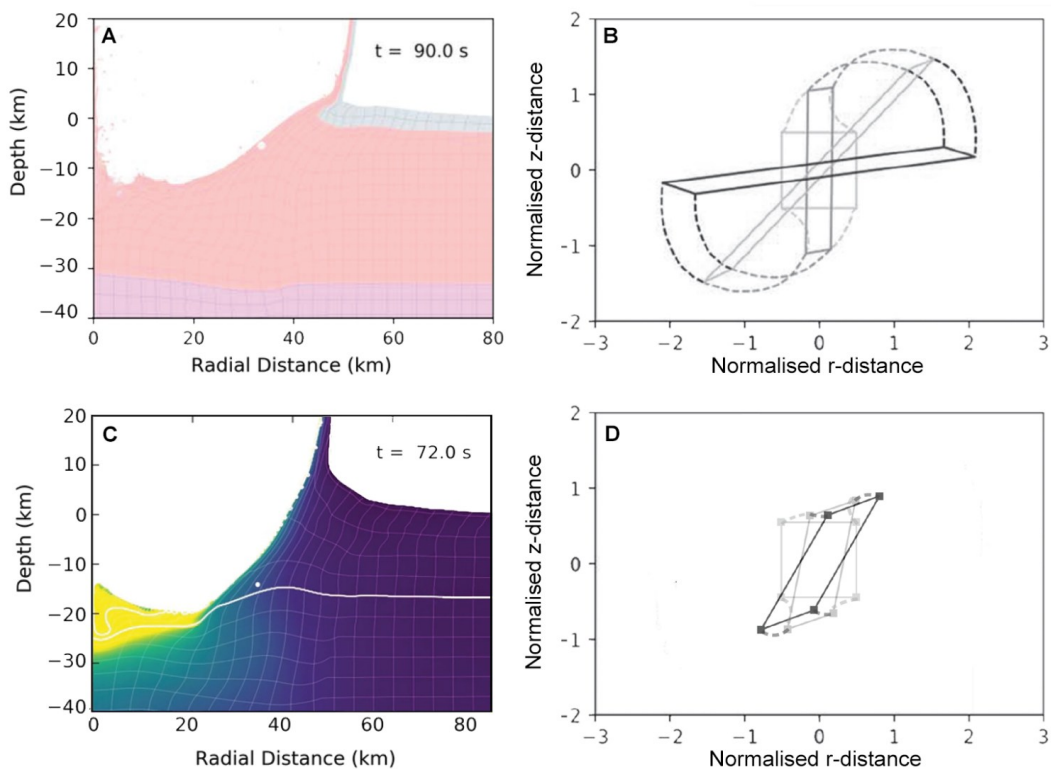


Figure 0.6: Comparative image of the crater morphology and position of the rock volume at the end of Modification Stage 1 in Chixulub and Vredefort impact structures (A) and (C), respectively. The white tracer particle depicts the location of the modelled lagrangian pseudo-cell. The thick white line in (C) marks the boundaries between the 17 km deep crystalline metavolcanic-granite contact. On the right is the shape of the lagrangian pseudo-cell arrangement (shown in black) in which R and Z are dimensionless distances normalised to 0 and 1. B) is for Chixulub, and (D) is for the Vredefort impact structures (B). The grey dashed lines show the pseudo-cell path from the previous steps/stages (grey solid lines). A and B are modified from ([Rae et al., 2019](#)).

When compared to the Chicxulub simulation of [Rae et al. \(2019\)](#), which has a unit cell slightly closer to the free surface but at a similar radial distance, one can see the effects of proximity to the free surface resulting in more significant deformation of the pseudo cell, as well as the degree of rotation that the unit cell experiences (Figure 0.6). The pseudo cell is significantly elongated radially to 4 times its original length during the inward transport during the modification stage and has rotated 90 degrees clockwise. At a similar stage in the Vredefort impact at greater depth, the unit cell has only gone through 40 degrees of clockwise rotation and been shortened radially to 0.6 times its original length. The distention for the Chicxulub unit cell is notably higher than that of the Vredefort unit cell, 1.03 vs 1.005; the distension in the upper levels is more gradually developed as compared to that in the lower levels, where the distention rapidly increases during the contact and compression phase. These changes are noteworthy as the impactor diameter for the Chicxulub crater (12 km) is only marginally smaller than that used in this simulation (15 km). It is also important to note the effect of target rock lithology. The model had a 60 CPPR and thus is more detailed than our simulation despite the same unit cell size.

2.4 Conclusions

1. The Contact and Compression phase is characterised by pure shear and horizontal compression with negligible rotation and volume reduction.
2. The Excavation Phase is characterised by a reduction in pressure as the volume of rock moves upward by 2.1 km and outward by 6.4 km.
3. Modification Stage 1 is characterised by simple shear.
4. Modification Stage 2 is characterised by the onset of significant rotation, deviatoric stresses S_1 and S_1 are approximately equal, and the rock volume moves approximately 14 km radially and 5.9 km vertically during this stage.
5. Modification Stage 3 is characterised by horizontal movement, simple shear, clockwise rotation and a decrease in strain rate as the crater stabilises.

Chapter 3

LARGE-SCALE SYMMETRY ASPECTS IN THE VREDEFORT DOME

3.1 Introduction

3.1.1 Symmetry in Impact Structures

Prior to the early years of the 21st Century, a consensus existed that, while impacts are expected to show a statistical mean angle of convergence between the bolide and its target of 45°, impact craters and their central uplifts overwhelmingly show circular shapes ([Gilbert, 1893](#); [Shoemaker, 1962](#)). Detailed structural mapping of central uplifts in partially eroded complex impact structures has revealed that these may contain evidence of asymmetry that is consistent with low angles of impact (< 15°), even as the crater-rim retains an essentially circular shape ([Gault and Wedekind, 1978](#); [Bottke et al., 2000](#)). Numerical modelling of impacts at angles of < 15° has provided general support for both the exocentric position of the central uplift and internal bilateral symmetry of secondary structures like folds and faults ([Gault and Wedekind, 1978](#); [Bottke et al., 2000](#); [Cox et al., 2021](#)). [Poelchau and Kenkmann \(2008\)](#) have used statistical analysis to investigate the bilateral symmetry of central uplifts inferred to have formed in low-angle impacts.

Maxwell's Z model ([Maxwell, 1977](#)) predicts a perfectly concentric strike for vertical impacts in a homogenous target. Final crater shapes should, thus, be circular for impacts at angles greater than 10° - 15° from the horizontal ([Gault and Wedekind, 1978](#); [Bottke et al., 2000](#)). This means that impact direction can generally not be determined from the shape of the impact structure as most impactors will strike the earth at 45° from the horizontal ([Gilbert, 1893](#); [Shoemaker, 1962](#)). As most impacts do not occur vertically or in homogeneous target rocks, the modified Z model was created. In the modified Z model ([Anderson et al., 2004](#); [Anderson and Schultz, 2006](#)), the internal and ejecta flow field centres shift downrange (in the direction away from the launch site and along the flight line of a meteorite) from uprange (the distance along a line that is 180 degrees to the downrange direction); this causes a bilateral non-radial symmetry in the ejecta for impacts < 45° ([Gault and Wedekind, 1978](#)).

Numerical simulations also show that the shock pressures during the contact and compression phases are asymmetric for oblique impacts (Figure 0.1). The flow velocity field of the TC wall is asymmetric until the TC collapses (rise in the TC floor), with higher velocities downrange

than uprange ([Artemieva and Ivanov, 2001](#)). Thus, asymmetry in the flow field decreases with time ([Anderson et al., 2003](#)). Downrange, the ejecta behaves concentrically, but substantial deviations occur in the crater rim between the downrange and cross-range sectors, forming two sharp corners due to abrupt strike changes on the edges of an overturned flap ([Gault and Wedekind, 1978](#)). The strike between the two corners marking the overturned flap is perpendicular to the force of the flow field ([Kenkmann and Poelchau, 2008](#)). This occurs at an impact angle below 45 degrees and becomes sharper with shallower angles. This is referred to as the “two-corner pattern” and is similar to the cardioid shape described by [Schultz et al. \(2007\)](#) (Figure 0.2). Radial ejecta trajectories occur in the downrange zone, whereas in the uprange zone, a zone of avoidance forms. This zone tends to be V-shaped and can also be called the forbidden zone ([Anderson et al., 2003](#)). Although these studies were primarily performed on ejecta, the method has since been applied to autochthonous bedding in the crater floor. This is necessary as ejecta blankets are not commonly preserved on Earth, so morphology, structure and geophysics are more commonly used to investigate potential oblique impacts ([Scherler et al., 2006](#)).

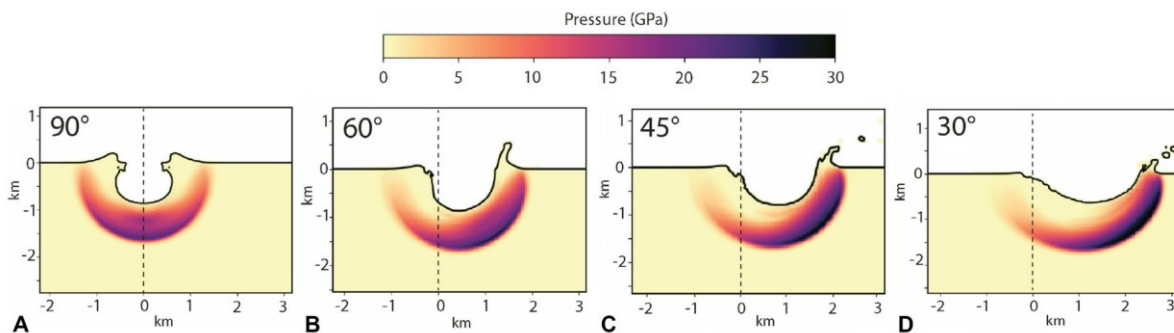


Figure 0.1: Numerical simulations of the shock pressure during the contact and compression stage in the Spider impact structure, using the iSALE-3D, at various impact angles. The shock wave changes from symmetric with an incident angle of 90° in (A) to increasingly more asymmetric with increased impactor obliquity. The impact direction is from left (uprange) to right (downrange), corresponding to a northwest-to-southeast transect in the field. The highest shock pressures migrate from directly below the impact with a 90° incident angle towards the downrange sector with increasing obliquity. A shallower incident angle reduces cratering efficiency, resulting in a shallower transient cavity. Modified from [Cox et al. \(2021\)](#).

Recent studies show that preferential folding and faulting of subsurface structures in central uplifts can be used to determine impact vectors ([Kenkmann et al., 2005](#); [Scherler et al., 2006](#); [Kenkmann and Poelchau, 2009](#)); however, early asymmetries are masked by symmetrical centripetal collapse. [Scherler et al. \(2006\)](#) reported that structural asymmetries are most prominent in the core of the central uplift and decrease centrifugally. This is due to centripetal

movement in the central uplift's roots ceasing earlier than the more distal regions, thereby preserving the large-scale circular shape of the first-order structural elements such as the crater-rim and the rim syncline. It is yet to be determined if preserved asymmetries are smaller in larger central uplifts that show more significant degrees of collapse. Thrust direction also reflects the prevailing lateral displacement direction during crater collapse and the direction of oblique impact ([Scherler et al., 2006](#)). This can be seen in the Spider structure, where imbricate thrust stacks have a prevailing top-to-south transport direction that coincides with the downrange transect ([Cox et al., 2021](#); Figure 0.3B). Upheaval Dome shows structural upliftment in the uprange sector. Near the centre of the central uplift, the bedding strike becomes more radial, associated with increased concentric shortening, and strata have shallower dips at deeper levels ([Scherler et al., 2006](#); Figure 0.3A and Figure 0.6). Mjolnir's crater displays evidence of oblique impact preserved by the elongation of the crater along the downrange transect, structural uplift in the uprange sector and an elongate gravity anomaly that follows the inferred impact direction southwest-northeast ([Tsikalas, 2005](#)). The low gravity anomaly makes a horseshoe shape that opens to the northeast (downrange sector), and the breached peak ring also occurs in the downrange sector ([Tsikalas, 2005](#)).

When attempting to investigate structural symmetry in an eroded crater, it is important to recognise that selecting a fixed, present-day, centre of impact on the map does not acknowledge that the centre of an oblique impact moves downrange with time. It is also worth remembering that, with only the most oblique impacts being able to cause marked central uplift asymmetry, asymmetry can reasonably be ignored for the majority of impacts. In the case of the Vredefort Dome, the great depth of erosion (8 - 10 km, [Gibson et al., 1998](#)) probably also means that asymmetry evidence is likely to be more difficult to see.

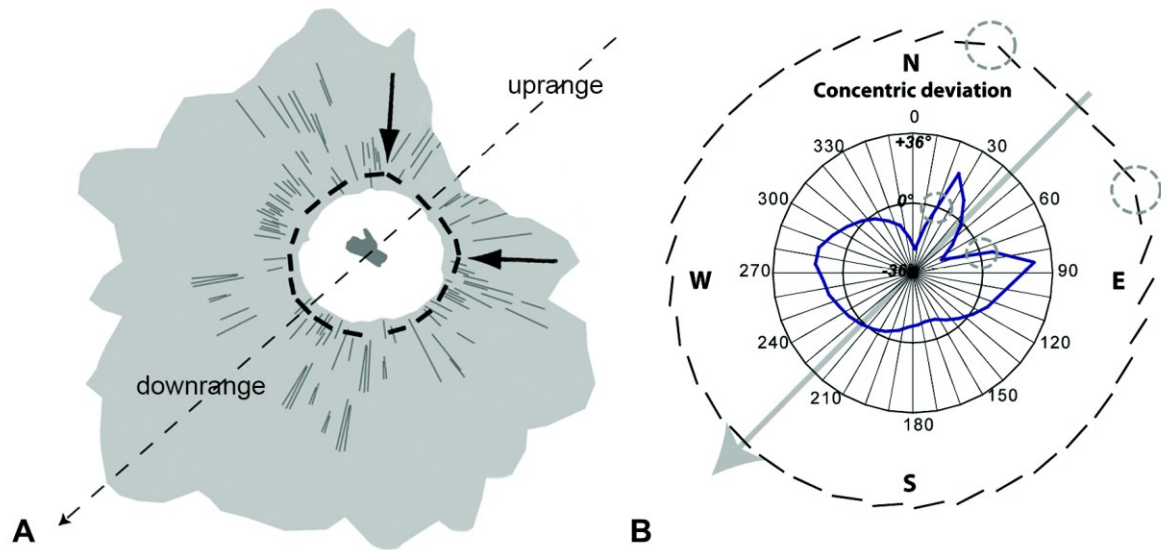


Figure 0.2: Example of the two-corner pattern formed in the ejecta and bedding for an oblique impact of less than 45° from the northeast based on a migrating flow field into a horizontal target. (A) The light grey lines represent ejecta trajectories and the black lines that display non-radial bilateral symmetry are assumed bedding strike lines constructed orthogonally to the ejecta trajectories. The black arrows show the two corners that mark the corners of the uprange area and the overturned flap. The dashed arrow shows the impactor trajectory. The grey area shows the ejecta blanket and displays a “V-shaped” zone of avoidance in the uprange zone and concentric radial ejecta trajectories in the downrange zone. (B) A polar plot showing the two-corner pattern. Black bars around the polar plot indicate the assumed bedding strike lines and are exaggerated by a factor of 3 to enhance the interpretation of the data. The grey arrow shows the impactor trajectory. The grey dashed circles indicate the sharp deviations in the strike that mark the two corners of the overturned flap of the uprange sector. Modified from [Poelchau and Kenkmann \(2008\)](#).

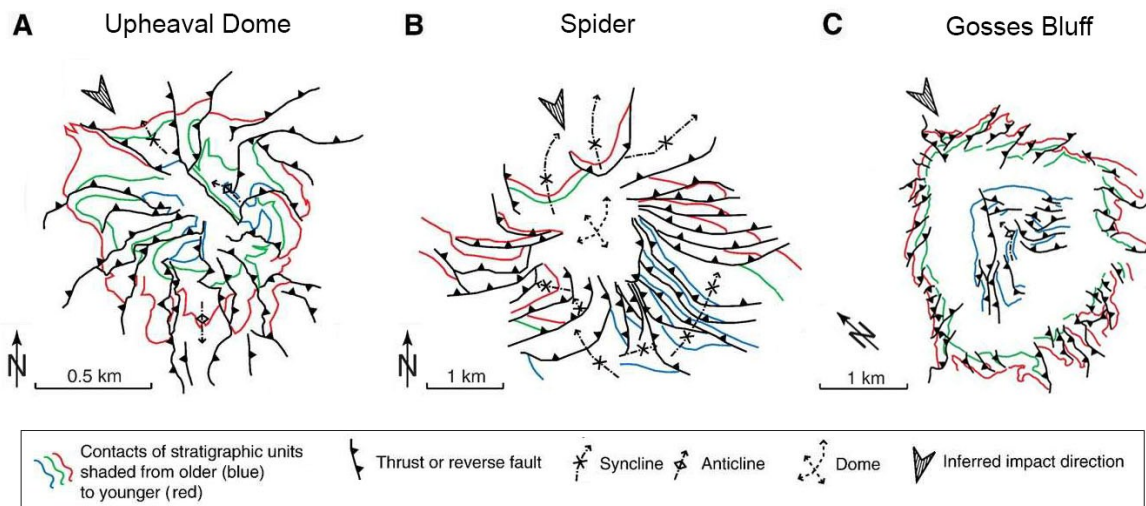


Figure 0.3: Simplified diagrams of the central uplifts of (A) Upheaval Dome, (B) Spider Structure, and (C) Gosses Bluff showing the central uplifts faults and contacts presumed to be created by an oblique impact. Modified from [Scherler et al. \(2006\)](#). The structures show enhanced shortening in the downrange direction with the thrusting direction in the top-to-downrange sector. In the case of Upheaval Dome and the Spider structure, the thrusting direction is top-to-southeast, and in Gosses Bluff, it is top-to-south.

3.1.2 Symmetry in the Vredefort Dome

The above studies inevitably involve small complex craters (mostly central peak types) that have undergone relatively limited erosion (Figure 0.3); however, due to the age of the Vredefort Dome and the depth of erosion, ejecta is not present to be studied. It will be demonstrated in [Chapter 4](#) that the collar of the Vredefort Dome shows evidence of prolonged deformation related to its rise and subsequent collapse. The question that has long been asked is whether the strong overturning of bedding ([Wieland et al., 2005](#); [Jahn and Riller, 2009](#)) is linked to a low-angle impact from the SE, placing the well-exposed NW sector of the dome in the downrange position. Others have suggested, however, that the dome was symmetric but that pre-impact strata now defining the collar may have been tilted prior to impact, creating a gross asymmetry ([Lana et al., 2003](#)); alternatively, extensive large-scale pre-impact faulting SE and S of the dome beneath the Karoo Supergroup cover may have played a significant role ([Friese et al., 1995](#)). [Andreoli and Hart \(1987\)](#) suggested that a significant NE-SW-trending fault intersects the structure, termed the southeast boundary fault, which accounts for its asymmetry. The fault was thought to have been formed at 2.8 Ga and been reactivated during impact, translating portions of the southern core northwestward ([Hart et al., 1990](#)). This superficially resembles the thrust sheets in the Spider impact structure ([Cox et al. 2021](#); Figure 0.3B). [Antoine et al. \(1990\)](#) and [Lana et al. \(2003\)](#) suggested that the dome could be divided into

polygonal straight segments separated by radial faults instead of being a circular feature. [Antoine et al.'s \(1990\)](#) interpretation was based on the patterns seen in geophysical imaging (Figure 0.5). [Wieland et al. \(2005\)](#) disproved this theory and reported that changes in bedding were on a smaller scale and that no long radial faults transecting the collar could be found at the proposed locations.

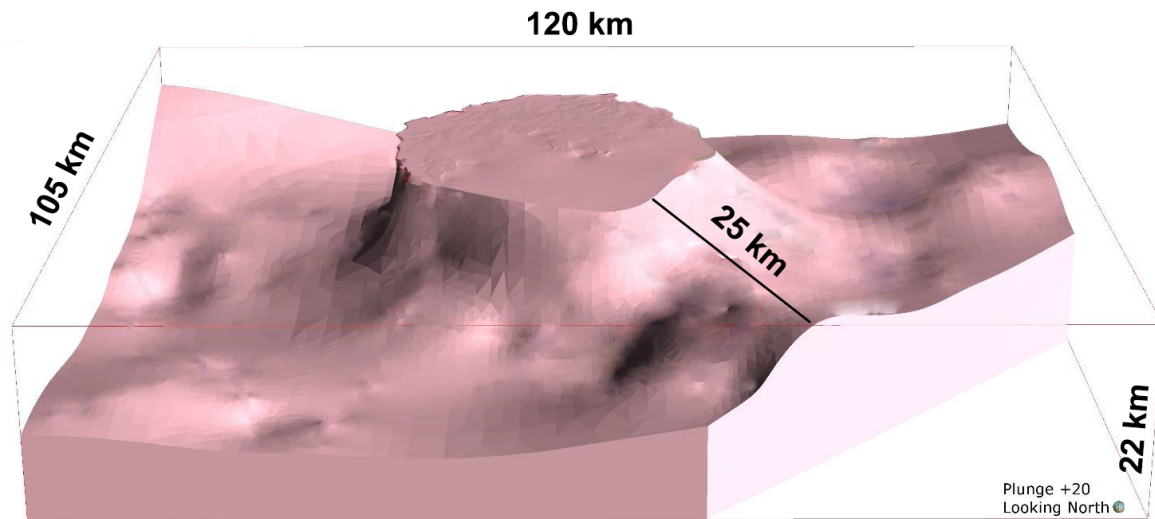


Figure 0.4: A leapfrog geological model volume created from borehole and seismic data of the Witwatersrand Basin modified from [Molezzi \(2018\)](#) showing the Q-shape in the basement-cover contact of the Vredefort Dome. The northern portion displays steep dips, whilst the southern portion shows moderate dips. The basement in the east is several kilometres shallower than in the West. The north-south diameter is ~ 46 km, whilst the east-west diameter is ~ 39 km. The basement contact ranges from 4 km to 16 km depth surrounding the dome. On the eastern margin, [Molezzi \(2018\)](#) proposed the increased elevation to result from a gentle asymmetric fold with a fold axis trending at 035°. The fold is less defined in the south as it is cross-cut by a listric fault. This fold is proposed by [Molezzi \(2018\)](#) to form during impact by the superposition of the rim syncline on the Vaal Dam Anticline formed in the pre-impact D4 Ukubambana fold-thrust belt event ([See Chapter 1.3.2](#)).

A major problem with the macrostructural analysis of the Vredefort Dome is that the southern half of the structure is buried. However, geophysical images of the subsurface reveal a “Q” or horseshoe shape of the central uplift (Figure 0.4 and Figure 0.5). This resembles some of the features in central uplifts formed by low-angle impacts characterising the uprange sector, such as structural upliftment (i.e., Upheaval Dome, [Scherler et al., 2006](#); Figure 0.3A; Figure 0.6). However, [Molezzi et al. \(2019\)](#) explained this feature by proposing that a major antiform (Vaal Dam Anticline) formed in the pre-impact D4 Ukubambana fold-thrust belt event (Transvaalide Events of [Alexandre et al. \(2006\)](#); see [Chapter 1.3.2](#)). The interference of the forming rim syncline with this anticline is proposed to form the feature ([Molezzi, 2018](#)). An alternative for

the elevated core-collar contact in the southeastern portion of the dome could be that there was a sizeable pre-impact horst similar to that found in the northern collar by [Mashabela \(2016\)](#) and Bush [\(2014\)](#) (see Figure 0.4).

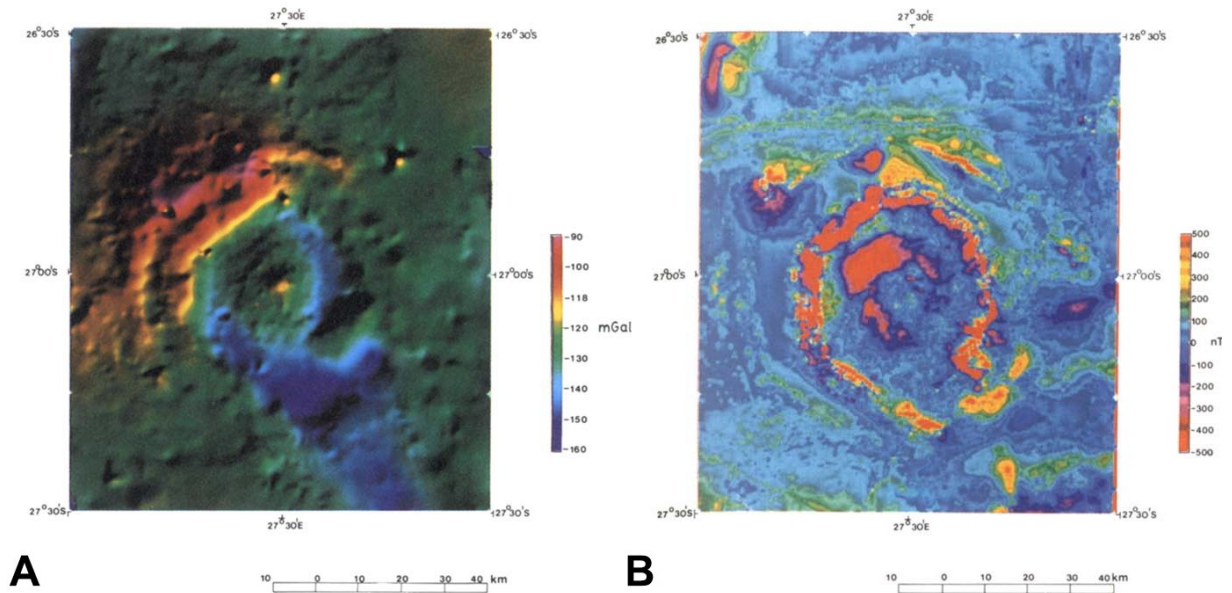


Figure 0.5: Potential field images of the Vredefort impact structure modified from [Antoine et al. \(1990\)](#). (A) Regional gravity colour image showing a southeast trending gravity low corresponding to the tail of the “Q” in the basement cover contact of the central uplift. (B) Total field magnetic intensity (TMI) map, note the polygonal shape marked by the high magnetic intensities of the Parktown Formation, West Rand Group and granulite-amphibolite transition in the basement rocks. The polygonal sectors are proposed to form by constriction during central uplift formation, with master faults forming at the corners.

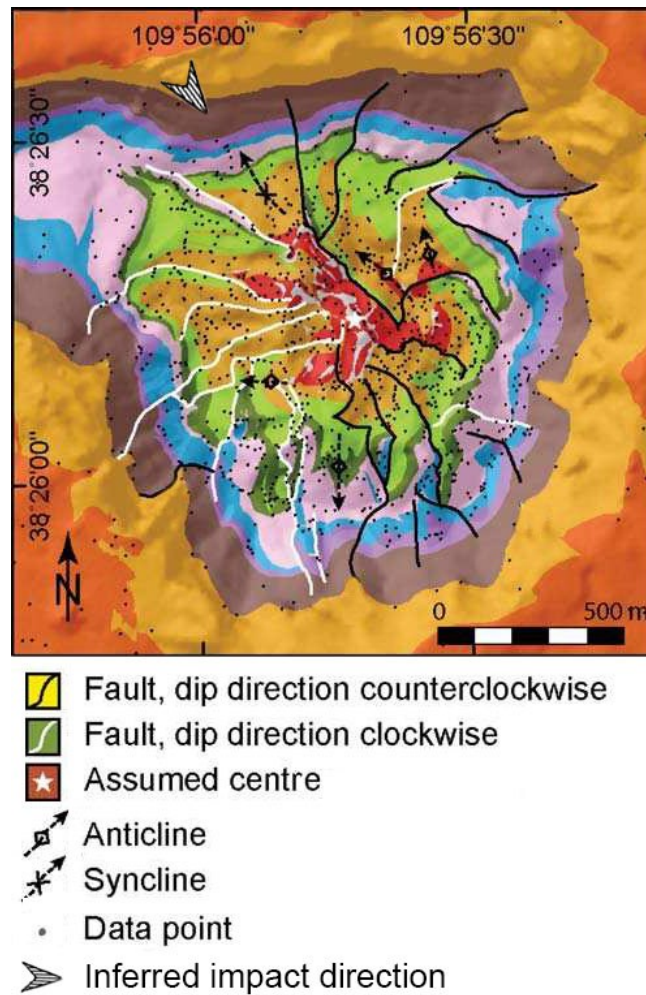


Figure 0.6: Geological map of Upheaval Dome’s central uplift showing the fault dip directions, synclines, anticlines, and the inferred impact direction from the northwest. The central uplift displays a “Q” shape, with the tail thereof denoting the uprange sector. The dip direction is categorised as clockwise or anticlockwise with respect to the centre of the dome.

3.1.3 Faults in the Vredefort Dome

Inward shallowly-dipping concentrically-striking faults occur in the rim syncline surrounding the Vredefort Dome ([Simpson, 1978](#); [Brink et al. 1997, 2000](#)) (Figure 0.7) whilst [Wieland et al. \(2005\)](#) described most of the faults in the inner collar of the dome as subvertical and striking radially and highly oblique to the collar strata, with some sector-based variation. Observed strike separations on these collar faults are up to hundreds of metres. [Wieland et al. \(2005\)](#) noted rare subhorizontal slickenlines associated with faults and concluded that the faults display essentially strike-slip movement in their present orientations. They linked the faults to the radial and oblique large-scale folds in the inner collar and, thus, to constrictional strain during the early stages of central uplift formation, i.e., slip would have been in a reverse sense and, if not true dip-slip, then steeply oblique.

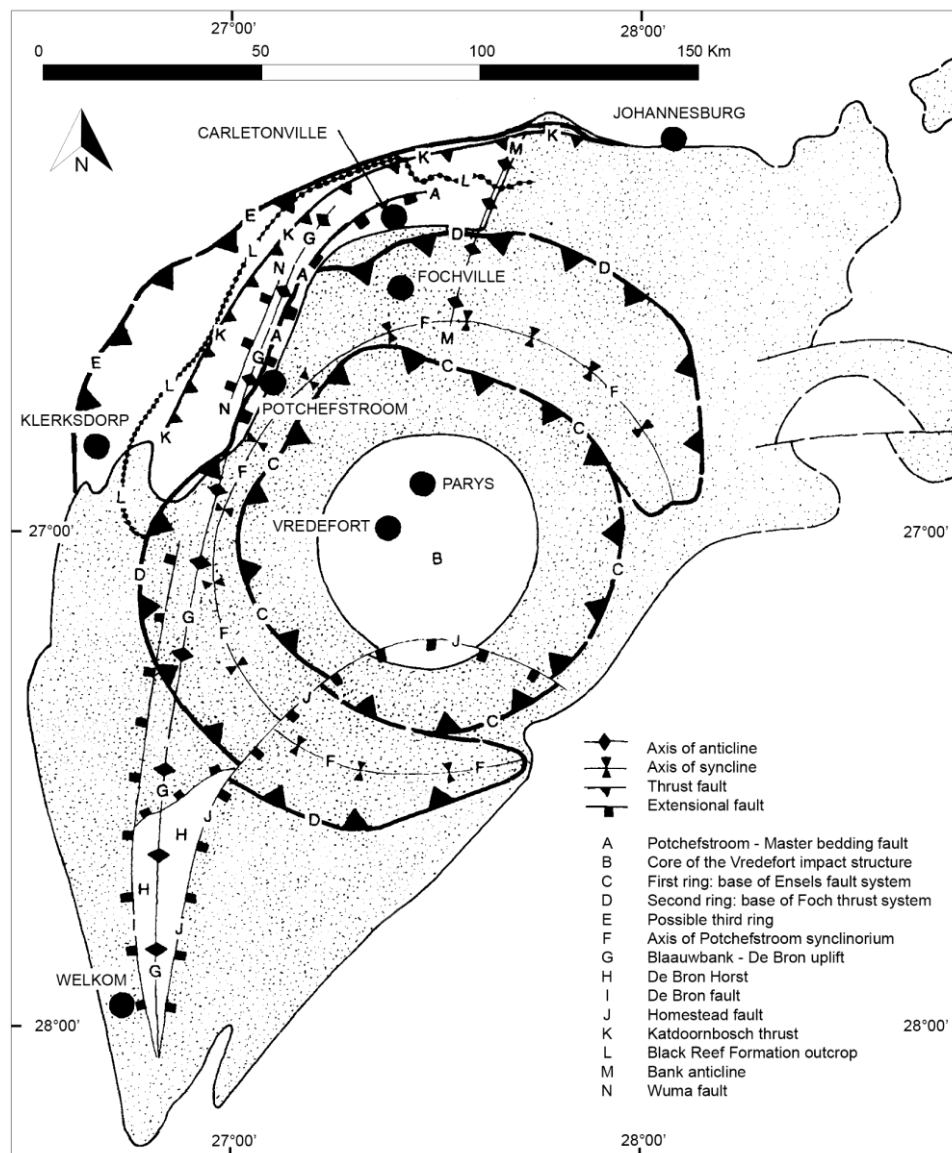


Figure 0.7: Images showing the concentric faults that dip shallowly inwards within the Vredefort rim syncline. Modified from [Brink et al. \(2000\)](#).

[Wieland et al. \(2005\)](#) excluded from their analysis of impact-related faults a conjugate fault system with km-scale strike separations in the Witwatersrand and Ventersdorp supergroups, which they interpreted as a rotated pre-impact, Ventersdorp-age, half-graben system (see also [Brink et al., 1997](#); [Mashabela, 2016](#)). The pre-impact listric faults indicate east-west extension and north-south shortening in their current geometries, while pre-impact orientations indicate a vertical shortening and tangential east-west extension. This extension is consistent with the Ventersdorp-age D3 extension/Hlukana-Platberg Event (see [Chapter 1.3.2](#)). However, [Jahn and Riller \(2009\)](#) interpreted these faults as a transpressional impact-related fault system. [Jahn and Riller \(2009, 2015\)](#) proposed that the radial faults displace older impact-related tangentially-striking, radially outward-dipping faults that show radial inward thrust

displacements. They interpreted the tangential faults as originally inward-dipping normal faults formed during the modification stage inward collapse of the TC wall that were rotated to their present orientations by the rise and ensuing outward collapse of the central uplift during the latter modification stages (Figure 0.6).

No large-scale impact-related radial or tangential faults of extensional origin have previously been identified in the inner collar of the dome. However, the Vredefort granophyre dykes (impact melt rock) are believed to follow km-scale extensional fracture zones with these orientations. [Wieland et al. \(2005\)](#) did suggest that transpressional radial faults could have been reactivated during extensional collapse but primarily associated joints and shear fractures in the quartzites with tangential and radial collapse. [Mashabela \(2016\)](#) reported that bedding offsets along individual impact-induced faults are greater in the inner collar, signifying that rocks closer to the dome's centre were subjected to more deformation during impact. In contrast, pre-impact faults showed more significant offsets in the outer collar. The dominant strike separation along the radial faults in the collar and rim synclinorium is left-lateral ([Lilly, 1978](#); [Wieland et al., 2005](#)), with subordinate normal faults ([Simpson, 1978](#); [Wieland et al., 2005](#)). Normal faults in the rim synclinorium show concentric and radial strikes relative to the dome ([Simpson, 1978](#)). [Lilly \(1978\)](#) reported conjugate fault sets in the collar of the dome. [Mashabela \(2016\)](#) supported the findings of [Lilly \(1978\)](#) and [Wieland et al. \(2005\)](#) that impact-induced fault geometries, and apparent slip directions imply a tangential maximum compressive stress, i.e., tangential shortening. These faults are, thus, associated by these authors with constriction during the inward and upward rise of the central uplift. These radial to oblique faults cut across impact-induced folds, indicating they postdate ductile impact-induced deformation ([Lilly, 1978](#); [Wieland et al., 2005](#)). The West Rand and Central Rand groups are estimated to have been shortened by approximately 7 - 17 % ([Manton, 1965](#); [Lilly, 1978](#); [Wieland et al., 2005](#)) and 5 - 7 % ([Wieland et al., 2005](#)), respectively.

Fault planes in the collar of the dome are hardly ever exposed and have generally been inferred to be vertical. This assumption has been made because the fault strike lines do not change orientation with variations in relief ([Lilly, 1978](#)). However, the map of [Bisschoff \(2000\)](#) shows numerous curved fault traces that could indicate dipping fault planes, although some cannot be attributed to topography and must, thus, indicate curved faults. It is thus possible that due to the limitations of field geologists working in map view with limited exposures from which slickensides are preserved, pertinent structural orientation data has been missed, causing the oversimplification of fault orientations in the Vredefort impact structure.

3.2 Methodology

This chapter applies a digital statistical approach to bedding and fault orientations across the collar to investigate symmetry and the possibility that the impact was oblique. This method of analysis, referred to as concentric deviation, was established by [Poelchau and Kenkmann \(2008\)](#) (Figure 0.8). This is done recognising that only about half the structure is available for study, due to rocks of the Karoo Supergroup covering the southeastern portion of the dome.

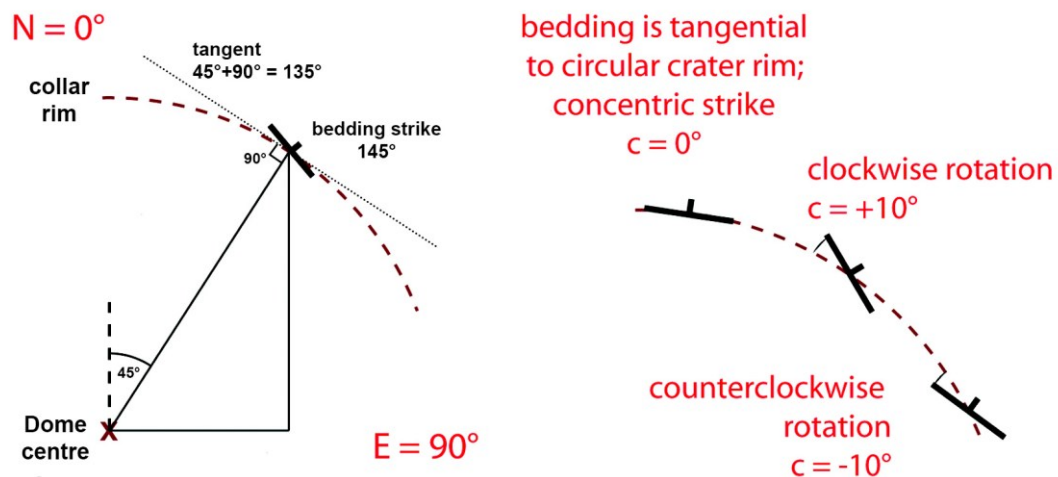


Figure 0.8: The method for analysis of concentric deviation, showing the negative concentric deviation for anticlockwise rotations and positive deviation for clockwise rotations. Modified from [Poelchau and Kenkmann \(2008\)](#).

The geological map of the Vredefort Dome ([Bisschoff, 2000](#)) was digitised and loaded into ArcGIS 10.3.1[®] where faults in the collar, as well as the Rhq3 of the WRG and Rt4 of the CRG bedding strike lines, were digitised (Figure 0.9). The bedding strike was approximated to be parallel with the mapped contacts of stratigraphic units. Segments of faults and bedding were delineated and measured using the longest possible straight-line approximation. The strike of the bedding or fault was then compared to the tangent of a circle centred at the geographic centre of the dome, at 27°1'48.35"S 27°27'27.03" E (Figure 0.8). This tangential orientation is called “concentric strike” ([Kenkmann et al., 2017](#)). The difference between the concentric and measured strikes results in a concentric deviation, which is negative for anticlockwise rotations and positive for clockwise rotations. The fault strikes were compared to both concentric deviation and radial deviation. The analysis process for the latter remains the same, but the reference angle now becomes the orientation relative to the local radial trend instead of the tangent. Clockwise rotations are positive, whilst anticlockwise rotations are negative.



Figure 0.9: Geological Map of the Vredefort Dome by [Bisschoff \(2000\)](#) with digitised fault traces shown in black. Visually, two corners appear in the collar-basement contact at 040° and 270°; however, these do not continue centrifugally. The formation of rollover anticlines associated with pre-impact D3 faults in the northern collar caused a pre-impact bedding anomaly ([see Chapter 1.3.2](#)).

The data was divided into 10 ° arc segments. Then, the data within the sector were averaged and weighted based on the length of the straight segment. The scatter in the data was further smoothed using the overlapping bin method ([Poelchau and Kenkmann, 2008](#)). The bin size used was 20° with a step of 10°. For instance, the data within a 20° range about true north were averaged from 350° - 010°, and that value was allocated to 0°. For the step size of 10°, the next value was calculated for all data between 000° - 020° and displayed at 010°, then from 010° -

030° for 020°, etc. The data were plotted as polar plots in which the negative values are located inside the x-axis and positive values outside. This plot distorts the visual depiction by exaggerating positive values and shortening negative ones ([Poelchau and Kenkmann, 2008](#)). Strike lines are drawn around the polar plots to depict fault traces and bedding strike lines. These lines are exaggerated by a factor of two, as was done in the original method by [Poelchau and Kenkmann \(2008\)](#) in order to interpret the data better. This exaggeration means the lines are twice their original size to highlight orientation changes.

3.3 Results

3.3.1 Bedding

The bedding analysis shows no noticeable throughgoing paired corners across the West Rand and Central Rand groups, nor does it show an axis of symmetry; however, there is a single sharp corner at 270° - 250° in the Rhq3 quartzite and a minor corner at 020°. The southeastern portion has sparse data due to the extensive Karoo cover sequence. A few outcrops provide data for this sector; however, this may not be representative, and the data is reduced as the overlapping bin method requires data sets on either side to generate a smoothed resultant value. Rt4 shows less deviation than Rhq3.

Concentric Deviation

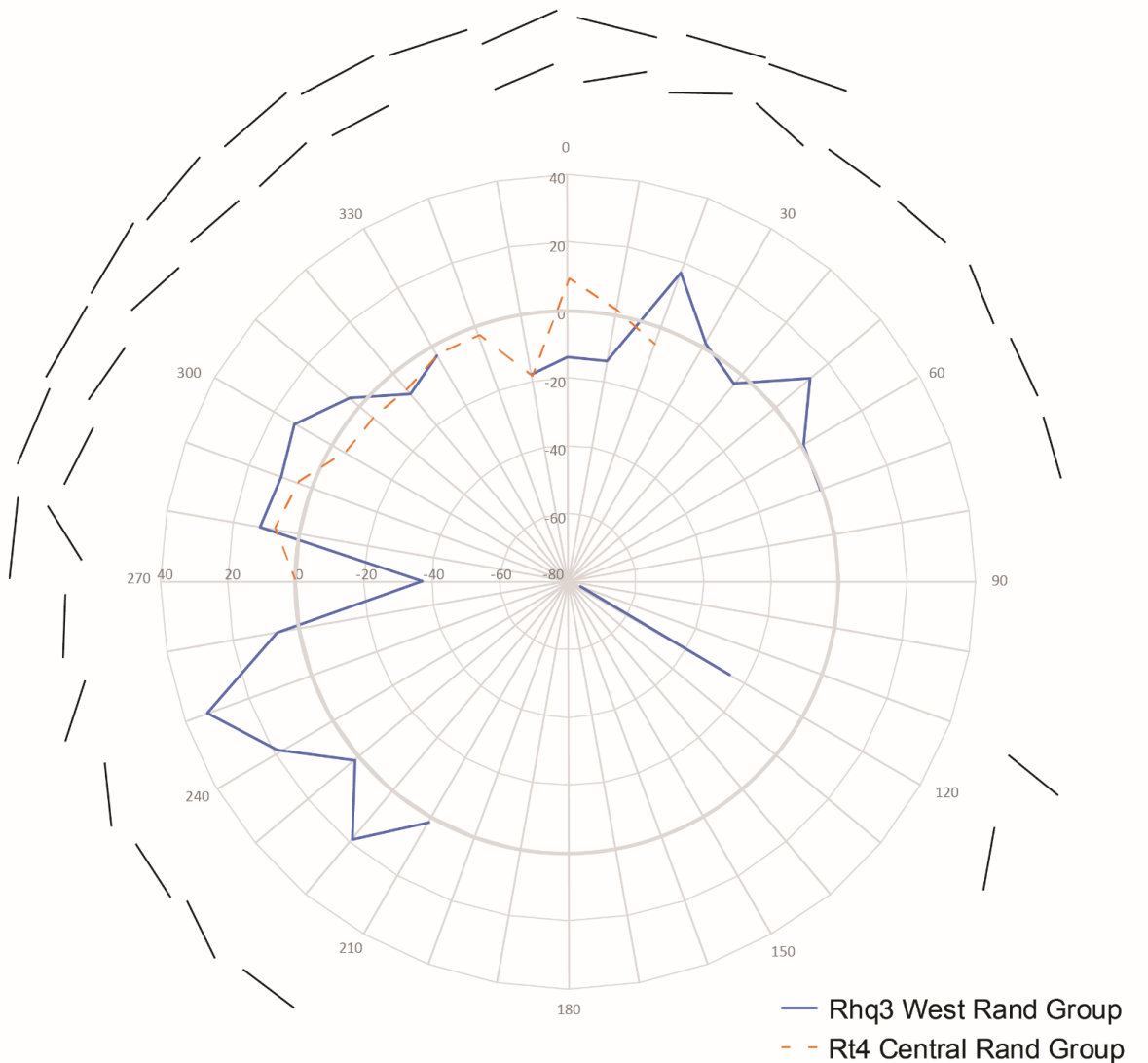


Figure 0.10: Polar plot of bedding strike deviation from the tangential orientation. Bedding strike lines are drawn around the polar plot to depict the bedding strike and are exaggerated by a factor of 2. Negative values indicate an anticlockwise rotation versus positive values, which depict clockwise rotations. The data is smoothed using an overlapping bin method (bin size 20°, with a step of 10°; see text for details). No noticeable paired corners or axis of symmetry is visible; however, there is a single sharp corner at 270° - 250° in the Rhq3 quartzite and a minor corner at 020°. The southeastern portion has sparse data due to the expansive Karoo cover sequence. Rt4 shows less deviation than Rhq3.

3.3.2 Faults

In the collar of the dome, the ratio of faults with a radial to tangential/concentric orientation is 1.2 based on fault trace length, as depicted on the geological map produced by [Bisschoff \(2000\)](#). This contradicts [Wieland et al. \(2005\)](#) and other workers who implied that there were not many bedding-parallel faults in the collar. Overall, the faults show no noticeable pattern;

however, a single sharp change in orientation is noted at 310° . The faults in the collar do not show a conjugate pattern as proposed by [Lilly \(1978\)](#). However, this analysis method would not be helpful for conjugate fault patterns as the oblique faults would destructively interfere to give a radial trend.

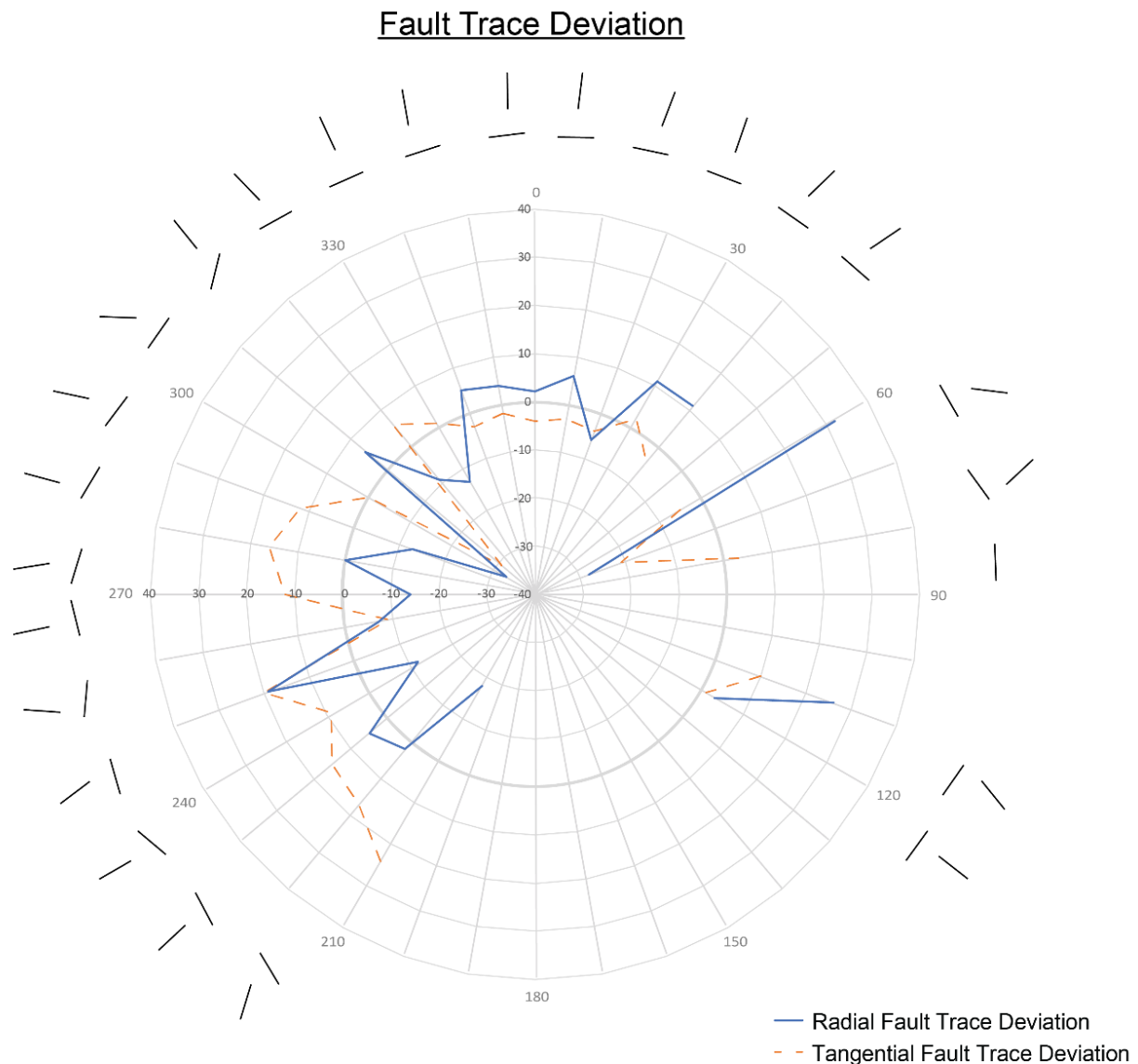


Figure 0.11: Polar plot of fault trace deviation from the radial and tangential. Black fault strike lines are drawn around the polar plot to depict the fault traces visually and are exaggerated by a factor of 2 (See text for details). Negative values indicate an anticlockwise rotation, whereas positive values depict clockwise rotations. The data is smoothed using an overlapping bin method (bin size 20° , with a step of 10° ; see text for details). No noticeable paired corners or axis of symmetry is visible. However, there is a single sharp corner at 310° . The deviation from the radial fault orientation is greater in the northwest than in the north.

3.4 Discussion

Symmetry studies of the Vredefort Dome have often been dismissed based on the erosional depth of the dome, the cover in the SE and complications due to pre-impact faulting. However, the dome has long been viewed as asymmetric based on geophysics and the extrapolation of a few right-way-up dips in the very limited outcrop in the SE sector. A suspected two corners in the core-collar contact might also be construed as indicating some sort of asymmetry. [Lilly \(1978\)](#) suggested that different sectors show different fault patterns - all encouraging signs it could be asymmetric. Comparison of the Q shape and dip asymmetry with what is known from smaller asymmetric central uplifts would lead to an inference of a SE-to-NW impact as shown in Figure 0.12.

Given that, for an asymmetric central uplift, the centre of the impact moves downrange with time, the centre of the impact could be adjusted uprange; this may improve the symmetry aspects in the bedding data. A larger bin size would further smooth the data, possibly obscuring the corners. In contrast, a smaller bin size creates too much noise in the data, making it challenging to identify corners accurately.

Maxwell's Z model ([Maxwell, 1977](#)) holds for the Vredefort Dome. Given the available outcrop at an erosional depth of 10 km, it is difficult to ascertain whether this is a result of the exposed depth in the central uplift or whether it can be attributed to impact angle. It is not possible to assess the modified Z model as, at depth, there is no ejecta material; however, no bilateral non-radial symmetry is present in the exposed collar. There is no "2 corner pattern" or cardioid shape evident within the collar rocks.

Bedding dip can also affect symmetry, although it has not been analysed as part of this chapter. Most of the bedding in the collar is overturned, with right-way-up dips reported in the SE ([Bisschoff, 2000; Wieland et al., 2005](#)) and localised right-way-up dips where there is folding (see Chapter 4). [Jahn and Riller \(2015\)](#) suggest variable dip angles in a radial direction, whilst [Wieland \(2006\)](#) reported consistent radial dips of $\sim 60^\circ$. Consistent dips around different sectors of the collar, favour symmetry, and thus, it is concluded that overturning is created by collapse as opposed to asymmetry.

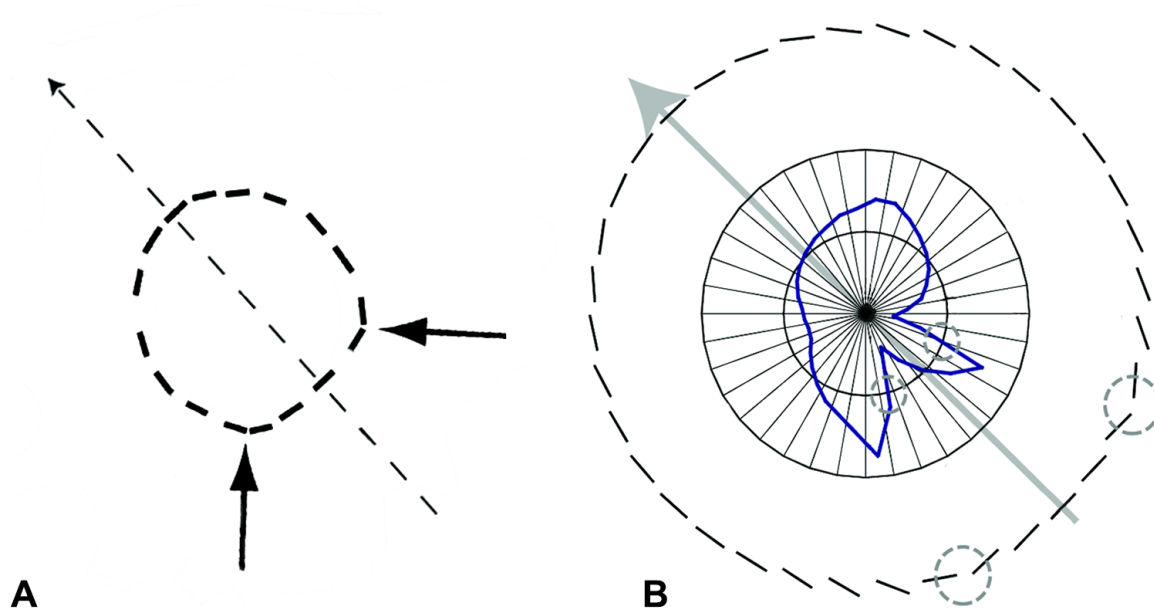


Figure 0.12: Expected bedding pattern for the Vredefort Dome assuming a low-angle impact from the SE. (A) Black lines are assumed strike lines that display non-radial bilateral symmetry. The black arrows show the two corners that indicate the uprange area. The dashed arrow shows the impactor trajectory from the SE. (B) Polar plot showing the expected two-corner pattern for a low-angle impact from the SE. Black bars around the polar plot indicate the bedding strike. The grey arrow shows the impactor trajectory from the SE.

3.4.1 Bedding Strike

The bedding strike of the thicker, quartzite-dominated CRG shows less concentric deviation than the less competent, thinner interbedded Rhq3 quartzite of the WRG (Figure 0.10). This could be due at least in part to the ability of the more heterogeneous and thinly bedded WRG to accommodate layer-parallel shortening by buckling, thereby causing more variable bedding orientations, but also to the higher amounts of constructional strain in the WRG ([Wieland et al., 2005](#); [Chapter 4](#)). No noticeable two corners or axis of symmetry is visible (Figure 0.10). There is, however, a single sharp corner at 270° - 250° in the Rhq3 quartzite and a mild corner at 020° . Deviations to the concentric bedding in the northeastern sector between 0° and 30° can be related to pre-impact faulting along the half grabens as described in [Chapter 1.3.2](#) and [Chapter 4](#); however, the overall pattern of large-scale faulting in the area is regular. The strike data available in the dome's southern portion is highly irregular and sparse. The strike of the layers in the unconformably covered section can be obtained by extracting data from the 3D leapfrog model by [Molezzi \(2018\)](#); however, the density of the control points and resolution of the model may not be sufficient to yield comparable results to those used in this study. The lack of several consistent bedding orientations over several contiguous bins or portions of the

collar contradicts [Antoine et al.'s \(1990\)](#) interpretation of a polygonised structure. The present disposition of bedding is the product of significant collapse that may have disguised or smoothed out any initial asymmetry ([see Chapter 4](#)).

3.4.2 Faults

One challenge to this analysis is that the large faults (strike length > 500 m) are not well exposed in the dome, whereas the smaller ones (< 500 m) are somewhat exposed in the quartzites. Thus, there is little control over the geometries of the lineaments drawn by [Bisschoff \(2000\)](#). The fault traces in the northern portion of the dome tend to show less deviation from both the radial and tangential orientations than the results from the limited exposures in the southern portion (Figure 0.11). The regular northern pattern is surprising as pre-impact normal faults that underwent > 90° of rotation during the impact ([Mashabela, 2016](#)) dominate in terms of length and displacement in the area. These pre-existing faults would have been expected to have affected the nature of faulting in the region, causing a distinction from the areas where prior faulting was absent. This is not apparent and may be because pre-impact metamorphism annealed and re-sealed the fault zones. Even though the pre-impact faults are conjugate, the small bin size does not cause them to average out to a radial orientation. It is challenging to ascertain a pattern for the southeastern portion of the dome due to the minimal outcrop.

[Jahn and Riller \(2009\)](#) reported that the fault pattern changes systematically in a radial sense across the collar, with radial orientations being replaced by concentric orientations towards the centre of the dome due to changes in sedimentary layer thickness and pre-impact tectonics; this study supports this. Tangential/concentrically striking faults are likely under-represented as they are more difficult to identify as they run parallel to bedding. However, the data suggests that the dominant fault orientation is still tangential by 20 % (i.e., for each 100 m of radially oriented fault strike, one can find 120 m of tangentially oriented fault strike). Approximately a third of the faults, by length, in the northwestern portion of the dome can be attributed to pre-impact faulting; however, because of their curvilinear nature, they tend to be equally distributed between the tangential and radial orientation and thus do not distort the data dramatically. Given an impact from the SE, it would be expected to find major radial faults at the ends of relatively straight segments of bedding. [Antoine et al. \(1990\)](#) proposed that the radial master faults at the corners of polygonal segments can also be ruled out as there is no evidence of long radial faults dominating in any given bin. Since faults are three-dimensional features, this two-dimensional analysis is a very limited attempt at comparing symmetry, as the third dimension

is significant in impact structures. The preserved faults are a combination of both constriction and collapse (See [Chapters 4](#) and [6](#)), and analysing the two sets of faults together may obscure the results. [Lilly \(1978\)](#) reported conjugate fault sets in the collar of the dome. This study does support the formation of conjugate fault sets in places ([see Chapter 6](#)), yet the overall fault patterns are not conjugate. [Lilly \(1978\)](#) also noted fault strike orientations in the WRG that were either radial or oblique to the core-collar contact but proposed that each type dominated in a different sector of the collar (N and NW, respectively).this study supports these findings as Figure 0.11

3.5 Conclusions

Based on bedding strike or fault patterns, no bilateral symmetry can be detected in the Vredefort Dome. This may be due to the lack of data in the southwestern portion of the dome that is covered by the Karoo cover sequence.

The collar does not display a polygonal pattern as proposed by [Antoine et al. \(1990\)](#).

No radial master faults can be detected at corners, as proposed by [Antoine et al. \(1990\)](#).

Overall, faults within the collar do not show conjugate patterns, as [Lilly \(1978\)](#) proposed.

Tangential faults dominate within the collar.

Chapter 4

PAPER I: DYNAMICS OF COLLAPSE OF AN IMPACT CENTRAL UPLIFT: EVIDENCE FROM FOLDS AND FAULTS IN THE COLLAR OF THE VREDEFORT DOME, SOUTH AFRICA

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My contribution to this paper as the first author involved conceptualising the dynamic collapse model based on the structural data I collected and geological mapping done in the field. I also did the necessary rotations to test the polyphase folding hypothesis and modelled the field areas in 3D using Leapfrog Geo® 5.1. I created the geological maps of the study areas and compiled most of the text and images in the paper with the assistance of my supervisors. The only modifications made from the original paper to maintain consistency with the thesis are updating text section numbering, figure numbering, and the spelling of various words that have been changed from American English to British English, and consolidation of the references into the main [Reference List](#). The spelling of pseudotachylite has also been updated to pseudotachylyte to be consistent with the rest of the thesis (see [Chapter 6](#)). To maintain the integrity of the paper, some diagrams used in other chapters are duplicated, with new figure numbers.

Abstract

Structural analysis of overturned metasedimentary strata of the lower Witwatersrand Supergroup in the inner collar of the Vredefort Dome reveals the presence of tangential folds and faults associated with the 2.02 Ga impact. The folds are distinct from previously identified subradially oriented, vertical to plunging-inclined, gentle folds that are interpreted as the products of convergent flow (constriction) during the initial stages of central uplift formation.

The tangential folds comprise disharmonic, open, asymmetric, horizontal to plunging-inclined anticline-syncline pairs with centripetally dipping axial planes and right-way-up intermediate limbs. They display centripetal-down vergence (anticline radially outward of the syncline) that is consistent with steep inward-directed shear of the overturned strata. We attribute this kinematic pattern to subvertical collapse of the Vredefort central uplift during the latter stages of crater modification. The folds are cut by pseudotachylyte-bearing steep to vertical tangential faults that display centre-down slip of < 10 m up to ~ 150 m. Both the tangential folds and the faults suggest that the large-scale overturning of strata related to outward collapse of the Vredefort central uplift was accompanied by a component of inward-directed collapse via layer-parallel shearing and folding, followed by faulting. Subradially oriented faults with conjugate strike separations of 1-2 km in the NNE collar of the dome suggest penecontemporaneous tangential extension of the inner collar rocks. This evidence indicates that second order structures in the metasedimentary collar of the Vredefort Dome preserve a complex, multistage record of evolving strain associated with both initial convergent and upward flow (constriction) related to central uplift rise and later divergent and downward flow (flattening) linked to its collapse, and that centripetally directed collapse features may be important components of the structural inventory of very large central uplifts.

4.1 Introduction

Detailed mapping of the internal structure of partially eroded large terrestrial impact craters provides important insights into the complex kinematics and dynamics of crater formation. Whereas the distal parts of such craters, which are not subjected to significant shock pressures and that experience primarily radially inward collapse, display relatively predictable (albeit locally complicated) fault displacements and block rotations, the proximal volume experiences a more extreme, prolonged and complex stress-strain history ([Kenkmann et al., 2014](#); [Rae et al., 2019](#)). Owing to the high strain rates associated with crater formation, it is expected that much of the strain response is accommodated by brittle failure and faulting at a range of scales; however, numerous crater structures also preserve evidence within their central domains of folded layering (e.g., [Kriens et al., 1999](#); [Kenkmann, 2002](#); [Lana et al., 2003](#); [Kenkmann and Poelchau, 2009](#); [Kenkmann et al., 2010, 2011](#)) that implies a quasi-ductile strain component.

Folds with wavelengths up to several hundred metres to kilometres have been described from the central uplifts of several partially eroded complex craters (e.g., Sierra Madera, [Wilshire and Howard, 1968](#); Upheaval Dome, [Kriens et al., 1999](#); Araguinha, [Lana et al., 2008](#); Serra de

Cangalha, [Kenkmann et al., 2011](#)). Their broadly radial orientations and their decreasing wavelength and increasing amplitude and tightness towards the centre of the impact structure indicate a constrictional strain field driven by the convergent mass flow that forms the central uplift ([Kenkmann et al., 2014](#)). In several crater structures, non-radial fold and thrust fault orientations are a crucial component in defining bilaterally symmetric central uplifts that form as a result of highly oblique impacts (e.g., Matt Wilson, [Kenkmann and Poelchau, 2009](#); Jebel Waqf as Suwwan, [Kenkmann et al., 2010, 2017](#)).

In smaller (central peak) complex craters constrictional strains driven by convergent flow may persist throughout much of the rise of the central uplift. However, in larger craters, vertically-directed buoyancy forces play an increasingly significant role in central uplift evolution (e.g., [Collins et al., 2004, 2008](#); [Morgan et al., 2016](#); [Rae et al., 2019](#)). These forces drive the central uplift above equilibrium height, ultimately leading to collapse that produces an hourglass-shaped central uplift profile as a result of radial outward flow of the upper parts of the central uplift. The surface manifestation of this collapse is a peak ring ([Kenkmann et al., 2000](#); [Collins et al., 2004](#); [Morgan et al., 2016](#); [Rae et al., 2019](#)). Geophysical interpretations, structural observations and modelling in the Chicxulub crater (e.g., [Riller et al., 2018](#); [Rae et al., 2019](#)) have confirmed the considerable outward lateral translation of mechanically weakened peak ring material in such large craters, assisted by low-angle extensional structures. However, in general, only passing reference is made in the literature to outcrop-scale structures that can be unequivocally linked to central uplift collapse ([Osinski and Spray, 2005](#); [Lana et al., 2008](#); [Kenkmann et al., 2011](#)), possibly owing to the limited number of well-exposed peak-ring craters in layered targets that are capable of preserving such evidence.

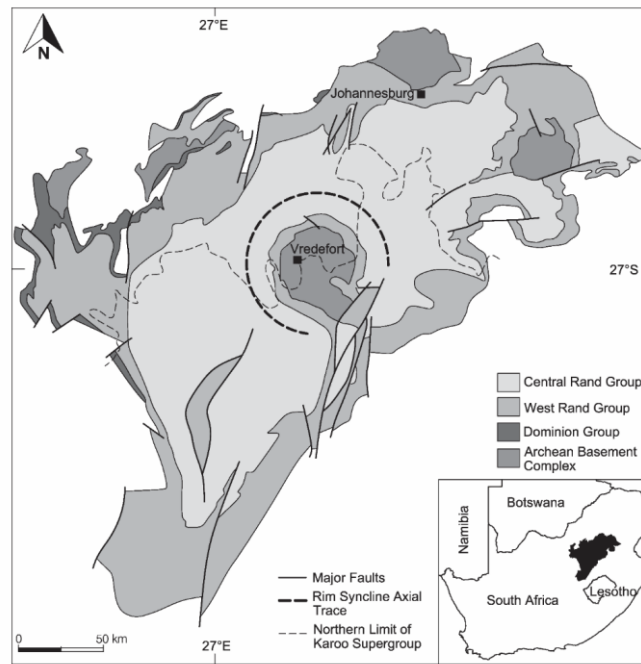


Figure 0.1: Geological map of the Witwatersrand Basin, stripped of all cover sequences, showing the location of the Vredefort Dome ~ 120 km southwest of Johannesburg. The axial trace of the rim syncline (Potchefstroom Synclinorium) and the northern limit of the Karoo Supergroup are also shown. Modified after [McCarthy \(2006\)](#).

In this paper we describe fold and fault structures from the interior of the world's largest and most deeply eroded central uplift, the Vredefort Dome, whose kinematic interpretation is consistent with formation during central uplift collapse. Contrary to the general view that collapse-flow involved only overturning related to subhorizontal, centrifugal escape, we infer from the orientation and kinematic indicators associated with these folds and faults that outward-directed overturning in the inner parts of the Vredefort central uplift was also accompanied by a component of steep inwards collapse that was at least in part channelled by the steeply overturned target layering.

4.1.1 Geological Setting

The centre of the Vredefort Dome is located ~ 120 km southwest of Johannesburg, South Africa (Figure 0.1). The dome (Figure 0.2) is the largest exposed impact central uplift on Earth, although its southern half is obscured beneath younger sediments. It comprises an ~ 40 km wide core of high-grade Mesoarchean gneisses that is surrounded by a 20 - 25 km wide annular collar of mainly Neoarchean to Paleoproterozoic supracrustal strata, metamorphosed at low to medium grades, which display generally steeply overturned to subvertical dips. The dome is surrounded by an asymmetric, ~ 50 km wide, rim syncline (Figure 0.1) with a steeply dipping

overturned to right-way-up inner limb and a shallow inward-dipping outer limb ([Simpson, 1978](#); [Grieve et al., 2008](#)). Reflection seismic imaging (e.g., [Henkel and Reimold, 1998](#); [Molezzi et al., 2019](#)) indicates that the unconformity marking the core-collar contact reaches a maximum depth of 12 km in the synclinal trough. The crater has undergone significant erosion, estimated at between 8 and 10 km ([McCarthy et al., 1990](#); [Gibson et al., 1998](#)). Consequently, the original crater diameter is unknown; however, estimates range from ~ 180 km ([Grieve et al., 2008](#)) to 250 - 300 km ([Therriault et al., 1995](#); [Henkel and Reimold, 1998](#)).

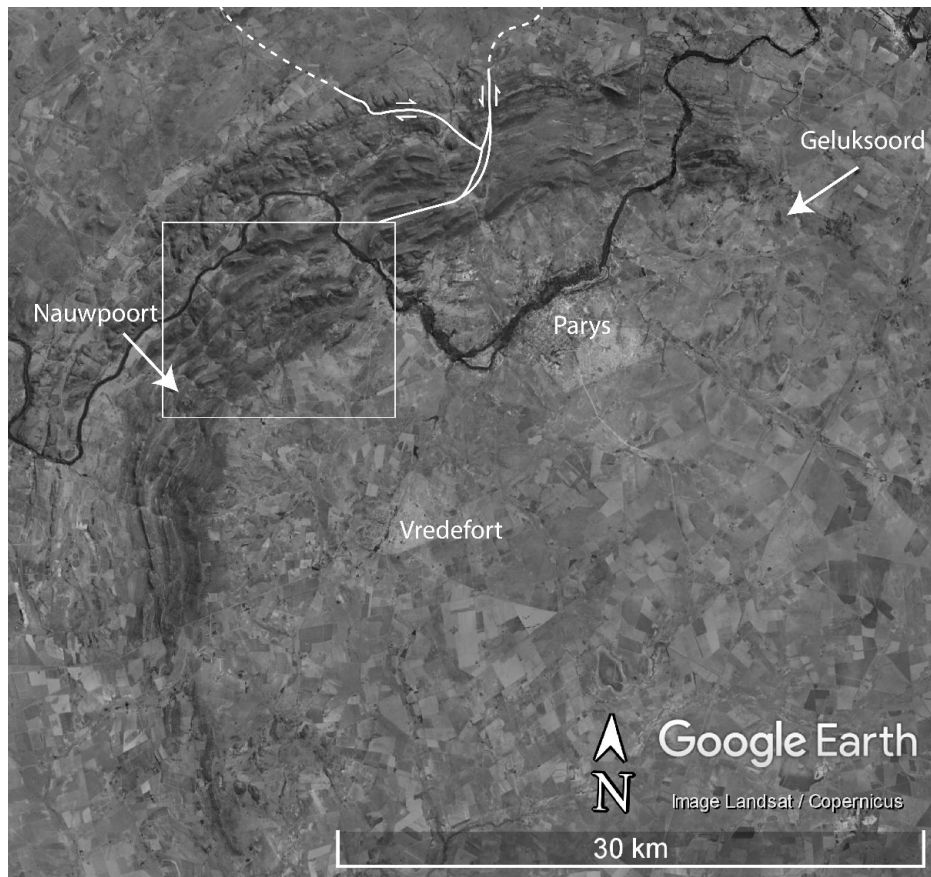


Figure 0.2: Google Earth image of the exposed portion of the Vredefort Dome showing the broadly polygonal geometry of the inner collar defined by alternating ridges and valleys. The topographically rugged collar surrounds the more subdued core of the dome, which is underlain by granitoid gneisses. The ridge-and-valley topographic expression of the West Rand Group (alternating quartzites, metapelites and meta-ironstones) in the innermost collar differs from that of the massive quartzites of the Central Rand Group (rugged highly dissected ridge with narrow valleys) and the mafic lavas of the Ventersdorp Supergroup (subdued topography, top left). Also shown (from [Mashabela, 2016](#)) are the principal faults of a conjugate fault system in the northern collar interpreted as a rotated Neoproterozoic half-graben by [Brink et al. \(1997\)](#), [Wieland et al. \(2005\)](#) and [Mashabela \(2016\)](#) but as an impact-related transpressional fault system by [Jahn and Riller \(2009\)](#). The area covered in Figure 0.3 is indicated by the rectangle and the locations of the Nauwpoort and Geluksoord study sites are shown. (Imagery date 07/10/2019)

The supracrustal sequence in the collar of the dome comprises an ~ 15 km stratigraphic thickness of sedimentary and volcanic rocks, with some mafic and granitic intrusions ([Bisschoff, 2000](#); [Gibson and Reimold, 2008](#)). The oldest unit consists of basalts (Dominion Group; 3.07 Ga; [Armstrong et al., 1991](#)). It was followed by the predominantly sedimentary (quartzite, shale, ironstone $\pm$ conglomerate) Witwatersrand Supergroup (2.99 - 2.85 Ga; [Kositcin and Krapež, 2004](#)), the basaltic Ventersdorp Supergroup (2.79 Ga; [Gumsley et al., 2019](#)), and the 2.67 - 2.05 Ga ([Hofmann, 2011](#)) Transvaal Supergroup (dolomite, quartzite, shale). The Witwatersrand Supergroup comprises the lithologically heterogeneous WRG and the quartzite-dominated CRG that give rise to the distinctive annular topography of the inner collar (Figure 0.2, Figure 0.3).

Evidence of impact in the Vredefort Dome includes shatter cones (e.g., [Hargraves, 1961](#); [Albat and Mayer, 1989](#)), planar deformation features in quartz (e.g., [Lilly, 1981](#); [Fricke et al., 1990](#); [Grieve et al., 1990](#)) and zircon ([Kamo et al., 1996](#); [Gibson et al., 1997](#)), high-pressure polymorphs of silica ([Martini, 1978, 1991](#)) and zircon ([Kovaleva et al., 2019](#)), and recrystallised diaplectic glass ([Gibson and Reimold, 2005](#)). Maximum shock pressures exceeded 30 - 35 GPa in the central core ([Gibson and Reimold, 2005](#)) and reached 10 - 15 GPa in the inner collar ([Lilly, 1981](#)). Impact-related melt rocks include ([Reimold and Gibson, 2006](#)) rare, volumetrically minor, shock melt veins ([Martini, 1991](#)), allochthonous, kilometres-long dykes of impact melt rock (e.g., [Therriault et al., 1996](#); [Reimold et al., 1990](#); [Koeberl et al., 1996](#); Reimold et al., this volume, and references therein) and ubiquitous mm- to decametre-wide pseudotachylytic breccias ([Dressler and Reimold, 2004](#); [Gibson and Reimold, 2008](#)). A 2.023 ± 0.004 Ga age for the impact was determined from single-zircon U-Pb dating of these melt rocks ([Kamo et al., 1996](#)).

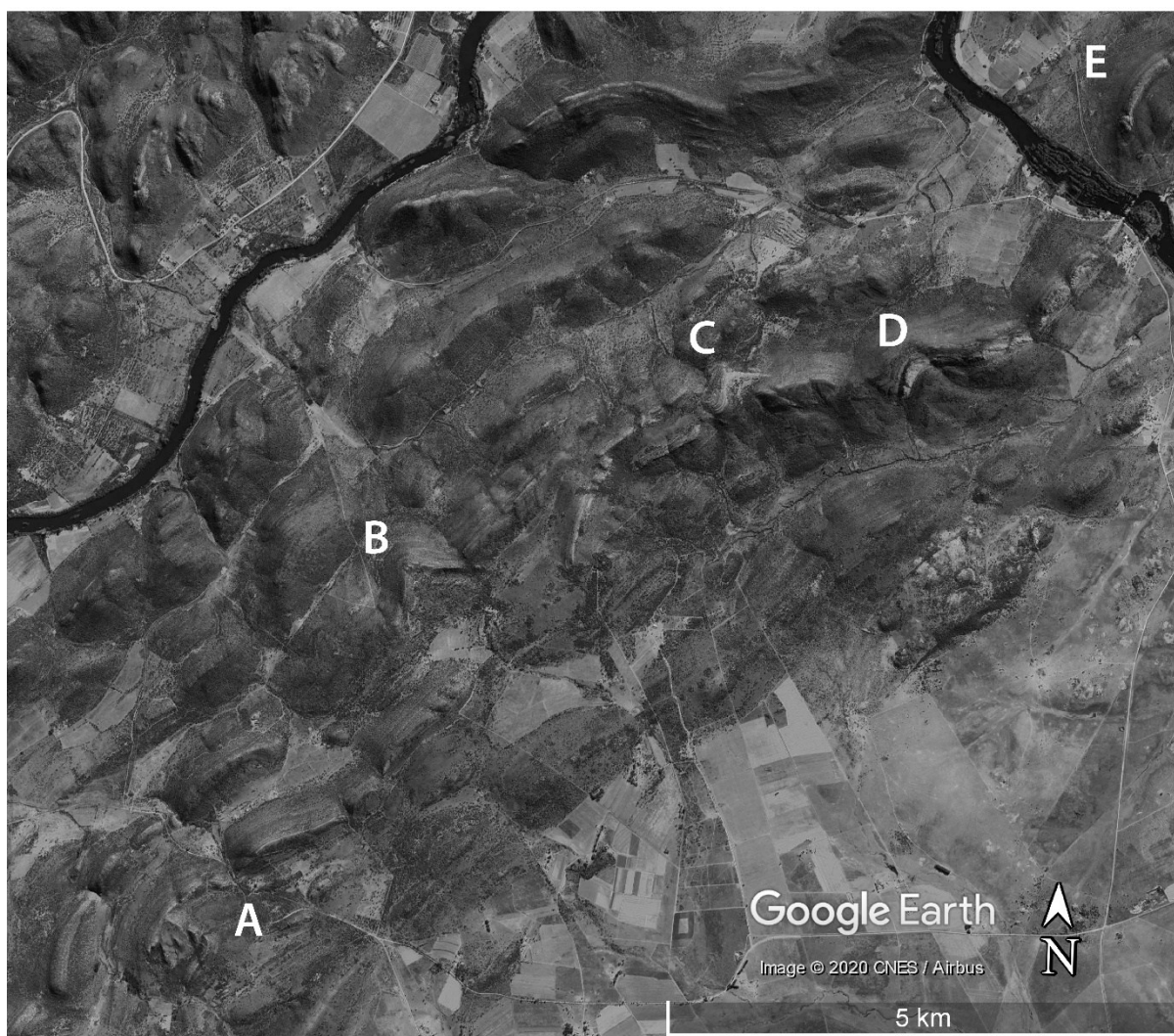


Figure 0.3: Google Earth image of the northwestern portion of the inner collar of the Vredefort Dome, showing general overturned centripetal dips of the supracrustal collar strata and large-scale folds in the Rhq3 (A, C, D, E) and Rgq3 quartzites (B) of the upper Hospital Hill and lower Government subgroups, respectively. Folds labelled A-E were studied by [Wieland et al. \(2005\)](#); B-E are shown in Figure 0.4 whereas A corresponds to the Nauwpoort locality discussed in this study. The relatively straight quartzite segments that individually stretch over several kilometres are disrupted by minor faults and displaced by up to several hundreds of metres by larger radial and oblique faults. (Imagery date 07/10/2019)

4.1.2 Structural Features of the Vredefort Central Uplift and Rim Syncline

In a comparative analysis of the Vredefort, Sudbury and Chicxulub impact structures, [Grieve et al. \(2008\)](#) interpreted most of the large-scale outcrop and geophysically-identified structural elements in the Vredefort Dome and its rim syncline as being consistent with a deep section through an ~ 180 km wide peak ring crater. They proposed that inward-dipping thrust faults in the outer collar of the dome ([Brink et al., 1997](#)) and outward-dipping thrust faults originating as splays off shallowly inward-dipping normal-slip faults beyond the rim syncline hinge meet

in a zone of convergence at $R \sim 40 - 50$ km that marks the root of the peak ring (see Figure 0.12A, B). The Transvaal Supergroup rocks in the vicinity of the rim syncline hinge display gentle to open, km-scale, slightly outward-verging, tangential folds with upright to inward-dipping axial planes that are cut by vertical to shallowly inward-dipping tangential faults ([Simpson, 1978](#)).

Within the collar of the dome, differential erosion highlights the principal structural features of the central uplift, particularly in the lithologically heterogeneous WRG in the inner collar (Figure 0.2, Figure 0.3). Topographic ridges underlain by quartzites emphasize the overall polygonal geometry, with relatively straight to slightly curved segments up to 10 - 15 km in length separated by radial to oblique faults showing up to hundreds of metres of strike separation. Most strata are overturned, dipping at $\sim 60^\circ$ towards the centre, with no obvious variation in dip angle noted between the sectors ([Wieland et al., 2005](#)), although steep normal outward dips, and overturned dips as shallow as 30° , are found locally within individual sectors ([Bisschoff, 2000](#); [Wieland et al., 2005](#); [Jahn and Riller, 2009](#)). Overall, overturning appears to increase outwards across the inner collar, and with increasing topographic elevation ([Bisschoff, 2000](#); [Jahn and Riller, 2009](#)). The inflexion point below which the collar strata show general right-way-up dips is estimated to lie ~ 5 km below the present surface ([Jahn and Riller, 2015](#); [Molezzi et al., 2019](#)).

Based on the reconstruction of the stratigraphic arc length at different radial distances in the collar, [Wieland et al. \(2005\)](#) estimated that tangential shortening decreased from $\sim 17\%$ in the lower WRG in the inner collar to $\sim 5\%$ in the CRG. This trend is corroborated by [Simpson's \(1978\)](#) observation that there is no evidence of bulk tangential shortening in the lower Transvaal Supergroup at the outer edge of the collar. Similar values for the inner collar were obtained by [Jahn and Riller \(2015\)](#) but are slightly lower than the 25% estimate obtained from numerical modelling based on general structural, shock and metamorphic parameters ([Ivanov, 2005](#)). [Wieland et al. \(2005\)](#) proposed that this tangential shortening in the inner collar was accommodated by km-scale radial folds and radial to oblique faults (Figure 0.3, Figure 0.4). [Jahn and Riller \(2015\)](#) linked the folds and localised differential block rotations to a series of inferred radial transpression zones within the collar that they proposed were the principal mechanism by which the tangential shortening occurred. They also proposed that these radial transpressional faults truncate several outward-dipping thrust faults in the collar (see Figure 0.12B). They interpreted these tangential faults as original inward-dipping normal faults that formed during collapse of the TC and that became rotated to their present orientation during

central uplift rise and collapse. [Jahn and Riller \(2009\)](#) identified several broadly radial faults with km-scale offsets in the northern part of the collar as the largest example of these impact-induced transpression systems. This interpretation differs from that of [Brink et al. \(1997\)](#), [Wieland et al. \(2005\)](#), and [Mashabela \(2016\)](#), who noted that these faults do not cut the Transvaal Supergroup and, consequently, interpreted them as part of a regional NNE-striking, half-graben fault system of Neoproterozoic age that became rotated into its present geometry by the impact.

[Lana et al. \(2003\)](#) used foliation patterns in basement gneisses to show that the overturning seen in the collar strata continues up to ~ 8 km into the core of the dome, but that the gneisses in the central core ($R \leq 12$ km) were not significantly rotated. This differential rotation is reproduced in numerical modelling of the Vredefort impact ([Ivanov, 2005](#)). In the absence of evidence of significant fault zones in the gneisses, [Lana et al. \(2003\)](#) speculated that differential movements and rotations were accommodated within a 3D network of mm- to cm-scale pseudotachylytic breccias and that gross ductile strain may also have been further facilitated by high post-shock temperatures (> 550 °C; [Gibson and Reimold, 2008](#)). However, [Lieger et al. \(2009\)](#) showed that most of the large pseudotachylytic breccia dykes in the core of the dome show geometries consistent with late-stage radial and tangential extension of the core of the dome, raising questions about the exact timing of melt formation/emplacement.

[Wieland et al. \(2005\)](#) and [Jahn and Riller \(2009, 2015\)](#) proposed that radial faults in the collar show predominantly strike-slip movement related to constriction, although [Wieland et al. \(2005\)](#) did not rule out subsequent collapse-related reactivation. [Wieland et al. \(2005\)](#) attributed joints and shear fractures in the collar rocks primarily to this late-stage radial and tangential collapse, although they acknowledged that the joints and shear fractures might equally reflect post-impact crater adjustments. Central uplift collapse and/or post-cratering adjustment was also proposed to explain the radial and concentric orientations of the vertical impact melt dykes in the core of the dome (e.g., [Therriault et al., 1996](#); [Gibson and Reimold, 2008](#), and references therein).

In summary, fault and fold features in the collar of the Vredefort Dome have been primarily interpreted as reflecting convergent flow during the rise of the central uplift, which produced moderate amounts (≤ 20 %) of constrictional strain in the innermost collar, and which decreased outwards. These estimates may be complicated by evidence of large-scale pre-impact extensional faulting in the northern collar. Features proposed as being related to collapse of the central uplift include overturning of strata, low-angle inward-dipping thrusts

inferred from seismic sections, and centrifugally-verging, km-scale, folds with tangential axial traces in the hinge of the rim syncline. Pseudotachylytic breccia and impact melt dykes in the core of the dome and radial joints and tangential, listric, outward-dipping shear fractures in quartzites in the inner collar are attributed to extensional radial and tangential strains in the latter stages of, and possibly also following, central uplift formation.

4.2 Background and Methods

As part of a larger analysis of impact-related structures within the Witwatersrand Supergroup in the Vredefort Dome, detailed structural mapping was performed on portions of the farms Nauwpoort and Geluksoord, respectively in the WNW and NNE sectors of the inner collar of the dome (Figure 0.2). These sites were selected because they show evidence of shallowly-dipping, right-way-up strata, which is highly unusual in the collar of the dome. The former locality was only partially studied by [Wieland et al. \(2005\)](#), owing to limited access at the time, whereas the latter has not been described previously, although [Bush \(2014\)](#) described highly faulted subhorizontally-oriented quartzite bedding 2 km to the north of the Geluksoord site studied here.

The two localities lie between 21 and 22 km from the centre of the dome and are located in the Hospital Hill Subgroup, the lowermost portion of the WRG (lower Witwatersrand Supergroup; [McCarthy, 2006](#)). The Hospital Hill Subgroup comprises a layered succession of quartzites, metapelites and meta-ironstones that were metamorphosed to amphibolite facies prior to the impact, but which also show a lower amphibolite facies overprint linked to the impact ([Gibson and Wallmach, 1995](#)).

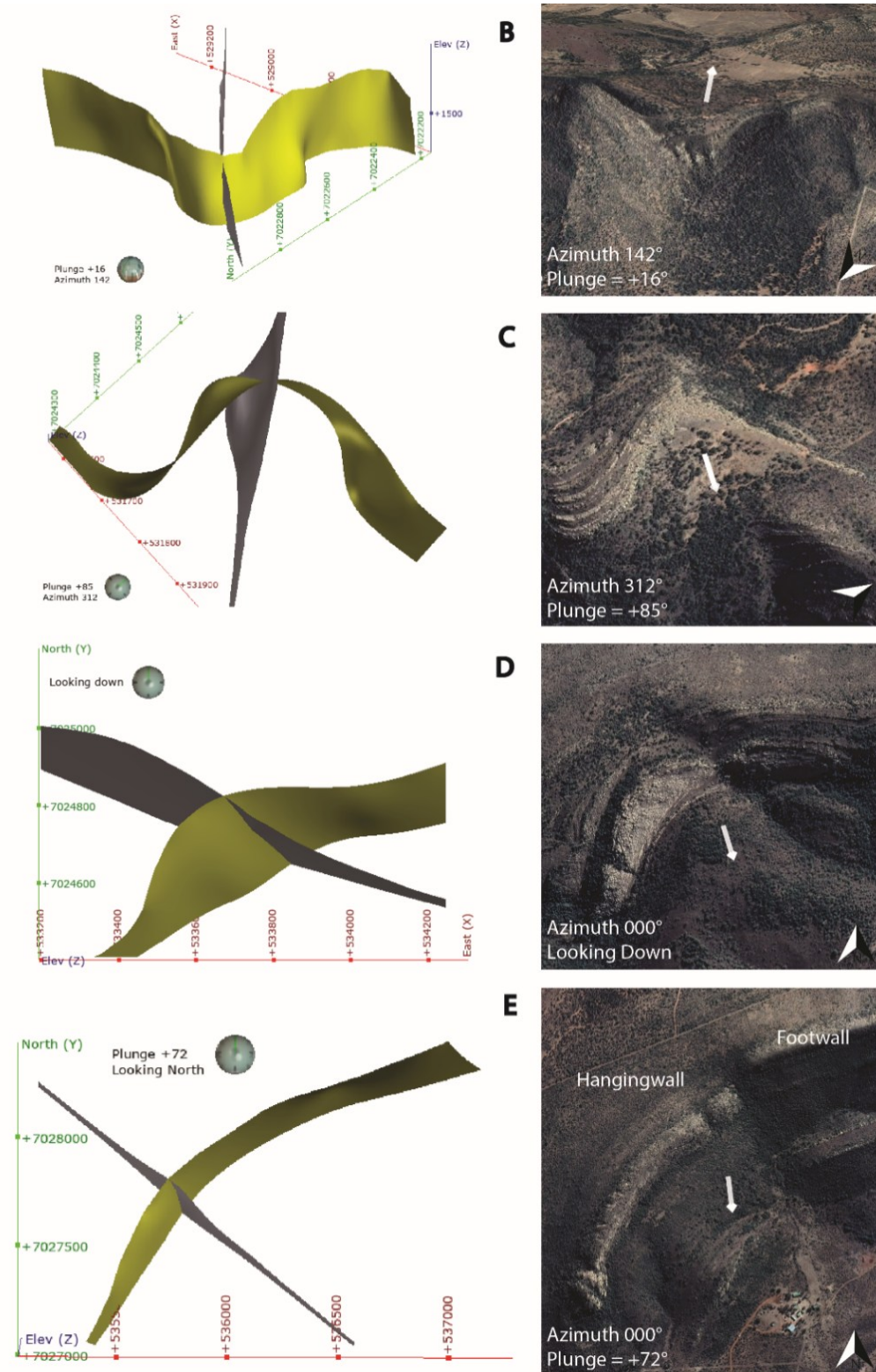


Figure 0.4: 3D Leapfrog Geo® representations of the folds B-E described by [Wieland et al. \(2005\)](#), based on their data. Symmetric synformal anticlines B and C display centripetal, steeply-plunging, approximately radial hinges and vertical, radially-oriented axial planes. Homoclinal folds D and E were interpreted as sinistral-verging asymmetric folds with steeply-plunging hinges and axial planes that lie oblique to the overall collar layer strike. Fold E is disrupted by several faults with left-lateral offset that [Wieland et al. \(2005\)](#) proposed were part of an oblique sinistral fault system. Google Earth imagery (Imagery date 07/10/2019).

Despite the metamorphism, the quartzites retain well-preserved sedimentary structures such as asymmetric ripples and planar and trough cross-bedding that indicate younging direction ([Bisschoff, 2000](#); [Gibson and Reimold, 2008](#)). The sedimentary units are intruded by Neoproterozoic (Venterdorp-age) mafic sills that are also metamorphosed. Quartzite makes up ~ 30 % of the WRG and is present as discrete units ranging in thickness from a few metres up to ~ 100 m ([Bisschoff, 2000](#)). The 50 - 100 m thick Rhq3 unit in the Bonanza Formation of the upper Hospital Hill Subgroup is the main unit of interest for structural measurements at Nauwpoort. At Geluksoord, facies variability and greater disruption of the stratigraphy creates some uncertainty as to whether the 50 m thick quartzite corresponds to the stratigraphically slightly lower Rhq1 or Rhq2 quartzites of the Bonanza Formation, although regional stratigraphic patterns suggest that it is the former ([Bisschoff, 2000](#)).

Bedding measurements using a Brunton Tru Arc compass-clinometer were collected as strike and dip, corrected for a magnetic declination of 20° W. Lithological contacts were mapped, although bedding measurements were particularly focused on the quartzite at both localities owing to the poor exposure and preservation of other units. Station points were UTM (WGS 84) georeferenced using a Garmin Etrex®20 Global Positioning System (GPS). The collected measurements and mapped contacts were used to create representative geological maps by overlaying UTM (WGS 84) GPS-referenced field station points onto georeferenced Google Earth Pro® images using ArcGIS 10.3.1®. The two study areas were then divided into domains based on the structural data. Thereafter, all the measurements from a single domain were plotted onto an equal-area lower-hemisphere stereonet (Schmidt net) using Stereonet 9.9.4® to determine structural trends and analyse fold geometries.

The Nauwpoort study area (Figure 0.3, Figure 0.5, Figure 0.6) was initially divided into 5 structural domains as identified by desktop studies. Bedding plane measurements (N = 685; Figure 0.6, Figure 0.7) were taken using a 10 m measurement spacing. The bedding plane measurements were used to further subdivide the field area into 15 discrete domains (A-O), each characterised by an approximately constant bedding orientation (Figure 0.6). Thereafter, each domain's bedding measurements were plotted onto a stereonet and filtered to remove outliers produced by, for instance, rotation adjacent to faults. In most cases bedding orientation was unchanged across faults, removing the need for further subdivision (Figure 0.6). The second iteration of filtration was performed visually on the stereonet to provide the smallest standard deviation and tightest cluster of data points from which to obtain a robust average value for each domain. Reasons for the spread in the data may include but are not limited to,

rotated blocks caused by small-scale faults or mass wasting, small-scale (< 20 m wavelength), second-order gentle folds, or original bedding surface variability caused by, for instance, large-scale trough cross-bedding. An error of $\pm 5^\circ$ strike and $\pm 2^\circ$ dip for bedding measurements was accepted. This resulted in a desired standard deviation in the bedding average of 10° for strike and 4° for dip. The Geluksoord study area (Figure 0.2, Figure 0.8) was approached in the same way, except for a coarser measurement density owing to the poorer outcrop. The Geluksoord structure is constrained by 45 bedding measurements, and 3 domains were identified, corresponding to the respective fold limbs (Figure 0.8).

In order to assist in 3D visualisation of the fold geometries and to allow comparison with the folds identified by [Wieland et al. \(2005\)](#), 3D structural models were constructed using Wieland et al.'s (2005) bedding data for their folds B-E (Figure 0.4) and bedding data for the Nauwpoort study area (Figure 0.10). The topography for the Nauwpoort study area was generated by digitising the 20 m contour interval from the 1:50 000 National Map Series in Leapfrog Geo® as polylines. A surface geological map was imported into Leapfrog Geo® from ArcGIS 10.3.1® and draped over the topography. The model required a 3D block boundary to confine the limits of the interpolations (mathematical links/extrapolations between data points that combine to create the 3D surfaces). The horizontal x- and y-axis limits correspond to the extent of each field mapping area. The z-axis limits were extended to 850 m above and below the mean topographic elevation to allow for extrapolation of the straight fold limbs, evident in the uniform bedding orientation observed for > 1 km along strike of the mapping localities (Figure 0.3), to better show the fold geometry. Lithological contacts were digitised from the geological maps as polylines and further used to constrain the geometry of the folds. Georeferenced field measurements of strike, dip and dip direction were imported as planar structural data and then plotted on the topography and allocated to the relevant lithological layer. The outcrop bedding data were further used to construct a 3D plane of the same orientation on which to place additional structural points to constrain bedding dip at depth. The mapped layers are conformable. Thus, where structural measurements were not available in the poorly-exposed non-quartzitic lithologies to construct surfaces in 3D, offset surfaces were made from the quartzite layer that had the best 3D surface exposure. After several iterations, faults were excluded from the final model, as net displacements were negligible (Figure 0.5, Figure 0.6) and no reliable dip measurements could be made in outcrop.

4.3 Results

4.3.1 3D Fold Analysis

Figure 0.4 shows 3D models for four of the five folds described by [Wieland et al. \(2005\)](#) in the northwestern inner collar (Figure 0.3), constructed using their data. [Wieland et al. \(2005\)](#) described the folds as centrifugally-closing km-scale folds separated by segments of the collar comprising relatively straight to slightly warped bedding. Although they identified all folds as occurring in the Rhq3 quartzite in the Hospital Hill Subgroup ([Bisschoff, 2000](#)), re-mapping of their Fold B suggests that it is located in the Rgq3 unit of the overlying Government Subgroup (Figure 0.3), which significantly reduces the estimated 1 - 2 km displacements along the flanking faults. [Wieland et al. \(2005\)](#) classified the folds as either symmetric or homoclinal and interpreted them as brittle-ductile features with faulted hinges and limbs. The impact timing and partially brittle character of the folds was supported by the reported presence of voluminous, 20 - 30 m wide, irregular pseudotachylytic network breccias in their hinge zones ([Wieland et al., 2005](#)).

The interlimb angles of these folds generally become more open towards the northeast (Figure 0.3, Figure 0.4). The absence of corresponding antiformal synclines supports a constrictional origin wherein the folds were able to amplify in a centrifugal/upward direction. The sinistral-verging homoclinal folds show steeply-plunging hinges and axial planes that lie oblique to the overall bedding strike in the collar, and are associated with pseudotachylyte-bearing faults that also appear to lie oblique to the strike of the collar strata (Figure 0.4E). [Wieland et al. \(2005\)](#) suggested that these faults also reflect constriction and, thus, that they formed penecontemporaneously with the symmetric folds.

Figure 0.4 shows that several of the fold axial planes and hinges depart quite significantly from radial trends. [Wieland et al. \(2005\)](#) suggested that these deviations might reflect subsequent fault block rotations.

Several additional, smaller, gentle, single synformal anticlines or anticline-syncline fold pairs with wavelengths up to several hundred metres and amplitudes up to 100 m were identified in less prominent quartzite units in the Hospital Hill Subgroup (e.g., [Bush, 2014](#); this study, see Figure 0.11). These also display broadly vertical radial to oblique axial planes and steeply centripetally-plunging fold hinges and are, thus, consistent with constriction.

4.3.2 Nauwpoort

The Nauwpoort locality lies in an ~ 6 km long segment of the inner collar, in which strata of the Bonanza Formation of the Hospital Hill Subgroup show an overall, slightly sinuous, 040° - 060°, strike and steep centripetal (60°) dip, with minor disruption by radial faults (Figure 0.3). To the south, bedding strikes 035°. [Wieland et al. \(2005\)](#) proposed that this locality shows a sinistral-verging box fold with a moderately (30°) SSW-plunging fold hinge and a shallowly (10° - 30°) SW-dipping axial surface, and that it formed in response to constriction. The Bonanza Formation comprises interbedded quartzite, metapelite, metapsammite and meta-ironstone (Figure 0.5; [Bisschoff, 2000](#); [McCarthy, 2006](#)). It is locally intruded by dolerite, which is also metamorphosed. The metapelites and meta-ironstones are variably magnetic and form poor, rubbly, outcrops. As a consequence, although the overall distribution of pelitic and ironstone units confirmed the fold axial trace, reliable structural data are mostly confined to the Rhq3 quartzite or the metapelites close to the quartzite contact (Figure 0.5). In addition to these limitations, the relative locations and sizes of the folds do not create a recognisable fold interference pattern or unequivocally establish their relative chronology; consequently, the proposed F1 - F2 chronology is based on consideration of the strain evolution during central uplift formation (see [Discussion](#)).

Folds

Three folds are interpreted as belonging to the F1 phase of folding (hinges labelled 1 - 3 in Figure 4.4). All are concentric folds that show similar sizes (wavelength of ≤ 400 m and amplitude of ~ 25 m) and tightness (gentle interlimb angles). Hinges plunge steeply (folds 1 and 2) to gently (fold 3) SSE and axial planes are upright to steeply inclined and strike SSE. Folds 1 and 2 form a fold pair on the overturned northeastern limb of the large F2 anticline, whereas fold 3 is found on the shallowly-dipping, right-way-up, intermediate limb of the F2 fold pair (Figure 0.5, Figure 0.6). Folds 2 and 3 are cut by broadly N-S trending, pseudotachylyte-bearing, faults (Figure 0.5, Figure 0.6). There is no evidence of an axial planar cleavage related to F1 and the absence of outcrop in the non-quartzitic lithologies restricts investigation of the extent of the folds.

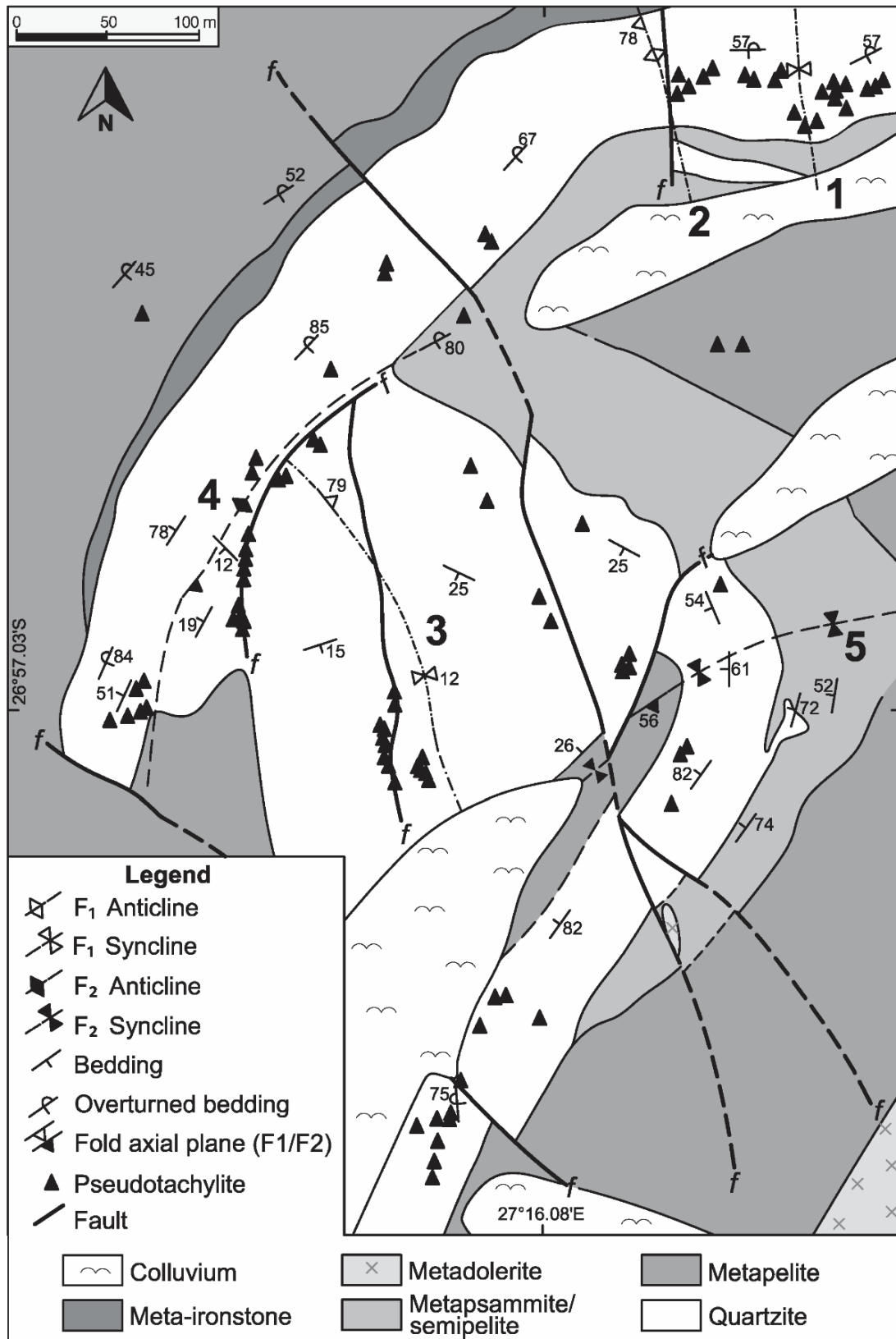


Figure 0.5: Geological map of the Nauwpoort fold locality showing fold axial traces, pseudotachylite occurrences and faults. The most voluminous pseudotachylite dykes occur along or adjacent to faults that, nonetheless, show minimal to small displacements.

Fold 1 is an antiformal syncline consisting of limbs A (mean bedding orientation $062^{\circ}/57^{\circ}/\text{SSE}$) and B ($091^{\circ}/57^{\circ}/\text{S}$), both of which are overturned (Figure 0.5, Figure 0.6). It is an asymmetric gentle fold with an interlimb angle of 156° and has a steeply plunging fold axis ($167^{\circ}/56^{\circ}$), and an upright axial plane ($167^{\circ}/90^{\circ}$) (Figure 0.7). The fold hinge is brecciated and contains significant amounts of pseudotachylyte, which extends parallel to bedding rather than along the fold axial trace. Semi-continuous, cm- to dm-wide dykes or dyke arrays, up to tens of metres long and with wedge-shaped offshoots, occur subparallel to bedding in the fold limbs (Figure 0.5).

Fold 2 is a synformal anticline consisting of limbs B (mean bedding orientation $091^{\circ}/57^{\circ}/\text{S}$) and C (mean bedding orientation $044^{\circ}/67^{\circ}/\text{ESE}$), both of which are overturned (Figure 0.5- Figure 0.7). It is an asymmetric gentle fold with an interlimb angle of 138° and has a steeply plunging fold axis ($183^{\circ}/57^{\circ}$), and a steeply inclined axial plane ($164^{\circ}/78^{\circ}/\text{WSW}$) (Figure 0.7). The hinge is cut obliquely by a N-S trending vertical fault that is marked by an ~ 40 m wide breccia zone where it cuts the quartzite. Despite the width of brecciation, the fault does not significantly offset the quartzite in a horizontal sense, although its proximity to the fold hinge, brecciation and abundant scree hamper precise location of the wall rock contacts.

Fold 3 is a synformal syncline consisting of limbs FG ($055^{\circ}/13^{\circ}/\text{SE}$) and DE ($118^{\circ}/26^{\circ}/\text{SSW}$), both of which display right-way-up bedding (Figure 0.5 - Figure 0.7). It is a symmetric gentle fold with an interlimb angle of 157° and has a gently plunging hinge ($144^{\circ}/12^{\circ}$), and a steeply inclined axial plane ($146^{\circ}/79^{\circ}/\text{NE}$) (Figure 0.7). A fault with minor offset of bedding but containing a vertical pseudotachylyte dyke up to 40 cm wide and > 100 m long cuts diagonally across the hinge.

The F2 phase of folding is represented by fold hinges 4 and 5 that form an anticline-syncline pair (Figure 0.5, Figure 0.6). These F2 folds are open and have an estimated wavelength of ~ 800 m and amplitude of ~ 100 m. They have gently plunging hinges and moderately inclined axial planes that strike broadly tangentially to the dome and dip centripetally. The folds are asymmetric and noncylindrical. Outcrop patterns and poorly preserved bedding in the metapelitic and meta-ironstone units lying stratigraphically below the Rhq3 quartzite confirm that they are also folded by F2 (Figure 0.5, Figure 0.6); however, folding is not evident in the stratigraphically overlying and underlying quartzites (Figure 0.3), indicating a disharmonic character. There is no evidence for an axial planar cleavage related to F2.

Fold 4 is a relatively sharp-hinged anticline consisting of a steeply dipping overturned limb N (mean bedding orientation $022^{\circ}/84^{\circ}/\text{ESE}$) and a shallowly-dipping right-way-up limb F (mean bedding orientation $074^{\circ}/13^{\circ}/\text{S}$, with allowance for bedding variation caused by fold 3) (Figure 0.6, Figure 0.7). It is an asymmetric open fold with an interlimb angle of 77° and has a gently plunging hinge ($200^{\circ}/10^{\circ}$), and an inclined axial plane ($031^{\circ}/47^{\circ}/\text{SE}$). Obtaining reliable readings of bedding in the hinge is difficult as the quartzite is extremely recrystallised, outcrop is poor, and the hinge is locally exploited by a fault (Figure 0.5).

Fold 5 is a concentric syncline consisting of limbs E (mean bedding orientation $118^{\circ}/26^{\circ}/\text{SSW}$) and H (mean bedding orientation $034^{\circ}/82^{\circ}/\text{WNW}$), which are both right-way-up (Figure 0.6, Figure 0.7), although bedding rotates to the more general steep overturned dips farther southwest (domain I; Figure 0.6). Fold 5 is an asymmetric open fold with an interlimb angle of 95° . It has a gently plunging hinge ($218^{\circ}/26^{\circ}$), and an inclined axial plane ($057^{\circ}/56^{\circ}/\text{SE}$) (Figure 0.7).

The geometry of, and the geometric relationships between, the F1 and F2 folds are shown in the 3D model in Figure 0.10. Contrary to [Wieland et al. \(2005\)](#), the geometry of the folds 4 and 5 pair is not consistent with tangential shortening and must, thus, be linked to some other phase of central uplift formation. Fold 3 was not described by [Wieland et al. \(2005\)](#) and alternative origins and timings were thus considered. The origin of Fold 3 after F2 via a third folding phase would require renewed oblique-tangential shortening, which appears inconsistent with modelled central uplift strain histories (e.g., [Rae et al., 2019](#)), and no corresponding synformal hinge was found in the steep limbs of either fold 4 or fold 5 to suggest that it postdates F2. Backrotation of the fold 3 hinge and axial plane by 103° around the fold 4 hinge and by 85° around the fold 5 hinge (Figure 0.7) results in variable, broadly radial vertical to reclined orientations that are not dissimilar to the orientations of some of the radial folds described by [Wieland et al. \(2005\)](#) (Figure 0.4). The backrotated orientation around the fold 5 hinge is in broad agreement with the orientation of folds 1 and 2. However, given the noncylindrical natures of both the F1 and F2 fold sets, the faulted F4 hinge, and the significant differences between the backrotated orientations for fold 3, these results are of limited value in trying to ascertain the relative timing of the F1 and F2 fold generations. Based on form, we propose that fold 3 is part of the same generation as folds 1 and 2 and, thus, is considered to have been rotated to its present orientation by F2.

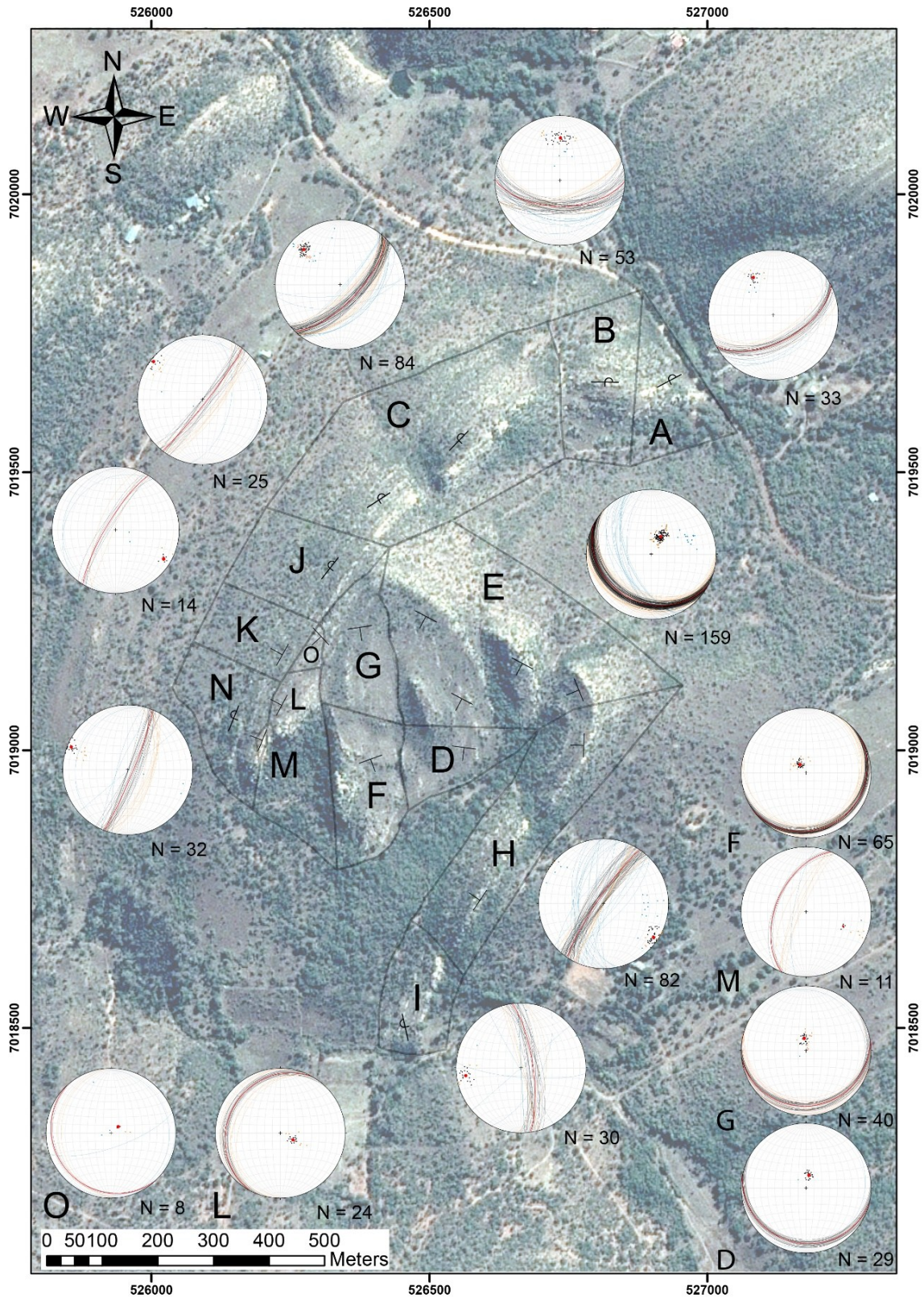


Figure 0.6: Google Earth Pro® image of the Nauwpoort locality showing the structural domains in the Rhq3 quartzite. Structural domains were identified on the basis of approximately constant bedding orientation. The stereonets show bedding measurements both as great circles and poles. (Imagery date 07/10/2019).

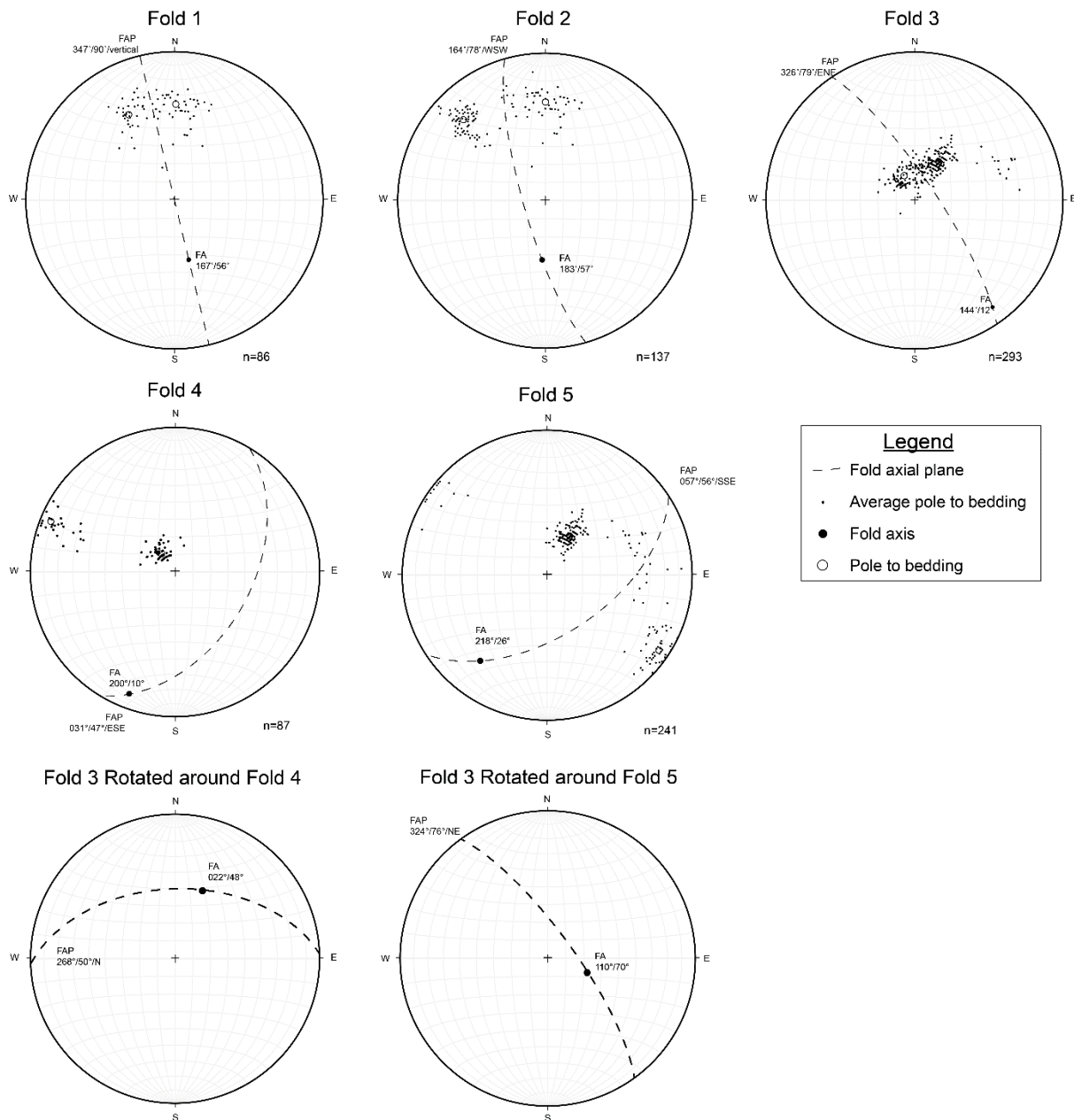


Figure 0.7: Equal-area lower hemisphere stereonet showing poles to bedding in folded Rhq3 quartzite at the Nauwpoort locality. Data are grouped geographically relative to folds 1 - 5 (see text and Figure 0.5 for details). The final two stereonet show the orientation of the fold 3 axial plane and axis after respectively rotating the shallow limb of fold 4 to the same orientation as the steep limb of fold 4 around the fold 4 axis and the steep limb of fold 5 around the fold 5 axis (see text for discussion).

Faults

Faults are primarily identified where they intersect the Rhq3 quartzite, although they are locally delineated by vegetation patterns, watercourses or, more rarely, clusters of pseudotachylyte boulders in areas underlain by metapelite and meta-ironstone (Figure 0.5, Figure 0.6). Most strike N-S to NNW-SSE, with a secondary NE-SW orientation (Figure 0.5). In the fault zone cutting fold 2 in the northern part of the area, angular, dm- to m-scale, quartzite clasts are embedded locally in a pseudotachylytic matrix. Despite the extensive brecciation in the quartzite, unambiguous slickenside surfaces with movement indicators could not be confirmed; however, rare slickenlines and grooves on joint surfaces in quartzite adjacent to the fault zones show general dip-slip to steep oblique-slip orientations. Given the visible offsets along faults cutting the shallow-dipping intermediate F2 limb and lack of similar offsets in the steep limbs (Figure 0.5, Figure 0.6), the faults are thus interpreted as dip-slip, with the outcrop pattern in the shallowly-dipping limb supporting east-side down individual movements of ≤ 10 m. The spatial coincidence of pseudotachylyte with the faults is striking (Figure 0.5), as are the exceptional volumes. Dykes up to 40 cm wide can be traced for tens to hundreds of metres in continuous to semi-continuous outcrop along faults that show minimal displacement, even where bedding lies at a high angle to the inferred slip direction. Planar dykes commonly show wedge-shaped offshoots ranging from decimetres up to 3 m in length.

4.3.3 Geluksoord

The Geluksoord locality in the NNE collar is located in a more complex structural environment than Nauwpoort (Figure 0.2), with the collar strata of the Geluksoord outcrops lying in a 2.5 - 4 km wide fault-bounded block that is displaced centripetally by 1 - 2 km relative to the main collar outcrops (Figure 0.11). Given this structural complexity and magnitude of the displacements, the Geluksoord fold orientation data may need to be regarded with more caution than those from Nauwpoort; nonetheless, these findings corroborate the F2 fold patterns seen at Nauwpoort.

The Geluksoord fold pair is inferred to be defined by the Rhq1 quartzite and adjacent metapelites and meta-ironstones ([Bisschoff, 2000](#)), with a thick dolerite sill intruding the lower metapelite (Figure 0.8; A in Figure 0.11). To the north, the quartzite displays an overturned, centripetal, $\sim 65^\circ$ dip and a sinuous N to NNW strike, whereas to the south it dips steeply ($\sim 80^\circ$) centrifugally and is right-way-up, but with more variable NW to NNW strike in broken outcrop. Similar sinuous strike on a scale of kilometres marks the WRG strata elsewhere in the area

(Figure 0.11). [Bush \(2014\)](#) noted an open, concentric, overturned anticline - syncline pair (wavelength ~ 500 m, amplitude ~ 100 m) with vertical, radial, axial planes and steeply radially plunging fold hinges in the Rhq1 quartzite ~3 km NW of the Geluksoord locality (B in Figure 0.11), indicating that F1-style folds are present in this part of the collar and have a distinctly different orientation and appearance to the Geluksoord fold pair. He also described subhorizontal to shallow N-dipping bedding in the Rhq3 quartzite in a small, strongly faulted area of exposure north of the present locality (Figure 0.11) but was not able to conclusively show that it was fold-related.

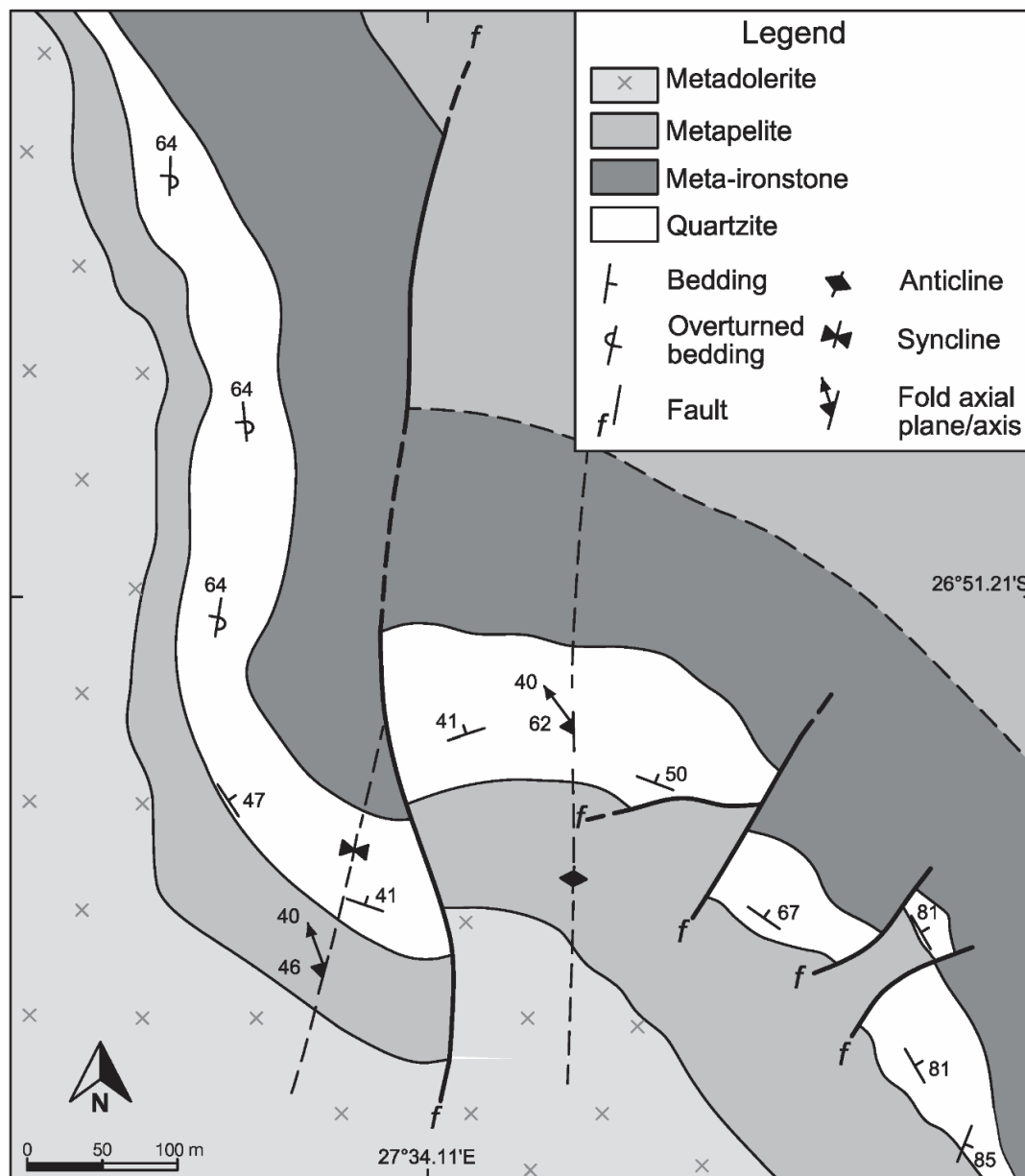


Figure 0.8: Geological map of the Geluksoord fold locality showing an inclined-plunging anticline-syncline pair with a faulted synclinal hinge.

Folds

At the Geluksoord locality, the quartzite defines an asymmetric anticline-syncline pair with a right-way-up intermediate limb (Figure 0.8). No axial planar cleavage was observed.

The anticline consists of a steeply-dipping to vertical right-way-up eastern limb (mean bedding orientation $140^{\circ}/74^{\circ}/\text{NE}$) and a more gently-dipping right-way-up western limb (mean bedding orientation $076^{\circ}/41^{\circ}/\text{NNW}$). It is an asymmetric open fold with an interlimb angle of 119° and has a moderately plunging fold axis ($334^{\circ}/41^{\circ}$) and an inclined axial plane ($001^{\circ}/62^{\circ}/\text{W}$) (Figure 0.9).

The syncline consists of a moderately-dipping right-way-up eastern limb (mean bedding orientation $076^{\circ}/41^{\circ}/\text{NW}$) and overturned western limb (mean bedding orientation $178^{\circ}/64^{\circ}/\text{W}$) (Figure 0.9). It is an asymmetric gentle fold with an interlimb angle of 147° and has a moderately plunging fold axis ($335^{\circ}/40^{\circ}$), and an inclined axial plane ($030^{\circ}/46^{\circ}/\text{NW}$) (Figure 0.9).

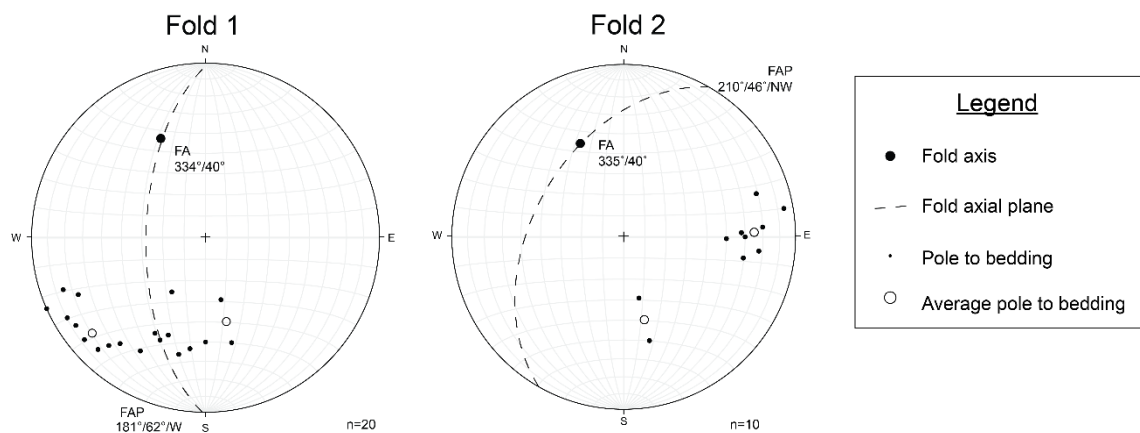


Figure 0.9: Equal-area lower-hemisphere stereonet projections showing poles to bedding in the folded quartzite at the Geluksoord locality. Fold 1 is the anticline and fold 2 is the syncline.

Faults

The syncline is cut by a slightly sinuous N-trending fault, which is not exposed in outcrop (Figure 0.8). When viewed in conjunction with the offsets of the meta-ironstone, the outcrop pattern of the quartzite is most compatible with dip-slip involving west-side-down displacement of $\sim 130 - 150$ m. This is considerably larger than the displacements along the faults at Nauwpoort. Although the dip of the fault could not be precisely determined, the centre-down sense of slip is consistent with that of the smaller faults at Nauwpoort. The SE limb of the anticline is disrupted by several NE- to ENE-trending faults that are also not exposed but

whose strike is well defined by the limits of the quartzite outcrops. The opposing senses of strike separation of subvertical bedding along faults with similar strike (Figure 0.8) suggests that these faults are also most likely dip-slip. In contrast to Nauwpoort, only a single cm-scale pseudotachylyte dyke was observed, but it shows no clear link to any of the mapped faults.

The broadly N and NE trends of the faults at the Geluksoord locality conform to the pattern of larger-scale faults bounding the centripetally-displaced Geluksoord block (Figure 0.11; [Bisschoff, 2000](#)). The NE-trending fault along the southeastern edge of the block shows a left-lateral separation of the core-collar contact of ~ 1 km whereas the N-trending fault along the western edge of the block shows a right-lateral separation of ~ 2 km. Bedding strike adjacent to the N-S fault shows a 45° clockwise rotation, from E-W in the western block to NW-SE in the Geluksoord block (Figure 0.2, Figure 0.11), making this a significant structural feature. [Bush \(2014\)](#) proposed that the scale and sense of slip along the N-trending fault would be consistent with an antithetic (E-side-down) fault to the Neoproterozoic rift fault system in the northern collar and that it was passively rotated during impact to its present orientation. However, an alternative explanation, given the post-fold timing of the main fault at Geluksoord (Figure 0.8), is that this km-scale faulting is a late impact feature. The implications of this are discussed in more detail below.

4.4 Discussion

4.4.1 Origin of Impact-related Folds in the Vredefort Dome

Compared with the central uplifts of several small- to mid-sized impact craters in layered targets that display a highly folded internal structure (e.g., Matt Wilson, [Kenkmann and Poelchau, 2009](#); Jebel Waqf as Suwwan, [Kenkmann et al., 2010, 2017](#); Serra de Cangalha, [Kenkmann et al., 2011](#)), the folds in the collar rocks of the Vredefort Dome (Figure 0.2, Figure 0.3) appear relatively minor in relation to the overall size of the central uplift. [Wieland et al. \(2005\)](#) attributed this to a combination of the rapid radial outward decline in measured constrictional strain, from 17 % to < 5 % in the inner collar, as well as the prevalence of relatively massive competent layering (particularly the CRG and Ventersdorp Supergroup) that occupy large parts of the collar. It is not uncommon for layering to show slight ($\leq 10^\circ$) sinuosity in strike on the 5-10 km scale (Figure 0.2, Figure 0.3) that would be consistent with very limited buckling. A further factor suppressing a fold response may be the 40 km wide crystalline basement core of the dome. [Jahn and Riller \(2009, 2015\)](#) suggested that radial transpressional

faults, rather than folds, were the principal mechanism by which constrictional strain was accommodated in the Vredefort Dome.

Our reassessment of fold structures in the innermost collar of the dome has revealed that the heterogeneously-layered WRG actually contains evidence of two fold generations, both of which show signs of a brittle-ductile origin and a close timing relationship with pseudotachylyte, both of which are interpreted to signify an impact origin. The vertical, radially-oriented F1 folds that affect the steeply dipping quartzite at Nauwpoort and steeply-dipping quartzites in close proximity to the Geluksoord locality ([Bush, 2014](#)) have similar orientation and form to the larger symmetric and asymmetric synformal anticlines described by [Wieland et al. \(2005\)](#). Their orientation supports them being the product of a constrictional strain pattern, which would indicate formation during the initial convergent flow related to rise of the central uplift. Their presence near the asymmetric F2 folds, which display distinctly different size and form to the F1 folds, rules out the latter being F1 folds reoriented by unusual fault block rotations. The early timing of F1 constrains F2 to a later stage within the evolution of the central uplift. While simple backrotations of the intermediate limb of the F2 fold pair do not restore the F1 fold 3 to exactly the same orientation as folds 1 and 2, the results (Figure 0.7) are consistent with the constrictional folds described by [Wieland et al. \(2005\)](#).

The local rotation of bedding in the Hospital Hill Subgroup quartzites and adjacent lithologies to shallow right-way-up at the Nauwpoort and Geluksoord localities from the almost ubiquitous overturned dips that dominate the collar of the Vredefort Dome ([Wieland et al., 2005](#)) requires that F2 folding occurred around subhorizontal to shallowly plunging, tangentially-aligned, hinges. Figure 0.12D and E show that, notwithstanding the steeper plunge of the Geluksoord fold pair, fold asymmetry at both localities is consistent with a centre-down shear sense (anticline located radially outward of syncline). As discussed further below, we argue that the kinematics responsible for this F2 fold pattern relate to collapse of the central uplift.

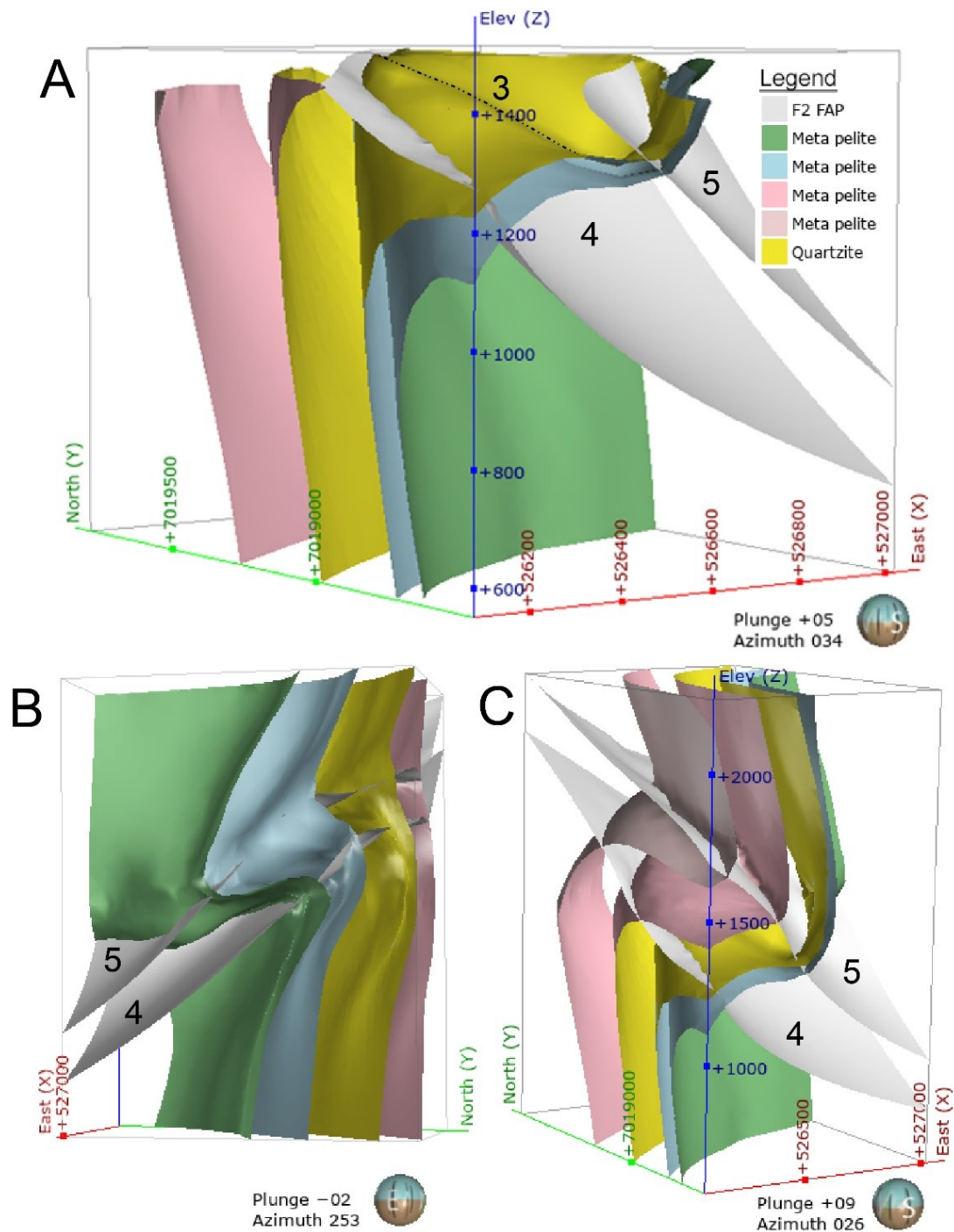


Figure 0.10: 3D Leapfrog Geo® model showing different perspectives of the fold geometries at the Nauwpoort locality. For simplification, faults have been excluded as displacements are minor at the scale of interest. F2 fold axial planes are indicated by grey/white surfaces and F1 fold hinges by stippled lines. Lithological contacts have been extended by ~ 850 m above and below the current topography to allow for extrapolation of the straight fold limbs to better show the F2 fold geometry. (A) View looking NE with a plunge of 05° cropped at the current topographic level. (B) View looking WSW depicting the overall geometry of the F2 folds. (C) View looking NNE depicting the overall geometry of the F2 folds.

Folding mechanisms

The F1 syncline (fold 1) at Nauwpoort shows the same relationships with respect to pseudotachylyte as those described by [Wieland et al. \(2005\)](#) for the radial km-scale folds in the northwestern sector of the dome, namely, a brecciated hinge occupied by an irregular pseudotachylyte network, with planar pseudotachylyte dykes with high length:width ratios subparallel to bedding in the limbs. This supports a brittle-ductile buckling mechanism and possibly an element of bedding-parallel flexural slip in generating the F1 folds, although the limited slip expected in the limbs of such gentle folds appears incompatible with the significant melt volumes observed in individual dykes (see also [Wieland et al., 2005](#)).

In contrast to F1, F2 involved tighter (open) folding, but brecciation or preferential localisation of pseudotachylyte in the F2 fold hinges was not observed. Whether this relates to slightly slower strain rates during F2, leading to a more ductile buckling response, or to different geometries of the F1 and F2 fold hinges relative to dilational sites at the time of pseudotachylytic melt ponding, or the generally poor exposure of faults, could not be determined. No conclusive evidence was found to indicate that the faults affecting the F2 anticline hinge at Nauwpoort and the synclinal hinge at Geluksoord developed simultaneously with the folds. However, even if this was the case, at both localities bedding shows smooth rotations in orientation across at least one hinge, consistent with a buckling mechanism.

Given the exceptional strain rates accompanying crater formation ([Ivanov, 2005](#); [Rae et al., 2019](#)), plastic strain *sensu stricto* is not a viable option to accommodate the relative movements and rotations accompanying buckling. [Kenkmann \(2002\)](#) proposed that tight, dm-scale folding in a thinly-bedded dolomite-siltstone sequence in the Crooked Creek impact structure involved primarily shear and microbrecciation along a mm- to cm-spaced shear fault network, accompanied by secondary folding. Analysis of outcrops and thin sections of the Rhq3 quartzite from around the Nauwpoort syncline hinge revealed no similar axial planar microbreccias displacing bedding; however, post-shock heating to $T > 400\text{ }^{\circ}\text{C}$ in these rocks (K. Beaton, pers. comm., 2020) facilitated extensive recrystallisation in the quartzite. This might obscure any evidence of grain boundary slip or cataclasis in the quartzite matrix. A distinctive feature of the recrystallisation is that, whereas planar fluid inclusion trails typical of decorated PDF are found in some sedimentary quartz grains, the grains more generally show distinctive planar zones up to 100 μm wide that are locally continuous with PDF arrays, but which comprise a relatively coarse (up to 50 - 100 μm) granoblastic polygonal quartz mosaic. [Huffman and Reimold \(1996\)](#) demonstrated that wider PDF lamellae might develop in quartz

shocked at elevated temperatures; however, the WRG rocks would not have been at temperatures above mid-greenschist facies immediately before the impact ([Gibson and Jones, 2002](#)). Any evidence of the reason why recrystallisation was enhanced in planes parallel to PDF has been destroyed by the new grain growth; however, one possibility is that buckling strains were at least partly accommodated by microshear along PDF glass lamellae and in the shock-damaged inter-PDF regions of the grains, leading to wide intragranular process zones that were subsequently more susceptible to annealing. Such intragranular deformation working together with grain boundary slip, and possibly facilitated by the elevated post-impact temperatures, could have enabled an apparent plastic strain response to F2 buckling.

It is possible that at least some of the faulting affecting the Nauwpoort F2 anticline could have been related to overtightening of the hinge; however, the high degree of recrystallisation in the hinge has obscured any potential evidence of microbrecciation or cataclasis.

F2 folds related to collapse of the Vredefort central uplift

Figure 0.12D and E demonstrate identical basic geometries of the F2 folds at Nauwpoort and Geluksoord relative to the centre of the dome. Both are asymmetric, disharmonic folds with the anticline positioned radially outward of the syncline, inward-dipping axial planes and subhorizontal to moderately-plunging hinges. The Geluksoord fold pair shows similar inclined axial plane dips, but steeper fold hinge line plunges compared to the Nauwpoort pair (Figure 0.7, Figure 0.9). Reasons for this might include the non-cylindrical and disharmonic nature of the folds or subsequent moderate fault block rotation, particularly as the Geluksoord sector is intensely faulted (see below).

Tangential folds attributed to collapse of central uplifts have been described previously from the Serra de Cangalha central uplift ([Kenkmann et al., 2011](#); Figure 4E) and from the Araguainha impact structure, where [Lana et al. \(2008\)](#), Figure 8e-f) located them within the annular basin but described bedding as generally outward dipping, which would be consistent with the outer edge of the central uplift. At Araguainha, dm-scale open to tight folds verge radially outwards (syncline radially outward of anticline) and are attributed by [Lana et al. \(2008\)](#) to predominantly horizontal centrifugal collapse of the central uplift. The closest equivalent to these folds in the Vredefort impact structure would be the slightly asymmetric km-scale folds in the rim syncline hinge described by [Simpson \(1978\)](#), in which subhorizontal bedding underwent brittle-ductile upright folding in response to lateral compressional stresses, although the Araguainha folds are far more asymmetric and show more shallowly-dipping axial

planes. At Serra de Cangalha, on the other hand, meter to decametre-sized gentle folds with recumbent to shallowly-inclined axial planes are common in the overturned NW sector of the central uplift ([Kenkmann et al., 2011](#)). The folds are gentle and display a slight asymmetry consistent with centre-down vergence. [Kenkmann et al. \(2011\)](#) suggested that they formed in response to load stresses generated by the overlying central uplift mass as the strata were tilted through vertical, before coming to rest in their present overturned orientation. The F2 folds at both Nauwpoort and Geluksoord are considerably larger, tighter and more asymmetric than those described at Serra de Cangalha, but they show the same overall steep centre-side-down vergence (Figure 0.12C).

In both the Serra de Cangalha and Araguinha cases, folds are located in sedimentary units with cm- to dm-scale layering. The Hospital Hill Subgroup is a metasedimentary sequence that comprises layering on a scale of tens to hundreds of metres with both a high competence contrast and highly competent layers (quartzites) that are both comparatively thin and make up a minor component of the packages. These characteristics favour a buckling response if layering is favourably oriented with respect to the principal compressive stress axis. In addition to being related to the amount of strain, fold wavelength is a function of competence contrast, the thickness of the competent unit, and the spacing between the competent units. All of these factors may have played a role in the localisation of both the F1 and F2 folds in the Hospital Hill Subgroup quartzites, which are not only comparatively thick (~ 50 m) but are also flanked by the thickest pelitic and ironstone units in the WRG ([Bisschoff, 2000](#); [McCarthy, 2006](#)). On a wider scale, the WRG is flanked by massive crystalline granitoid gneisses constituting the core of the dome and the mechanically homogeneous, quartzite-dominated CRG and metabasalt-dominated Ventersdorp Supergroup. The differential collapse of these more rigid flanking formations could have preferentially channelled strain into the WRG lithologies in the inner collar. The centre-down movement sense deduced from the F2 fold asymmetry is kinematically incompatible with rise of the central uplift, but fits vertical collapse in a mechanically heterogeneous, anisotropic, target where shear is preferentially channelled along steeply inclined layering (Figure 0.12C).

4.4.2 Fault Kinematics

In contrast to the shallowly-dipping concentrically-striking faults seen in the rim syncline ([Brink et al. 1997, 2000](#); Figure 0.12A, B), [Wieland et al. \(2005\)](#) described most of the faults in the inner collar of the dome as subvertical and radial to highly oblique to the collar strata,

with some sector-based variation (Figure 0.12B). Observed strike separations are up to hundreds of metres. [Wieland et al. \(2005\)](#) noted rare subhorizontal slickenlines associated with faults and concluded that the faults display essentially strike-slip movement in their present orientations. They linked the faults to the radial and oblique large-scale folds in the inner collar and, thus, to constrictional strain during the early stages of central uplift formation. In contrast, while also suggesting that radial faults were related to the constriction stage, [Jahn and Riller \(2015\)](#) proposed that radial fault orientations are largely absent from the lower WRG.

[Wieland et al. \(2005\)](#) excluded from their analysis a conjugate fault system with km-scale strike separations in the Witwatersrand and Ventersdorp supergroups, which they interpreted as a rotated pre-impact, Ventersdorp-age, half-graben system (see also [Brink et al., 1997](#)) (Figure 0.2). However, [Jahn and Riller \(2009\)](#) interpreted these faults as a transpressional impact-related fault system. [Jahn and Riller \(2009, 2015\)](#) proposed that the radial faults displace older impact-related tangential faults that dip radially outwards and that show radial inward thrust displacements (Figure 0.12B). They interpreted the tangential faults as originally inward-dipping normal faults in the TC wall that were rotated to their present orientations by the rise and ensuing outward collapse of the central uplift.

Almost all faults in the collar contain traces of voluminous pseudotachylyte. [Gibson et al. \(1997\)](#) showed that small pseudotachylyte-bearing faults with mm to cm displacements display geometries consistent with both constriction and tangential extension in metapelitic lithologies in the inner collar. However, no larger-scale impact-related radial or tangential faults of extensional origin have previously been identified in the inner collar of the dome. [Wieland et al. \(2005\)](#) did suggest that transpressional radial faults could have been reactivated during extensional collapse but primarily associated joints and shear fractures in the quartzites with tangential and radial collapse. The only significant impact-related extensional features previously noted in the dome are the vertical, broadly tangential and radial, impact melt dykes ([Therriault et al., 1995](#); [Reimold et al.](#), this volume and references therein) and large volume pseudotachylytic breccia dykes ([Lieber et al., 2009](#); [Reimold and Gibson, 2006](#) and references therein) in the core of the dome (Figure 0.12C), but neither show compelling evidence of slip displacements along their margins. The exact timing of intrusion of the impact melt dykes (and, thus, of their hosting fracture systems) is unknown, as the overlying impact melt sheet is likely to have remained partially molten for at least tens of thousands of years ([Ivanov, 2005](#)). However, the pseudotachylytic melts are unlikely to have survived more than a few seconds to hours, depending on volume and the degree of superheating ([Wieland, 2006](#)), and so their

geometries must be linked to strains during central uplift collapse ([Lieger et al., 2009](#)). A similar argument can, thus, be made for the pseudotachylyte-hosting tangential faults at Nauwpoort. Unlike the core dykes, however, the Nauwpoort faults show displacements of marker beds that provide clues to the nature of the late-stage collapse kinematics.

The folded units at Nauwpoort are displaced by a broadly tangentially-striking fault network that contains voluminous pseudotachylyte (Figure 0.5, Figure 0.12C, D). The wide brecciation zones along the faults relative to their slip magnitudes resemble the extensional faults described by [Osinski and Spray \(2005\)](#) in the Houghton structure. Field relations (Figure 0.5) suggest that the faults are subvertical, accommodating radial extension via centre-side-down dip slip. No pseudotachylyte was observed in the vicinity of the main fault cutting the Geluksoord syncline, which has an oblique strike relative to the centre of the dome; however, bedding offset across it is also consistent with centre-side-down slip (Figure 0.8, Figure 0.12E).

[Wieland et al. \(2005\)](#) speculated that semi-continuous pseudotachylyte dykes extending for kilometres parallel to quartzite bedding with relatively uniform tangential strike (Figure 0.12C) were generated by bedding-parallel dip-slip movements, but they were unable to confirm displacement owing to the absence of cross-cutting relationships and the generally poor exposure of faults in the collar. In this context, the deviation in bedding orientation in the intermediate limbs of the Nauwpoort and Geluksoord folds provides a rare opportunity to observe such fault displacements (Figure 0.12D, E). Their identification extends the outcrop-scale structural inventory of central uplift collapse in the inner collar beyond the joints and shear fractures identified by [Wieland et al. \(2005\)](#).

In the NNE collar, the radial to oblique-radial orientations and opposing senses of displacement of the faults bounding the Geluksoord block (Figure 0.11, Figure 0.12F) suggest a conjugate relationship that is consistent with tangential extension. Smaller-scale fault separations within the lower Hospital Hill Subgroup in the block ([Bisschoff, 2000](#); Figure 0.11) are also consistent with tangential extension, although slip vectors are not visible. [Kenkmann and Von Dalwigk \(2000\)](#) proposed that outward collapse of a central uplift could generate outward-widening radial transtension troughs. In the case of the Geluksoord block, the relative fault displacements suggest that the flanking collar segments experienced greater outward collapse than the Geluksoord block (Figure 0.12F). If extrapolated to higher structural levels, such displacements would suggest that the Geluksoord block would manifest as a radially-oriented horst, rather than a trough (Figure 0.12F). At the present erosion level, the Geluksoord faults suggest that

inferring a general transpressional origin for all faults cutting the collar of the Vredefort Dome at moderate to high angles should be reassessed.

We have argued that the fortuitous subhorizontal orientation of bedding at the Nauwpoort and Geluksoord sites is instrumental in revealing vertical tangential faults and their centre-down kinematics in the Vredefort central uplift. However, slip magnitudes are small and, although slip displacements at Geluksoord appear larger, these offsets are dwarfed by the major, shallowly inward-dipping, normal fault system identified by [Simpson \(1978\)](#) in the outer collar of the dome that has a cumulative 2.3 km of centre-down slip (Figure 0.12C). A single fault in this system displays 1.5 km of slip ([Simpson, 1978](#)). Subhorizontal to inward-dipping seismic reflectors in the collar of the dome were previously exclusively interpreted as outward-directed thrusts by [Brink et al. \(1997\)](#) and [Grieve et al. \(2008\)](#), although the latter proposed that these thrusts may rotate to steeply outward-dipping normal fault geometries towards the centre of the dome (Figure 0.12A). Together with the Nauwpoort and Geluksoord observations, [Simpson's \(1978\)](#) data suggest that both steep and shallow-dipping collapse-related extensional faults are more common in the Vredefort Dome than have been previously recognised. Both vertical and shallowly dipping faults are consistent with vertical shortening of the Vredefort central uplift and support a component of inward collapse, rather than exclusively outward late-stage fault displacements within the central uplift, such as proposed by [Grieve et al. \(2008; Figure 0.12A\)](#).

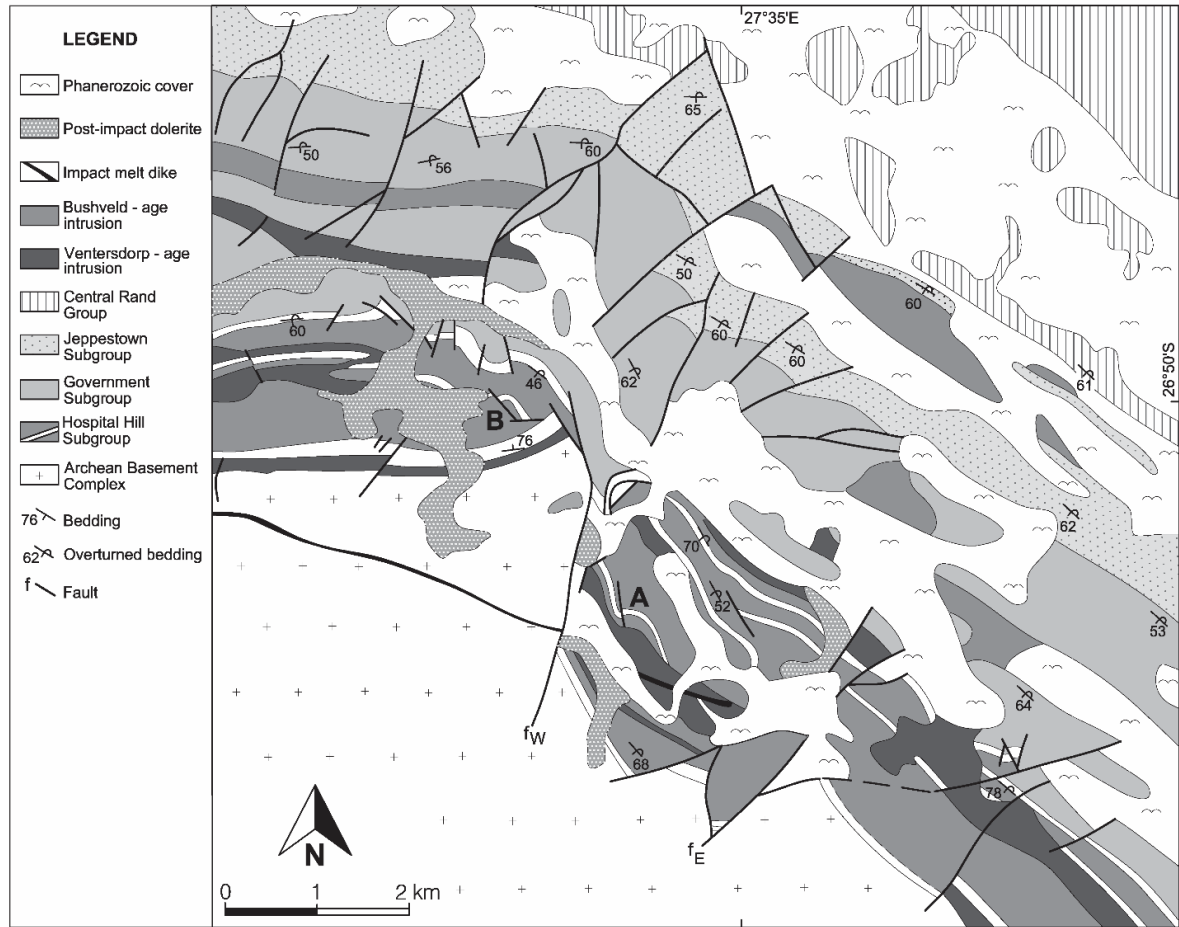


Figure 0.11: Simplified geological map of the inner collar in the NNE sector of the Vredefort Dome, modified from [Bisschoff \(2000\)](#). The proposed bounding faults of the Geluksoord block are labelled f_W and f_E . These splay into smaller faults away from the centre of the dome. (A) Geluksoord fold locality (this study): (B) F1 fold identified by [Bush \(2014\)](#). Although the relationship of the tangential impact melt dyke to f_W is obscured by alluvium, the dyke continues east of the fault with no obvious offset, constraining the timing of the faulting.

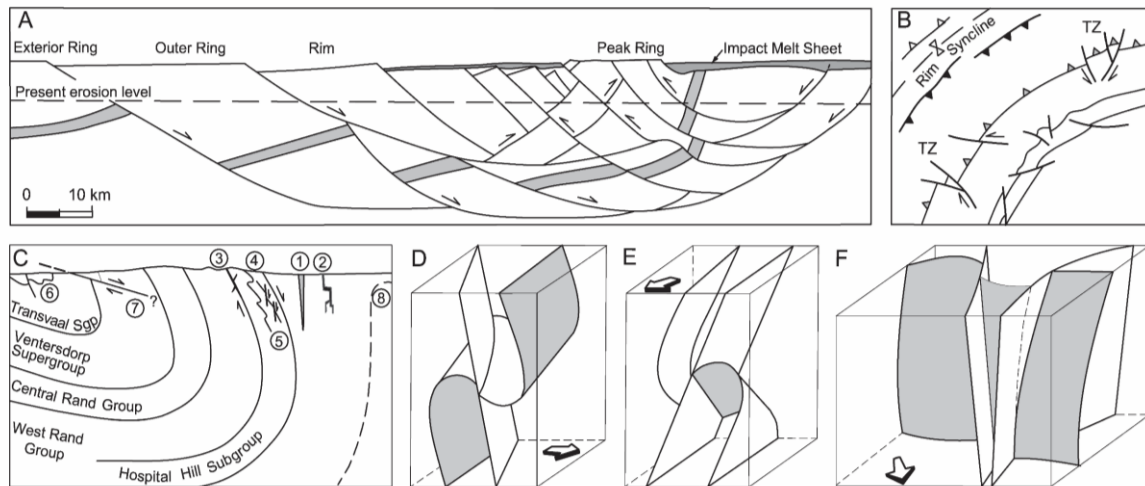


Figure 0.12: Composite schematic diagram showing the main structural elements attributed to the Vredefort impact. (A) Schematic cross-section of [Grieve et al. \(2008\)](#); modified herein) from peak-ring basin model showing the approximate erosion level for the Vredefort structure. Outward-dipping faults in the central uplift with centrifugal slip interfere with inward-dipping faults in the outer parts of the structure with centripetal slip to produce the structurally complex peak ring. (B) Schematic structural map of the northwestern sector of the Vredefort Dome showing the principal impact-related structural elements identified by previous authors in the collar. Open triangles—inward-verging thrust fault splays off outer rim normal faults ([Grieve et al., 2008](#)). Black triangles—outward-verging thrust faults ([Brink et al., 1997](#); [Grieve et al., 2008](#)). Grey triangles—rim-collapse normal faults rotated to thrust geometries in central uplift ([Jahn and Riller, 2009](#)). TZ—transpressional fault zones ([Jahn and Riller, 2009](#)). Innermost collar—constrictional folds and radial to oblique faults in Hospital Hill Subgroup ([Wieland et al., 2005](#)). (C) Schematic radial cross-section of the Vredefort Dome showing structural elements related to central uplift collapse (in addition to overturning; see text for details): 1—tangential impact melt dyke; 2—tangential pseudotachylytic breccia dyke network; 3—bedding-parallel shear surfaces and pseudotachylyte dykes; 4—folds; 5—normal faults with centre-down displacement; 6—upright to centrifugally verging folds and reverse-slip faults; 7—normal fault system with centre-down displacement; 8—gneissosity in the Archean basement complex showing inner limit of rotation of pre-impact features at radius (R) ~12 km. Sgp—Supergroup. (D) Block diagram showing geometry of collapse-related F2 folds and faults at Nauwpoort locality; arrow indicates dome centre. (E) Block diagram showing the geometry of collapse-related F2 folds and faults at Geluksoord locality; arrow indicates dome centre. (F) Block diagram showing relative displacement of the core-collar contact in the Geluksoord fault block; arrow indicates dome centre.

4.4.3 General Implications

Considerable advances have been made over the past two decades in systematically documenting the internal structural inventories of, particularly, the central uplifts of partially eroded impact structures. These inventories have proven invaluable in placing constraints on the kinematics and dynamics of complex crater formation and allow testing of numerical simulations of the response of target crust to large impacts (e.g., [Kenkmann et al., 2014](#); [Rae](#)

[et al., 2019](#)). While the precise mechanisms by which target weakening occurs remain debated, a key revelation has been the degree to which structural coherence in large parts of the central uplift volume appears to be retained despite the substantial displacements and prevailing high strain rates.

Deciphering cratering kinematics and dynamics is best achieved where a target contains well-defined layering that retains coherence on significant length-scales during the cratering process; consequently, the most informative structural studies of complex impact structures have come from small- to mid-sized craters in sedimentary basins (e.g., [Kenkmann et al. 2010, 2011](#)). In such crater structures, the total vertical displacement of the central uplift volume - and thus its potential to overheighten and collapse - is limited. In larger structures there is a greater likelihood that the central uplift will be cored by crystalline, massive, lithologies that lack useful markers for tracking strain evolution, yet it is the central uplifts of these craters that experience the greatest vertical displacements and that should, thus, show the greatest internal evidence of collapse.

In central uplifts formed in response to moderate to high-angle impacts, symmetric overturning of the flanks of the central uplift is identified as the primary evidence of dynamic outward collapse of the central uplift as a result of overheightening. Such collapse produces a peak-ring crater and has received considerable recent attention in relation to the Chicxulub structure (e.g., [Morgan et al., 2016](#); [Riller et al., 2018](#); [Rae et al., 2019](#)). [Kenkmann et al. \(2014\)](#) list three additional smaller-scale features related to central uplift collapse - outward-directed buckle folding and kinking; tangential, outward-dipping, normal dip-slip faults, and radial transtensional troughs.

The Vredefort Dome is the world's largest identified central uplift, with maximum structural uplift (after collapse) estimated at 25 - 27 km ([Ivanov, 2005](#); [Gibson and Reimold, 2008](#)). The entire collar, between 20 and 45 km from the centre of the dome, is overturned and, although it is more difficult to confirm overturning in the core of the dome, [Lana et al. \(2003\)](#) proposed that the gneissic foliation beyond $R = 12$ km is sharply rotated by at least 90° (Figure 0.12C). Despite erosion levels of between 8 and 10 km, the narrowest part of the core-collar contact in the dome is estimated to lie a further 5 km below the surface ([Molezzi et al., 2019](#)). From this large-scale geometry it is obvious that the presently exposed levels of the dome would be expected to record additional features related to radial and tangential extensional collapse (flattening strain field). As a first-order approximation, projecting an average 60° inward dip of the core-collar contact to 5 km depth suggests that the exposed inner collar experienced ~

10 % radial and tangential extension during central uplift collapse. These strains would have varied vertically, diminishing downwards (Figure 0.12C).

Given the almost universally poor exposure of faults in the Vredefort Dome, it is beyond the scope of the present study to re-evaluate the entire collar to ascertain how many faults might be extensional/transensional, rather than compressional/transpressional, in origin, or whether early-formed impact faults show signs of extensional reactivation. However, the magnitude of the final collapse indicated by the extent of the overturning and the fault evidence related to the Geluksoord block suggest that caution should be exercised in claiming collar faults with radial to subradial orientations at the present level of erosion as exclusively the product of constriction. Moreover, the interpretation of shallowly inward-dipping subsurface reflectors in the outer collar as outward-directed thrusts (e.g., [Brink et al., 1997](#); [Grieve et al., 2008](#)) is contradicted by [Simpson's \(1978\)](#) field evidence of a low-angle normal fault system. The kinematics implied by this geometry seem counterintuitive, given the dynamic collapse model, numerical models and interpretations of the subsurface structure of the Chicxulub peak ring (e.g., [Riller et al., 2018](#)), but are supported by both the F2 folds and tangential faults documented at Nauwpoort and Geluksoord, which are both consistent with centre-down displacements.

These fault displacements differ from the only other well-described fault system linked to central uplift collapse, in the Haughton crater, where the faults dip outwards and displacement is thus outer-side-down ([Osinski and Spray, 2005](#)). However, it receives support from several other large crater structures. Drilling in the Manicouagan structure has established large variations in impact melt sheet thickness that [Spray and Thompson \(2008\)](#) attributed to normal block faulting within the crystalline crater basement radially inward of the proposed peak ring. Although, like the kilometres-long fractures hosting the impact melt dykes in the Vredefort Dome, these faults could also potentially have formed during post-impact crustal adjustments within the crater, the preservation of a thick melt sheet over the centre of the central uplift at Manicouagan implies an element of centre-down collapse flow. Similar footwall relief beneath the Sudbury impact melt sheet of up to 1500 m (varying by up to 400 m locally; [Dreuse et al., 2010](#)) would also be consistent with a block-faulted central uplift core. Central topographic depressions preserving thick melt and/or impact breccia sequences have also been described from Popigai ([Masaitis et al., 1999](#)). While the continuation of such radial extension at depth has not been established in any of these structures, the evidence from Vredefort suggests that collapse is not restricted to shallow levels of the central uplift.

[Simpson's \(1978\)](#) observations extend this inward-collapse kinematics across much of the Vredefort central uplift. They also suggest that, at least in the outer parts of the central uplift, inward collapse postdates the outward-directed overturning-collapse. The asymmetry of the F2 folds in the inner collar suggests at least the latter parts of their formation occurred after rotation of bedding through the vertical. Steep bedding surfaces could have preferentially channelled steep radially inward-directed slip owing to the lower critical resolved shear stress necessary to activate such favourably-oriented bedding surfaces during vertical collapse (Figure 0.12C). One manifestation of large bedding-parallel slips could be the remarkably persistent bedding-parallel pseudotachylite dykes observed in the inner collar ([Wieland et al., 2005](#)). However, such a mechanism would mean that much of the evidence of this collapse could be hidden in the collar, except where unusual bedding orientations are found. With collapse-driven displacements also occurring along radial to oblique faults to form structures such as the Geluksoord fault block, a more complex, multi-stage, history of second-order structure development related to central uplift kinematics emerges in the Vredefort Dome.

The combination of its large size and the deep levels of erosion make the Vredefort Dome an ideal site in which to investigate the kinematics and dynamics of the interior of a central uplift in which the contribution of vertical impact-related displacements and forces is enhanced relative to convergent radial flow and forces (e.g., [Collins et al., 2004](#)). The observations presented here of vertically-driven, inward, collapse flow challenge models that invoke primarily horizontal radial compressional stresses extending deep within the interior of large central uplifts ([Kenkmann et al., 2000](#); [Grieve et al., 2008](#)). At the very least they limit the depths and radial distances to which such subhorizontal, outward-verging structures (and, thus, kinematics) may be dominant, or they reflect a late-stage phenomenon not yet recognised in the formation of dynamically collapsing central uplifts. Although the vertical displacements represented by the F2 folds and tangential faults in the two study areas are relatively small compared to the outward translation represented by the overturning in the 20 - 25 km wide collar of the Vredefort Dome, large amounts of vertical collapse-related displacement could be hidden by activation of the predominantly tangentially-oriented, steep, bedding planes. However, the 2.3 km of inward slip in the outer collar fault zone noted by [Simpson \(1978\)](#) suggests that rotational outward flow on the flanks of the central uplift was closely followed by inward-and-downward return flow in the interior of the central uplift, even at a depth of 8-10 km below the crater surface.

In their discussion of the limitations of their analog model of central uplift collapse, [Dörfler and Kenkmann \(2020\)](#) point out that inertia likely plays a key role in the kinematics of central uplift collapse and the ultimate collapse morphology. With the outer parts of the central uplift collapsing outwards before the upper part of the central uplift reverses its rise and begins to collapse vertically, and with the outer crater floor still pushing inwards after vertical collapse commences, a significant portion of the upper levels of the central peak may be drawn radially outwards, thus thinning the shallow upper levels of the central uplift. Also, as the acoustic vibrations driving mechanical weakening in the central uplift volume decay first at depth, the roots of the central uplift will strengthen before the more freely moving shallow levels, allowing transmission of stress over larger distances and consequent development and preservation of large-scale collapse-related folds and faults.

4.5 Conclusions

Tangentially-oriented, asymmetric, large-scale buckle folds in layered metasedimentary strata within the inner collar of the Vredefort Dome are attributed to subvertical collapse of the central uplift of the Vredefort impact structure. Fold asymmetry is consistent with centre-down shear along the steeply overturned bedding. Tangential vertical faults with centre-down slip sense that overprint the folds indicate an evolution from brittle-ductile to brittle deformation under the same vertical flattening strain field. Additionally, radial to oblique faults in the NNE sector of the dome display km-scale offsets consistent with tangential extension after overturning and folding. The folds and faults extend the inventory of macroscopic impact-induced structures observed in the Vredefort Dome and suggest that vertical collapse-related features may be more common than previously recognised. Vertical collapse of the inner parts of central uplifts in large complex craters has implications for crater floor morphology and the distribution of impact melt in complex craters. The presence of these collapse-related folds and faults in rocks originally buried between 8 and 10 km below the surface of the Vredefort crater indicates that strain effects linked to central uplift collapse may persist to significant depths within the central uplift volume and are not limited to the near-surface environment.

4.6 Acknowledgments

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Chapter 5

IMPACT-RELATED FRACTURE PHENOMENA

5.1 Introduction

5.1.1 Fractures

Fracturing is the irreversible, rapid propagation and connection of grain-scale cracks. These cracks are called Griffith cracks and are sub-microscopic flaws that pre-exist and can be induced in grains or along grain boundaries ([Griffith, 1920](#)). These cracks have a flattened and elongated ellipsoid shape. Fracturing initiates when the differential stress exceeds the rock's yield strength. The fractures spread spontaneously from their tips, connecting with nearby cracks which coalesce to form a fracture. Fractures can be broadly divided into two categories: 1) Extensional fractures, which show extension perpendicular to the walls resulting from separation of two previously adjoining surfaces; the displacement is thus orthogonal to the fracture plane, parallel to σ_3 ; and 2) Shear fractures, where displacement is along the fracture plane (fracture - parallel); these fractures form obliquely ($20^\circ - 30^\circ$) to σ_1 ([Fossen, 2010](#)).

Fracture spacing can be described as the average orthogonal distance between adjacent fracture planes. In massive rocks, fracture spacing follows an approximately log-normal frequency distribution ([Ruf et al., 1998](#)). In layered rocks, fracture spacing is proportional to layer thickness and differs according to lithology, structural position and strain (e.g., [Ferrill and Dunne, 1989](#); [Evans and Dunne, 1991](#); [Bai and Pollard, 2000](#); [McGinnis et al., 2017](#)). Fractures are more closely spaced in thinner beds as they release tensile stress over a lateral distance proportional to the fracture length. Stronger (more brittle) rocks have more closely spaced fractures than weaker (less brittle) rocks of the same thickness (e.g., [Ramsay and Huber, 1987](#); [Twiss and Moores, 1992](#); [Pollard and Fletcher, 2005](#); [Fossen, 2010](#)). Fracture spacing-to-layer thickness ratios - known as the fracture saturation index (FSI) - range from ≤ 0.1 to 10 ([Bai et al., 2002](#)). As tensile stress increases, more fractures sequentially fill in between earlier adjacent fractures. [Gross \(1993\)](#) described decreasing fracture spacing as sequential infilling. Sequential infilling occurs until fracture saturation is obtained ([Kelly and Tyson, 1965](#); [Price, 1966](#)). Fracture saturation occurs once the critical ratio (usually FSI=1) has been attained and pressure shadows form around the fractures ([Bai et al., 2002](#)). After that, the local stress fields around the fractures transition from tensile to compressive, forming orthogonal fractures ([Bai and Pollard, 2000](#); [Bai et al., 2002](#)) (Figure 0.1). Orthogonal fractures in tectonic settings can

be achieved through a 90° regional stress field inversion or sequential infilling and fracture saturation (Figure 0.1).

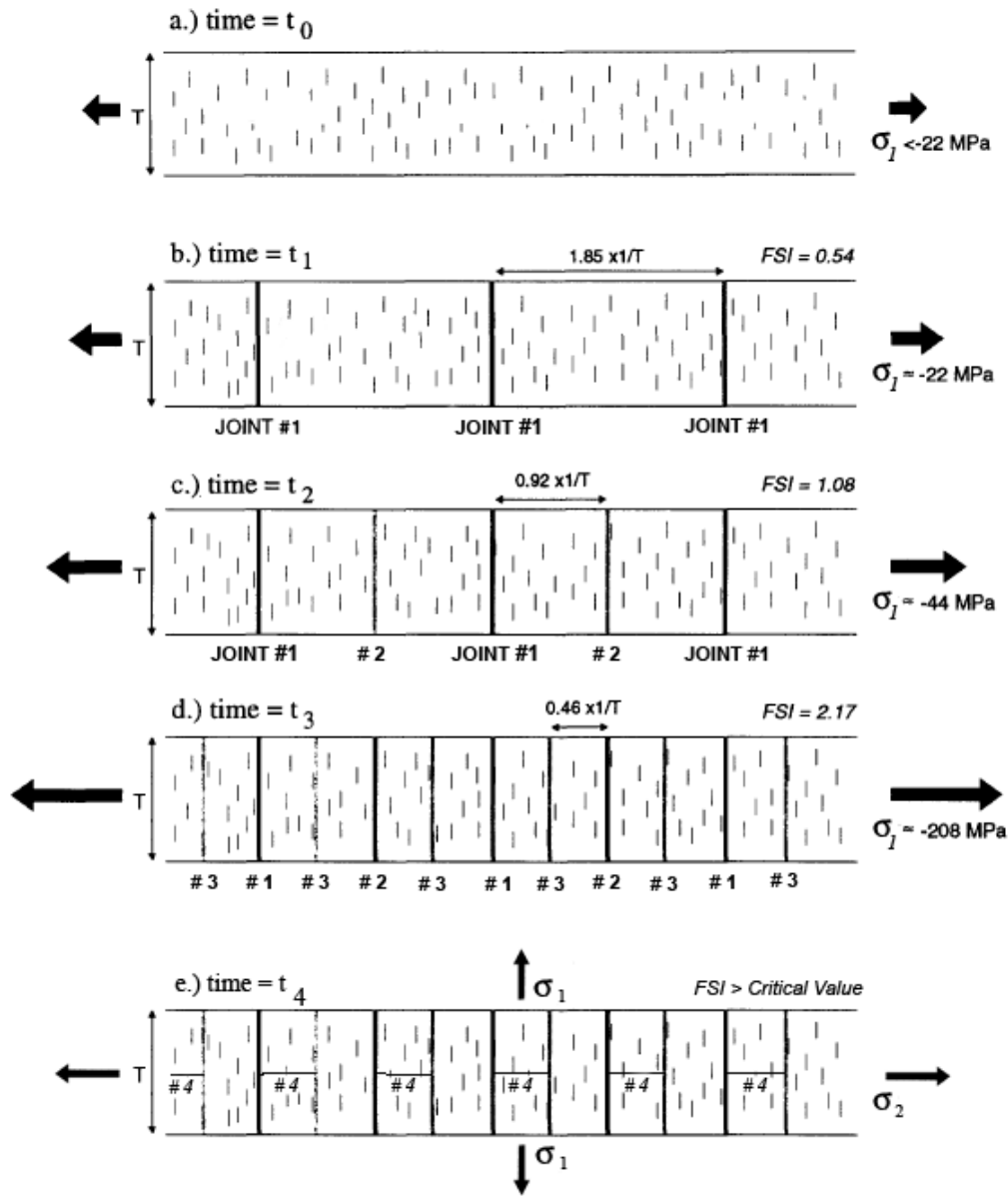


Figure 0.1: Model for fracture saturation leading to the formation of orthogonal fractures under a consistent stress field. Modified from [Gross \(1993\) figure 9](#). Sequential infilling occurs as stress increases over time in a tectonic setting. σ_1 denotes the maximum tensile stress. At a critical ratio of fracture spacing to layer thickness (FSI), the layer reaches fracture saturation, and the stress field undergoes local stress field inversion. Pressure shadows form around the fractures, facilitating a stress inversion. Further stress is accommodated by fractures forming in the orthogonal direction. The application of such a mechanism in a hypervelocity impact setting is unproven.

When more than one fracture set occurs in the same volume of rock, determining the relative timing of fractures, it is noteworthy that new fractures can be assisted by the fact that new

fractures will grow pre-existing fractures at a 90° angle ([Fossen, 2010](#)) (Figure 0.2). Later fractures are also shorter than earlier fractures and are commonly truncated against the primary fracture in a T or Y intersection ([Pollard and Fletcher, 2005](#)).

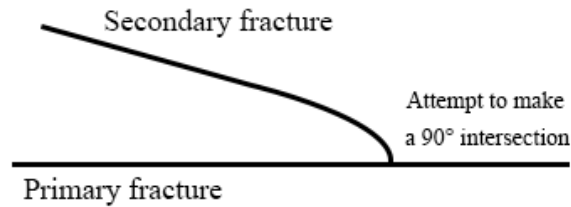


Figure 0.2: Schematic showing the abutting relationship between secondary and primary fractures.

A fracture propagating too fast has stress oscillations at the Griffiths wing cracks (microscopic defects or cracks in the crystal lattice) that cause it to bend. The resulting features are hackles: long, narrow planes that form oblique to the main fracture (plume) axis (Figure 0.3) ([Fossen, 2010](#)). They initiate at an angle of $\sim 30^\circ$ to the main fracture axis and curve gradually to $\sim 70^\circ$ at the margins ([Bahat et al., 2005](#)). Plume axes develop parallel to the main propagation direction. The array of hackles forms a plumose (feather-like) structure (Figure 0.3). Arrest lines (ribs) occur because the fracture forms in pulses. These features mark places of minimum fracture propagation velocity. Rib marks form perpendicular to hackles.

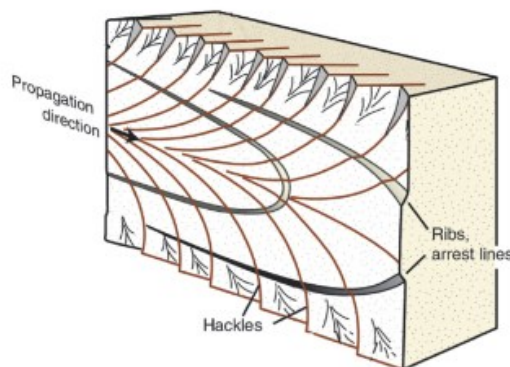


Figure 0.3: Schematic of a plumose fracture/joint structure showing hackles and ribs/arrest lines. Modified from [Fossen \(2010\)](#).

5.1.2 Branching Fractures

Branching fractures form spontaneously when the ratio of the fracture propagation velocity to Rayleigh wave velocity (V_R) reaches the critical threshold of $0.3 V_R$ for the medium it is travelling through (e.g., [Lee and Freund, 1990](#); [Kulander and Dean, 1995](#); [Rosakis and Ravichandran, 2000](#); [Sagy et al., 2001](#); [Bhat et al., 2012](#)). Rayleigh waves are surface waves

that move in an elliptical motion. They produce both a vertical and horizontal motion component in the direction of wave propagation ([Braile, 2001](#)). Materials return to their original shape after the wave passes ([Braile, 2001](#)). At the given threshold, a single fracture bifurcates on heterogeneities (micro flaws) that act as a nucleation point by scattering waves and amplifying the tensile stress ([Griffith, 1920](#)). In addition, weak interfaces between beds may encourage fracture branching (e.g., [Xu et al., 2003](#); [Chalivendra and Rosakis, 2008](#); [Sundaram and Tippur, 2016](#)). The bifurcation minimises the energy of the system ([Meyers, 1994](#)). Branching fractures are tensile fractures ([Bahat, 1982; 1991](#)). They open in the direction of fracture propagation with their apices pointing towards the energy/wave source ([Hossain, 2015](#)). Branching fractures have been successfully reproduced in dynamic tensile experiments and high-impact cratering experiments (e.g. [Sagy et al., 2004](#); [Wilk and Kenkmann, 2016a](#)). [Sagy et al. \(2001\)](#) proposed that dynamic fracturing at high strain rates ($> 4 \times 10^{-3} \text{ s}^{-1}$) should not be thought of in terms of the theoretical framework of fracture formation. Instead, they should be considered a new mechanism for developing high fracture densities. Branching fractures might indicate past seismic events in the rock ([Sagy et al., 2001](#); [Rowe and Griffith, 2015](#)).

In meteorite impacts, branching fractures are proposed to form as a result of tensile stresses that form in the back end of the shock wave during the contact and compression phase ([Sagy et al., 2004](#)). According to [Baratoux and Melosh \(2003\)](#), the magnitude of the tensile stresses could be as much as half the value of the compressive pressure of the shock wave front. Horsetail fractures are branching fractures that form when the fracture propagation is between 0.3 and 0.9 times the VR in the given medium ([Sagy et al., 2006](#)) (Figure 0.4). Horsetail fractures occur in brittle-ductile shear zones and impact settings and are not unequivocal indicators of impact. Irrespective of the scale or setting of the horsetail fractures, there is no evidence of slip or measurable shear displacements ([Hossain, 2015](#)). Pseudotachylyte is not seen on horsetail fractures, indicating an absence of shearing ([Hossain, 2015](#)). Instead, they show diverging pairs of grooves, indicating their extensional nature based on geometry ([Hossain, 2015](#)). The horsetail's long axis is the direction perpendicular to the lines of intersection of the different fractures and opposite to the branching direction of the horsetails (Figure 0.4).

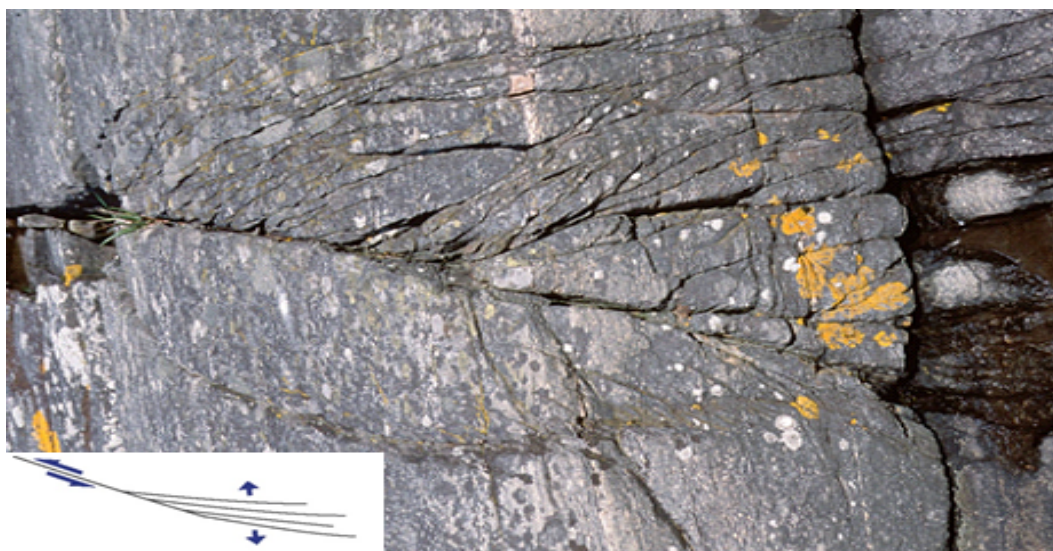


Figure 0.4: A tectonic horsetail fracture in gneiss at the end of a shear fracture. The horsetail itself does not display any shearing. Modified from [Fossen \(2010\)](#). The horsetail's long axis is the direction perpendicular to the lines of intersection of the different fractures and opposite to the fanning direction of the horsetails.

5.1.3 Shatter Cones

Shatter cones comprise a range of curved to curvilinear fractures with divergent striations on their surfaces (Figure 0.5) ([Baratoux and Reimold, 2016](#)). The striations form ridges and valleys that radiate from an apex or apical area on a smooth, well-rounded conical feature (e.g., [Nicolaysen and Reimold, 1999](#)). The apical areas, of a few millimetres to centimetres, may be crudely round to polygonal in profile and delimited by fractures (Figure 0.5) (e.g., [Hasch et al. 2016](#)). Shatter cones may be covered locally with thin films of melt rock ([Gibson and Spray, 1998](#); [Nicolaysen and Reimold, 1999](#); [Pittarello et al., 2015](#)). Shatter cones occur in clusters pervasive at an outcrop scale (Figure 0.5) (e.g., [Hasch et al. 2016](#)). They are restricted in their natural occurrences solely to meteorite impact sites and appear to form at pressures between 2 and 30 GPa (e.g. [French, 1998](#); [Osinski and Spray, 2006](#); [Ferriere and Osinski, 2010](#)). Consequently, they are regarded as an unequivocal indicator of impact ([Dietz, 1960](#)). [Manton \(1965\)](#) and [Nicolaysen and Reimold \(1999\)](#) noted that shatter cones occur in close proximity to an intense sub-planar to curvilinear jointing, referred to as multiply striated joint sets (MSJS). [Sagy et al. \(2002\)](#) reported that shatter cones are not discrete, isolated bodies within the rock volume but are formed along larger coherent fractures. This is visible in Figure 0.5, whereby the axes of shatter cones associated with a single large fracture are aligned over decametres in what is referred to as a “cone-in-cone” network. Individual cones occur on planar fractures with the conical surfaces branching away from the main fracture. Shatter cones are

generally studied in geographic and geometric isolation from other features. Shatter cones have been found in most crustal lithologies but are developed best in finer-grained rocks ([Baratoux and Reimold, 2016](#)). They show a positive correlation between size and layer thickness ([Baratoux and Reimold, 2016](#)). Shatter cones have been reproduced in explosion experiments at pressures between 2 - 6 GPa ([Bunch and Quaide, 1968](#); [Milton, 1977](#); [Roddy and Davis, 1977](#)) and in shock experiments at pressures up to 70 GPa ([Schneider and Wagner, 1976](#); [Kenkmann et al., 2012](#); [Wilk and Kenkmann, 2015a](#); [2015b](#); [2016](#)).

The orientation of shatter cones is attributed to the orientation of stresses during or shortly after shock, and variability was explained by reflections of the shock wave (e.g., [Wieland et al., 2006](#); [Osinski and Ferrière, 2016](#)). [Rae et al. \(2021\)](#) showed that stress and strain vary spatially and with time during the shock wave passage, even before accounting for reflections. They demonstrated that the shatter cones form parallel to the direction of maximum instantaneous shortening/minimum instantaneous stretching (ISA₃), or finite strain, as opposed to the maximum principal stress orientation, with the shatter cones forming early in the shock deformation history. Historically, shatter cones were said to point to the centre of the impact sites ([Milton, 1977](#); [Roddy and Davis, 1977](#)). [Manton \(1965\)](#) proposed that, after backrotation of bedding to a horizontal orientation, shatter cones point to the centre of the Vredefort impact structure; however, this was challenged by [Wieland et al. \(2006\)](#), who found shatter cones with variable apex orientations, from parallel to normal to bedding planes, on an outcrop scale. [Milton et al. \(1996\)](#) found that, once reoriented to pre-impact orientations, the shatter cone apices at Gosses Bluff converge at the same location in the centre of the impact structure but at different altitudes. However, [Rae et al. \(2021\)](#) reconciled this by showing that shatter cones should not converge to a single focal point because the shock wave varies spatially and temporally. Shatter cone orientations are also expected to be asymmetric in craters formed by oblique impacts ([Pierazzo and Melosh, 1999](#); [2000](#); [Dahl and Schultz, 2001](#); [Cox et al., 2021](#)). The interaction between an asymmetric shock wave and the rarefaction wave will create even further misorientation from the radial orientation, as presumed by [Manton \(1965\)](#).

Several models exist to explain shatter cone formation. Most authors propose they form in the decompressional part of the shock wave because of the shock wave interacting with heterogeneities in the rock ([Baratoux and Melosh, 2003](#); [Sagy et al., 2004](#); [Wieland et al., 2005](#); [Dawson, 2009](#); [Baratoux and Reimold, 2016](#); [Osinski and Ferrière, 2016](#)).

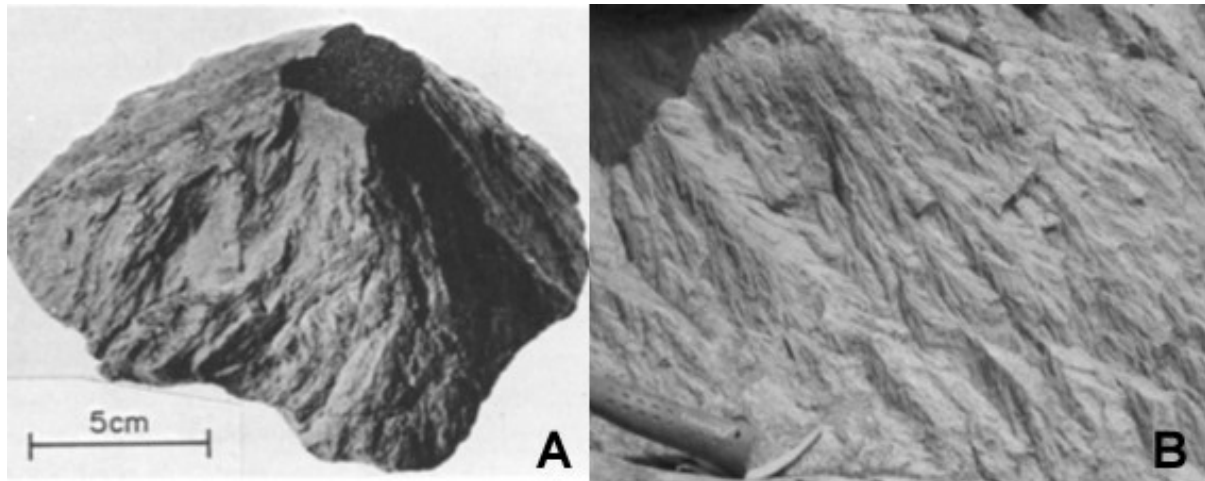


Figure 0.5: Shatter cone examples displaying perfect habits. (A) An example of a complete shatter cone in metabasalt from the Vredefort Dome, the apical area $\sim 2\text{cm}$ wide is polygonal and delimited by fractures ([Albat and Mayer, 1989](#)). (B) Typical cone-in-cone fracture network branching off a single large fracture within quartzite in the Vredefort Dome from the Schoemansdrift locality studied in this chapter ([Sagy et al., 2002](#)).

[Johnson and Talbot \(1964\)](#) studied the effect of a scattered elastic wave due to heterogeneities in the rock. The elastic precursor wave (elastic longitudinal wave) and scattered wave combine to exceed the HEL in a conical region, with the heterogeneity at its apex. The material within this region undergoes permanent plastic deformation. Outside of the conical region, the superposition of the waves reduces the stress, keeping it below the HEL. The rock in this region undergoes elastic deformation and then returns to its initial state. The two regions part to form a cone.

[Gash \(1971\)](#) purported that shatter cones are tensile fractures. They were proposed to form by superimposing the shock wave (plastic wave) generated below the earth's surface and the reflected tensile waves from the free surface. Shatter cones would then develop along the conical trajectories of the principal compressive stress. The result would be shatter cones developing directly beneath the point of impact at apical angles near 90° .

[Sagy et al. \(2002; 2004\)](#) suggested that the curved or spoon-like shape and network of fractures resemble tensile branched fractures, as depicted in dynamic fracture experiments ([Sharon et al., 1995](#); [Sagy et al., 2001](#)). Branched fractures require fracture propagation velocities above $0.9 V_R$. They also suggested that the striations can be explained using experimental rapid fracture analysis, whereby a rapidly moving elastic front wave creates a pair of traces on the fracture surface when encountering a heterogeneity ([Sharon et al., 2001; 2002](#); [Fineberg et al., 2003](#)). Thus, the angle between the two striations should reflect the fracture propagation

velocity, which depends on the strength of the tensile stress and, consequently, the local intensity of the shock wave. According to them, the decreasing intensity of the tensile stress with distance from the point of impact can potentially be used to predict the locations of the various types of branching features in impact sites relative to the point of impact ([Sagy et al., 2006](#)).

The [Baratoux and Melosh \(2003\)](#) hypothesis relies on constructive interference between the shock wave and the scattered wave formed at a heterogeneity in the rock. This interference creates a localised excess tensile stress if the inclusion has a lower density than the surrounding rock—the excess tensile stress forms as a curved surface. If the material's strength in tension is exceeded, rupture occurs along the curved surface, forming a shatter cone.

[Dawson \(2009\)](#) reported that shatter cones form from adiabatic shear bands during maximum shock compression due to material heterogeneities. [Wright \(2002\)](#) describes an adiabatic shear band as a planar region of significant shearing caused by thermal softening due to shear strain-induced heating. This theory is supported by the presence of pseudotachylite on the shatter cone surfaces (e.g., [Hargraves and White, 1996](#); [Gibson and Spray, 1998](#)). Adiabatic shear banding was modelled numerically, producing a conical feature akin to shatter cones. [Kenkmann et al. \(2016\)](#) took a phenomenological approach. They explained the horsetailing effect as a phenomenon whereby fractures propagating at velocities approaching the VR bifurcate on curved fracture planes, where their intersection lineations form the diverging striations of shatter cones. This model explains their formation without the pre-requisite of a material heterogeneity.

5.1.4 Shatter Cleavage

[Milton \(1977\)](#) and [Milton et al. \(1996\)](#) described a fracture network they termed “shatter cleavage” in the Carmichael Sandstone in the Gosses Bluff impact structure, Australia (Figure 0.6A), although they did not present any further data on the feature. Similar intersecting joint sets were also identified in the Huronian metaquartzite of the Sudbury South Range (Figure 0.6B) ([Nicolaysen and Reimold, 1999](#)). [Kenkmann et al. \(2014\)](#) noted closely spaced fracture networks in the quartzite of the Matt Wilson Crater, which they recorded as displaying stair-stepping displacements (Figure 0.6D). [Albat and Mayer \(1989\)](#) completed a detailed study of such shatter cleavage, which they referred to as s-surfaces, in the Vredefort Dome (Figure 0.6C). They were the first and only authors to attempt to link shatter cones and these fractures geometrically. The term shatter cleavage will be used throughout this work to describe what

[Albat and Mayer \(1989\)](#) referred to as s-surfaces; their observations are summarised below. The fracture cleavage is not considered a shock diagnostic feature despite only being reported in impact settings.

[Albat and Mayer \(1989\)](#) recorded pervasive single and conjugate, densely spaced, systematic arrays of shatter cleavage in the quartzites of the WRG of the Witwatersrand Supergroup in the Vredefort Dome (Figure 0.6C). They were reportedly found throughout the dome, although rarely observed in the argillaceous units and meta-ironstones. These shatter cleavages are planar, although locally gently curved. The cleavage plane spacing is typically 1 - 1.5 cm (Figure 0.6C), but spacing of as little as a few millimetres has been observed. The two cleavage sets are not equally well developed. The shatter cleavages are commonly enhanced by weathering on the outcrop surface but become indistinct welded planes deeper into the sample ([Albat and Mayer, 1989](#)). Microscopic examination reveals they represent planar zones of fine-grained recrystallised quartz in a medium-to-coarse-grained quartzite ([Albat and Mayer, 1989](#)).

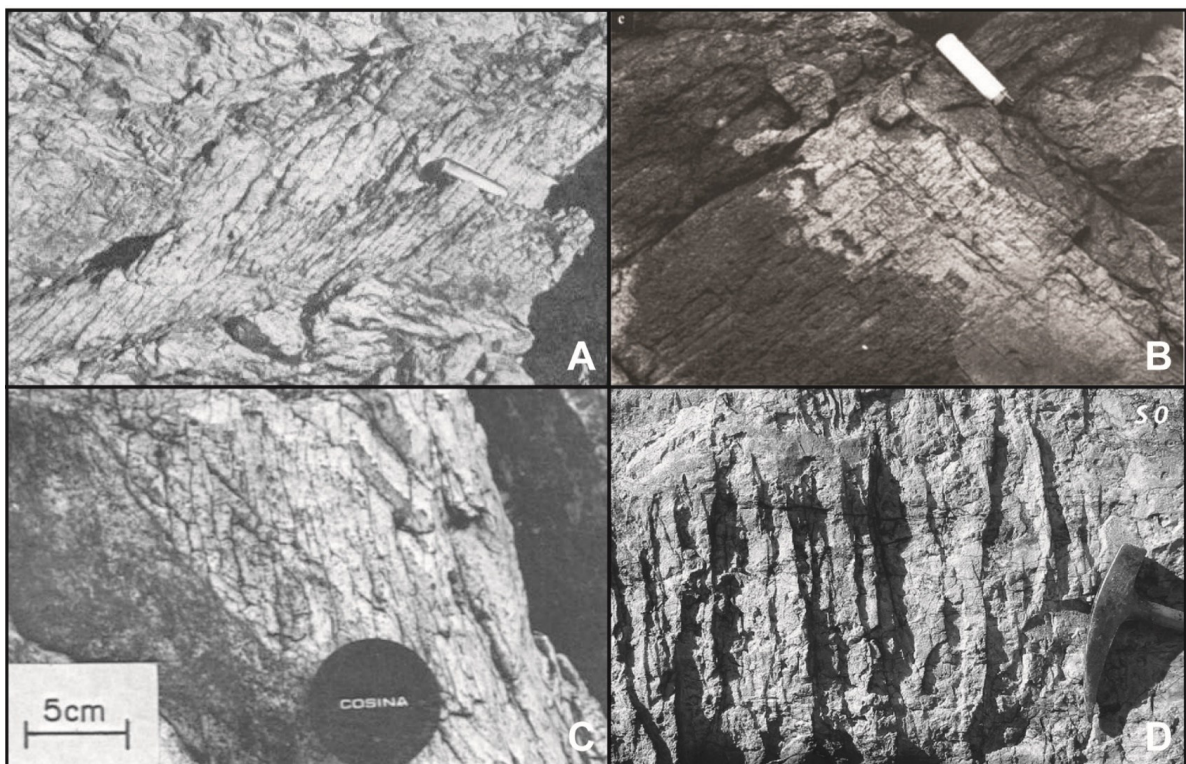


Figure 0.6: Examples of shatter cleavage from various impact structures. (A) Centimetre-spaced shatter cleavage found in the Carmichael Sandstone in the Gosses Bluff impact structure, Australia, lighter for scale ([Milton, 1977](#)). (B) Dense cross-cutting joint sets in the Huronian metaquartzite of the Sudbury South Range, Lighter for scale ([Nicolaysen and Reimold, 1999](#)). (C) Centimetre-spaced s-surfaces, herein referred to as shatter cleavage, found in quartzite of the Hospital Hill Subgroup in the Vredefort impact structure, South Africa ([Albat and Mayer, 1989](#)). (D) Fracture network found in the quartzite of the Matt Wilson Crater, with stair-stepping displacements and geological hammer for scale ([Kenkmann et al., 2014](#)).

[Albat and Mayer \(1989\)](#) proposed that shatter cleavages possess a geometric relationship with a so-called master shatter cone axis. The master cone theory ([Dietz, 1963, 1968](#); [Milton, 1977](#); [Roddy and Davis, 1977](#); [Simpson, 1981](#)) suggests that the apices of multiple shatter cones in a volume of rock create a larger-scale cone of best fit. The master cone theory has lost popularity as whole cones are seldom preserved, and shatter cones have been found to point in opposite and highly variable directions within one volume of rock, making it an unlikely hypothesis. However, [Albat and Mayer \(1989\)](#) proposed that the two shatter cleavages encase the master shatter cone. The master cone axis is said to be perpendicular to the intersection of the shatter cleavages and bisects the obtuse angle of the set. They measured the mean angle between the two shatter cleavages at 80° (standard deviation of 5.9°). The mean angle between the two shatter cleavages was thus complementary to the mean apical shatter cone angle of 100° . Based on this relationship, they proposed that the shock wave produced the cleavages before overturning the collar rocks. This early formation is further supported by their findings of consistent angular relationships between bedding and the shatter cleavages, indicating a consistent bedding orientation at the time of shatter cleavage formation. Based on a simplistic palinspastic reconstruction of the horizontal, they show that the shatter cleavages had initially steep dips, with their intersections plunging steeply towards the centre of the dome. According to them, these shatter cleavages can thus act as a marker in time to constrain pre- and post-shock wave features and may be used to identify other impact structures ([Albat and Mayer, 1989](#)).

[Nicolaysen and Reimold \(1987\)](#) described intersecting, closely spaced sets of multiple-striated joint surfaces (MSJS) with spacing from less than 1 mm to several mm. MSJS are fractures characterised by a distinctly curvilinear geometry that can occur in up to 12 orientations within a single outcrop that can vary from outcrop to outcrop, having no preferred orientation. [Nicolaysen and Reimold \(1999\)](#) concluded that the s-surfaces described by [Albat and Mayer \(1989\)](#), herein referred to as shatter cleavage, were two preferentially selected orientations of MSJS. They determined that the MSJS were directly related to shatter cones in the Vredefort Dome. They supported this by noting the presence of pseudotachylyte on shatter cones and MSJS, as well as the surfaces being both striated and found near one another. They noted shear displacements of up to several hundred micrometres at the grain scale along these joints. [Gibson et al. \(1997\)](#) suggested that the cm-spaced fractures that hosted pseudotachylyte in the metapelites of the WRG constituted a fracture cleavage. However, they did not link it to the shatter cleavage described by [Albat and Mayer \(1989\)](#).

5.2 Results

Impact-related fracture phenomena were investigated in study areas 1 and 2, as well as the Schoemansdrift locality, as part of the broader analysis of the chronology of deformation associated with the Vredefort impact. As [Wieland \(2006\)](#) and [Wieland et al. \(2005, 2006\)](#) reported extensively on large-scale joints and shear fractures that they attributed to the final stages of relaxation of the Vredefort crater, these are not considered here. [Gibson et al. \(1997\)](#) reported pseudotachylyte associated with fractures that they reported locally having the spacing intensity of a fracture cleavage in metapelitic units of the Hospital Hill Subgroup in the northern collar; however, this relationship was not seen in the various study areas investigated; instead, pseudotachylyte dykes are associated with faults ([Chapter 6](#)) and cut the shatter cleavage.

5.2.1 Horsetail Fractures

Several examples of macroscopic branching fractures were identified at Nauwpoort (Study Area 1; Figure 0.7). The branching fractures are spatially associated with a single shatter cleavage orientation (CLA, see next section). The shatter cleavage is the primary fracture that branches into first-, second- and higher-order fractures. The master fracture and long axis lie parallel to the shatter cleavage, with the long axis perpendicular to the lines of intersection of the different fractures and opposite to the branching direction of the horsetails. The propagation direction of the fracture system can be determined using the branching direction (Figure 0.7). The individual branching fractures range from a few millimetres to several decimetres, but only those longer than 1 cm were sketched and analysed. Up to 8th-order branching was counted beginning from a single primary fracture. The branching fractures show a preferred fracture orientation roughly parallel to the long axis/primary fracture. The branching fractures do not show any macroscopic evidence of slip or shear displacements but could only be viewed on the bedding surface, which provided no clear displacement markers.

Horsetail fractures were not reported by [Manton \(1965\)](#), [Albat and Mayer \(1989\)](#), [Nicolaysen and Reimold, 1999](#), [Wieland \(2006\)](#), etc. The reason may be the inaccessibility of bedding surfaces owing to the general steep dip.

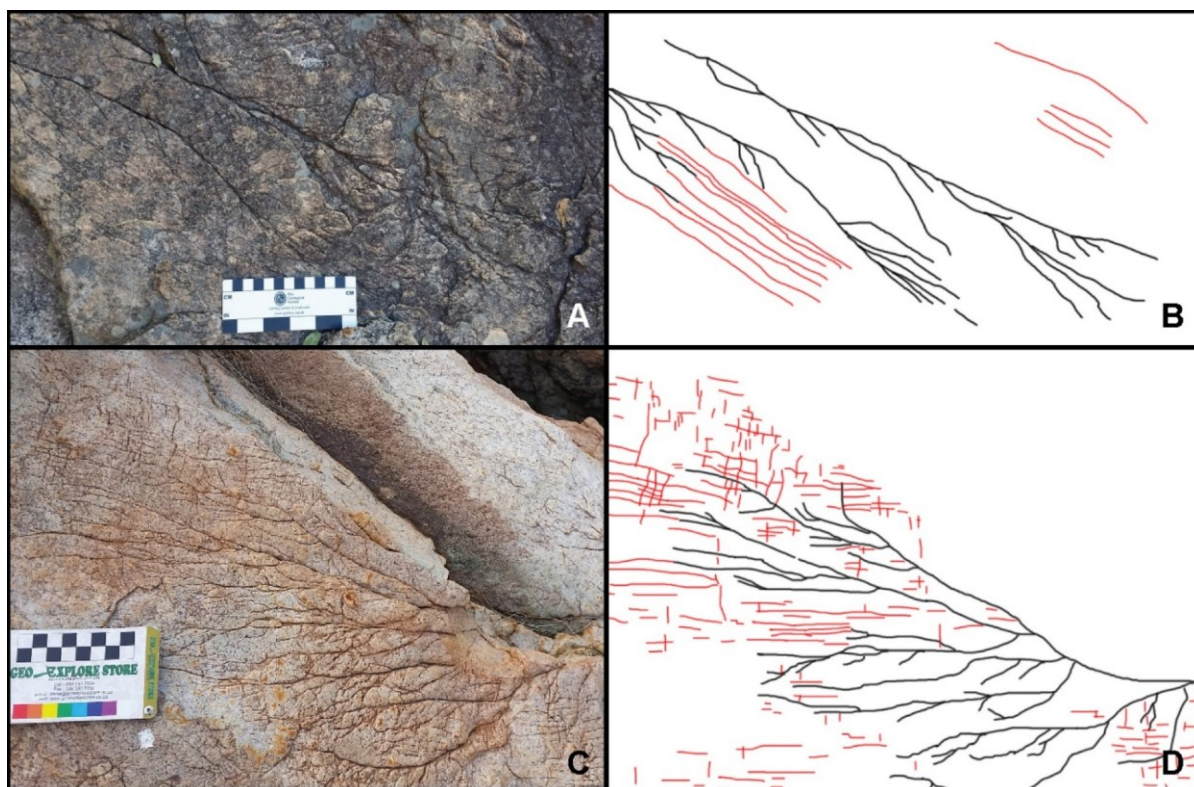


Figure 0.7: Horsetail fractures preserved in Rhq3 quartzite at the Nauwpoort locality. Note that the orientation of the horsetail fracture's long axis corresponds to a shatter cleavage seen more clearly further away from the horsetail fractures, as shown in B and D. The bearing to the dome's centre at this outcrop is 115°. (A) Horsetail fracture oriented 288°/42°, bedding orientation 044°/67°/SE. Photo location UTM 35J 526641 E 7019566 S. (B) Illustration of the horsetail fracture in black and shatter cleavage in red. (C) Horsetail fracture 146°/14°, bedding orientation 142°/16°/W. Photo location UTM 35J 0526585 E 7018982 S. (D) Illustration of the horsetail fracture in black and shatter cleavages in red.

5.2.2 Shatter cones

Shatter cones were identified in all the study areas (**Figure 0.8**). The shatter cones varied in size between 10 and 15 cm in length. At the Schoemandsdrift locality, they display cone-in-cone structures (**Figure 0.8A**). The striae diverge from an apex curving to form cones, conforming to that of conventionally studied shatter cones. Surfaces are typically coated with black Fe and Mn oxides and hydroxides and were thus not investigated for evidence of glass as reported by e.g. [Gibson and Spray \(1998\)](#), [Nicolaysen and Reimold \(1999\)](#), [Hasch et al. \(2016\)](#), [Wilk et al. \(2018\)](#). At Schoemandsdrift, the shatter cone orientations formed a partial mastercone which was used to constrain the apex location for the mastercone used in the orientation analysis (see [Chapter 5.2.4](#)).



Figure 0.8: (A) ~10 cm long shatter cones displaying a cone-in-cone structure preserved in the Rjo1 quartzite ([Bisschoff, 2000](#)) of the Turffontein subgroup CRG at the Schoemansdrift locality (Figure 0.2). Mean shatter cone axis orientation $162^{\circ}/30^{\circ}$. Photo location UTM 35J 520756 E 7016812 S. Bearing to the centre of the dome 105° . (B) ~15 cm long shatter cones preserved in the Rhs4 metapelite ([Bisschoff, 2000](#)) at the Geluksoord locality (Study Area 2; Figure 0.2) with the long axis oriented $010^{\circ}/82^{\circ}$. The bearing to the centre of the dome from this outcrop is 210° . Photo location UTM 35J 0556774 E 7029048 S. (C) Poorly formed shatter cones, ~ 10 cm long, preserved in Rgs2 metapelite ([Bisschoff, 2000](#)) at the Nauwpoort locality (Study Area 1; Figure 0.2). Bearing to the centre of the dome 116° . Photo location UTM 35J 526203 E 7019391.

5.2.3 Shatter Cleavage

The shatter cleavage is shown by [Albat and Mayer \(1989\)](#) to be pervasive across the dome in the quartzite units. However, this study analysed it in detail in three areas: Schoemansdrift, Nauwpoort (Study Area 1; Figure 0.2) and Geluksoord (Study Area 2; Figure 0.2). In these study areas, a pervasive, densely spaced, orthogonal, systematic array of shatter cleavages is found in the quartzites, metapelites and meta-ironstones (Figure 0.9 - Figure 0.11). The cleavage was not recorded in the meta-banded iron formations or metadolerite intrusions. The shatter cleavages are present as near-orthogonal, straight fractures, although microscopically, they are seen to be slightly curved, sinuous and undulating. The shatter cleavages are unequally developed. The dominantly developed set is referred to as “CLA”, and the less dominantly developed set is called “CLB”. The cleavages strikes can vary as much as 20° in a single outcrop (Figure 0.10D and Figure 0.12). Considering all domains, the mean angle between the two shatter cleavages is 97°, with a standard deviation of 7.7° within error of [Albat and Mayer’s \(1989\)](#) $100^{\circ} \pm 5.5^{\circ}$.

CLA and CLB are not confined to a particular bedding layer in the quartzites but cross-cut multiple layers on a cm- to dm scale. CLB locally displays abutting relationships with CLA, whereby it is truncated against CLA, although more commonly, they share cross-cutting relationships. The CLA spacing is typically between 0.5 - 1 cm, although it is found in places to be as dense as 1 mm (Figure 0.9A). The CLB spacing is typically between 5 - 15 cm, although it can be as dense as 2 mm (Figure 0.9). However, CLB is difficult to identify as it is discontinuous in places where it occurs at a spacing as large as 45 cm or is not present. Irrespective of layer thickness, the fracture spacing remains consistent (Figure 0.11). This may indicate that the spacing is a function of differential stress. The shatter cleavage CLB is not well developed near the two branching fractures (Figure 0.7). The branching fractures form at high angles to CLA (Figure 0.7). The fractures connecting CLB within the branching horsetail fractures tend to be irregular and discontinuous, as seen in Figure 0.7. This may indicate that the horsetail fractures form before CLB.

The shatter cleavages are accentuated where weathering has preferentially removed material and deposited oxides. Deeper into the samples, they are indistinct and tightly welded. Where bedding is near horizontal, weathering exploits the cleavage as planes of weakness forming rod-like fingers of rock (Figure 0.11C).

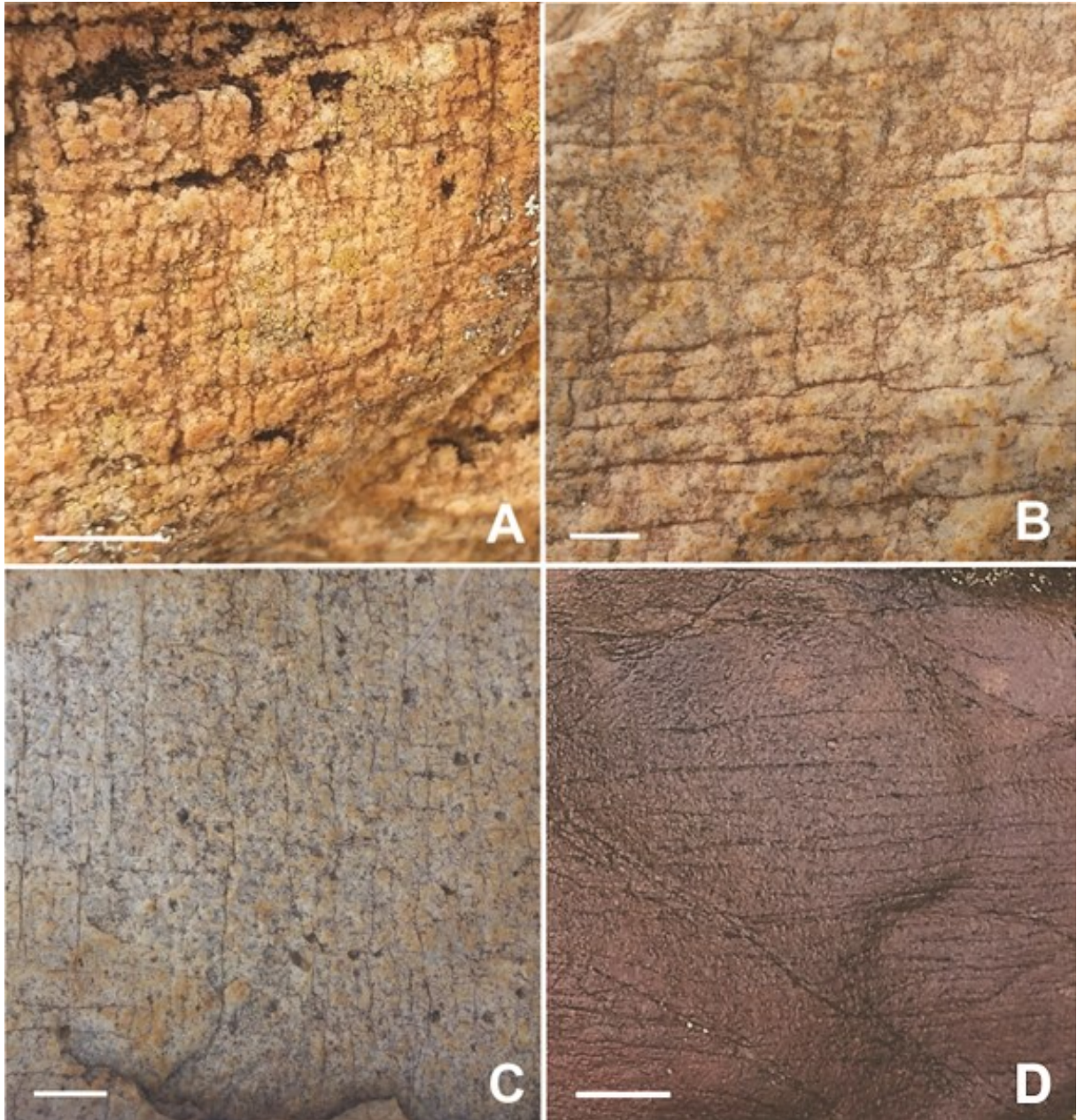


Figure 0.9: Various ways the shatter cleavages are preserved in Study Area 1. The white line in each of the above photographs represents 1 cm. (A) Millimetre spaced cleavage in quartzite (UTM 35J 0526746 E 7019609 S). (B) Centimetre scale CLA cleavage running left-to-right and a 3 cm spaced CLB cleavage running top-to-bottom in quartzite (UTM 35J 0526780 E 7019615 S). (C) Millimetre-spaced cleavage in quartzite (float; Sample RAS-01; no GPS location available). (D) Millimetre-spaced CLA cleavage in metapelite (UTM 35J 0526144 E 7019092 S).

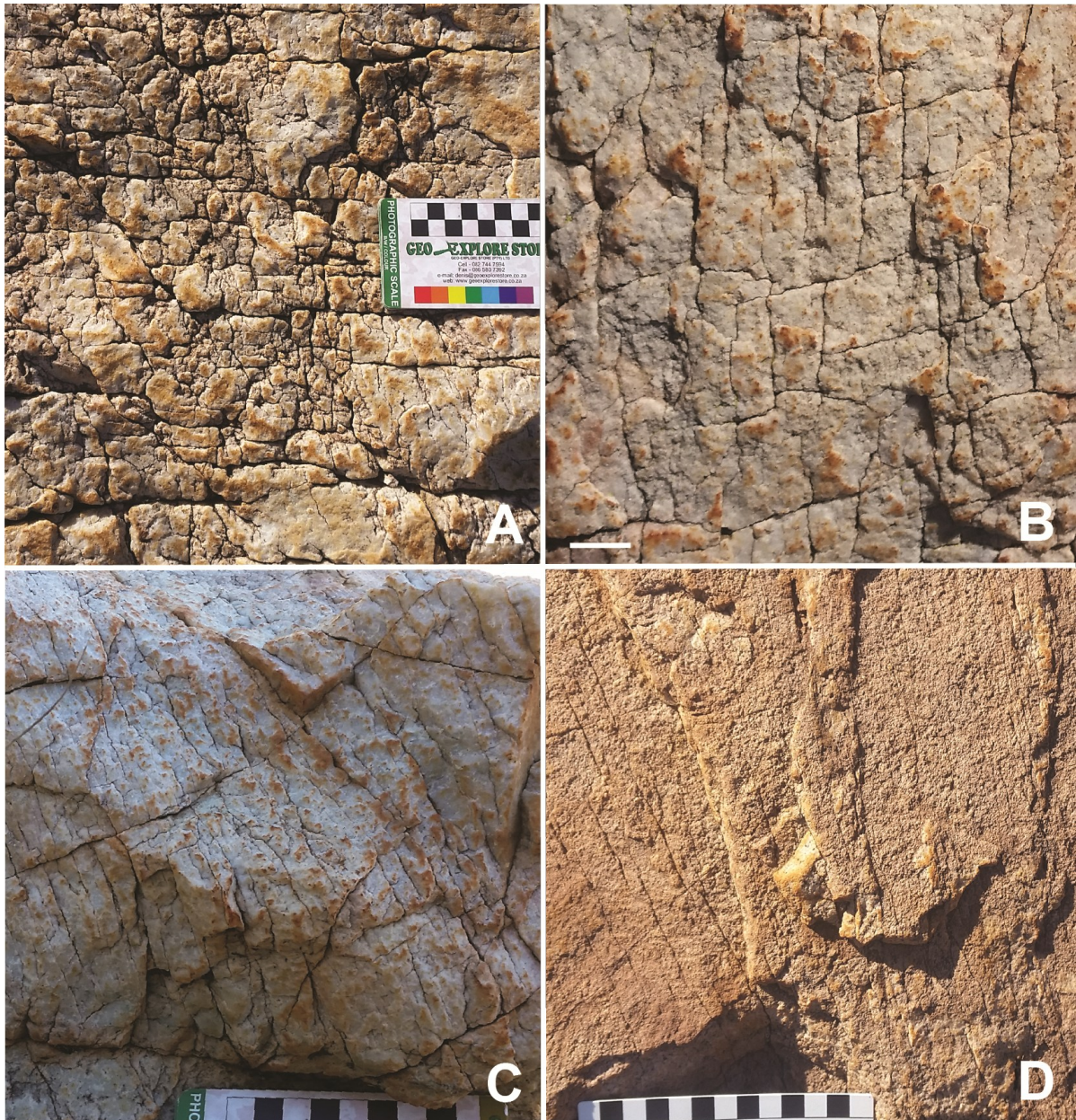


Figure 0.10: Various ways the shatter cleavages are preserved at Study Area 1. (A) Crocodile skin textured appearance caused by irregularly spaced cleavage in highly recrystallised quartzite where different layers' cleavage responds differently to weathering (UTM 35J 0526662 E 7019131 S). (B) Rhombohedral shatter cleavages in quartzite. The white line represents 1 cm (UTM 35J 0526386 E 7018820 S). (C) Very regular centimetre-spaced CLA cleavage running top-to-bottom with CLB running left-to-right with a 5 cm spacing in quartzite (UTM 35J 0526680 E 7019045 S). (D) Centimetre-spaced CLA cleavage in quartzite showing various orientations in the strike in one outcrop (UTM 35J 526300 E 7019198 S).

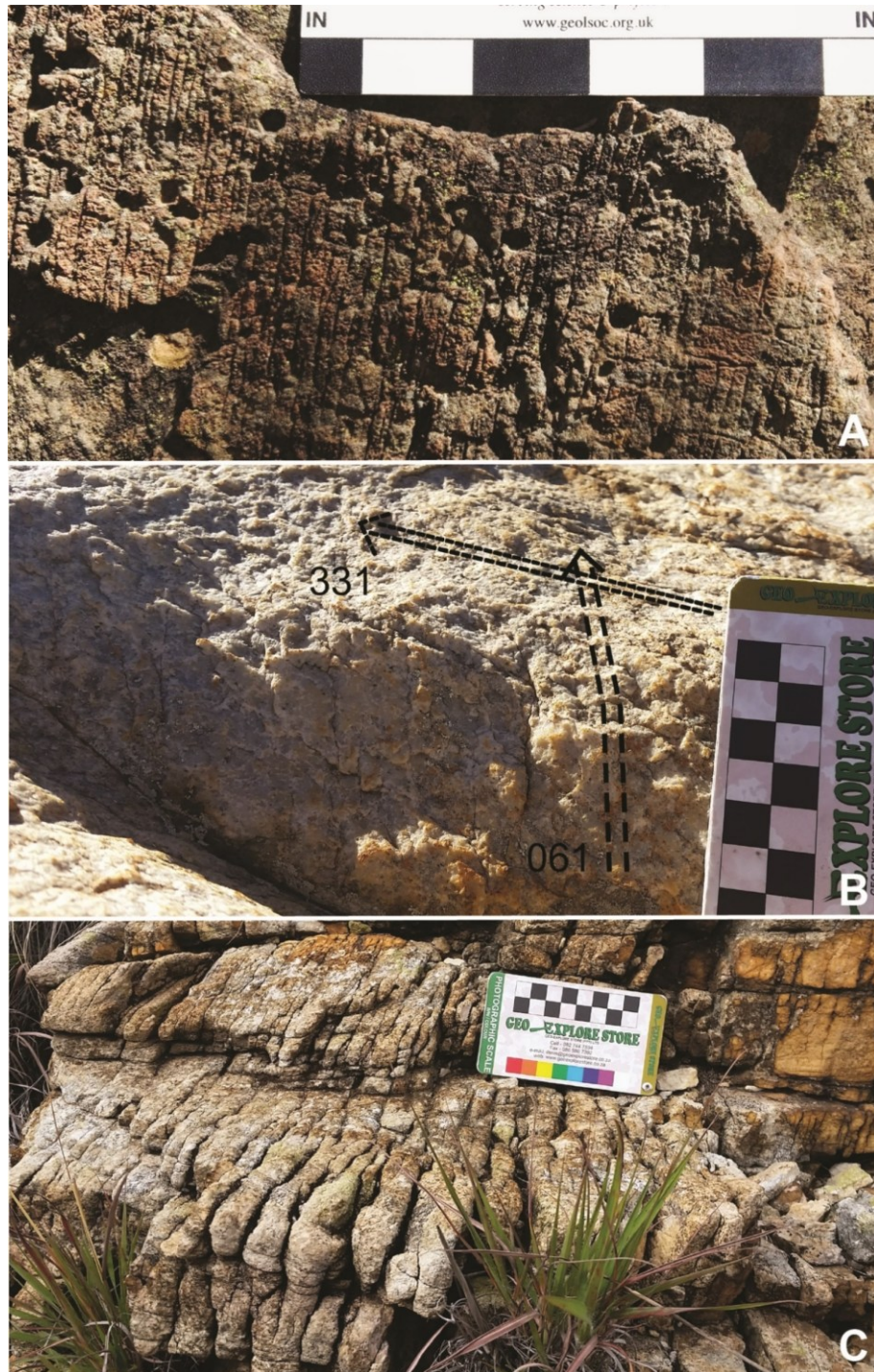


Figure 0.11: Various ways the shatter cleavages are preserved in Study Area 1. (A) Two sets of perpendicular shatter cleavage in metapelite. The cordierite porphyroblasts have weathered out to form circular voids (Photo location UTM 035J 528945 E 7022613 S). (B) Centimetre scale CLA cleavage running top-to-bottom present as faint welded planes making an inter-cleavage angle of 90° with CLB (Photo location UTM 35J 0526678 E 7019067 S). (C) Centimetre-spaced CLA cleavage (vertical) transecting multiple sedimentary beds being exploited by weathering, with horizontal sedimentary layering (Photo location UTM 35J 0526473 E 7019123 S).

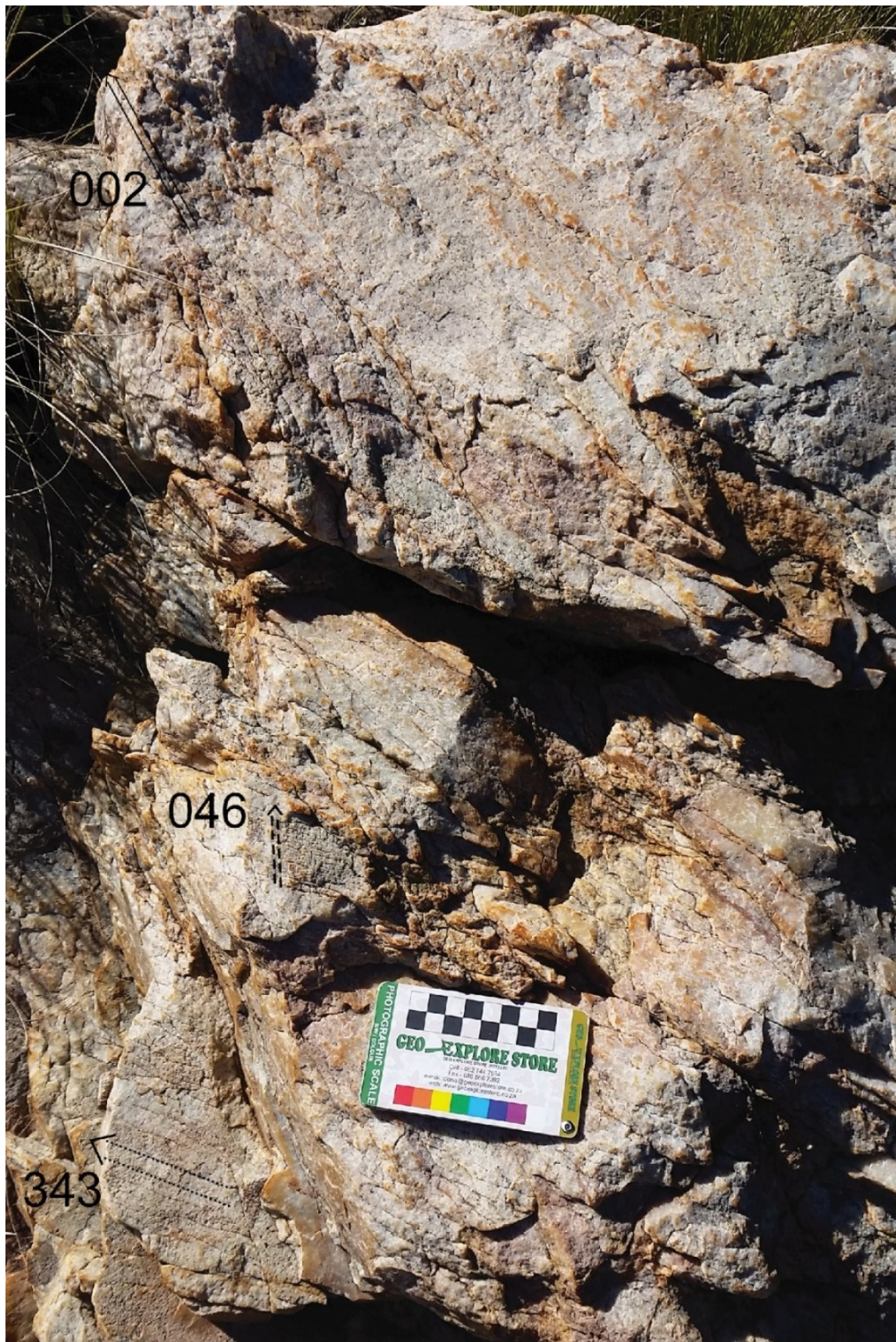


Figure 0.12: Centimetre-spaced cleavage in quartzite showing a variation in the strike of 17° - 34° in one outcrop (UTM 35J 0526680 E 7019045 S). Arrows indicate the orientation of the cleavage in the given location. Bearing to the centre of the dome is 116° .

Petrographic Analysis of Shatter Cleavage

Microscopic examination of 17 thin sections cut in multiple orientations reveals that the shatter cleavages are defined by narrow, $\sim 500\ \mu\text{m}$ wide subplanar zones of fine-grained ($125 - 250\ \mu\text{m}$) recrystallised quartz in a coarser-grained ($250 - 500\ \mu\text{m}$) quartzite matrix (Figure 0.13) or as fractures in a medium-grained ($250 - 500\ \mu\text{m}$) equidimensional interlocking granoblastic quartzite (Figure 0.13). The cleavage is best seen in DF and PPL by low reflectance areas. The orthogonal shatter cleavages is also occasionally marked by open fractures (Figure 0.14). CLA has an approximate spacing of 1 cm in a highly recrystallised fine-grained quartzite. CLB abuts against CLA. CLB fractures are more open than CLA. There are nor signs that the PDF's increase in intensity or abundance toward the shatter cleavage. Quartz grain boundaries are highly irregular indicative of dynamic recrystallisation. The grain size associated with the cleavage is smaller indicating a higher internal differential stress (Figure 5.14, Figure 5.16).

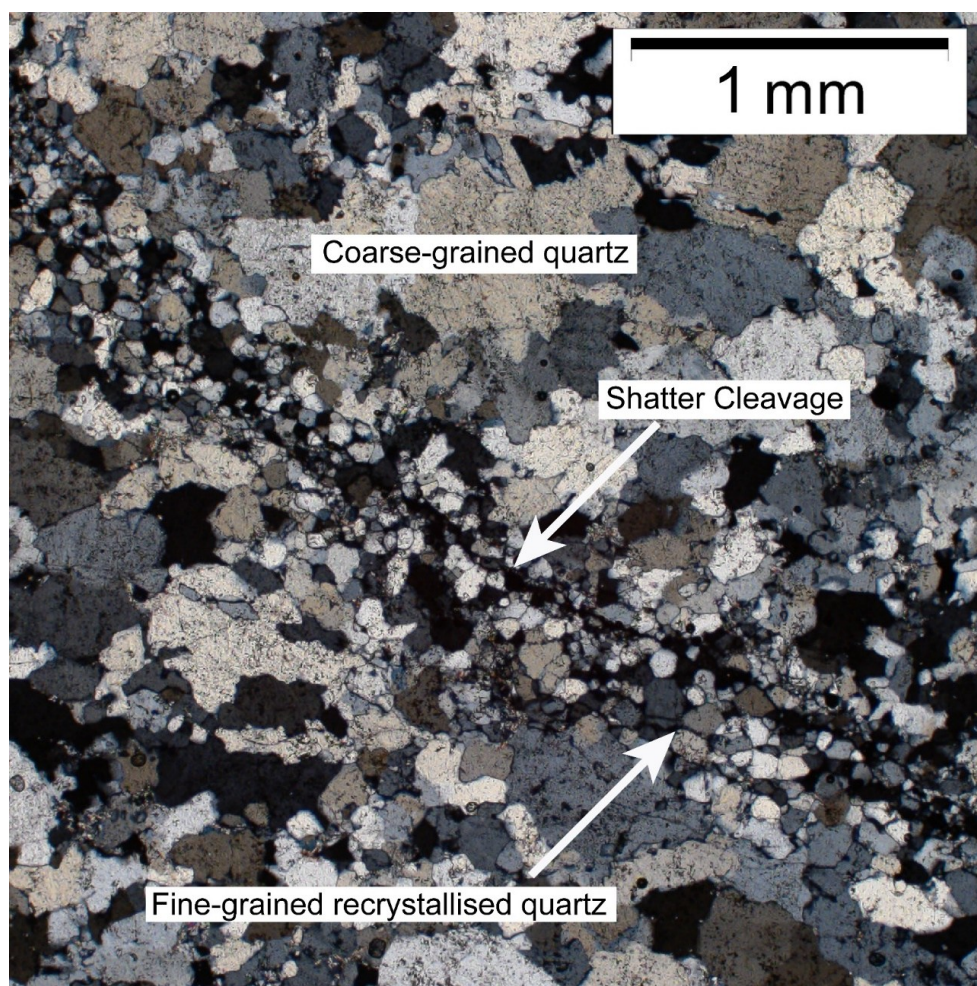


Figure 0.13: Microphotograph showing the shatter cleavage defined by narrow $\sim 500\ \mu\text{m}$ planar zones of fine-grained ($125 - 250\ \mu\text{m}$) recrystallised quartz in a medium-grained ($250 - 500\ \mu\text{m}$) granoblastic quartzite. Sample SM-2h location UTM 35J 0526660 E 7019144 S.

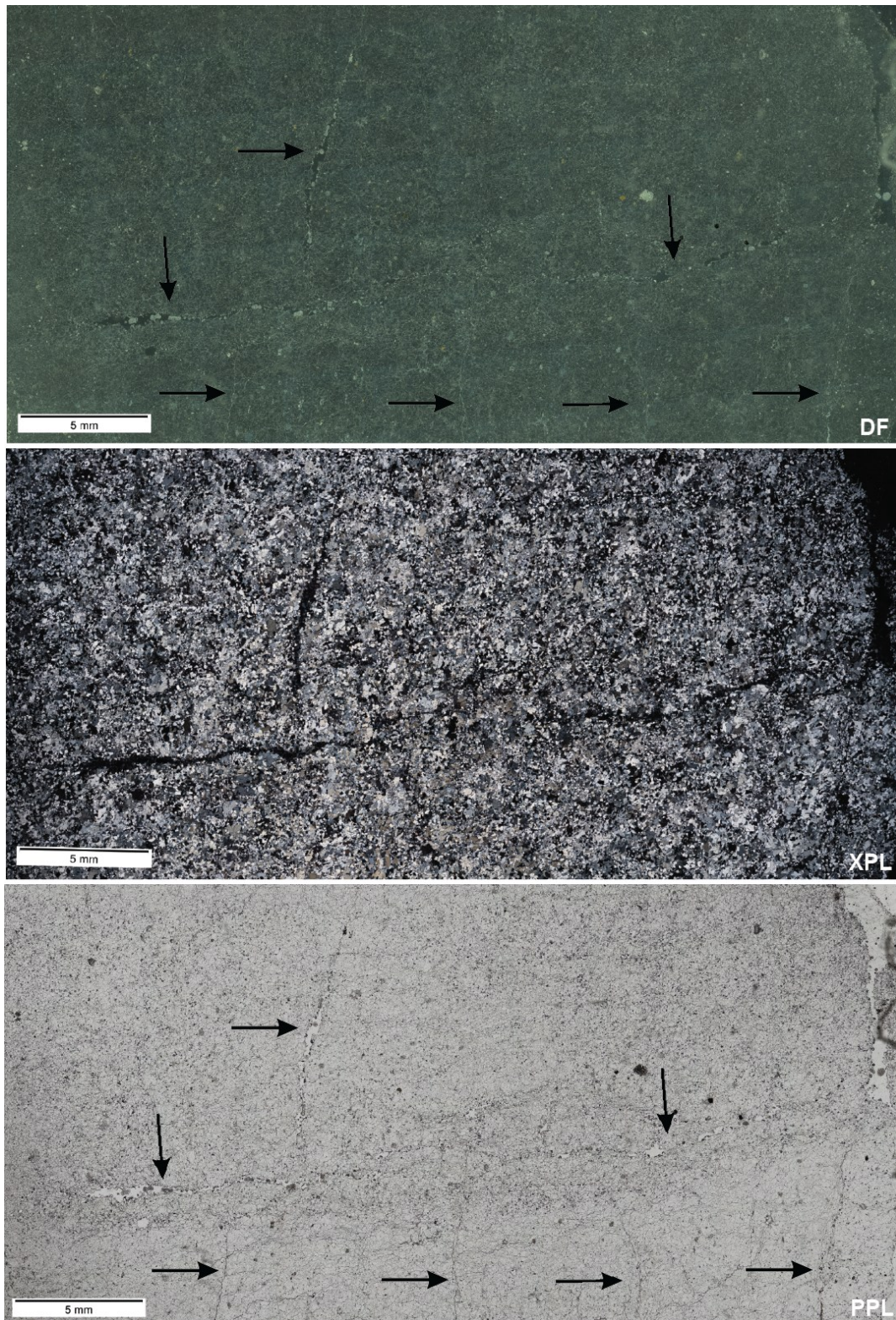


Figure 0.14: Thin section SMV-1 showing orthogonal shatter cleavages. CLA (running top-to-bottom) has an approximate spacing of 1 cm in a highly recrystallised fine-grained quartzite. The shatter cleavages in the sample are marked by fractures. The black areas in (B) and the corresponding white areas in (C) represent plucked areas along the fractures. Sample location UTM 35J 0526660 E 7019144 S.

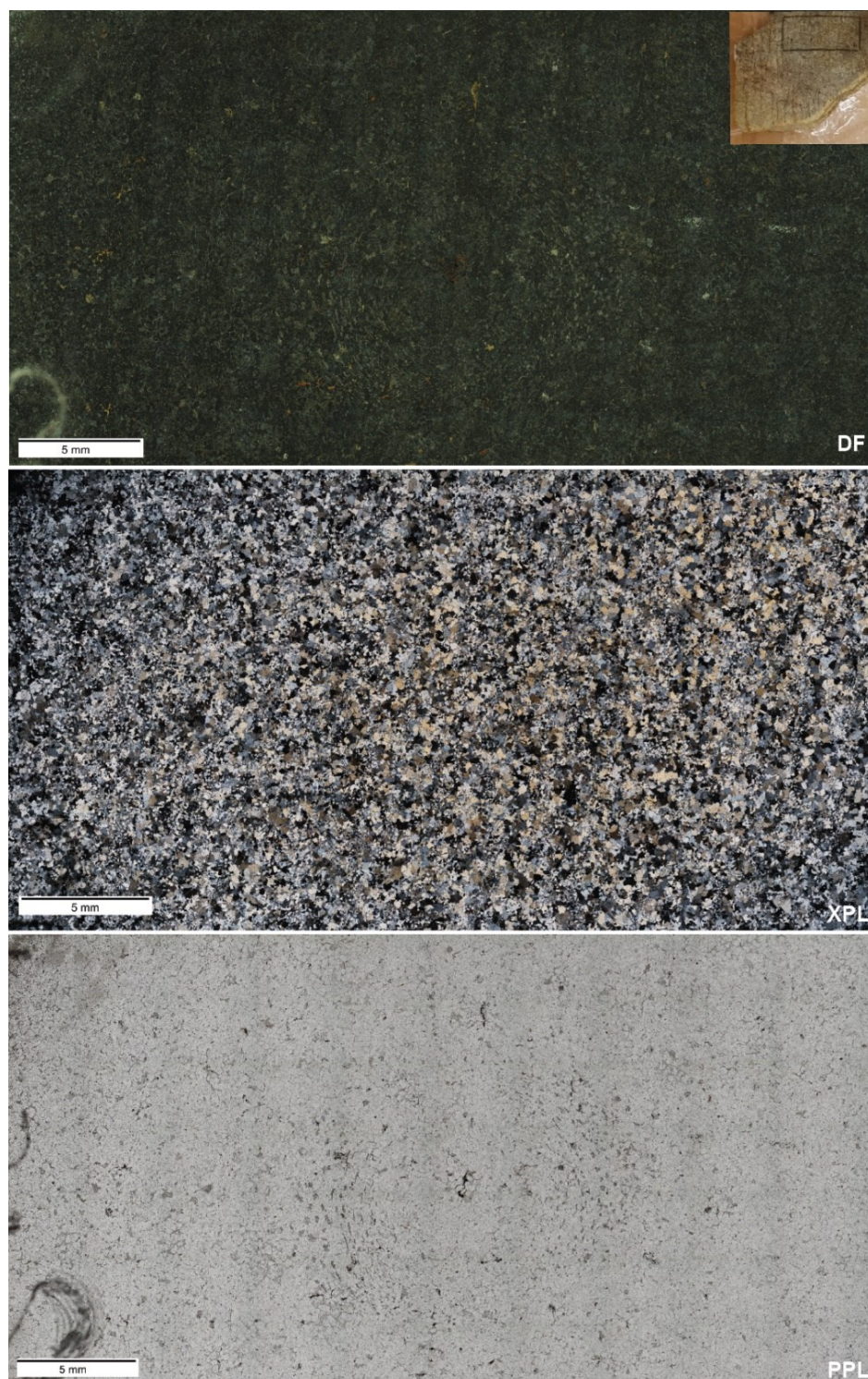


Figure 0.15: Thin section RAS-1 showing orthogonal shatter cleavages presented by fine-grained (125 – 250 μm) recrystallised quartz in $\sim 500 \mu\text{m}$ bands in a medium-to-coarse-grained quartzite (0.25mm - 1 mm). The finer-grained bands are evident by their lower birefringence. The inset in the top right corner shows the position from where the thin section was taken relative to the shatter cleavage orientations. There is no sample location available as it was a piece of float. This thin section is comparable to Figure 0.9C. The cleavage is best seen in DF and PPL by low reflectance/high absorption/darker areas. CLA (running top to bottom) is more intensely developed / closely spaced than CLB, running right to left. The sample is highly recrystallised, and a fracture does not mark the cleavage.

5.2.4 Orientation Analysis

Geometric constraints and cross-cutting relationships suggest that, like the shatter cones, the horsetail fractures and shatter cleavages formed early in the impact event and, thus, backrotation is necessary to investigate the stress and strain orientations prevailing during their formation. As laid out in [Chapters 2](#) and [4](#), the rocks containing these fractures have experienced polyphase deformation linked primarily to the modification stage of the Vredefort impact. The strong continuity of the 2.96-2.1 Ga supracrustal sequence in the exposed collar and minor thickness variation along strike supports the WRG having been horizontal at the time of impact. Study Areas 1 and 2 have the added complication of late-stage fold rotations that have differentially rotated strata and provide rare examples of subhorizontal bedding orientations. Contrary to earlier studies (e.g. [Wieland et al., 2006](#)), the bedding at the Schoemansdrift locality was also determined to be right-way-up and, therefore, anomalous for the inner collar.

Nauwpoort

The Nauwpoort locality preserves two phases of folding. Three folds are interpreted as belonging to the F1 phase of folding (hinges labelled 1 - 3 in Figure 0.5). They are concentric folds that show similar sizes (wavelength of ≤ 400 m and amplitude of ~ 25 m) and tightness (gentle interlimb angles). Hinges plunge steeply (folds 1 and 2) to gently (fold 3) SSE and axial planes are upright to steeply inclined and strike SSE. Folds 1 and 2 form a fold pair on the overturned northeastern limb of the large F2 anticline, whereas fold 3 is found on the shallowly-dipping, right-way-up, intermediate limb of the F2 fold pair (Figure 0.5, Figure 0.6). The F2 phase of folding is represented by fold hinges 4 and 5 that form an anticline-syncline pair (Figure 0.5, Figure 0.6). These F2 folds are open and have an estimated wavelength of ~ 800 m and amplitude of ~ 100 m. They have gently plunging hinges and moderately inclined axial planes that strike broadly tangentially to the dome and dip centripetally. The folds are asymmetric and noncylindrical.

The first horse tail fracture (CH) is located in domain C, whilst the second (EH) is located in domain E (Figure 0.6). The domains in the study area undergo a sequential rotation to a single orientation before being rotated back to the horizontal. The chosen orientation of domain J is chosen based on its strike being the closest to tangential as well as its central location in the study area.

- 1) Domain A is rotated to B resulting in domain AB
- 2) Domain AB is rotated to C resulting in domain ABC.
- 3) Domain N is rotated to K resulting in domain NK
- 4) Domain NK is rotated to J resulting in domain NKJ
- 5) Domain ABC is rotated to NKJ resulting in domain ABCNKJ
- 6) Domains DE and FG are unfolded about an inclined axis
- 7) Domain M to L resulting in domain LM
- 8) Domain LM to O resulting in domain LMO
- 9) Domain LMO to DEFG resulting in domain LMODEFG
- 10) Domain I to H resulting in domain HI
- 11) Domain HI to DEFG resulting in domain HIDEFG
- 12) Domain DEFGHILMO to ABCJKN resulting in domain ABCDEFGHIJKLMNO
- 13) Domain ABCDEFGHIJKLMNO rotated clockwise to the horizontal

These rotations can be found in the Appendix.

Based on the location of the horsetail fractures CH undergoes rotations 5, and 13. Horsetail fracture EH undergoes rotation 6,12 and 13 (Figure 0.16). The rotation reveals pre-folding azimuths of the main fractures of 098° and 103° , with an average of 101° , verifying that a consistent stress orientation was experienced across the Nauwpoort locality at the time of impact (Figure 0.16). The primary horsetail fractures have a similar strike as the strike of one of the shatter cleavages (CLA - $092^{\circ}/79^{\circ}/S$) and tend to splay off from this orientation (Figure 0.7; and Figure 0.20).

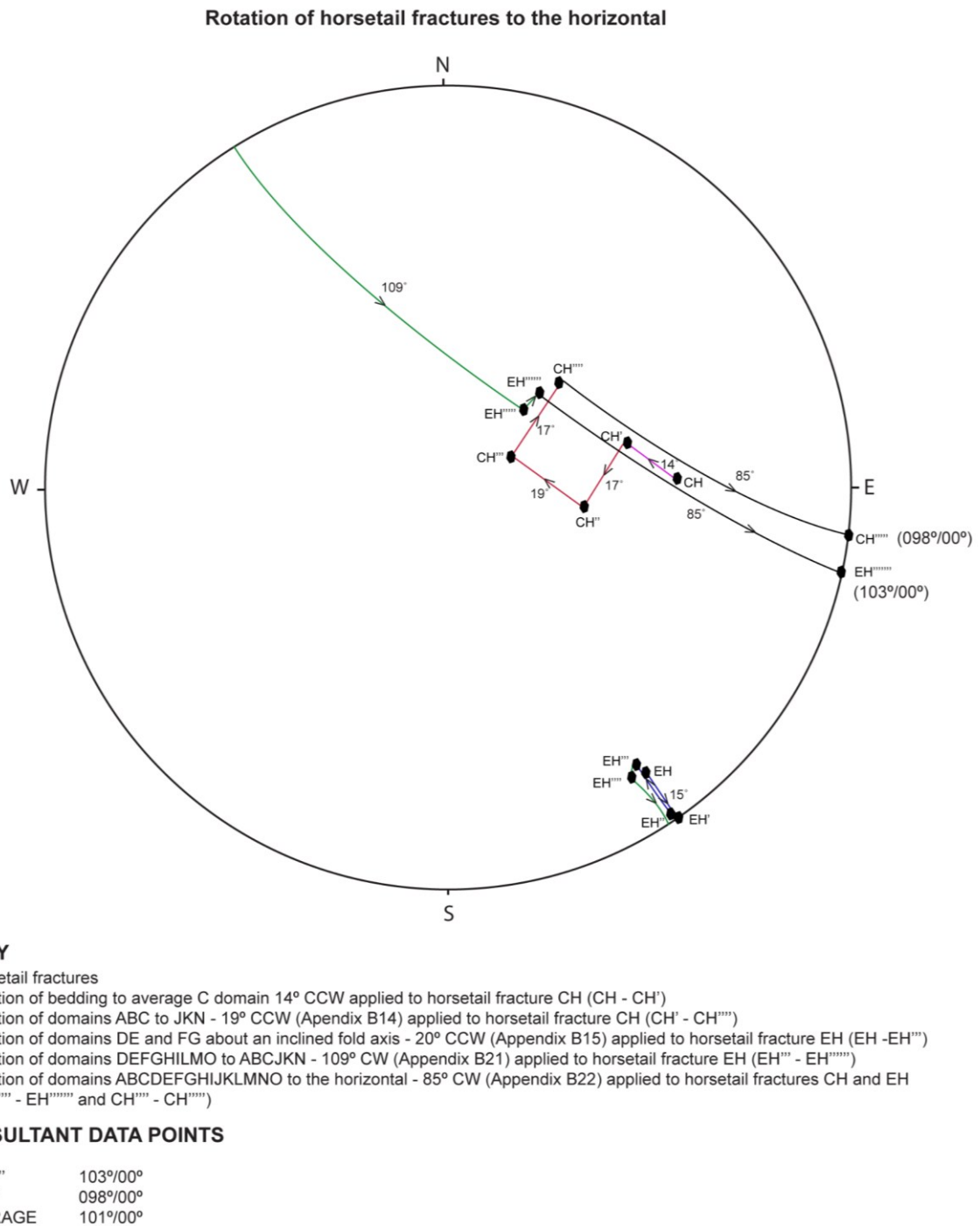


Figure 0.16: Equal area stereonet showing the rotations performed on the horsetail fractures CH and EH to rotate their host bedding to a pre-impact horizontal orientation. The entire set of rotations performed on the rocks at the Nauwpoort locality can be found in [Appendix B: Figures B10 - B22](#). The resultant horsetail fracture orientation at the time of impact is 101°/00° as depicted in the summary stereonet in Figure 0.20.

The shatter cleavages were mapped in detail at Nauwpoort and are represented as poles and planes on equal area stereonet (CLA: Figure 0.17 and CLB: Figure 0.18). The cleavage measurements were subdivided into 15 discrete domains at the Nauwpoort locality (A - O) based on the domains established by the bedding measurements in [Chapter 4.3.2](#) ([Appendix B: Table B1](#)). After that, each domain's measurements were plotted into equal area stereonet and filtered to remove outliers (Figure 0.17 and Figure 0.18). Measurements showing rotation along faults or in fold hinge zones were removed in the first iteration of filtration (blue). The second iteration of filtration (orange) was done visually on the stereonet to provide the least standard deviation and the tightest cluster of data points to obtain a robust average value for each domain (red). Reasons for the spread in the data may include but are not limited to, irregular bedding surface, rotation or diffraction in the shatter cleavages (Figure 0.12), measurement error, and/or rotated blocks caused by small-scale faults or mass wasting. Considering an error of $\pm 5^\circ$ strike and $\pm 5^\circ$ dip for cleavage measurements, the desired standard deviation in the cleavage average was 10° strike and 10° dip. Based on the data in Table 5.1 and Table 0.2 the average orientations of CLA and CLB on the northeastern subvertical fold limbs are $293^\circ/54^\circ/\text{NNE}$ and $180^\circ/56^\circ/\text{WNW}$, respectively, whereas on the shallow-dipping intermediate limb of the F2 fold, they are $331^\circ/73^\circ/\text{ENE}$ and $232^\circ/64^\circ/\text{NNW}$. [Appendix C](#) displays these results visually on equal area stereonet ([Appendix C: Figures C6 - C11](#)). The summarised data is displayed as rotated cleavage symbols on a Google Earth Pro[®] map image using ArcGIS 10.3.1[®] in Figure 0.19.

The cleavage data is then taken through the same back rotation as described above for the horstail fractures (Individual rotations can be found in the Appendix). The cleavage orientations for each domain were then weighted based on the number of measurements taken in order to obtain a single orientation for CLA $092^\circ/79^\circ/\text{S}$ and CLB $003^\circ/89^\circ/\text{E}$ (Figure 0.20; Table B2 in the appendix). The cleavage bisector of interest is $138^\circ/83^\circ/\text{SW}$. The horsetail fractures lie 11° off of CLA. The bearing to the centre of the dome is 111° which is approximately the same the CLA and Horsetail fracture orientation.

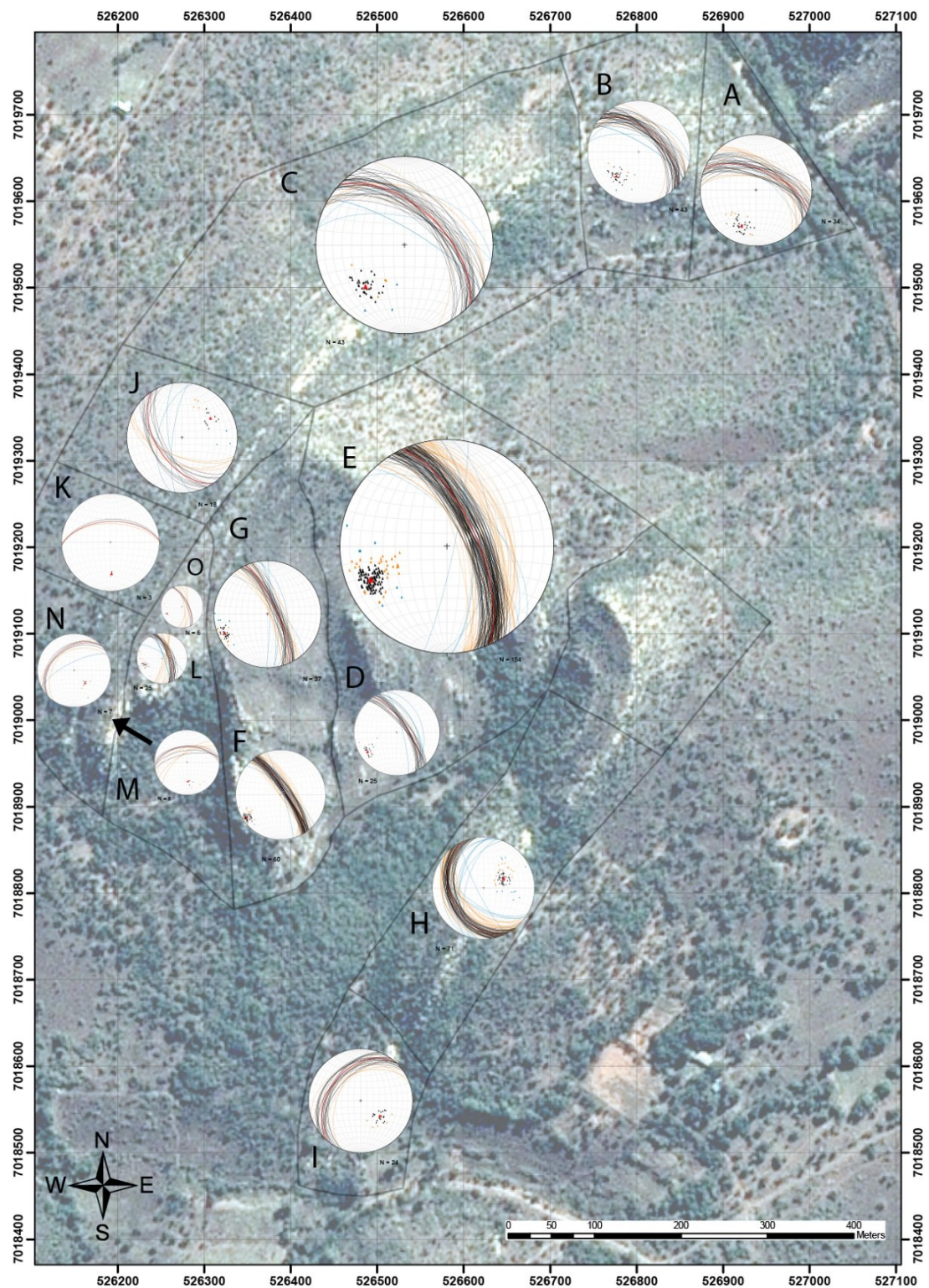


Figure 0.17: Google Earth Pro[®] image showing CLA in the study area divided into 15 discrete domains (A-O). Blue measurements were removed in the first iteration of filtration as they recorded bedding rotation along faults or in hinge zones. The second iteration of filtration (orange) was done visually on the stereonet to provide the least standard deviation and the tightest cluster of data points to obtain a robust average value for each domain (red). Stereonet size is not a reflection of the number of measurements taken. For the number of measurements taken in each domain, see Table 5.1.(Imagery date 07/10/2019).



Figure 0.18: Google Earth Pro[®] image showing CLB in the study area divided into 15 discrete domains (A-O). Blue measurements were removed in the first iteration of filtration as they show bedding rotation along faults or in hinge zones. The second iteration of filtration (orange) was done visually on the stereonet to provide the least standard deviation and the tightest cluster of data points to obtain a robust average value for each domain (red). Stereonet size is not a reflection of the number of measurements taken. For the number of measurements taken in each domain, see Table 0.2. (Imagery date 07/10/2019).

Table 0.1: CLA results for each domain at the Nauwpoort locality, showing the number of measurements taken (N), standard deviation (STD), dip direction (DD) and the resultant values for each successive iteration of filtration.

		Average CLA orientation before filtration					Average CLA orientation after first filtration					Average CLA orientation after second filtration											
Domain	Limb	Strike	Dip	DD	STD	N	Strike	Dip	DD	STD	N	Strike	Dip	DD	STD	N							
A	Northeastern Limb	295	60	NNE	12	8	34	294	60	NNE	12	8	33	292	60	NNE	9	8	26				
B	Northeastern Limb	311	55	NNE	12	9	43	313	54	NNE	10	7	39	313	55	NNE	9	7	35				
C	Northeastern Limb	276	51	NNE	11	10	77	276	50	NNE	8	9	66	276	48	NNE	6	7	57				
D	Middle Eastern Limb	329	67	ENE	8	9	25	327	70	ENE	6	4	20	326	71	ENE	6	4	19				
E	Middle Eastern Limb	337	64	ENE	8	9	154	337	64	ENE	7	9	147	336	66	ENE	5	6	113				
F	Middle Western Limb	326	77	ENE	5	7	60	326	77	ENE	5	7	58	326	78	ENE	4	5	46				
G	Middle Western Limb	337	76	ENE	9	6	37	335	76	ENE	6	6	32	336	75	ENE	5	4	23				
H	Southern Limb	155	39	WSW	18	11	71	152	37	WSW	12	9	60	156	35	WSW	10	5	41				
I	Southern Limb	222	40	WNW	14	8	24	219	38	WNW	11	7	19										
J	Northwestern Limb	145	55	WSW	24	10	18	137	53	WSW	17	8	15	146	52	WSW	9	8	11				
K	Northwestern Limb	265	59	NNW	6	9	5	269	54	NNW	2	6	3	269	54	NNW	2	6	2				
L	Northwestern Limb	344	68	ENE	16	10	25	347	66	ENE	14	9	21	339	63	ENE	6	7	15				
M	Northwestern Limb	256	56	NNW	15	8	8	266	52	NNW	4	5	5										
N	Northwestern Limb	232	44	NNW	11	13	7	232	44	NNW	11	13	4										
O	Northwestern Limb	347	70	ENE	15	7	6											229	37	NNW	6	4	4
																		335	67	ENE	3	4	3

Table 0.2: CLB results for each domain on Nauwpoort locality, showing the number of measurements taken (N), standard deviation (STD), dip direction (DD) and the resultant values for each successive iteration of filtration.

		Average CLB orientation before filtration					Average CLB orientation after first filtration					Average CLB orientation after second filtration							
Domain	Limb	Strike	Dip	DD	STD	N	Strike	Dip	DD	STD	N	Strike	Dip	DD	STD	N			
A	Northeastern Limb	178	51	WNW	10	5	14					182	49	WNW	7	3	10		
B	Northeastern Limb	204	61	WNW	11	6	12					204	61	WNW	11	6	12		
C	Northeastern Limb	154	63	WSW	13	9	44	152	62	WSW	11	9	43	153	60	WSW	7	6	29
D	Middle Eastern Limb	227	63	NW	11	10	10					225	59	NW	11	6	8		
E	Middle Eastern Limb	235	67	NNW	10	8	103	234	66	NNW	7	6	84	235	66	NNW	6	6	73
F	Middle Western Limb	231	68	NNW	10	7	44	230	67	NNW	7	5	36	228	68	NNW	5	2	25
G	Middle Western Limb	241	67	NNW	14	10	18	243	65	NNW	11	8	17	243	63	NNW	11	6	15
H	Southern Limb	270	55	N	19	16	38					270	49	N	12	14	22		
I	Southern Limb	093	58	SSW	9	8	12					92	60	SSW	5	5	9		
J	Northwestern Limb	364	56	NNW	8	7	9					262	54	NNW	3	7	6		
K	Northwestern Limb	142	44	WSW	8	12	12					137	48	WSW	4	4	4		
L	Northwestern Limb	063	71	SSE	16	9	16	065	71	SSE	13	8	15	060	66	SSE	8	5	8
M	Northwestern Limb	148	50	WSW	18	11	10	143	49	WSW	9	11	9	137	48	WSW	4	8	5
N	Northwestern Limb	121	58	SSW	22	13	31	123	54	SSW	18	10	23	123	53	SSW	8	10	12
O	Northwestern Limb	251	67	NNW	12	10	5					247	69	NNW	1	6	3		

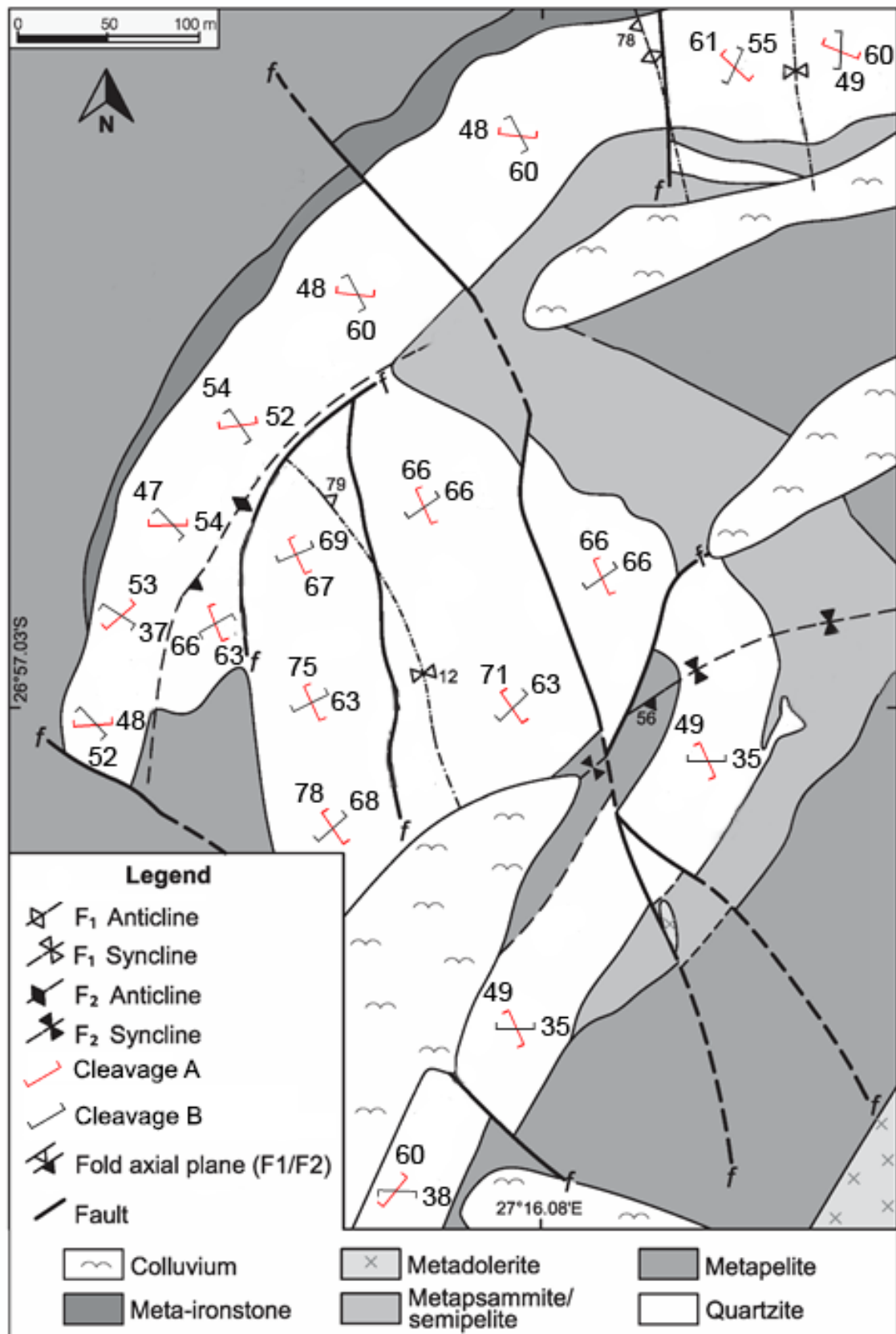


Figure 0.19: Geological map of Nauwpoort locality showing the summarised shatter cleavage data.

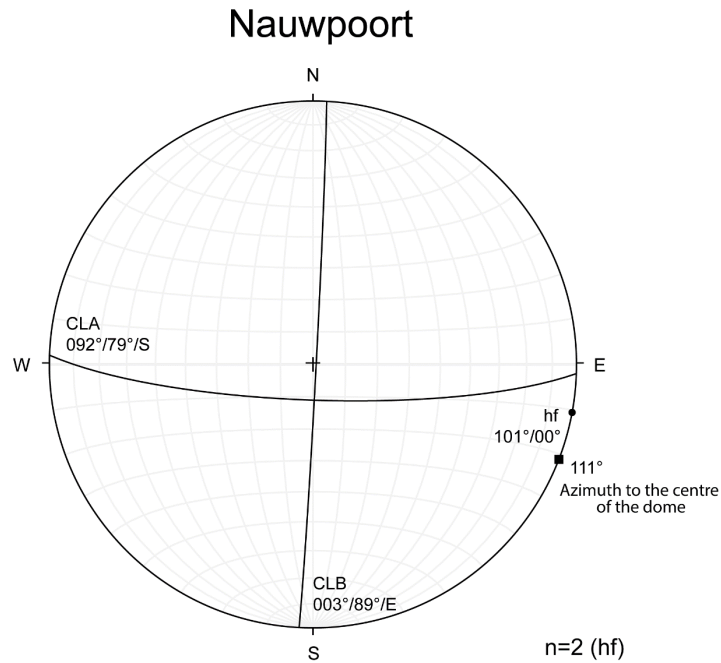


Figure 0.20: Equal area stereonet shows the results of backrotations to the horizontal of various fractures within the Study Area 1. Horsetail fractures long axis (hf) at the Nauwpoort locality (26°56'59.81" S 27°16'1.44" E) as well as the shatter cleavages (CLA and CLB). The long axis of the horsetail fractures lies on or near the CLA shatter cleavage. The bearing to the centre of the dome is 111°.

Geluksoord

The cleavage data from the Geluksoord locality are displayed in the geological map in Figure 0.18, and the details of the measurements are displayed in Table 0.3. The cleavage measurements were subdivided into nine discrete domains on the Geluksoord locality (A - I) based on the domains established by the bedding measurements in [Chapter 4.4.3](#) ([Appendix B: Table B3](#)). The data went through successive rotation to the orientation 300°/46°/NE which was selected as it best represents the average bedding in the area if the unit had not been folded. There after it was rotated to the horizontal.

- 1) Domain I is rotated to C resulting in domain CI
- 2) Domain CI is rotated to B resulting in domain CIB
- 3) Domain CIB is rotated to A resulting in domain CIBA
- 4) Domain CIBA is rotated to 300°/46°/NE
- 5) Domain F is rotated to G resulting in domain FG
- 6) Domain FG to H resulting in domain FGH
- 7) Domain FGH to D resulting in domain FGHD

- 8) Domain FDHG to E resulting in domain FDHGE
- 9) Domain FDHGE to 300°/46°/NE
- 10) All data rotated to the horizontal.

Once the rotations were performed the data was weighted based on the number of measurements in order to obtain a single orientation for each cleavage. CLA had an average orientation of 332°/50°/E and CLB 208°/52°/W (Figure 0.22). The step-by-step rotations applied to the data can be found in Appendix C. The orientation to the centre of the dome from Geluksoord is 210° which is approximately the orientation of CLB (Figure 0.22). The bisector of the cleavages is 043°/82°/SE.

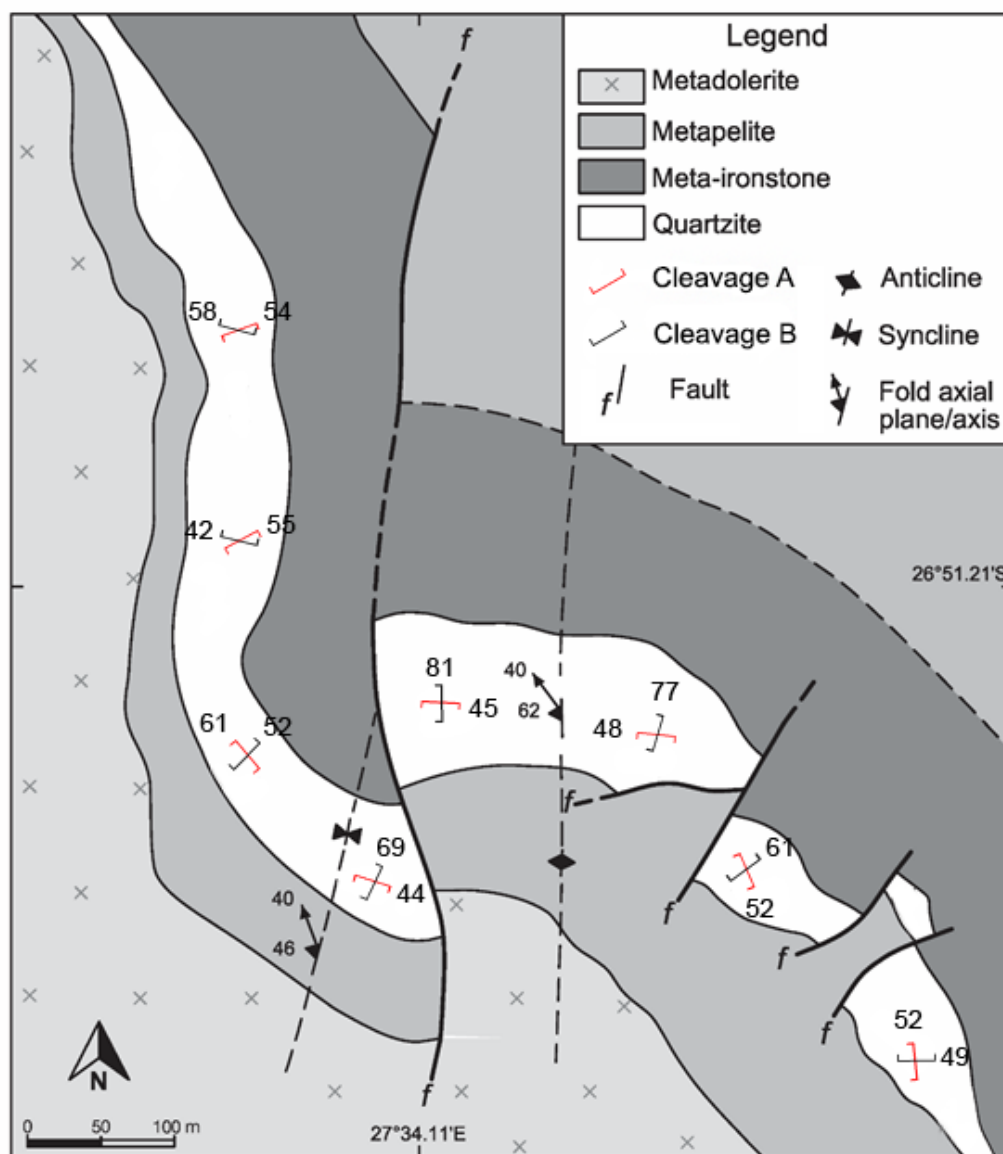


Figure 0.21: Geological map of Geluksoord locality showing the summarised shatter cleavage data.

Table 0.3: CLA and CLB results for each domain on Geluksoord locality (Study Area 2), showing the number of measurements taken (N) and the dip direction (DD).

Domain	CLA				CLB			
	Strike	Dip	DD	N	Strike	Dip	DD	N
A	196	77	W	6	96	48	S	6
B	157	52	W	8	233	61	NW	7
C	269	52	N	6	174	49	W	6
D	205	69	NW	5	108	44	S	5
E	93	45	S	2	179	81	W	2
F	70	54	S	6	285	58	N	6
G	61	55	SE	3	284	42	N	3
H	140	61	SW	7	224	52	NW	7
I	269	53	N	2	126	50	SW	2

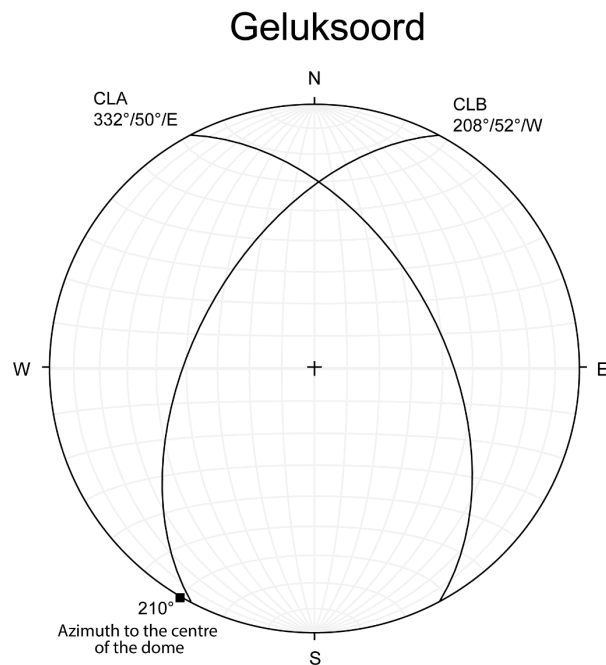


Figure 0.22: Equal area stereonet shows the results of backrotations to the horizontal of shatter cleavages within Geluksoord Study Area 2. The bearing to the centre of the dome is 210°.

Schoemansdrift

Shatter cleavage and the average shatter cone apex at Schoemansdrift (26°58'8.58" S 27°12'32.13" E) (Figure 0.2 and Figure 0.8A) were rotated by 50° counterclockwise about the strike to account for the right way up bedding required to restore bedding to horizontal (Figure 0.23 and Figure 0.24). This results in a shatter cleavage (CLA) oriented at 312°/80°/NE, (CLB) 020°/90° vertical, shatter cone apex of 170°/14°, master cone axis 286°/18° and cleavage bisector 346°/84°/E. These results show that the shatter cones occur approximately on the

shatter cleavage bisector. The master cone axis showed no clear relationship with the shatter cleavage (Figure 0.23).

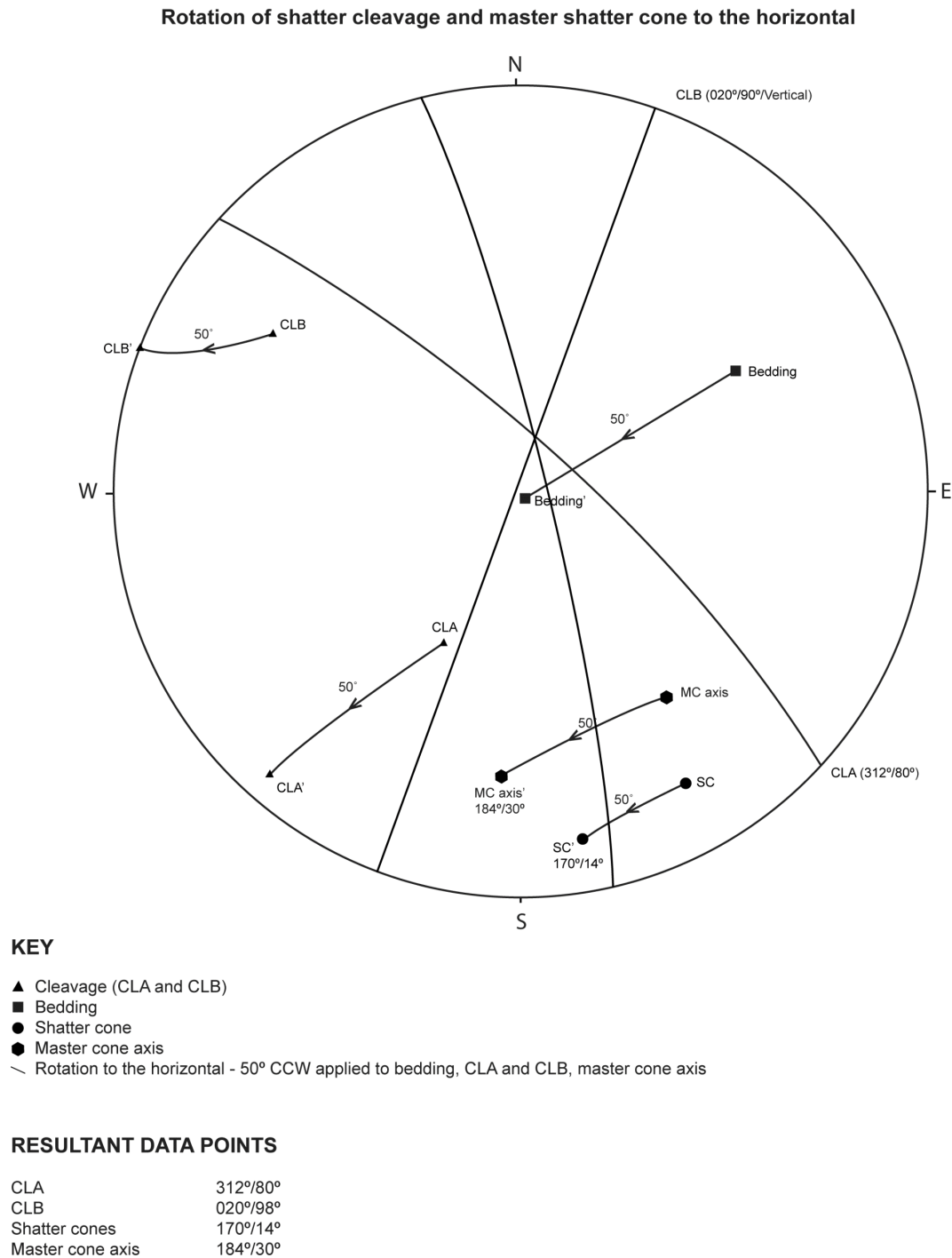


Figure 0.23: Equal area stereonet showing the rotation of shatter cleavage, the master cone and the bedding to the horizontal at the Schoemansdrift locality. The raw data and rotations are in [Appendix C](#) and S. A portion of the master cone was compiled from the shatter cone apex measurements taken at the outcrop. Then, the apex of that cone was located. The Master cone axis and the shatter cone axis differ significantly. Some authors suggest that the bedding at Schoemansdrift is overturned; the corresponding rotation can be found in [Appendix C](#): Figure [C25](#). However, I proved it right-way up, hence this counterclockwise rotation by 50°.

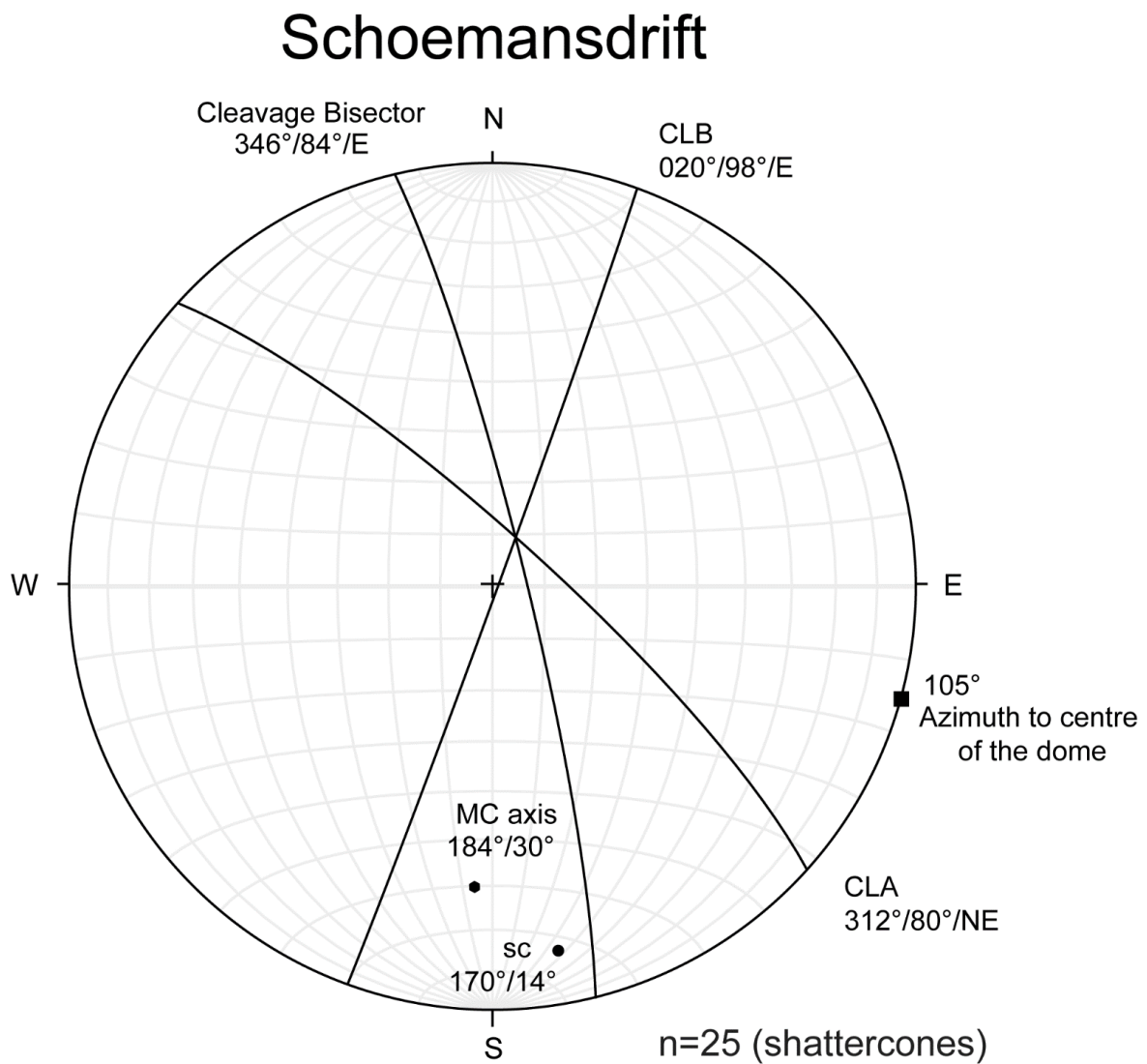


Figure 0.24: Equal area stereonet showing the results of backrotations to the horizontal of various fractures within the Schoemansdrift locality (26°58'8.58" S 27°12'32.13" E; Figure 0.2). The master shatter cone apex orientation (sc), as well as the shatter cleavages (CLA and CLB), are shown. Azimuth to the centre of the dome is 105°.

5.2.4 Testing the Albat and Mayer (1989) Hypothesis

Table 0.4: Shatter cleavage data used in the construction of Figure 0.25 and Figure 0.26. Location numbers come from [Albat and Mayer \(1989\)](#). Nauwpoort and Geluksoord are new locations from this study.

	Location	Strike deviation from tangential	Bearing to centre	CLA	CLB	Cleavage bisector	Deviation of CLA/CLB from radial	Deviation of cleavage bisector from radial
Rotation to the horizontal given the ideal tangential strike	3		60	035/65/E	308/80/N	084/79/S	25	24
	4		66	015/64/E	282/69/N	060/87/S	36	6
	6		62	051/78/S	336/78/E	014/75/E	11	48
	13		84	064/60/S	332/72/E	111/82/S	20	27
	18		124	274/77/S	356/79/E	125/88/S	30	1
	19		138	210/84/W	317/74/E	262/72/N	1	56
	21		154	112/70/S	020/57/E	332/81/E	42	2
	22		156	282/84/N	197/88/W	330/87/E	41	6
	37		218	224/73/W	131/79/S	177/70/W	6	41
	38		234	248/78/N	174/78/W	211/75/W	14	23
	39		236	218/68/W	134/74/S	267/86/N	18	31
	41		314	204/86/W	296/72/N	341/80/E	18	27
	Geluksoord		210	332/50/E	208/52/W	000/89/E	46	30
	Nauwpoort		11	092/79/S	003/89/E	138/83/SW	19	27
	Average						23	25
Rotation to the horizontal given the actual strike	3	21	60	027/66/E	300/79/NE	076/81/S	33	16
	4	1	66	023/64/E	289/68/N	067/87/S	43	1
	6	23	62	308/76/N	022/80/E	254/87/N	40	12
	13	2	84	291/58/N	020/72/E	242/80/N	27	22
	18	9	124	094/77/S	16/80/E	145/88/W	30	21
	19	26	138	359/74/E	250/84/N	303/72/E	17	15
	21	8	154	104/71/S	013/56/E	325/80/E	40	9
	22	2	156	282/84/N	197/88/W	330/87/E	41	6
	37	31	218	177/76/W	084/79/S	221/88/N	41	3
	38	65	234	135/79/S	069/82/S	101/79/S	15	47
	39	37	236	150/59/W	068/74/S	106/60/S	12	50
	41	27	314	244/90	153/86/W	108/87/S	19	26
	Geluksoord	9	210	355/73/E	269/86/N	043/82/Se	35	13
	Nauwpoort	17	11	340/84/E	078/78/S	119/86/S	33	8
	Average	20					30	18

Previous studies have suggested a relationship between the shatter cones, MSJS, and the shatter cleavages (see [Chapter 5.1](#)) shatter cones form off the shatter cleavages, and the master cone axis is perpendicular to the intersection of the cleavages ([Albat and Mayer, 1989](#)). The bisector of the two cleavages and the cleavage strike from across the dome were plotted to note any systematic changes that may be present prior to the overturning of the strata.

The data from [Albat and Mayer \(1989\)](#), as seen in Table 0.4, were used to perform a dome-wide orientation analysis. Cleavage bisectors (Figure 0.25A and Figure 0.26A) and cleavage strike (Figure 0.25B and Figure 0.26B) were plotted as rays from their respective geographic locations around the dome (represented by a circle) after the bedding had been rotated to the pre-impact horizontal around the respective bedding strikes. Numbers signify the corresponding field location of [Albat and Mayer \(1989\)](#), whereas Geluksoord and Nauwpoort

indicate the locations of data collected in this study. The cleavage bisectors were plotted to assess their proposal that these originally pointed towards the source of the shock wave at the centre of the dome. The bisector that was closest to the radial orientation was selected. Thus in some cases it is the acute bisector and in others the obtuse bisector. The strike of the cleavages in relation to the centre of the dome was plotted to evaluate if either cleavage strike is radial or whether they are symmetric about a radial bisector. The cleavage whose strike most closely resembles the radial orientation, pointing to the dome's centre, was chosen.

Given that the cleavages are roughly perpendicular to one another, the inter-cleavage angles tend to be roughly equal, and thus, choosing the acute or obtuse angle does not prove a helpful determinate. In Figure 0.25, the rotation of the cleavage strike and bisector was made to the horizontal about the strike of the bedding. In Figure 0.26 the bedding strike was rotated to the ideal tangential orientation before the cleavage strike and bisector were rotated to the horizontal about the idealised bedding strike. This was done to account for any block faulting. The individual rotations for each locality of [Albat and Mayer \(1989\)](#) are provided in [Appendix C: Figures S1 – S22](#). The best results for comparing the cleavage to the centre of the dome are provided by the rotation of strike of the bedding to the ideal tangential orientation and then the horizontal using the bisector of the cleavage (Figure 0.26A). This result is in support of the theory laid out by [Albat and Mayer \(1989\)](#) that the bisector of the cleavages points to the centre of the dome where the shock wave originated as opposed to that set out within this chapter that one of the shatter cleavages point to the centre of the dome.

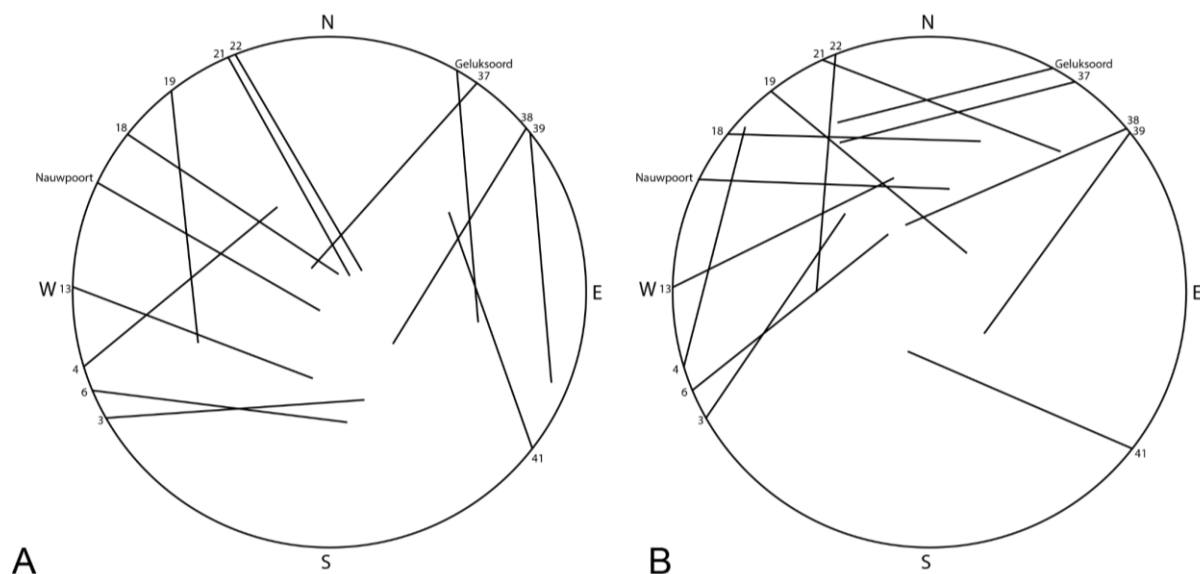


Figure 0.25: The strike lines of the cleavage bisectors (A) and cleavage strike (B) are plotted as a ray from their geographic location around the dome (represented by the circle) after the rotation of the strike to the horizontal. Numbers signify the corresponding field location of [Albat and Mayer \(1989\)](#), whereas Geluksoord and Nauwpoort indicate the locations of data collected in this study (see [Chapter 4](#)).

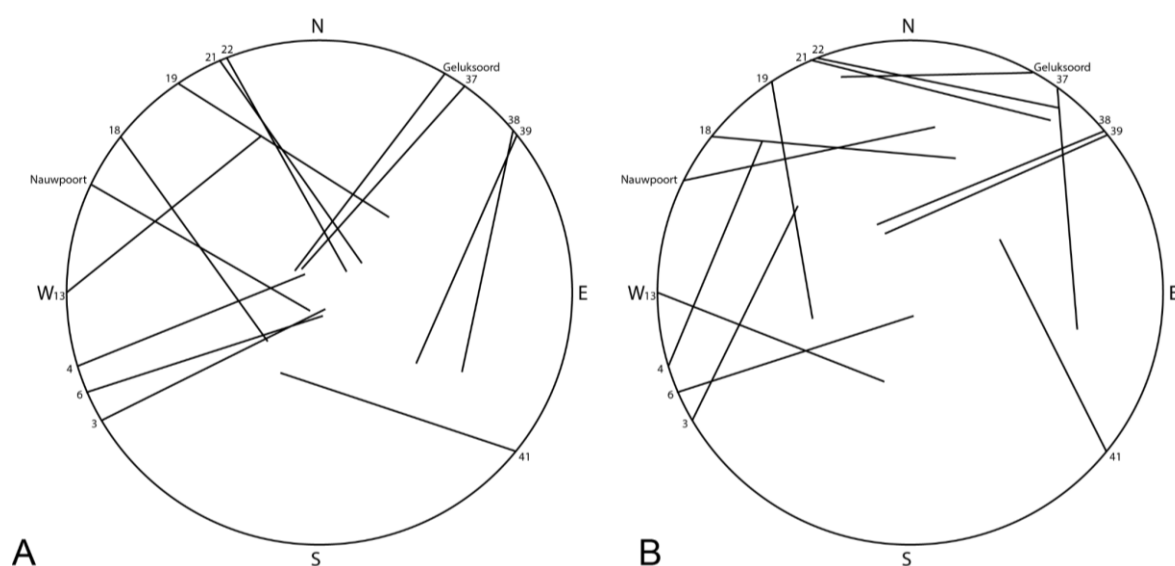


Figure 0.26: The strike lines of the cleavage bisectors (A) and cleavage (B) are plotted as a ray from their geographic location around the dome (represented by the circle) after the rotation of the bedding strike to the tangent before bedding is rotated to the horizontal. Numbers signify the corresponding field location of [Albat and Mayer \(1989\)](#), whereas Geluksoord and Nauwpoort indicate the locations of data collected in this study (see [Chapter 4](#)).

5.3 Discussion

5.3.1 Is shatter cleavage impact related and is it a shock feature?

The shatter cleavage is similar to that found in other impacts (Gosses Bluff impact structure, [Milton \(1977\)](#); Sudbury, [Nicolaysen and Reimold \(1999\)](#); Matt Wilson Crater, [Kenkmann et al. \(2014\)](#)). In the Vredefort Dome [Nicolaysen and Reimold \(1999\)](#) and [Albat and Mayer \(1989\)](#) linked them to shatter cones which are bona fide shock features. Their occurrence is localised to the dome and not observed elsewhere in the lower WRG. Rotation analysis indicates that the cleavage predates impact related folding. PDFs show no signs of being enhanced in the shatter cleavage or towards shatter cleavage planes. Modeling indicates post impact temperatures of ~480°C which would support annealing. The shatter cleavages are annealed indicating they formed pre-cooling. The high FSI and close spacing of the fractures suggests high differential stress such as would be expected in the early stages of impact such as the Contact and Compression Phase.

5.3.2. Shatter Cleavage Genesis

The shatter cleavages orthogonal relationship raises questions about genesis and timing. They are not conjugate shear fractures, and no displacement is noted macro- or microscopically. The fractures are annealed and alteration rules out any melt evidence that would support shear. There was no evidence of coesite-stishovite in the veins (e.g. [Spray and Boonsue \(2018\)](#)). The vein geometry does not fit stylolite - joints as they are orthogonal and require them to form by two different stress fields. The shatter cleavages have an FSI of 20 - 100, given a modest inter-cleavage spacing of 1 cm and using an average layer thickness based on the average 0.2 - 1 m bed thickness. This can increase to an FSI of 40 - 200 if one uses a fracture spacing of 5 mm. This is significantly higher than any recorded FSI, and thus, it can be expected that there would be local stress field inversion, as proposed by [Bai et al. \(2002\)](#). Whether stress field inversion due to fracture saturation applies to fracturing during impact tectonic is open to discussion as strain rates and differential stresses are significantly higher. At present, there are no other theories on how these shatter cleavages form. The narrowly spaced fractures, or shatter cleavages, are said to form in nonporous rocks at shock pressure of between 0 - 8 GPa, whereas shatter cones form between 1 and 20 GPa ([Stöffler, 1984](#); [Stöffler and Langenhorst, 1994](#); [French, 1998](#)).

In impact tectonics, although the stress field associated with the shock wave is always compressional ([Rae et al., 2021](#)); at the radial distance and depth of the WRG the maximum

stress rotates 50 degrees to accommodate the refraction wave. Thus, although it is unlikely that we will be dealing with a 90° stress field inversion within the shock wave, 3 seconds later the rarefaction wave causes the stress field to rotate by up to 50°. This can thus potentially be the source of CLB. This supports [Winkler et al. \(2018\)](#) who describe the formation of microstructures in terms of superimposing regimes created by hypervelocity impact. The first regime (I) is above the HEL (see [Chapter 2.2.1](#) for more details on HEL). Above the HEL, a rock behaves strengthless, like liquids, and will undergo a shock-metamorphic overprint. The second regime (II) occurs as the rarefaction wave encroaches on the shock wave, and pressures fall below the HEL, creating a concentric hoop stress that forms radial-oriented features. The third regime (III) occurs when the rarefaction wave creates radial tensile stress that produces concentrically oriented fractures. This can thus be described as an approximately orthogonal fracture set when observed on a local scale.

Based on the relationship with the branching fractures, the CLA cleavage is believed to be formed by the passage of the shock wave in regime II - Contact and Compression phase ([Winkler et al., 2018](#)). The rotation of the CLA strike is evidence of the refraction and interference of the shock waves with different layer boundaries, as seen in Figure 0.10D and Figure 0.12. The CLB is proposed to form by combining fracture saturation stress field inversion, as described by [Bai et al. \(2002\)](#) and the radial tensile stress in regime III, as described by [Winkler et al. \(2018\)](#) and [Rae et al. \(2021\)](#). The timing of CLB is fractionally later than CLA by no more than a few seconds.

The term shatter cleavage used here is perhaps misleading as it is formed by tensional stresses and, at the time of formation, best represented by a mode I extensional fracture, as opposed to compaction cleavage under compression. However, its genetic relationship to shatter cones has deemed its name suitable. The formation of this dense orthogonal shatter cleavage may be a viable mechanism for acoustic fluidisation as large volumes of rock are deconstructed into mm³ and cm³ blocks, enabling the movement between the blocks.

5.3.3. Does geometric analysis work?

In both Study Areas 1 and 2, restoration of bedding to the horizontal demonstrates that the two cleavages are the same on both fold limbs. Thus, the cleavages predates folding, which is shown in Chapter 4 to be impact-related. As seen in the outcrop, the local rotation in the cleavage strike appears to be due to diffraction within varying layers of the meta-sedimentary rocks however should sufficient data points be collected this does not pose a problem.

Schoemansdrift locality shows that the shatter cones forms approximately along the acute bisector of the two cleavages.

The geometric analysis to examine if the cleavage strike or bisector showed a radial/tangential orientations revealed that the shatter cleavage bisector can roughly locate the centre of the dome or source of impact. Shatter cleavage bisectors can accurately determine the centre of an impact if given a large enough and widespread data set. Thus, there is a geometric relationship between the shatter cleavage bisector, the centre of the dome, and the impact location. The inconsistencies can be explained by more complex 3D rotations either pre- or syn-impact, or by shock wave reflection/refraction as per [Wieland et al. \(2006\)](#). [Rae et al. \(2021\)](#) modelled that the shock stresses and strains are highly complex and not truly radial. Based on the data provided by [Rae et al. \(2021\)](#) (see [Chapter 1.3.3](#); Figure 0.5) in conjunction with the data obtained through numerical modelling ([Chapter 2](#)), given an impactor of 14 km diameter, initial radial distance for the field area of 26.3 km, and depth of 15.5 km the angle of the σ_1 principal stress axis at peak shock pressure would have been 10° steeper than true radial. The orientation of σ_1 during shock release would be 40° shallower than radial. The radial orientation for the given location is 30° . Thus, for peak shock pressure, the angle of σ_1 would be 40° outwards and for shock pressure release, it would be 10° towards the centre of the dome. Given these potential challenges the convergence of the rays are very good and the cleavages seems remarkably uniform over hundreds of metres. Based on the numerical modelling ([Chapter 2](#)). The shock wave orientation changes as a result of the rarefaction wave from the moment of peak pressure onwards ([Rae et al., 2021](#)). Thus when the peak pressure arrives at 3 s σ_1 is 40° outwards and then has 6 seconds to rotate by 50 degrees to be oriented 10° inwards for rocks in the lower WRG. Caution should thus be taken not to compare shock feature from different radial distances as this will introduce error. Thus, perhaps excluding location 18 from this analysis would have assisted in yielding better results as it is located within the CRG.

The accuracy of the result was improved by correcting fault block rotations by adding rotation to correct the bedding strike to the concentric orientation. Even given this added rotation, some data points still show significant deviations from the centre of the dome, which could be due to significant block rotations that are not accounted for accurately in 2D.

The shatter cleavage strike shows a more substantial deviation from the centre of the dome than the cleavage bisector. This is unexpected at the strike of the dominant cleavage, hypothesised based on its relationship to the shatter cones, as seen at Schoemansdrift in [Chapter 5.2.1](#), to point to the centre of the dome. Thus, given the overall dome-wide approach, this is invalid.

The bisector of the cleavage is more strongly affiliated with the orientation to the centre of the dome.

5.3.4. Horsetail fractures

The horsetail fractures seem to be geometrically and genetically linked to CLA, and therefore, early impact. The fractures connecting CLB within the branching horsetail fractures tend to be irregular and discontinuous, as seen in Figure 0.7. This may indicate that the horsetail fractures form before CLB. The increased width between fractures in the branches reduces the FSI, resulting in a locally less well-developed CLB. Horsetail fractures can be used to establish the sense of shock wave directionality in a given volume of rock, much like shatter cones. This, in conjunction with geographical locations, can also be used to determine if the volume of rock is overturned where sedimentary way-up indicators are absent. The occurrence of these fractures also indicates that the ratio of the fracture propagation velocity of the shatter cleavage to Rayleigh wave velocity was between $0.3 \geq 0.9$ VR in the Rhq3 quartzite on Nauwpoort was insufficient to form shatter cones. The fracture propagation velocity should decrease radially outwards from the point of impact; however, we find the shatter cones at the Schoemansdrift locality at a radial distance of 25.5 km. This can be explained by the CRG Rjo1 quartzite ([Bisschoff, 2000](#)) being stratigraphically higher than the Rhq3 quartzite ([Bisschoff, 2000](#)) of the WRG at the time of impact. It was thus closer to the point of impact before overturning. It was also nearer to the free surface at the time of impact. Rayleigh wave velocity also increases with depth, which results in a smaller fraction of $VR < 0.9$ and thus is less likely to result in shatter cone formation. In addition, the homogeneity of the host needs to be considered as branching fractures are proposed to require heterogeneities on which to nucleate.

5.4 Conclusions

The shatter cleavages resemble the “shatter cleavage” of the Gosses Bluff impact structure in Australia (Figure 0.6A) ([Milton, 1977](#); [Milton et al., 1996](#)) and does, therefore, suggest that similar fractures could be found in other impact structures. Shatter cones form approximately along the acute bisector of the two shatter cleavages. Horsetail fractures are geometrically and genetically linked to CLA.

Chapter 6

PAPER II: PSEUDOTACHYLYTE GENESIS LINKED TO FRICTION MELTING IN LARGE IMPACT-RELATED FAULT ZONES IN THE VREDEFORT DOME, SOUTH AFRICA

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My contribution to this paper as the first author involved conceptualising the Kopjeskraal faults system model based on the structural data I collected and geological mapping done in the field. I also did the necessary rotations to test the fault timing hypotheses. I modelled the Western Fault in 3D using Leapfrog Geo® 5.1. I created the geological maps of the study area and compiled most of the text and images in the paper with the assistance of my supervisor. The only modifications made from the intended journal manuscript to maintain consistency with the thesis are updating text section numbering, figure numbering, and the spelling of various words that have been changed from American English to UK English, and consolidation of the references into the main Reference List. The spelling of pseudotachylite has also been updated to pseudotachylyte to be consistent with the rest of the thesis. To maintain the integrity of the paper, some diagrams used in other chapters are duplicated, with new figure numbers.

Abstract

Impact cratering involves significant structural disruption and differential movements within the affected target rock volume that occur at high strain rates. Such conditions tend to favour strain localisation along faults; however, few studies have been able to rigorously constrain fault kinematics in the interior parts of large craters through field observation. A related question is the nature and timing of pseudotachylyte formation during a cratering event and whether the voluminous clast-bearing melt bodies found in several large impact structures can be unequivocally linked to faults. In this study, we constrain the spatial, geometric and kinematic relationships between pseudotachylyte and a pair of large-displacement faults within

the central uplift of the Vredefort impact structure that links pseudotachylyte genesis to slip-induced frictional sliding. Kinematic indicators obtained from the displacement of bedding and quartz veins, and outcrop and microstructural observations of structural fabrics, indicate that the two faults form a conjugate set with respective strike separations of approximately 300 m right-lateral and 500 m left-lateral. In their present geometry, this conjugate system is consistent with a shallow NNW-plunging (radial) σ_1 stress direction and shallow E-plunging (tangential) σ_2 orientation. However, iSALE-2D numerical modelling of the strain evolution of the rock volume in response to the impact suggests that the faults originated as normal dip-slip faults during the initial stages of the collapse of the transient crater walls (late part of Modification Stage 1 at approximately 65-69 seconds after impact) and that they were rotated to their present orientation during subsequent central uplift rise and collapse. This study provides a first constraint on the timing of melting to produce the large pseudotachylyte volumes observed in the Vredefort Dome, and ties their origin to frictional heating along large-slip faults.

6.1 Introduction

The term pseudotachylyte was first introduced by [Shand \(1916\)](#) in his description of cohesive, dark, aphanitic, clast-bearing, flinty veins, dykes and irregular masses in the granitoid gneisses of the Vredefort Dome. Although [Shand \(1916\)](#) favoured an origin for the dykes by some form of gas fluxing, the term pseudotachylyte has come to be almost universally associated with friction melts found in tectonic fault zones (Bates and Jackson, 1987). This has created a problem because studies in the type locality have repeatedly observed that the pseudotachylyte in the Vredefort Dome lacks any consistent association with faults of suitable size to account for their volume (see review in [Reimold and Gibson, 2005](#); [Gibson and Reimold, 2008](#)); moreover, the exceptional volumes of pseudotachylyte bodies in the dome – both individually and cumulatively – dwarf those found in tectonic fault settings. Following confirmation of an impact origin for the dome and identification of similar voluminous pseudotachylyte bodies in the Sudbury Structure, which was also formed by impact, several alternative models linking pseudotachylyte genesis to impact processes have been proposed. These are primarily based on studies of these two impact structures and the Manicouagan and Rochechouart craters that contain smaller pseudotachylyte bodies (Lambert, [1977, 1981](#); [Spray and Biren, 2021](#)). Among these mechanisms are shock melting with or without a frictional heating component ([Martini, 1978, 1991](#); [Fiske et al., 1995](#); [Spray and Boonsue, 2018](#)), flash replacement melting by explosive transfer of thermal shock energy (Dressler and Reimold, 2004), decompression

melting ([Reimold and Gibson, 2005](#)), a combination of shock and decompression melting ([Mohr-Westheide and Reimold, 2011](#)), drainage from a superheated impact melt pool into tensional fractures ([Lieber et al., 2009](#); [Riller et al., 2010](#)), frictional sliding along large-slip terrace faults related to the collapse of the transient crater (Spray and Thompson 1995; [Spray, 1997](#)) and oscillatory impact-induced seismic shaking following passage of the shock wave ([Garde and Klausen, 2016](#)). In reviewing the conflicting models for pseudotachylyte genesis in the Vredefort Dome, Reimold ([1995, 1998](#)) suggested the term “pseudotachylitic breccia” be used for occurrences in impact structures until their origin could be unambiguously confirmed. However, this did not find favour among the impact research community (see, e.g., [Dressler and Reimold, 2004](#)).

In the Vredefort Dome, almost all studies of pseudotachylyte have concentrated on large dykes and more irregular network breccias exposed in granitoid gneisses of the Archaean Basement Complex in its core (e.g., [Dressler and Reimold, 2004](#); [Lieber et al., 2011](#); [Mohr-Westheide et al., 2009](#); [Garde and Klausen, 2016](#); see [Gibson and Reimold, 2008](#) for older references). While some studies report evidence of small amounts of slip (usually < 20 cm, never more than 1 m), they inevitably point to the lack of scaling up of displacements associated with larger pseudotachylyte volumes ([Reimold, 1995, 1998](#); [Dressler and Reimold, 2004](#)). [Dressler and Reimold \(2004\)](#) claimed that large dykes showed no preferred orientation in the core of the Vredefort Dome, describing their geometry as a “random zigzag” (p. 27) with smaller dykes branching off “mother lode” dykes. However, [Lieber et al. \(2009\)](#) proposed that the large pseudotachylyte occurrences in the granitoid gneisses occupy dilational sites with preferred concentric and radial orientations. They argued that this pattern is consistent with the final centrifugal and tangential collapse stage of central uplift formation, but that the melts are allochthonous and that they could not conclusively demonstrate the mechanism by which the melt originally formed. [Garde and Klausen \(2016\)](#) proposed that the lack of evidence of large fault displacements associated with the pseudotachylyte bodies, noted by many workers, could be explained if cataclasis, comminution and melting were achieved by oscillatory seismic shaking following the passage of the impact shock wave, rather than unidirectional slip. [Dressler and Reimold \(2004\)](#) measured strike lengths of up to 2.6 km and widths of 120 m for pseudotachylyte dyke networks in the granitoid gneisses in the Vredefort Dome and favoured a steep to vertical orientation. They contrasted this with pseudotachylyte dykes with dimensions 3 cm x 500 m in the supracrustal rocks surrounding the gneissic core.

In this study, we describe pseudotachylyte occurrences associated with two km-scale faults that transect the boundary between the gneissic core and supracrustal collar in the northwestern sector of the Vredefort Dome. Kinematic evidence suggests that the faults acted as a conjugate pair, allowing constraints to be placed on the bulk stress pattern that led to their formation. We demonstrate that pseudotachylyte networks and dykes show spatial, geometric, and kinematic relationships consistent with an origin by friction melting within these fault zones. The implications of this timing for pseudotachylyte genesis during the impact cratering process are discussed.

6.2 Geological Setting

The Vredefort Dome, located ~ 120 km southwest of Johannesburg, South Africa (Figure 0.1A), formed in response to a 2.023 ± 0.004 Ga (2σ ; concordant U-Pb single zircon age; [Kamo et al., 1996](#)) impact event. The 80 - 90 km dome forms the central part of what is estimated to have been a 250-300 km diameter complex impact crater and corresponds to the remaining root of the central uplift of the crater following 8 - 10 km of post-impact erosion ([Gibson et al., 1998](#); [Gibson and Reimold, 2008](#)). Although numerical modelling by [Ivanov \(2005\)](#) suggested that the main features observed in the dome can be reconciled with a significantly smaller final diameter (160 km), the dome still presents a unique opportunity to investigate the crustal response to impact at deep levels beneath a large impact crater. The dome is well exposed in the northern and western sectors but is largely covered by the Phanerozoic Karoo Supergroup in the southern and eastern sectors ([Gibson and Reimold, 2008](#); Figure 0.1A). Its ~ 40 km wide core comprises > 3.08 Ga Archaean granitoid gneisses ([Armstrong et al., 1991](#)) and minor high-grade metamafic and metasedimentary gneisses, except in the southeast where a low-grade greenstone succession is exposed ([Gibson and Reimold, 2008](#)). This is surrounded by a 20 - 25 km wide collar of unconformity-bounded supracrustal strata that were deposited in a series of basins located on the Kaapvaal Craton between 2.96 Ga and 2.1 Ga ([McCarthy, 2006](#); [Paprika et al., 2021](#)), and which have been intruded by a variety of Neoarchean to Paleoproterozoic and younger intrusions ([Gibson and Reimold, 2008](#)). The collar rocks are generally overturned and arranged as a broadly concentric sequence of layers that young outwards from the core ([Gibson and Reimold, 2008](#)). The dome is flanked by the Potchefstroom Synclinorium, an ~ 50 km wide annular structural trough created during the collapse of the transient crater and rise of the central uplift (Figure 0.1A) ([Simpson, 1978](#)).

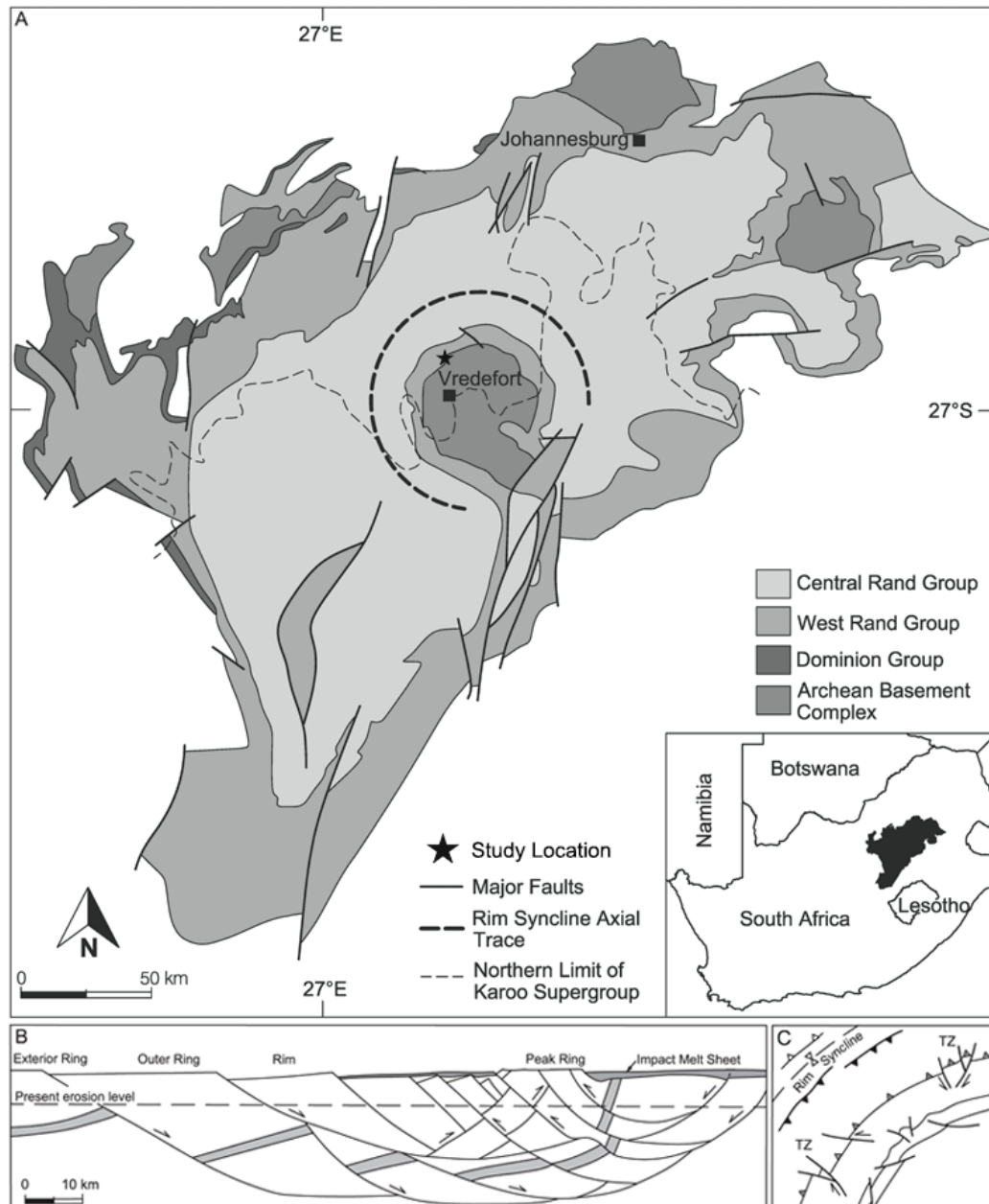


Figure 0.1: (A) Geologic map of the Witwatersrand Basin, stripped of all cover sequences, showing the location of the Vredefort Dome and the axial trace of the Potchefstroom rim syncline. The northern limit of the Karoo Supergroup is also shown. The figure is modified after [McCarthy \(2006\)](#). (B) Schematic cross-section of [Grieve et al. \(2008\)](#); modified herein) showing the approximate erosion level for the Vredefort structure of 8-10 km depth. Outward-dipping faults in the central uplift with centrifugal slip interfere with inward-dipping faults in the outer parts of the structure with centripetal slip to produce the structurally complex peak ring. (C) Schematic structural map of the northwestern sector of the Vredefort Dome showing the principal impact-related structural elements identified by previous authors in the collar, modified from [Manzi et al. \(2021\)](#). Open triangles—inward-verging thrust fault splays off outer rim normal faults ([Grieve et al., 2008](#)). Black triangles—outward-verging thrust faults ([Brink et al., 1997](#); [Grieve et al., 2008](#)). Grey triangles—rim-collapse normal faults rotated to thrust geometries in central uplift ([Jahn and Riller, 2009](#)). TZ—transpressional fault zones ([Jahn and Riller, 2009](#)). Innermost collar—constrictional folds and radial to oblique faults in Hospital Hill Subgroup ([Wieland et al., 2005](#)).

The supracrustal collar rocks comprise the 2.96 Ga $\pm$ 5 Ma Dominion Group ([Paprika et al., 2021](#)), the 2.96 - 2.79 Ga Witwatersrand Supergroup ([Kositcin and Krapež, 2004](#)), the 2.791-2.683 Ga Ventersdorp Supergroup ([Gumsley et al., 2019](#)) and the 2.67 - 2.05 Ga Transvaal Supergroup ([Hofmann, 2011](#)). The intrusive rocks comprise metadolerite of the Klipriversberg Group (2.791 - 2.779 Ga; [Gumsley et al., 2019](#)), alkali granite and mafic-ultramafic complexes of the 2.055 Ga Bushveld Complex ([Walraven, 1988](#); [De Waal, 2008](#)), and the 1.05 Ga Umkhondo age Anna's Rust dolerite ([Reimold et al., 2000](#)).

This study focuses on rocks of the Archaean Basement Complex, Dominion Group and lower WRG of the Witwatersrand Supergroup that straddle the core-collar contact in the NNW sector of the dome (Figure 0.1A). The supracrustal rocks show evidence of two metamorphic events that predate the impact ([Bisschoff, 1982](#); [Beaton et al., 2022](#)). The first event (M1; 500°C, 3.13 kbar) occurred at 2796.0 ± 1.5 Ma ([Beaton et al., 2022](#)) and varies from lower greenschist facies grade in the CRG to amphibolite facies in the WRG ([Bisschoff, 1982](#); [Gibson and Wallmach, 1995](#); [Beaton et al., 2022](#)). The second event (M2; 528°C, 4.95 kbar) is linked [by Beaton et al. \(2022\)](#) to heating associated with the Bushveld Large Igneous Province, resulting in minor textural modification of the M1 assemblage. Following the impact, the rocks experienced renewed metamorphism induced by residual shock heating and structural uplift, which led to the recrystallisation of pseudotachylyte and partial overprinting of the pre-impact metamorphic assemblages ([Gibson et al., 1998](#)). [Gibson et al. \(1998\)](#) estimated post-impact metamorphic conditions of 525 °C, 3 kbar for rocks in the study area, indicating that the present levels of exposure were located at a depth of ~10 km following the impact.

6.3 Methods

Detailed lithological and structural mapping was conducted on parts of Kopjeskraal, Eldorado, and Koedoesfontein farms, located 7 km northwest of Parys, during several visits between 2019 and 2024, totalling one week. Station points were georeferenced to Universal Transverse Mercator (UTM) World Geodetic System 1984 (WGS 84) using a Garmin Etrex®20 global positioning system (GPS). The collected measurements and mapped contacts were used to create representative geological maps by overlaying UTM WGS 84 GPS-referenced field station points onto georeferenced Google Earth Pro® images using ArcGIS 10.3.1®.

The orientations of all planar features (sedimentary bedding, lava flow tops and layering, quartz veins, planar pseudotachylyte dykes and significant dyke offshoots, faults and principal

fractures) were measured using a Brunton TruArc compass–clinometer, corrected for a magnetic declination of 20° W. The various fractures and features were grouped according to orientation and location and plotted onto an equal-area lower-hemisphere stereonet (Schmidt net) using Stereonet 9.9.4© to determine structural trends and analyse fault geometries.

The geometry of the poorly exposed Eastern Fault was obtained using Leapfrog Geo®. Topographic contours (20 m interval) were digitised from the 1:50,000 National Map Series (National Geospatial Information Directorate, 1995) as polylines, and the surface geological map was imported into Leapfrog Geo® from ArcGIS 10.3.1® and draped over the topography. The software then calculated the fault plane geometry from the fault trace. The model required a 3D block boundary to confine the limits of the interpolations (mathematical links/extrapolations between data points that combine to create the 3D surfaces). The horizontal *x*- and *y*-axis limits correspond to the extent of the field mapping area (Figure 0.2). To extrapolate the fault plane to depth, the *z*-axis limits were extended from a base of 1360 m elevation upwards to the topography, with a maximum elevation of 1575 m.

A total of 25 thin sections from a suite of 37 samples were studied under transmitted light for mineralogy, textures and microstructural fabrics. Two oriented thin sections from the metabasalt outcrops along the Western Fault were used to constrain fault kinematics from microstructural fabric analysis. Selected samples were subjected to backscatter electron imaging to investigate microcrystalline mineralogy and textures. The automated Tescan Integrated Mineral Analyser (TIMA) housed in the Wits Automated Mineralogy Lab (WAM-Lab) in the School of Geosciences was used to generate high-resolution element and phase maps and back scatter electron images of selected thin sections. The full capabilities of the TIMA are outlined in [Hrstka et al. \(2018\)](#).

Sample powders of 31 samples were analysed for major element compositions at the Earth Laboratory in the School of Geosciences at the University of the Witwatersrand. Fused disks with a latetrabate flux were analysed using the Panalytical Axios Max instrument, operating at 50 mA and 50 kV. Data reduction was performed using SuperQ software. Accuracies for the major element XRF analyses (all in wt. %) are: SiO₂ [0.5]; Al₂O₃ [0.1]; Fe₂O₃ [0.05], MgO [0.05], CaO [0.05], Na₂O [0.05] and K₂O [0.05]; TiO₂ [0.01], MnO [0.01] and P₂O₅ [0.01].

The strain history of the rock volume represented by the field area was modelled using iSALE-2D (impact Simplified Arbitrary Lagrangian Eulerian), which is based on the SALE hydrocode by [Amsden et al. \(1980\)](#). This multi-rheology, multi-material shock physics code includes

several modifications: 1) porous compaction model ([Wünnemann et al., 2006](#)); 2) fragmentation models ([Melosh et al., 1992](#)); 3) several equations of state ([Ivanov et al., 1997](#)); 4) elastoplastic constitutive model; and 5) dilatancy model ([Collins, 2014](#)). To achieve a 160 km final crater diameter obtained by [Ivanov \(2005\)](#), the impactor parameters selected for the model are a 14 km diameter spherical projectile with a velocity of 15 km/s and a density of 2.670 kg/m^3 , representative of asteroid densities ([Britt et al., 2002](#)). The vertical velocity component of 15 km/s was chosen due to the statistical probability that impact occurs at a 45° angle ([Shoemaker, 1962](#)) and that the impactor would be travelling at the average asteroid impactor velocity of 20.5 km/s ([Le Feuvre and Wieczorek, 2011](#)). A 3-layer target was chosen to represent, respectively, a 17 km thick upper crust dominated by supracrustals (represented by quartzite), a 28 km thick crystalline lower crust represented by granite, and the mantle that begins at a depth of 45 km, represented by dunite.

Tables generated by ANEOS ([Thompson and Lauson, 1974](#)) were used to constrain the thermodynamic behaviour of the materials in the model. The resolution cell width is 200 m, with 35 computational cells per impactor radius (CPPR), with a final rim-to-rim diameter of 160 km and a core-rim syncline boundary at a 40 km radial distance, as proposed by [Ivanov \(2005\)](#). The model was run in 2 s time steps for 600 s, which is enough time for the final crater shape to form. Stress and strain data were extracted for the volume of rock, hereafter referred to as the unit cell, that had its final position at a radial distance of 21.4 km and 10 km depth with respect to the surface following the impact, corresponding to the location of the study area.

The formation of complex impact craters can be divided into 3 phases that may partially overlap or vary spatially according to the location within the impact structure ([Gault and Wedekind, 1978](#); [Melosh, 1989](#); [Melosh and Ivanov, 1999](#)): 1) The Contact and Compression Phase; 2) The Excavation Phase; and 3) The Modification Phase. For the purposes of this paper the Modification Stage has further been subdivided into 3 substages: Stage 1) Inward movement of the TC wall; Stage 2) Upward movement of the central part of the TC floor; and Stage 3) Outward and downward movement during the collapse of the central uplift. These can be compared to the strain phases of [Rae et al. \(2019\)](#) figure 8: Inward Transport, Central Peak Entrainment, Outward Transport and Relaxation.

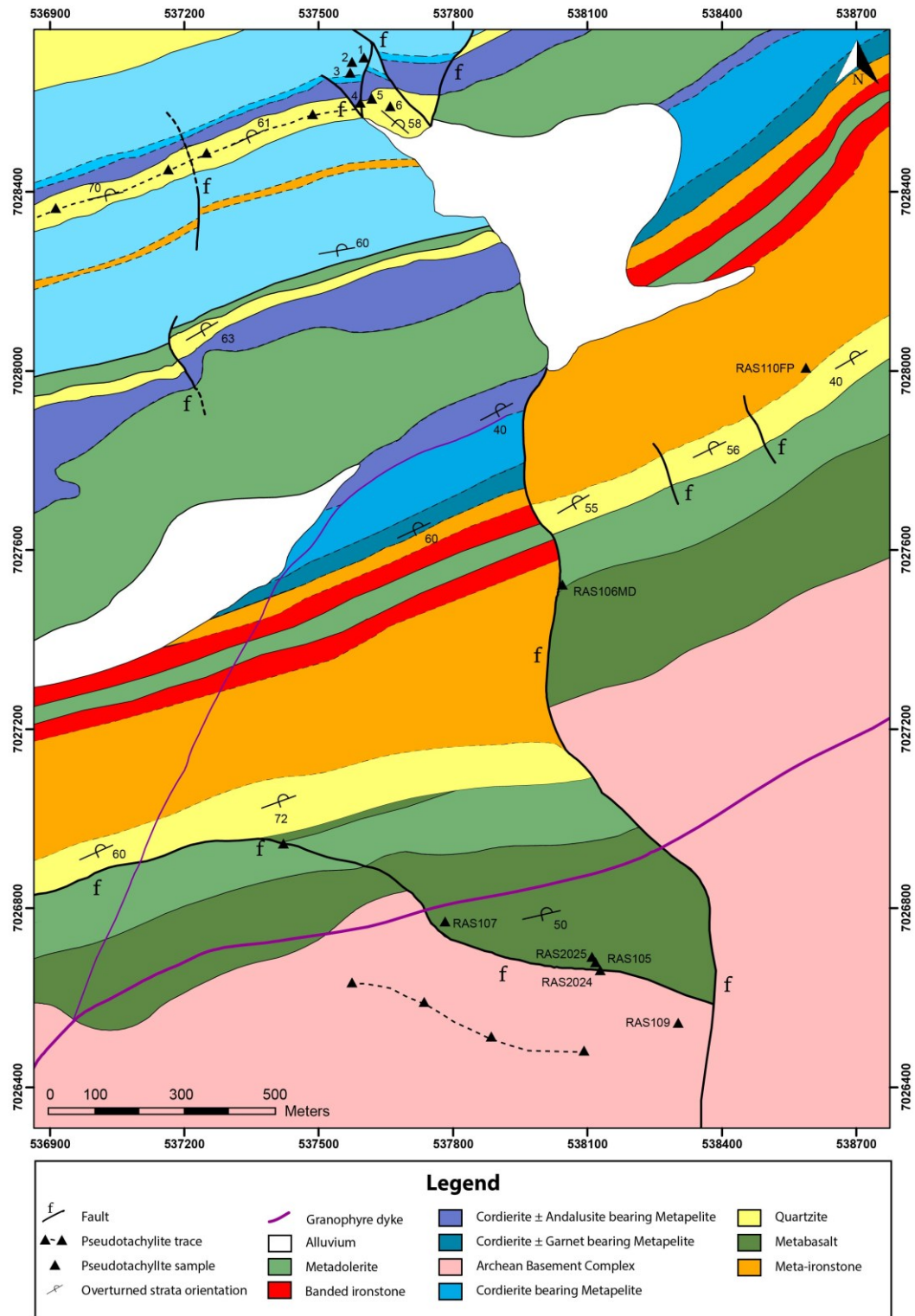


Figure 0.2: Geologic map of the study area. The Western Fault strikes approximately E-W with a dextral strike separation of 300 m. The Eastern Fault strikes approximately N-S, continuing beneath the alluvium and has a sinistral strike separation of 500 m. Noteworthy pseudotachylite sample locations are marked. 1 – RAS108MP, 2 – RAS103, 3 – RAS104FP, 4 – RAS100MP, 5 – RAS102QP, 6 – RAS101QP. A 550 m long pseudotachylite dyke that runs approximately parallel to the Western Fault is shown in the dashed lines. Both faults are cross-cut by granophyre dykes.

6.4 Results

6.4.1 Lithological Descriptions

The southern part of the study area (Figure 0.2) is underlain by the generally massive Archaean Basement Complex. To the north, the supracrustal strata show a general ENE strike and steep (40°-70°, average 60°) SSE dip, younging towards the north. Well-preserved sedimentary way-up criteria confirm that the sequence is overturned. Comprehensive lithological descriptions and discussion of metamorphic assemblages are available in [Mashabela \(2016\)](#) and [Beaton et al. \(2022\)](#). For the purposes of the present study, which focuses on pseudotachylyte genesis, the principal features of the host rocks and the pre-impact mineral assemblages are as follows:

1. The granitoid gneisses are weakly migmatitic and dominated by relatively K-poor varieties (trondhjemite) and granodiorite, with plagioclase more abundant than K-feldspar. They are coarse-grained (average grain size ~6 mm), locally megacrystic (2 cm), and leucocratic (<5% biotite). The foliation is variably but generally weakly developed and is defined by migmatitic layering, biotite knots and aligned feldspar megacrysts or quartz lenses. Exposures are excellent to the west, with relief of up to 25 m in the form of tors.
2. The Dominion Group metabasalt sequence is ~310 m thick in the study area ([Jackson, 1994](#)). It is generally well-exposed on the south-facing pediment of the ridge that exposes the Orange Grove Formation quartzite. Amygdales become more abundant stratigraphically upwards. Slightly irregular flow top breccias containing dm-scale lenses of amygdale-rich metabasalt allow confirmation of an overturned layering orientation of 077°/50°/SE. The metabasalt mineralogy is dominated by fine-grained (≤1 mm) amphibole and plagioclase with limited quartz and accessory ilmenite ± titanite ± garnet (Figure 0.3).

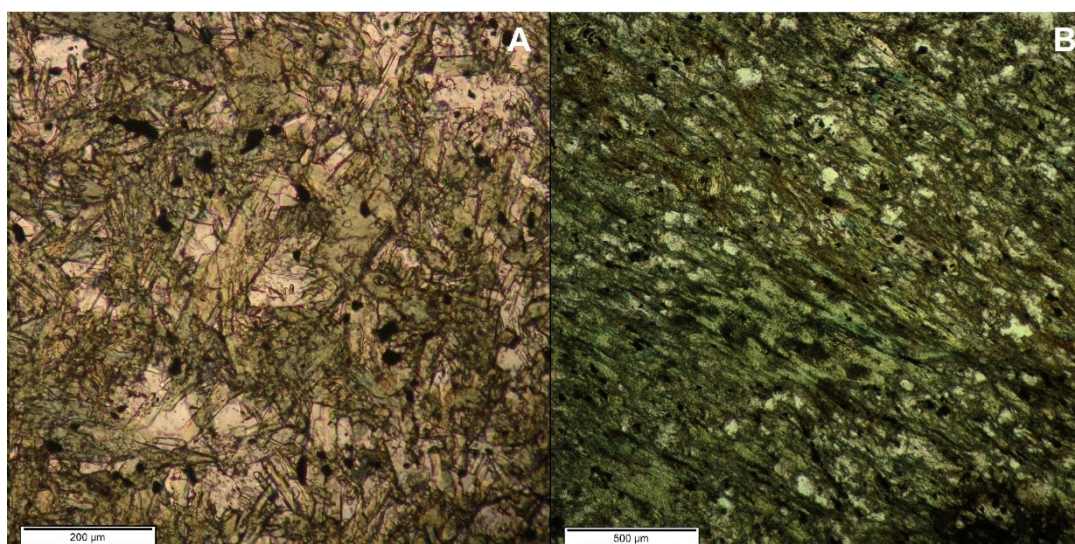


Figure 0.3: Plane-polarised light microphotographs of metabasalt sample RAS-2025 showing (A) Metabasalt groundmass away from the sheared zone displaying a fine-grained (average grain size $< 75\mu\text{m}$) randomly oriented groundmass consisting of 55% amphibole, 32% plagioclase, 9% quartz, 2% ilmenite, 2% other. (B) Foliation of the mylonitic shear zone representing the primary shear fault set ($287^\circ/54^\circ/\text{NE}$) represented by aligned amphibole that defines the oblique foliation. Amphibole 55%, plagioclase 41%, hypersthene 3% and 1% ilmenite. Sample location UTM 35J 0538109 E 7026691 S.

3. The Hospital Hill Subgroup comprises three coarsening-upward cycles of quartzite, separated by metapelite and meta-ironstone units (Figure 0.2). Peak metamorphic grade of mid-amphibolite facies ([Bisschoff, 1982, 2000](#); [Gibson and Wallmach, 1995](#); [Beaton et al., 2022](#)) has produced coarse porphyroblastic assemblages in the pelites and ironstones, but the pelites and quartzites still preserve excellent sedimentary structures. Ironstones are dominated by garnet-grunerite-magnetite-quartz-chlorite ([Bisschoff, 1982, 2000](#); [Mashabela, 2016](#)) and pelites by various combinations of cordierite-andalusite-staurolite-garnet-biotite-muscovite-chlorite-quartz-ilmenite and plagioclase, depending on bulk composition ([Bisschoff, 1982, 2000](#); [Mashabela, 2016](#)).
4. Two major – and several other thin - metadolerite sills intrude the basalt-sediment sequence. Based on their geochemistry, [Beaton et al. \(2022\)](#) correlated them with the lavas of the 2.79 Ga Klipriviersberg Formation. The largest sills, which are respectively 242 m and 536 m thick, retain vestiges of their igneous texture macroscopically, particularly in the coarse-grained middle to upper-middle levels. Microscopic analysis shows that the igneous mineralogy has been pseudomorphously replaced by hornblende/actinolite-plagioclase-clinzoisite-titanite-ilmenite $\pm$ chlorite. The smaller sills and large sill margins display more complete replacement by medium-to very coarse-grained acicular actinolite and plagioclase.

5. An up to 30 m wide impact melt rock dyke, trending 080°, extends across the study area, with a thinner NE-trending offshoot indicated by sporadic small outcrops and loose boulders (Figure 0.2). The dyke displays a positive topographic expression owing to its siliceous character. It comprises orthopyroxene-plagioclase-K-feldspar-quartz with trace amounts of ilmenite and biotite, and with K-feldspar and quartz intergrown in a granophyric texture. Exposed dyke margins display a thin aphanitic chill zone and strong jointing. Lithic clasts (predominantly cm- to dm-scale granite and quartzite) are concentrated within a few metres of the contact, whereas the dyke centre is clast-free but coarser-grained, with plagioclase laths up to 1-2 mm in the matrix. The dyke is the only unmetamorphosed lithology in the area and is the only lithology not displaced by the faults (Figure 0.2) or intruded by pseudotachylyte.

6.4.2 Structural Geology

The 1:50,000 maps of [Nel \(1927\)](#) and [Bisschoff \(2000\)](#) show only a single, broadly N-S trending fault (Eastern Fault) with a left-lateral strike separation of ~500 m in the study area. However, we have identified a second fault, trending ESE (Western Fault), which shows a right-lateral strike separation of the core-collar contact of ~300 m, but which curves parallel to the basal contact of the Orange Grove Formation towards the west (Figure 0.2). As is common throughout the collar, the fault zones are characterised by a general lack of outcrop; however, unlike in the core of the dome, the well-defined stratigraphy and the rugged topography (200-300 m elevation variation), as well as distinctive vegetation patterns linked to the bedrock, allow tight constraints (≤ 10 m) to be placed on the fault traces. In both the Eastern and Western Faults, the primary fault zone, marked by an absence of outcrop and, locally, negative relief, appears to be a maximum of 10-20 m wide. However, both are accompanied by wider damage zones in which secondary synthetic faults with smaller displacements provide key information. Both faults are cut by the ENE-trending impact-melt dyke (Figure 0.2), which provides an important time constraint.

Eastern Fault

Decimetre- to m-scale boulders rich in pseudotachylyte are found sporadically along the poorly exposed Eastern Fault trace. In some of these boulders, the pseudotachylyte is polymict and contains exotic clasts corresponding to wall rocks found tens to hundreds of metres away (see Figure 0.7A). Towards the northern end of the Eastern Fault, several smaller fault splays into the footwall west of the main fault show left-lateral strike separations of the Brixton Formation

quartzite of a few metres up to ~40 m (Figure 0.2). They are marked by negative weathering, polymict pseudotachylyte-rich boulders with dm- to m-scale randomly oriented lithic clasts of metapelite and quartzite (Figure 0.7) and irregular pseudotachylyte network breccias in adjacent outcrops. The splay faults isolate an ~100 m x 100 m block of Brixton Formation quartzite that is rotated clockwise by ~70° in the main fault's footwall, but which is also internally faulted (Figure 0.2). The quartzite displays at least one, and locally two, intense cm- to dm-spaced fracture cleavages that show mm- to cm-scale offsets of bedding.

Leapfrog Geo 3D modelling ([see Methods](#)) was used to obtain the orientation of the Eastern Fault (340°/26°/E) from its irregular outcrop trace. The truncation and left-lateral offset of a mafic dyke in the granitoid gneisses, established from a geophysical survey (S.J. Webb, pers. comm., 2024) suggests that the Eastern Fault extends at least a kilometre further south of the core-collar contact (Figure 0.2). No trace of the Western Fault was detected on the eastern side of the Eastern Fault in the geophysical survey.



Figure 0.4: Photograph with schematic showing bedding subparallel generation plane pseudotachylyte oriented at $110^{\circ}/72^{\circ}/S$ with a mean width of 11 cm. Offshoot veins branch off the generation plane as wedges oriented centrifugally towards the north. Bedding oriented at $120^{\circ}/58^{\circ}/S$. The offshoot veins are not always continuous in the plane of view and have a mean width of 3 cm. The generation plane pseudotachylyte is oriented 10° counterclockwise to the bedding orientation. Photo location UTM 35J 0537684 E 7028574 S.

Western Fault

A narrow, alluvium-filled, valley between outcrops of the granitoid gneiss and metabasalt metadolerite marks the Western Fault trace (Figure 0.2). A broad, well-exposed metabasalt pavement that extends more than ~50 m north of the valley towards its ESE end displays a prominent irregularly developed dm- to m-spaced fault-fracture pattern that includes mm- to cm-wide mylonitic shear faults oriented $287^{\circ}/54^{\circ}/NE$ (Figure 0.5) and which is associated with abundant pseudotachylyte dykes and network breccias. These faults show consistent brittle-to-ductile mm- to cm-scale right-handed offset of subvertical N-S oriented quartz veins (Figure 0.6), and the wider faults also display an oblique shear fabric and local asymmetric, mm-scale crenulations that indicate right-handed shear. Left-lateral offsets of 3 cm wide quartz veins are seen locally on a subsidiary exclusively brittle fault set oriented $237^{\circ}/62^{\circ}/NW$ (Figure 0.5 and Figure 0.6). Based on the similar, right-handed, large-scale offset of the metabasalt-gneiss

contact along the main fault zone, we infer that the main fault parallels the dominant shear fault orientation and, thus, exhibits an orientation of $287^{\circ}/54^{\circ}/\text{NE}$ (Figure 0.5). Although dense vegetation and the tor-like outcrop expression of the granitoid gneisses close to the fault trace hampers structural analysis, flat outcrop pavements up to 150 m south of the fault display a similar conjugate pattern of narrow (<1 cm) planar faults with both right- and left-handed cm-scale displacements of quartz veins.

The absence of any significant apparent offset of the Orange Grove Formation towards the WNW (Figure 0.2), but the presence of boulders of polymict pseudotachylyte containing metadolerite, metabasalt and quartzite clasts immediately south of the quartzite outcrop leads us to conclude that the fault zone strike is deflected into parallelism with bedding strike in the supracrustal strata towards the WNW. Unfortunately, no outcrop evidence of the dip angle of the fault could be found here.

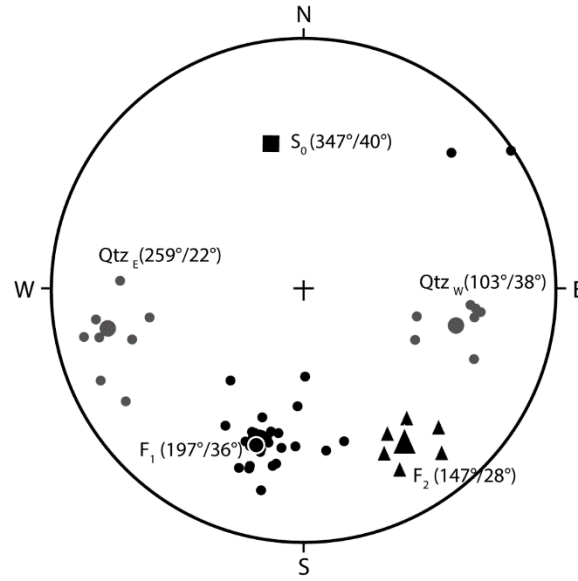


Figure 0.5: Equal area stereonet summarising all the structural data from the metabasalt outcrop adjacent to the Western Fault: S_0 denotes poles to metabasalt layering ($077^\circ/50^\circ/\text{SE}$); Qtz_W denotes poles of the westerly dipping quartz veins ($193^\circ/52^\circ/\text{W}$; $N=6$) and Qtz_E denotes poles of the easterly dipping quartz veins ($349^\circ/68^\circ/\text{E}$; $N=8$) (grey circles); primary right-handed fault set 1 ($287^\circ/54^\circ/\text{NW}$; $N=5$) and tabular fault-parallel, pseudotachylyte dykes F_1 (black circles), interpreted as the orientation of the West Fault ($287^\circ/54^\circ/\text{NE}$; $N=34$); F_2 (triangles), denotes the poles to left-handed fault set 2 ($237^\circ/62^\circ/\text{NW}$; $N=5$).

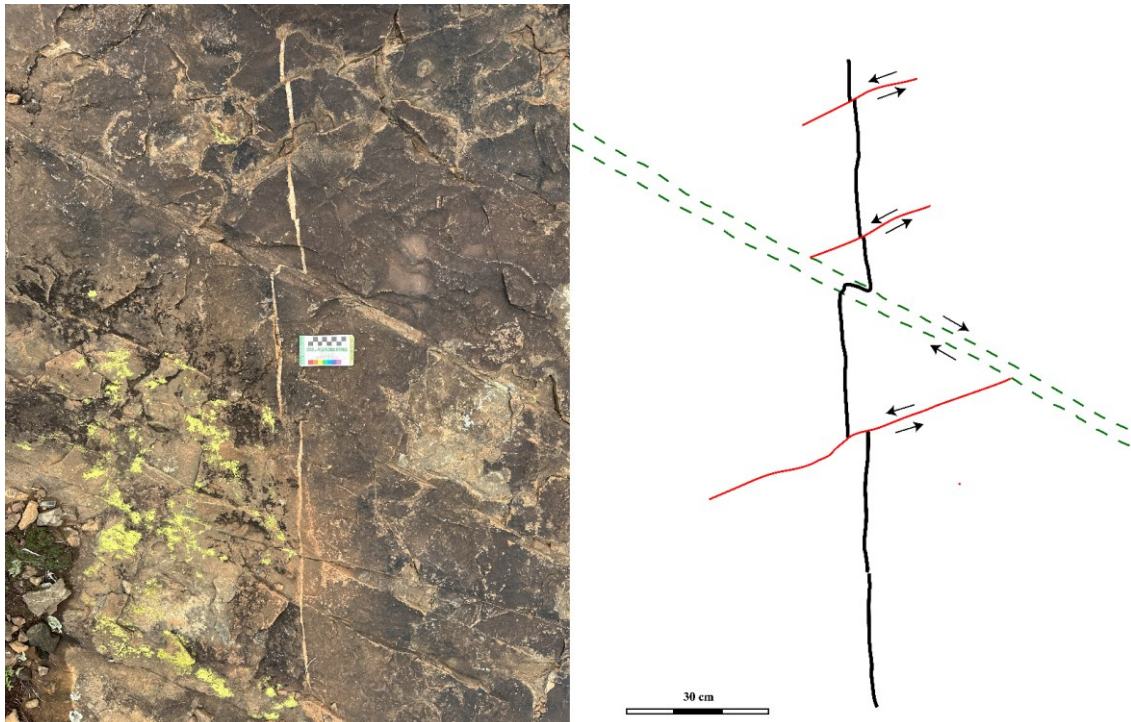


Figure 0.6: Outcrop photograph (left) and interpretive sketch (right) showing a right-handed mylonite (green dashed lines) oriented at $287^\circ/54^\circ/\text{NE}$, that displaces a quartz vein oriented $193^\circ/52^\circ/\text{W}$. Thin section RAS-2025A of Figure 0.3 is representative of this mylonite zone. While the displacement is ductile, brittle left-handed displacement occurs on a secondary fault set (red) oriented $237^\circ/62^\circ/\text{NW}$. Photo location UTM 35J 0538109 E 7026691 S.

6.4.3 Pseudotachylyte

Macroscopic Features

All the pre-impact lithologies within the field area contain pseudotachylyte; however, only a few, mm-to-cm wide, dykes were found in the meta-ironstones and none in the banded iron formations. Although dykes ranging from <1 mm up to a few centimetres in width are found in outcrops throughout the area (e.g., [Gibson et al., 1997](#)), the number and volume of individual dykes appears greatest in the vicinity of the two fault traces.

Pseudotachylyte dykes in the metabasalt outcrop are spatially closely associated with secondary, cm-scale mylonitic fault zones and reach widths of 10-20 cm and lengths of at least several metres. The number of dykes decreases away from the main fault trace to the north as the number of secondary faults reduces. Morphologically, the longest dykes occur parallel to the Western Fault orientation and either display parallel margins or pinch-and-swell features. Orthogonal, short, wedge-shaped offshoots of melt up to tens of centimetres long occur as offshoots from the main dykes, with flow folds indicating migration of melt from the fault-parallel dykes into the offshoot veins. The fault-parallel dykes locally merge with network breccias in which thin, highly irregular, branching networks of pseudotachylyte anastomose around slightly disrupted dm-scale basalt clasts. These network breccias occur either between two planar pseudotachylytes or in left- or right-stepping rhombic relay structures between partially overlapping faults. The pseudotachylyte is generally monomictic with only rounded lithic clasts of metabasalt (Figure 0.7A), although clasts of quartz derived either from pre-faulting veins or amygdalae are noted locally. A single boulder of polymictic pseudotachylyte containing metabasalt, metadolerite and quartzite clasts was observed at the western end of the fault close to the contact with the Orange Grove Formation.

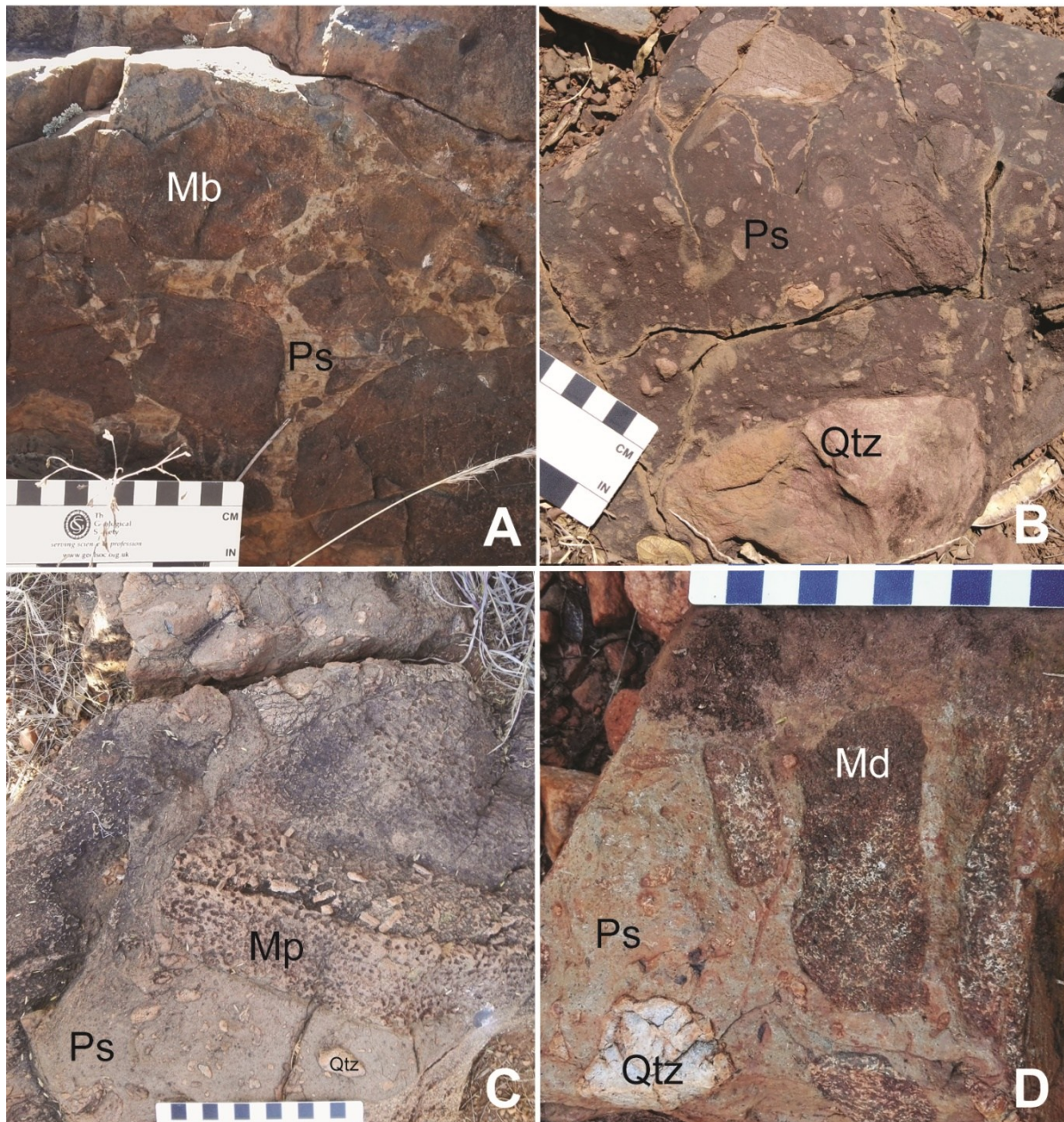


Figure 0.7: A) Pseudotachylyte in metabasalt, white and chalky in appearance. Sample RAS-105, UTM 35J 538183 E 7026673 S. B) Pseudotachylyte in metapelite containing clasts of metapelite and quartzite UTM 35J 528070 E 7022759 S. C) Pseudotachylyte in metapelite UTM 35J 0537852 E 7028681 S. D) Pseudotachylyte in metadolerite, green and grainy in appearance, with quartzite clasts. Sample RAS-106, UTM 35J 538043 E 7027523 S. Abbreviations: Qtz-quartzite, Mb-metabasalt, Md-metadolerite, Mp-metapelite, Ps-pseudotachylyte.

The matrix of the metabasalt pseudotachylyte is yellowish to dark brown and appears very fine-grained and homogenous on weathered surfaces. It is black and aphanitic on fresh surfaces. Contacts with the wall rock and clasts are sharp. Differential weathering enhances internal layering in some dykes; mm- to cm-scale asymmetric flow folding is found locally. Metabasalt-hosted dykes contain significantly fewer small clasts ($\leq 1\text{ cm}$) than pseudotachylytes in other lithologies, although the network breccias are commonly clast-rich (Figure 0.7).

No increase in volume or intensity of pseudotachylyte dykes was noted towards the fault in the granitoid gneiss outcrops to the south of the Western Fault, although these do display an irregular network of mm- to cm-wide dykes, locally reaching dm-widths, with a wide range of orientations. However, 170 m south of the fault trace and striking parallel to it is a $> 500\text{ m}$ long, continuously exposed dyke which has a maximum width of 17 m towards its eastern end and which pinches and swells as it tapers towards the west (Figure 0.2). The pseudotachylyte matrix is brown-grey on weathered surfaces and blue-grey on fresh surfaces and it displays a generally steep flow foliation parallel to the dyke margins that wraps around clasts (Figure 0.8A). Apart from granitoid gneiss, the only other clast type seen is vein quartz. Clasts are significantly more abundant than in the metabasalt-hosted dykes and range from mm-sized quartz and feldspar mineral clasts to dm to m-sized granitoid gneiss lithic clasts; the degree of rounding increases with clast size. Adjacent to the Eastern Fault, the gneisses north of the impact-melt dyke contain irregular clast-poor network breccias with local tension gash geometries indicative of a sinistral slip sense. Small boulders of polymictic pseudotachylyte containing cm- to dm-scale subrounded to subangular clasts of granitoid gneiss and metadolerite, and in one case, meta-ironstone, are found along the fault trace.

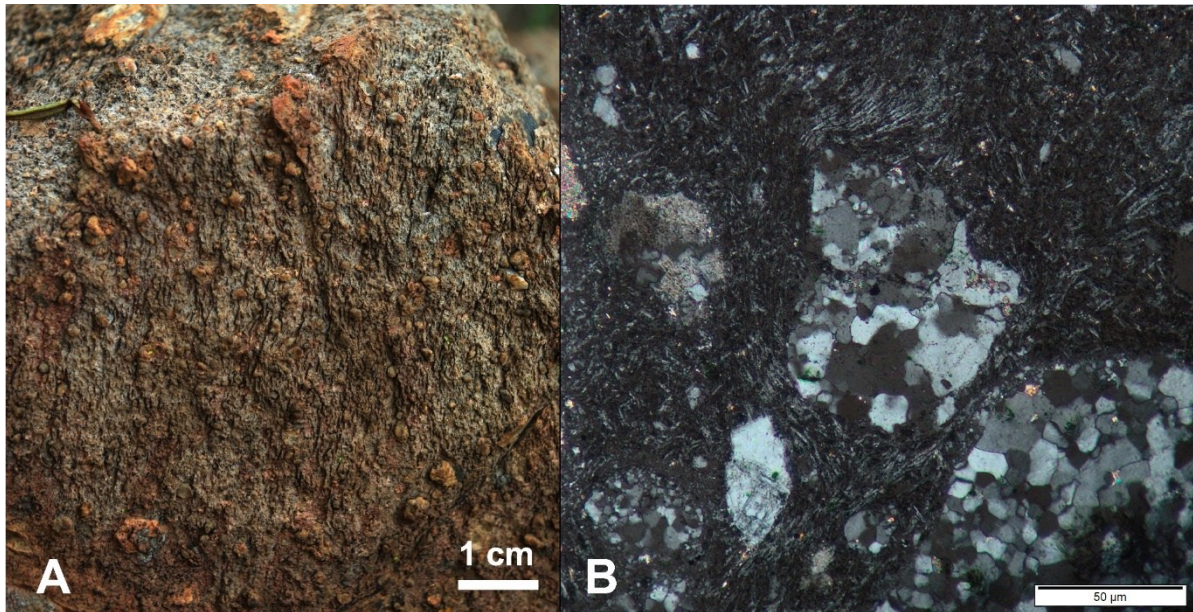


Figure 0.8: Characteristics of large pseudotachylyte dyke in the granitoid gneisses south of the Western Fault. (A) Outcrop photograph showing the strong subvertical foliation and small quartzofeldspathic lithic and mineral clasts. Photo location UTM 35J 0538060 E 7026489 S. (B) Photomicrograph showing aligned plagioclase microlites defining a flow foliation in the matrix surrounding recrystallised mono- and polymineralic clasts of quartz and/or feldspar (Crossed polars). Sample location UTM 35J 0538060 E 7026489 S.

Poor exposure of the metadolerite against both faults limits in situ study of the pseudotachylyte, but mm- to cm-wide dykes occur in association with small (<1 cm) slightly irregular to branching shear zones that consist of <1 mm wide sinuous shear planes that anastomose between lensoid low-strain domains. The dykes appear brown on weathered surfaces and black on fresh surfaces, and are relatively clast-poor. In addition to metadolerite clasts, the metadolerite-hosted pseudotachylyte RAS-106MD contains a quartzite clast (Figure 0.7D), despite the nearest quartzite outcrop being over 100 m distant (Figure 0.2)

In the north of the area, boulders of pseudotachylyte up to 50 cm in diameter along the main Eastern Fault trace contain dm-scale clasts of metapelite and quartzite that are generally angular to subangular and the splay fault region contains an irregular branching cm- to dm-wide dyke network with a similar polymictic lithic clast population, as well as mineral clast remnants of porphyroblastic aluminosilicates from the metapelites (Figure 0.7B, C). The pseudotachylyte is pale brown on weathered surfaces.

The Brixton Formation quartzite contains cm- to dm- scale pseudotachylyte dykes that are blue-grey on weathered surfaces but dark grey to black on fresh surfaces. Planar dykes that are oriented subparallel to bedding display wedge-shaped offshoots – some of which extend into

slightly sinuous tension gash arrays (Figure 0.9). Although in most cases the wall rocks adjacent to the planar dykes do not show any macroscopic evidence of shearing or cataclasis, Figure 0.10 shows a dyke within a cataclasite zone with a distinctive blue-grey colour relative to the host quartzite, confirming the temporal relationship between cataclasis and pseudotachylyte development (see next section).



Figure 0.9: Bedding-subparallel pseudotachylyte generation plane with wedge-shaped offshoots in cross-bedded Brixton Formation quartzite. Photo location UTM 35J 0537615 E 7028598 S.



Figure 0.10: Thin pseudotachylyte within cataclasite in Brixton Formation quartzite. Sample RAS-102. Photo location UTM 35J 0537659 E 7028591 S. The dark blue-grey pseudotachylyte shows flow banding and contains visible quartzite clasts. The lower dyke contact is more irregular and diffuse than the upper contact. The lighter grey marginal zones comprise cataclasite. These are flanked by wider zones of discoloured blue-white quartzite that transition into the white quartzite host rock and contain irregular spots of black iron oxide, which are absent from the host quartzite (see Figure 0.13).

A stratabound pseudotachylyte dyke network system can be traced almost continuously along strike in the Brixton Formation for over 2 km to the west of the Eastern Fault (Figure 2; [Mashabela, 2016](#)). Unlike the much more voluminous dyke in the granitoid gneisses, the bedding-parallel straight dyke segments that extend up to several tens of metres rarely exceed 20 cm in width (Figure 0.4). These segments are linked by irregular network breccias, similar to the geometry seen in the metabasalt. The planar dyke segments are characterised by a strong mm-scale planar layering that imparts a fissile character to the pseudotachylyte. The layering is parallel to the dyke margins and is strongly deflected around rare quartzite clasts (Figure 0.11), consistent with significant laminar flow. In contrast, pseudotachylyte layering in the network breccias is locally chaotically folded.

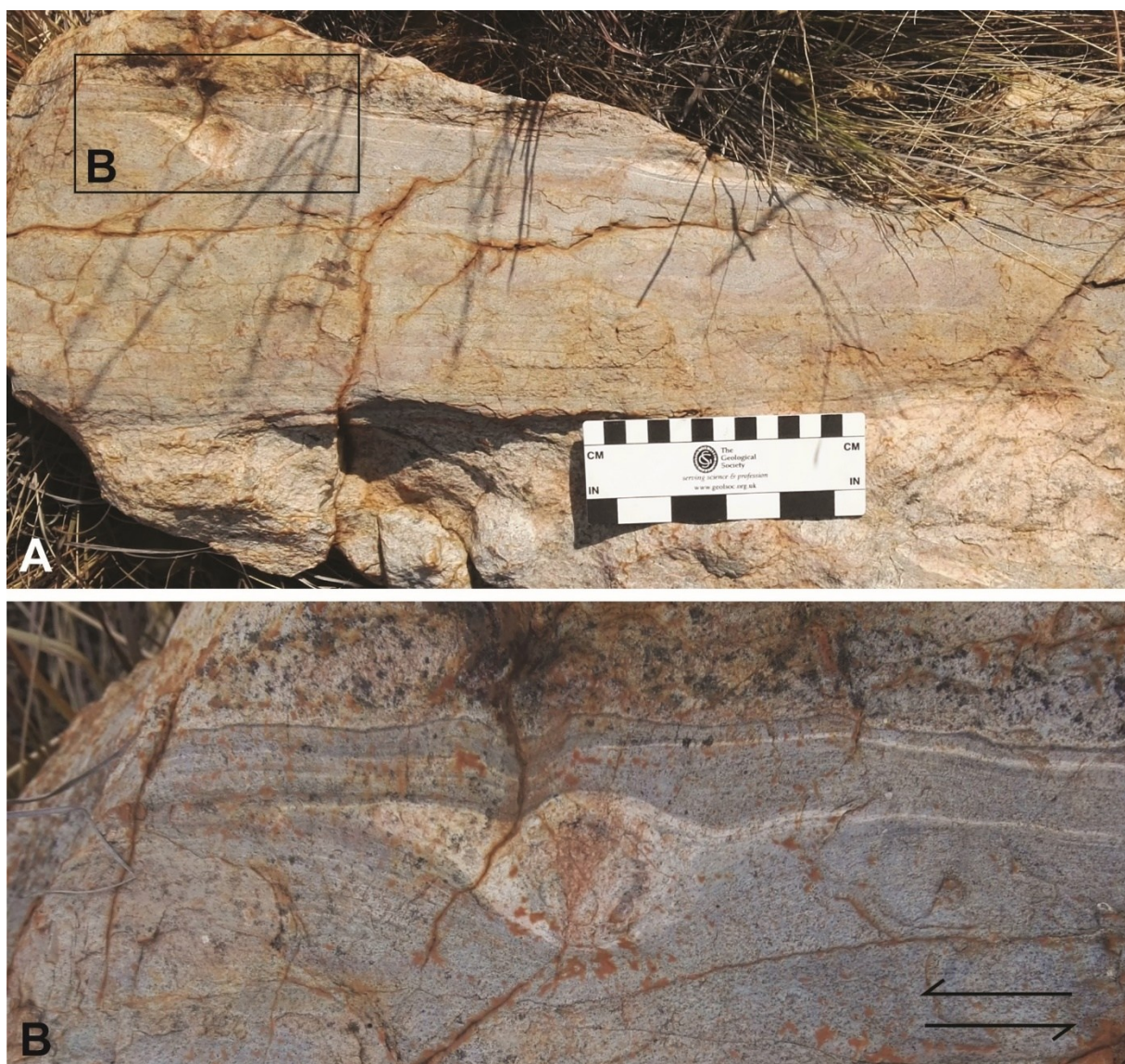


Figure 0.11: A) A decimeter-wide, almost bedding-parallel pseudotachylyte dyke in Brixton Formation quartzite showing layering being deflected around a cm-sized quartzite clast (B). The slightly asymmetric pressure shadow adjacent to the quartzite clast is consistent with a sinistral shear sense. A 1.5 cm marginal zone with black/blue speckles occurs on the margin of the pseudotachylyte (see Figure 0.10). Photo location UTM 35J 537589 E 7028579 S.

Microscopic Features

Microscopic analysis confirms that pseudotachylyte dykes and veins are characterised by a microcrystalline matrix. No glass was detected; however, plagioclase microlites in a granitoid-hosted dyke define a flow foliation (Figure 0.8B) that supports crystallisation from a melt, and plagioclase spherulites in a metabasalt-hosted dyke indicate either crystallisation from a melt or devitrification of glass (Figure 0.12). In all other samples, the matrix is granoblastic-microcrystalline, which we interpret as a metamorphic recrystallisation texture caused by the persistence of elevated temperatures following the impact (e.g., [Gibson et al., 1997](#); [Ivanov, 2005](#); [see Section 1.3.1](#)). This metamorphic overprint makes it difficult to tell if colour and grain size changes along the dyke margins represent a chill zone or slightly different composition.

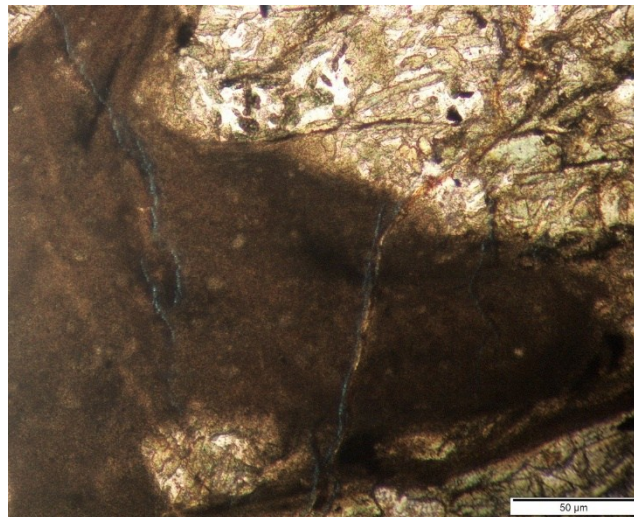


Figure 0.12: Microphotograph (plane-polarised light) showing ~10 µm plagioclase spherulites in the metabasalt pseudotachylyte sample RAS-2025. Sample location UTM 35J 0538109 E 7026691 S.

Dyke contacts are usually sharp but show microscopic evidence of embayments and cusped intrusions into, and rip-outs of, the wall rock. Wall rock may be undeformed or display a cataclastic microtexture, and cataclastic aggregates are locally entrained and variably assimilated into the pseudotachylyte matrix. Lithic clasts may show a gross angular shape but typically show rounded corners. Quartz mineral clasts are found mainly in the granitoid-hosted pseudotachylyte and may be angular to subrounded, with heterogranular polygonal-granoblastic recrystallisation (Figure 0.14). Feldspar clasts are also recrystallised but locally show plastically deformed lobate to wispy tails (Figure 0.14).

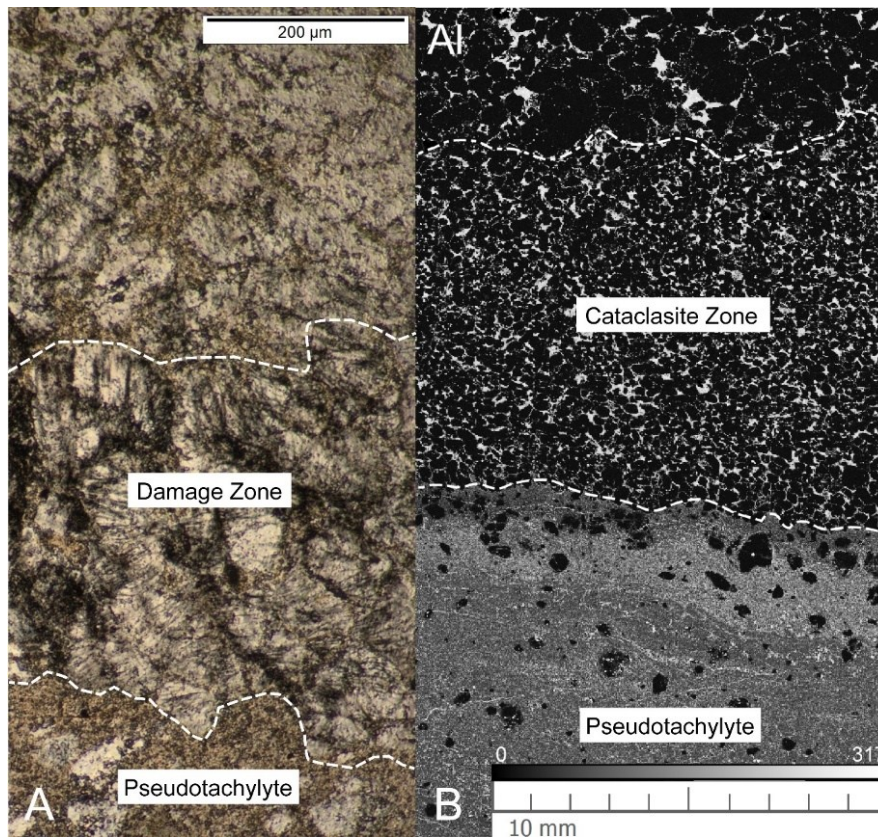


Figure 0.13: A) Transmitted light Microphotograph (plane-polarised light) of the cataclasite-hosted pseudotachylyte in the quartzite, showing damaged zones characterised by fracturing and iron oxides (Figure 0.10). The zone is marked by brittle deformation of the grains—undeformed quartz grains of the wall rock (top) and a pseudotachylyte matrix (bottom). B) TIMA element map showing the distribution of Al. The pseudotachylyte (bottom of image) shows internal compositional layering and quartz clasts (black) that decrease in abundance into the pseudotachylyte dyke. Well-rounded quartz sand grains are visible in the undeformed quartzite (top), but grain size has been significantly reduced in the cataclasite (middle). The Al enrichment is mirrored by K (not shown here), indicating a white mica phase. Sample RAS-101, location 35 J 0537618 E 7028608 S.

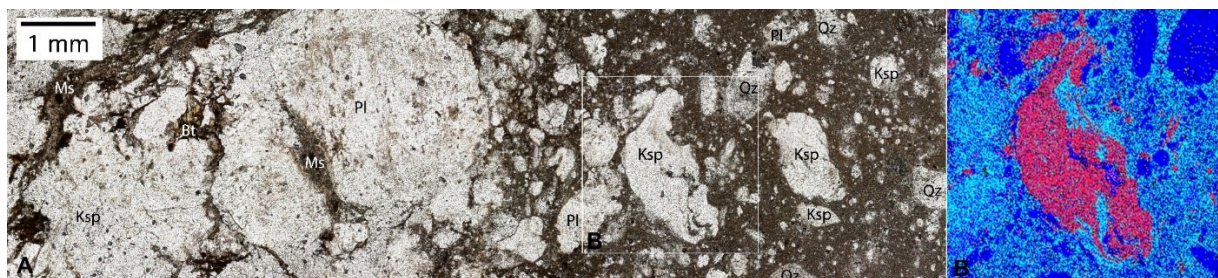


Figure 0.14: Microphotograph (plane-polarised light) of pseudotachylyte in granitoid gneiss. Quartz and feldspar clasts are abundant. The plagioclase clasts in the pseudotachylyte matrix tend to be small < 1 mm. The feldspar grains have lobate margins in contrast to the more angular quartz grains. The clast-rich matrix is indicative of pre-melting cataclasis. B) Inset of a TIMA phase map image of the lobate K-feldspar grain and surrounding matrix. Pink – K-feldspar, Blue – quartz, Cyan – plagioclase. Sample RAS-109, location 35J 538300 E 7026544 S.

Geochemistry

Fourteen paired samples of pseudotachylyte and adjacent wall rock were analysed for their major element compositions by XRF analysis, together with an additional pseudotachylyte sample (RAS-090), which had two different wall rocks (Table 0.1). While dm-sized boulders of locally polymictic pseudotachylyte were observed along the fault traces, these were not sampled as they could not be proven to be in situ. Care was taken in sample preparation to avoid visible clasts wherever possible to reduce skewing the analysis. This was most difficult in the granitoid gneiss-hosted pseudotachylyte, but all other types showed relatively clast-free matrix areas for representative geochemical sampling ([see section 6.4.1](#)).

Harker plots for host rock - pseudotachylyte pairs for selected major elements against SiO_2 are shown in Figure 0.15. TiO_2 , P_2O_5 and MnO were not plotted as these oxides show very little variation between paired samples (with the exception of MnO in metapelites; see discussion on metapelites below).

Table 0.1: Major element composition of pseudotachylyte veins and host rocks. H= Host rock, PT= Pseudotachylyte, LOI= Loss on ignition at 1000°C.

Lithology Sample Type	Metapelite								Meta-ironstone-Quartzite			Granite		Metadolerite	
	RAS-081		RAS-097		RAS-100		RAS-108		RAS-090			RAS-109		RAS-106	
	H	PT	H	PT	H	PT	H	PT	H _{Fe}	PT	H _{Qtz}	H	PT	H	PT
SiO ₂	55.03	60.83	47.44	60.84	61.01	61.67	67.78	65.39	49.61	63.88	94.4	73.89	71.91	51.53	51.78
Al ₂ O ₃	12.03	12.45	13.3	11.76	19.86	19.69	15.74	17.8	9.23	12.21	1.06	13.93	14.69	14.54	15.1
Fe ₂ O ₃	23.07	15.86	31.25	14.19	7.77	7.72	6.3	6.7	25.08	9.44	0.48	2.11	2.2	9.82	9.17
MnO	1.22	0.42	0.84	1.73	0.06	0.05	0.06	0.08	9.76	0.76	0.02	0.02	0.03	0.16	0.14
MgO	3.94	4.01	3.6	4.92	4.59	4.77	3.63	4.05	1.81	4.23	0.06	0.34	0.36	8.57	8.11
CaO	0.91	1.42	0.94	1.23	0.2	0.25	0.19	0.26	1.63	0.21	0.03	0.91	0.8	11.91	11.94
Na ₂ O	0.39	0.47	0.63	0.62	0.83	0.72	0.34	0.59	0.1	0.28	0.01	4.05	4.39	2.14	2.29
K ₂ O	1.86	1.58	1.93	1.66	2.91	2.78	2.42	2.64	0.08	1.26	0.21	3.9	3.78	0.21	0.25
TiO ₂	0.51	0.52	0.5	0.49	0.8	0.81	0.74	0.74	0.3	0.52	0.05	0.2	0.22	0.55	0.54
P ₂ O ₅	0.08	0.08	0.09	0.08	0.04	0.06	0.08	0.06	0.1	0.04	0.02	0.06	0.06	0.06	0.06
Cr ₂ O ₃	0.09	0.11	0.09	0.1	0.11	0.11	0.11	0.11	0.07	0.13	0.03	n.d.	n.d.	0.08	0.07
NiO	0.03	0.03	0.03	0.03	0.05	0.04	0.03	0.04	0.02	0.04	n.d.	n.d.	n.d.	0.03	0.02
LOI	0.86	1.52	0.61	1.8	1.78	1.67	2.01	1.79	1.91	5.18	0.32	0.56	0.71	0.96	0.98
TOTAL	100.01	99.31	101.24	99.44	100	100.34	99.43	100.23	99.68	98.18	96.68	99.96	99.15	100.54	100.45

Lithology Sample Type	Quartzite										Metabasalt				Meta-ironstone	
	RAS-079		RAS-080		RAS-091		RAS-094		RAS-098		RAS-105		RAS-107		RAS-104	
	H	PT	H	PT	H	PT	H	PT	H	PT	H	PT	H	PT	PT	PT
SiO ₂	94.15	58.01	95.34	85.25	84.23	75.45	87.82	69.34	89.76	78.43	52.95	53.9	53.06	54.22	63.4	67.89
Al ₂ O ₃	1.02	20.37	0.36	6.31	7.96	11.9	5.48	17.16	4.35	11.92	14.66	14.34	15.1	14.76	9.26	5.64
Fe ₂ O ₃	0.53	7.63	0.26	1.33	1.09	2.66	0.6	1.93	0.79	1.12	10.76	10.81	12.01	12.19	18.08	20.16
MnO	0.01	0.02	0	0.01	0.01	0.02	0.01	0.01	0	0.01	0.17	0.19	0.22	0.21	2.9	0.06
MgO	0.04	2.21	0.02	0.4	0.24	0.87	0.18	0.43	0.07	0.24	7.4	7.51	5.36	5.45	3.05	1.52
CaO	0.04	0.06	0.03	0.05	0.03	0.05	0.03	0.06	0.04	0.05	9.39	9.15	9.73	9.08	0.75	0.21
Na ₂ O	0.02	0.52	0.03	0	0.02	0.03	0.01	0.1	0.02	0.06	2.84	2.86	2.33	2.28	0.48	0
K ₂ O	0.25	3.11	0.08	1.34	2.53	3.15	1.75	4.29	1.17	3.16	0.17	0.17	0.26	0.21	1.37	0.68
TiO ₂	0.04	0.76	0.03	0.14	0.1	0.32	0.11	0.49	0.03	0.22	0.69	0.68	0.76	0.75	0.37	0.2
P ₂ O ₅	0.02	0.12	0.02	0.04	0.03	0.04	0.02	0.04	0.03	0.02	0.2	0.19	0.23	0.17	0.09	0.04
Cr ₂ O ₃	0.01	0.15	0.01	0.03	0.01	0.04	0.02	0.06	0.01	0.03	0.1	0.1	0.14	0.15	0.06	0.02
NiO	n.d.	0.04	n.d.	0.01	n.d.	0.01	n.d.	n.d.	n.d.	n.d.	0.03	0.03	0.04	0.04	0.02	0.01
LOI	0.51	6.05	0.52	2.09	1.41	3.08	1	3.94	0.94	2.54	0.81	0.66	1.24	1.66	0.98	4.09
TOTAL	96.63	99.06	96.69	97	97.67	97.62	97.03	97.85	97.21	97.79	100.16	100.59	100.49	101.16	100.81	100.51

Among all the samples, metadolerite-hosted pseudotachylyte shows the least (almost no) deviation in composition from the host rock, as might be expected given the massive nature and relatively simple mineralogy (amphibole + plagioclase) of the metadolerite. Both amphibole and plagioclase melt readily under conditions of frictional sliding ([Spray, 1997](#); [Lin, 2008](#)). Metabasalt, which has a similar basic mineralogy, also shows similarity between the pseudotachylyte and host rock, except for Na₂O (enriched) and MgO (depleted) in the RAS-107 sample. Slight enrichment in SiO₂ of ≤ 1 wt % in the metabasalt-hosted pseudotachylytes may reflect quartz microclasts derived from amygdaloids or quartz veins that are not necessarily

well represented in the bulk rock composition (Table 0.1). Other samples show more significant degrees of deviations between pseudotachylyte and wall rocks, which are discussed below.

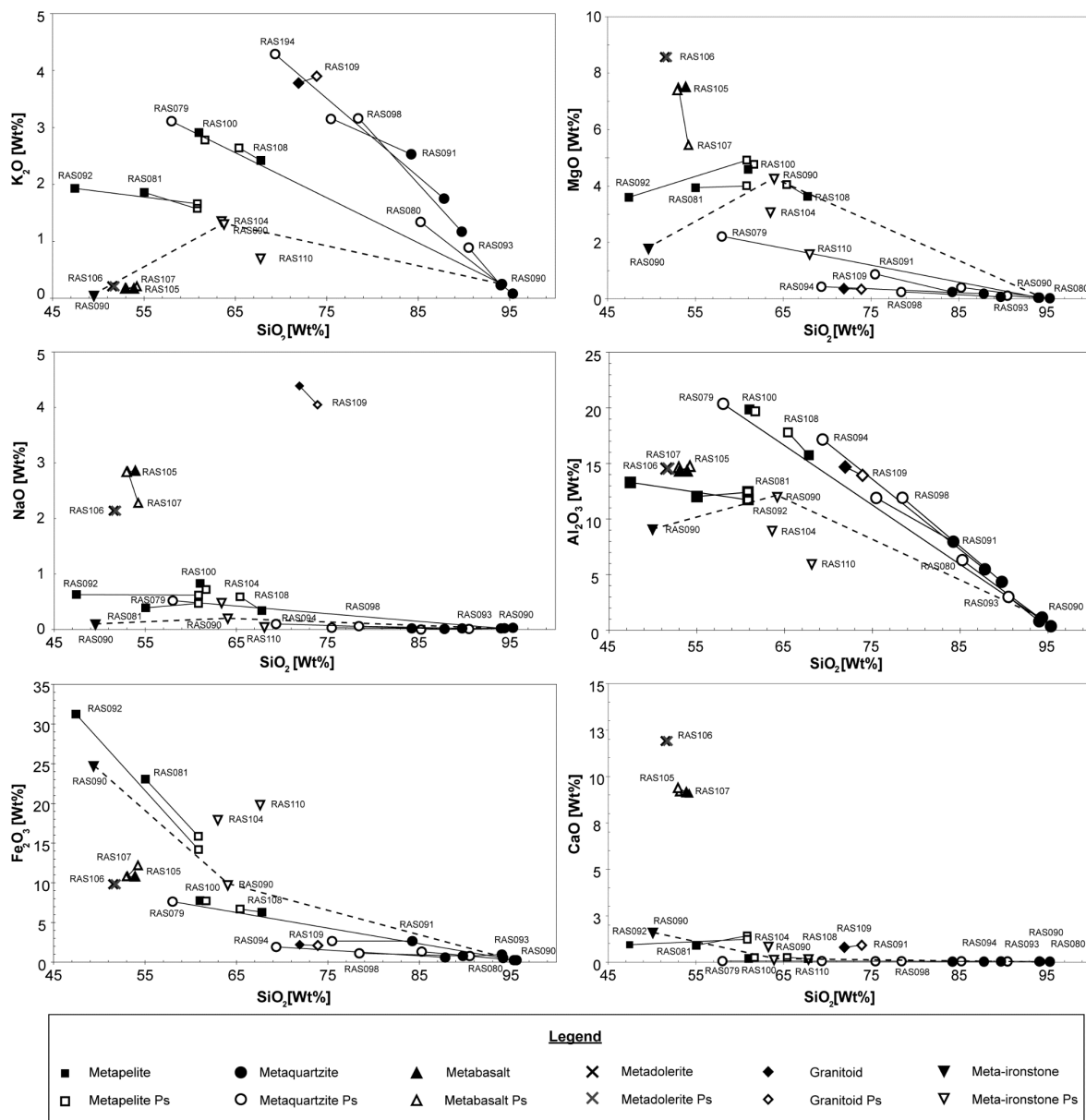


Figure 0.15: Harker plots of pseudotachylyte and host rocks in and adjacent to the Kopjeskraal Fault System. Open symbols are pseudotachylytes (Ps), closed symbols are wall rocks, and pairs are joined by tie lines.

The single granitoid gneiss sample pair shows a slight enrichment in SiO_2 and K_2O , and a depletion in MgO , Al_2O_3 and Na_2O . This slightly differs from the general trends observed by [Reimold \(1991\)](#) and [Lieger et al. \(2011\)](#) in granite-hosted pseudotachylytes by ~ 2 wt%. The SiO_2 enrichment, in particular, is counterintuitive, as the presence of abundant quartz clasts supports its resistance to melting, as predicted by experiments and theoretical considerations.

However, it may represent a nugget effect caused by quartz microclasts. Alternatively, the granitoid gneisses have a banded, migmatitic character ([Lana et al., 2004](#)), and so wall rock heterogeneity with small amounts of melt movement could be the cause.

The greatest discrepancy between dyke-wall rock pairs relates to the quartzite-hosted dykes. Quartzite-hosted pseudotachylytes show a significant depletion in SiO_2 (of up to nearly 40%) and enrichment in K_2O , Fe_2O_3 and Al_2O_3 (Table 0.1, Figure 0.15). Figure 0.13 shows that the quartzite does contain matrix phyllosilicate; however, the pseudotachylyte is clearly significantly enriched relative to the wall rock. Pseudotachylyte sample RAS-079 shows a strong correlation with the RAS-100 and RAS-108 metapelite and pseudotachylyte compositions. Although discernable exotic (pelitic) clasts were not observed in the pseudotachylytes the fact that pelitic and psammitic rocks are commonly intimately interlayered down to the cm scale, lends some plausibility to this option. However, an alternative - fluid-assisted introduction of these components, possibly from the adjacent metapelites - might also be a possibility. The significantly higher loss on ignition (LOI) in the quartzite-hosted dykes relative to the wall rock, and to the other rock pairs (Table 0.1) may support either scenario.

The metapelite-pseudotachylyte pairs show an unusual pattern in which host rock compositions show a wide scatter, and pseudotachylyte compositions tend towards intermediate values between these values (Table 0.1; Figure 0.15). We interpret this as a consequence of the mixing of melts of slightly different compositions derived from mm- to dm-scale layering in the metapelites, reflecting layered deposits of sand, silt and clay that have produced highly diverse mineral modes, together with small amounts of melt movement

The meta-ironstone pseudotachylytes comprise 60 - 70 wt % SiO_2 , ~ 20 wt % Fe_2O_3 and 5 - 10 wt % Al_2O_3 . Except for the metabasalt and metadolerite pseudotachylytes, all samples show substantial deviations from their wall rocks, indicating the introduction of an exotic melt, melt mixing or allochthonous melt formation as partial melting of local lithologies cannot account for the dramatic change in chemical composition.

Two pseudotachylytes – RAS-104 and RAS-110 – sampled within the Eastern Fault splay zone show elevated Fe:Mg and lower Al_2O_3 than the metapelites in which they are found. We conclude that these pseudotachylytes were derived from the melting of meta-ironstone and are thus shown as metaironstone symbols in Figure 0.15. Unfortunately, no meta-ironstone-hosted dykes were found along the Eastern Fault. However, a thin pseudotachylyte at the contact

between the Orange Grove Formation quartzite and Parktown Formation meta-ironstone was sampled, with its respective wall rocks, east of the Eastern Fault (Figure 0.2). Sample RAS090- H_{Fe} , H_{Qtz} and PT (Table 0.1) indicate melt mixing in the fault zone. The pseudotachylyte has an intermediate composition between the quartzite and meta-ironstone wall rocks for most oxides. However, it is enriched in K_2O , MgO and Al_2O_3 and has an elevated LOI relative to both wall rocks. Scattered boulders of thin, poorly exposed, metapelite occur along strike to the west of the sample locality between the quartzite and meta-ironstone, and may provide an explanation for this deviation. Alternatively, hydrothermal fluid movement may have modified the pseudotachylyte composition.

In summary, the pseudotachylyte dykes display a wide range of compositions that are broadly correlated with their respective wall rocks, supporting a local melting origin. No significant differences seen in the data require melt movement over more than a few metres to, at most, a few tens of metres. Although macro- and microscopic analysis, including elemental mapping, suggest internal compositional variation in some pseudotachylytes that might be interpreted as mixing of melts, TIMA element mapping (e.g., Figure 0.13) does not suggest significant compositional differences. Furthermore, the small-scale heterogeneity in wall rock compositions, in particular, in the metapelites and meta-ironstones, means that these heterogeneities might result from melts derived millimetres or centimetres apart. However, a high-spatial-resolution analysis of some of the polymictic pseudotachylyte boulders along the fault traces might be a useful further avenue of study to address the question of length scales of melt migration in the fault zones.

6.5 Discussion

6.5.1 Fault Kinematic Analysis in the Vredefort Dome

Previous work

While the 3D orientations of some faults in the annular trough (Potchefstroom Syncline) and outer parts of the central uplift (Vredefort Dome) have been obtained from a combination of field mapping (e.g., [Simpson, 1978](#)) and interpretation of borehole intersections and geophysical methods (e.g. [Henkel and Reimold, 1998](#); [Brink et al., 1997](#); [Ellis, 2005](#); [Grieve et al., 2008](#)), structural analysis of the faults identified by [Nel \(1927\)](#) and [Bisschoff \(2000\)](#) in the inner collar of the dome has generally been limited to their strike orientations. [Lilly \(1978\)](#) noted fault strike orientations in the WRG that were either sub-radial or oblique to the core-

collar contact but proposed that each type dominated in a different sector of the collar (N and NW, respectively). By proposing a vertical orientation and strike-slip movement sense for all faults, he concluded that the faults formed under fundamentally different stress orientations, which would imply different timing. However, as shown in [Chapter 3.3.2](#), there is no definitive sector variation in fault orientations.

[Wieland et al. \(2005\)](#) also assumed subvertical orientations and dominant strike-slip displacements for radially striking faults in the inner collar, but linked them to km-scale folds that they attributed to constrictional stresses (tangential σ_1) during the early stages of central uplift rise (Figure 0.1C). Based on 3D modelling, [Jahn and Riller \(2009\)](#) subdivided faults in the inner collar into an early set of rotated crater wall collapse faults with concentric strikes and a later set with radial – but locally strongly curvilinear – strikes and either normal or reverse displacement (Figure 0.1 B, C). They attributed the younger set to the spreading of the central uplift towards the end of the crater modification stage. Similar to [Grieve et al. \(2008\)](#), they proposed that the early set comprised initially low-angle, centripetally-dipping, normal faults formed during the initial collapse of the Vredefort transient cavity walls that were rotated to centrifugal to steep orientations during the rise of the central uplift, thereby manifesting as centripetally-directed thrusts. In contrast to the concentric faults, [Jahn and Riller \(2009\)](#) regarded the faults in their later set as still largely in their original orientation. Their study provided no field-based fault-related orientation or slip lineation data. [Brink et al. \(1997\)](#) and [Mashabela \(2016\)](#) demonstrated that the large fault system attributed by [Jahn and Riller \(2009\)](#) to the impact is an asymmetric Ventersdorp-age listric fault system with up to 3 km of vertical displacement, that lacks pseudotachylyte.

[Manzi et al. \(2021\)](#) identified subvertical pseudotachylyte-bearing faults with broadly tangential strike and centre-down displacement in the western collar, as well as a subradial conjugate fault pair in the NNE collar, both of which cut folds that they interpreted as related to late-modification stage collapse. Consequently, they proposed that both these fault sets formed by concomitant radial and tangential extension of the inner collar rocks related to the collapse of the central uplift (later stages of Modification Stage 3). The Vredefort granophyre dykes (impact melt rock) are interpreted to form even later than the faults, as km-scale Mode I extensional fractures that show no offset of wall rocks ([Wieland et al., 2005](#); [Huber et al., 2020](#)).

Kinematic Analysis of the Kopjeskraal Fault System

The moderate to moderately steep dips of the Eastern and Western Faults at Kopjeskraal contrast with the subvertical dips generally assumed for faults in the collar of the dome by previous workers (e.g., [Nel, 1927](#); [Lilly, 1978](#); [Bisschoff, 2000](#); [Wieland et al., 2005](#); [Jahn and Riller, 2009](#)). [Lilly \(1978\)](#) recognised only the Eastern Fault and inferred a sinistral strike-slip motion; on this basis, he inferred a subhorizontal NW-SE trend for the σ_1 stress direction, which constitutes a 15° anticlockwise deviation from the true radial trend relative to the centre of the dome. In recognising the equally large slip Western Fault with its right-lateral separation and its similar association with large volumes of pseudotachylyte, it is worth testing whether both faults could have formed simultaneously as a conjugate set. Figure 0.16 shows that, although the acute angle between them is slightly smaller (43°) than the predicted Andersonian angle of 60° , in all other respects, they conform to a conjugate system with a shallow NNW-plunging (radial) σ_1 stress direction ($344^\circ/24^\circ$) and shallow E-plunging (roughly tangential) σ_2 orientation ($087^\circ/25^\circ$). In this geometry, the slip lineation on the Eastern Fault is close to true strike-slip ($354^\circ/05^\circ$); however, the lineation on the Western Fault is highly oblique ($330^\circ/43^\circ$). This indicates respective sinistral strike-slip and reverse-dextral oblique-slip movements for the faults. In such a scenario, the apparent offset on the Western Fault represents only the strike component of the true slip, which would, therefore, be ~ 400 m.

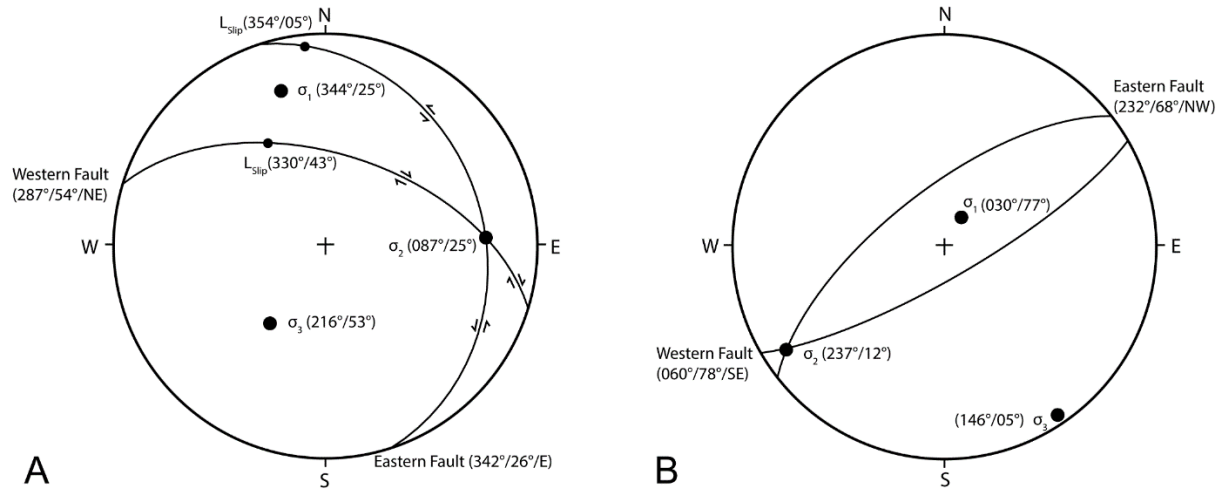


Figure 0.16: A) Equal area stereonet displaying the Eastern Fault (342°/26°/E) and Western Fault (287°/54°/NE) in their current orientations. Stress axis orientations (σ_1 - 344°/24°; σ_2 - 087°/25°; σ_3 - 216°/53°) and the slip lineations on the faults (Eastern - 354°/05°; Western - 330°/43°) are deduced assuming a conjugate fault pair. Although the dip angle of the large pseudotachylyte dyke in the granitoid gneisses could not be reliably determined, the dyke strikes approximately parallel to the acute bisector between the faults (306°/37°/NE), defining the σ_1 - σ_2 plane. B) Equal area stereonet displaying the backrotated orientations of the Eastern (232°/68°/NW) and Western (060°/78°/SE) faults. Stress axis orientations (σ_1 - 030°/77°; σ_2 - 237°/12°; σ_3 - 146°/05°). This assumes they formed prior to the rotation of the strata. The centre of the dome lies on a radial vector of 150° from the study area, which is within the error of the predicted σ_3 stress axis direction.

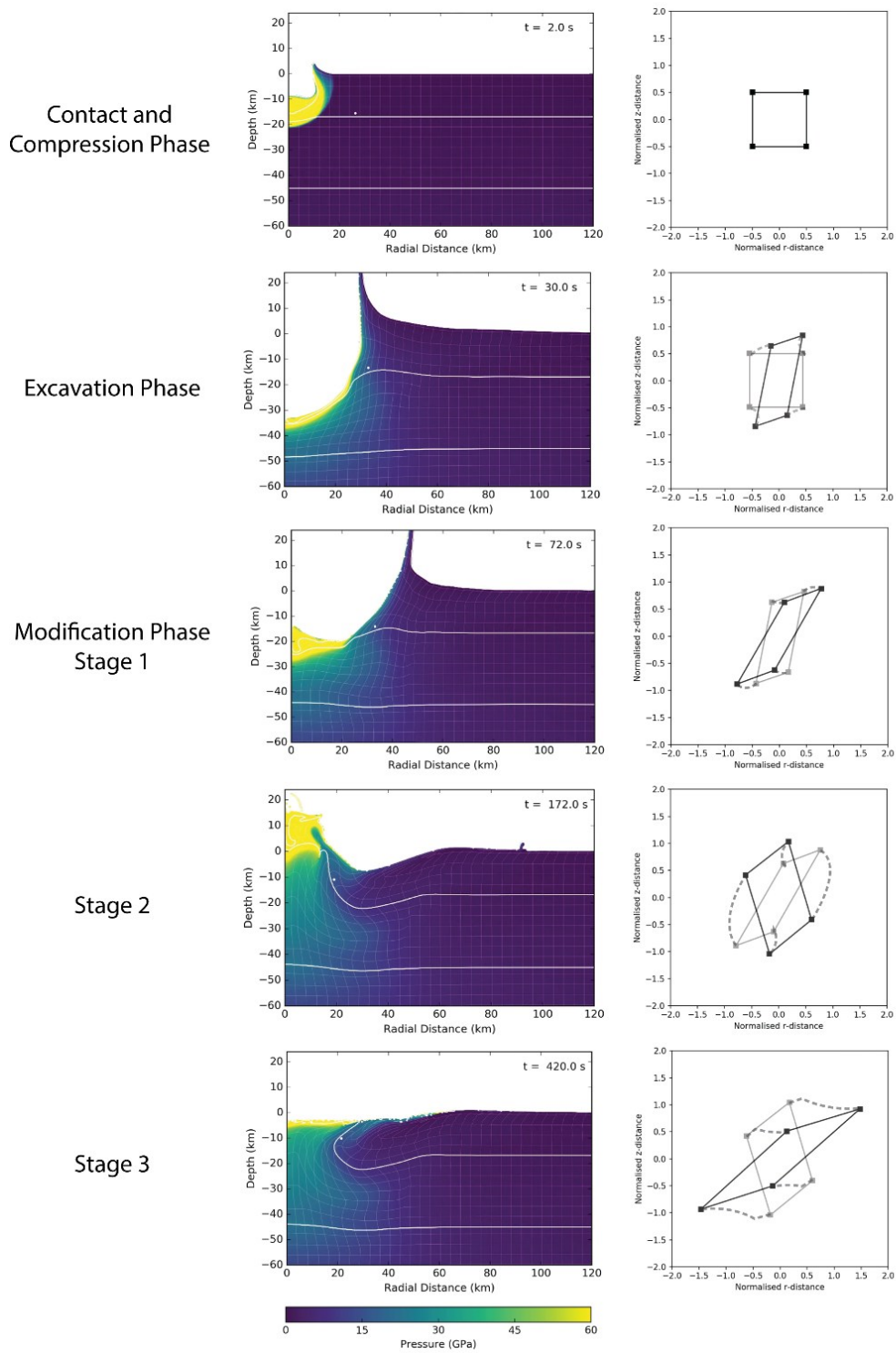


Figure 0.17: A 5-stage iSALE model that shows the crater morphology, shock pressure distribution and position of the inner collar rock volume at the end of each significant phase/stage in the Vredefort cratering process is shown on the left. The white tracer particle depicts the location of the 200 m x 200m lagrangian pseudo-cell. The thick white lines mark the boundaries between the granitic basement and the mantle at 45 km depth (lower) and the 17 km deep crystalline metavolcanic-granite contact (upper). On the right is the shape of the lagrangian pseudo-cell arrangement (shown in black) in which r and z are dimensionless distances normalised to 0 and 1. The grey dashed lines show the pseudo-cell path from the previous phase/stage (grey solid lines).

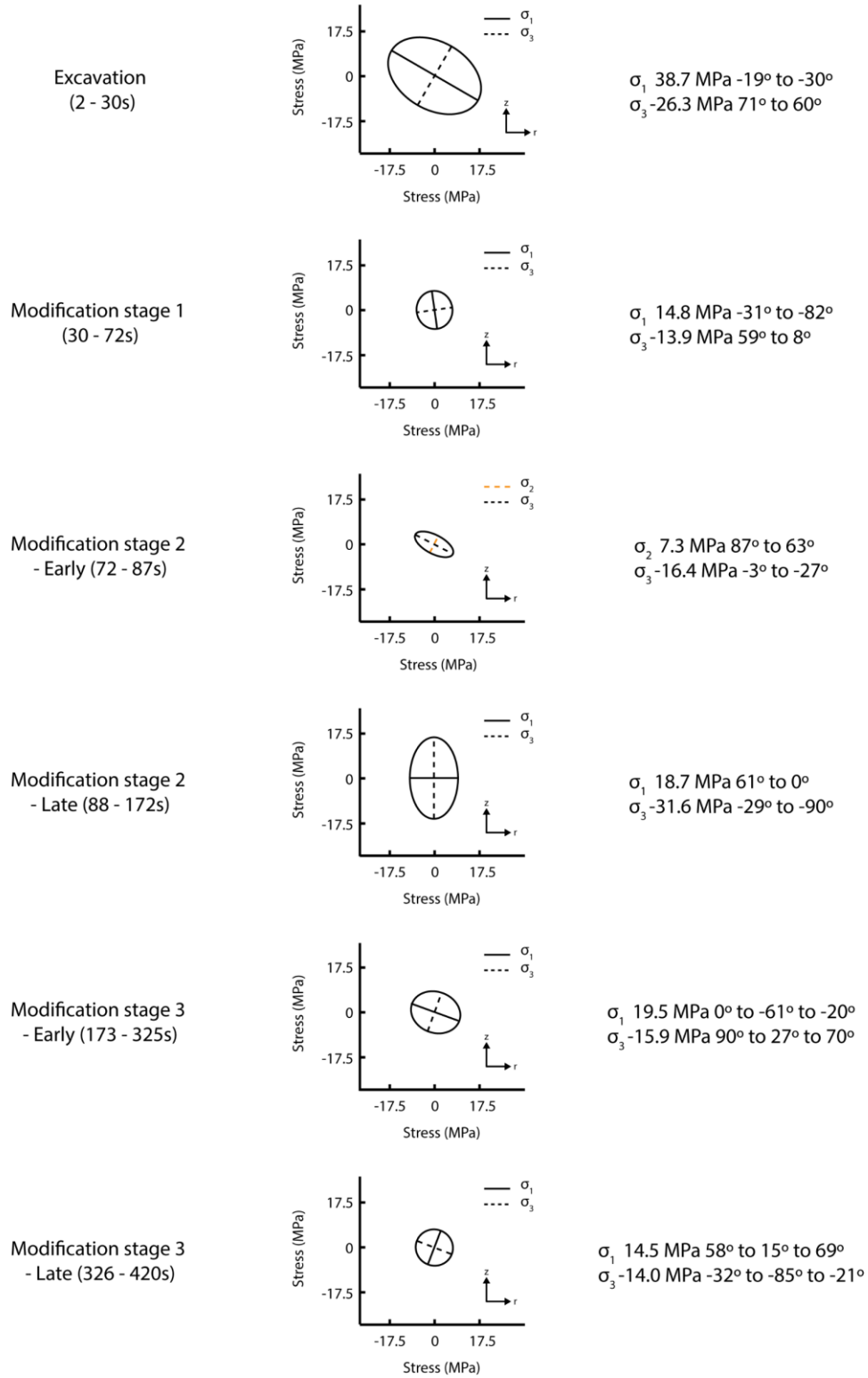


Figure 0.18: Sectional deviatoric Lamé's stress ellipses through the phases of cratering. The image displays the orientation at the end of each time step. On the right are the stress values

for the two stress axes oriented within the 2D space and the angles through which they rotate in the given time step.

The orientations of the principal stress axes do not correspond to one of the three general scenarios identified by [Anderson \(1951\)](#). Although this might not be unusual in the context of an impact cratering event, iSALE-2D numerical modelling can be used to investigate the stress and strain evolution of a rock volume corresponding to the Kopjeskraal area during crater formation and, thus, the most likely time for fault formation. Figure 0.17 and Figure 0.18 show that Modification Stages 2 and 3 involve significant rotation of the stress and strain axes, with σ_1 rotating from subhorizontal to subvertical during the final collapse of the central uplift. Evidence for Late Modification Stage 3 brittle strain, in the form of tangentially-striking normal faults and Mode I tangential and radial megafractures into which the impact melt dykes were emplaced ([Huber et al., 2020](#); [Manzi et al., 2021](#)) is consistent with an oblate strain pattern involving a vertical σ_1 and subhorizontal $\sigma_2 \approx \sigma_3$. However, as shown in Figure 0.2, the impact melt dyke cuts both faults, indicating that the faults developed under an earlier stress condition. A subvertical σ_2 and radial σ_1 are not compatible with the terminal stages of central uplift collapse. Any scenario that envisages the faults forming earlier in the crater evolution must include a component of body rotation (Figure 0.17). Figure 0.16B shows the result of rotating the faults by 120° about the strike of the collar strata, which restores the overturned WRG bedding to horizontal. This rotation produces a quasi-Andersonian normal dip-slip geometry for both faults, with a subvertical σ_1 stress orientation of $030^\circ/77^\circ$, a radial σ_3 stress orientation $146^\circ/05^\circ$ and a tangential intermediate stress axis orientation of $237^\circ/12^\circ$. This stress scenario closely resembles that predicted by the modelling during the latter parts of Modification Stage 1 (65-69 s; Figure 0.18). Both rotated faults show near-tangential strikes which, together with their opposing senses of normal dip-slip, provide compelling evidence that they formed during extensional collapse of the TC wall and that their present orientation reflects subsequent body rotation during the rise and ensuing collapse of the central uplift (Figure 0.17). The rather small 2θ angle of 43° between the Kopjeskraal faults reflects their steep initial dip angles; such a geometry is consistent with Mode III-type shear faulting, with a dilational component, which could be triggered by proximity to the free surface defined by the transient crater wall.

Fault slip magnitude correlates with fault length and spacing, which are, in turn, correlated with strain rate ([Melosh, 1989](#)). Under these conditions, Figure 0.19 shows that initial fault spacing (and therefore length and slip magnitude) would have been small during the contact and

compression and excavation phases of cratering. It is likely to have reached a maximum in the transition from Modification Stage 1 to Modification Stage 2, shortly after the strain rate reached its lowest value and began rising again (Figure 0.19). This period is also significant in two other important respects: first, the Kopjeskraal rock volume achieved its maximum displaced radial distance of 33.9 km at 59 s (Figure 0.20A), and simultaneously, the confining pressure reached its lowest value (Figure 0.20B), creating the most favourable conditions for fault nucleation.

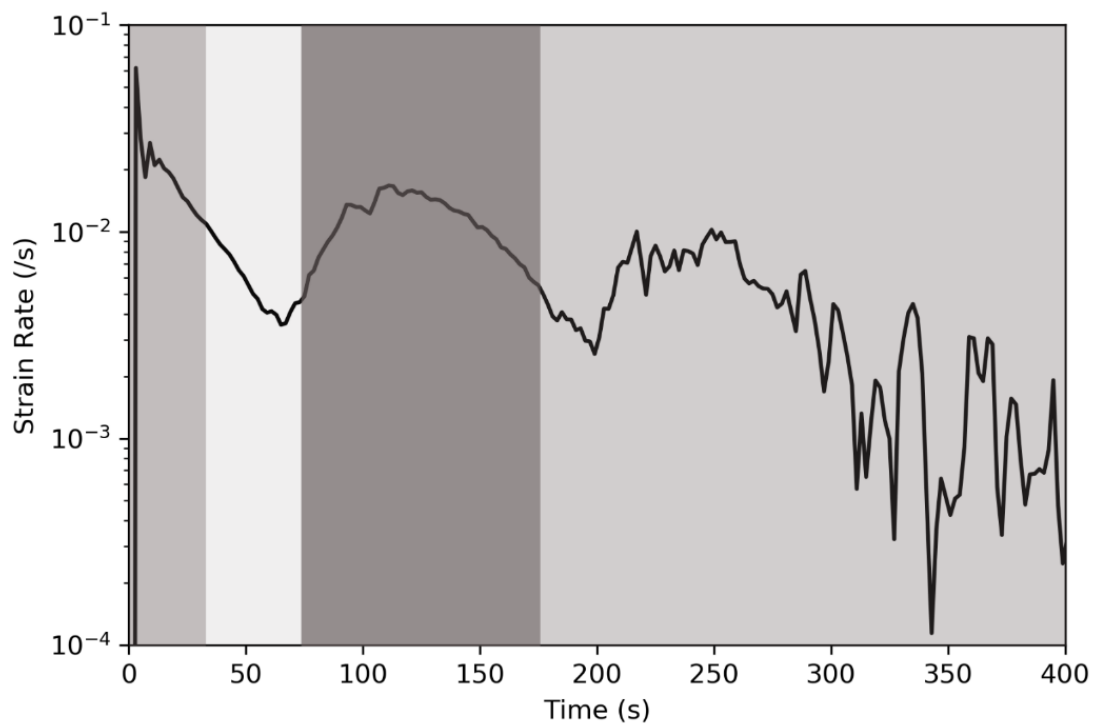


Figure 0.19: Graph of time vs strain rate during impact crater formation for the lower West Rand Group lagrangian unit cell. Overlain grey zones, from left to right: represent the Contact and Compression and Excavation phases (mid-grey), Modification Stage 1 (light grey), Modification Stage 2 (dark grey), and Modification Stage 3 (mid-grey). The strain rate decreases from a peak in the first two stages into Modification Stage 1. During Modification Stage 2, the strain rate increases, but approaches zero by the end of the stage as the central uplift reaches its peak and becomes gravitationally unstable. During Modification Stage 3, the strain rate oscillates as the crater material comes to rest and rotation halts.

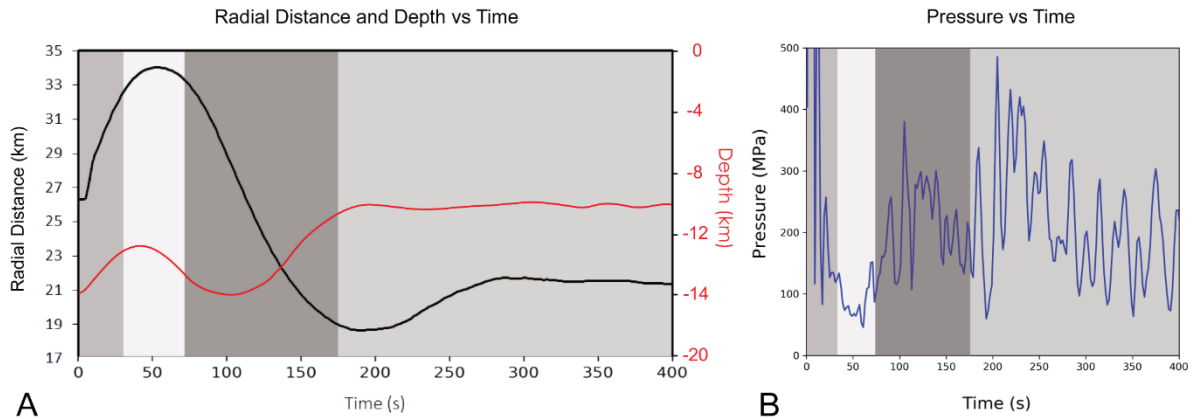


Figure 0.20: A) Graph showing the radial distance (black curve, left side axis) and depth (red curve, right side axis) of the lower West Rand Group lagrangian unit cell with respect to the centre of the dome or pre-impact surface, versus time. The unit cell originates at a radial distance of 26.3 km at $t = 0$ s. It reaches its maximum radial distance of 33.9 km at $t = 59$ s and a minimum radial distance of 19 km at $t = 193$ s before settling into its final radial distance of 21.4 km at 269 s. The unit cell rises from its original location of 15.5 km depth to 12.7 km at $t = 50$ s before dropping again to 15.9 km below the surface at $t = 103$ s. It is then exhumed to its minimum depth of 10 km below the surface at $t = 193$ s. B) Graph showing the pressure experienced by the lower West Rand Group lagrangian unit cell with respect to time. Overlain grey zones in both figures, from left to right, represent the Contact and Compression and Excavation phases (mid-grey), Modification Stage 1 (light grey), Modification Stage 2 (dark grey), and Modification Stage 3 (mid-grey).

As mentioned in the above section, the rotation of early-formed crater wall collapse faults into new orientations by the developing central uplift has been proposed for the Vredefort impact structure by [Grieve et al. \(2008\)](#) and [Jahn and Riller \(2009\)](#), however, neither study was able to show direct field evidence of such faults. In both cases these faults were portrayed as listric, intersecting bedding at a low angle (Figure 0.1B,C). Conversely, radial faults were portrayed as vertical and having formed under a different, younger, set of stress conditions. By unequivocally constraining the true 3D orientation of the Kopjeskraal faults our study illustrates the pitfalls of this interpretation. For instance, assuming a vertical dip for a radial fault means that even 120° of rotation around the bedding strike axis would not materially alter its orientation, although it would significantly rotate its slip vector. Thus, while both [Wieland et al. \(2005\)](#) and [Jahn and Riller \(2009\)](#) inferred present-day strike-slip displacements on radial faults, the former proposed that the radial faults formed as high-angle reverse faults during the constriction stage (Modification Stage 2, Figure 0.17) whereas the latter inferred late-stage dip-slip movements (Modification Stage 3, Figure 0.17). Not only do we propose an earlier timing for the Kopjeskraal faults than either of these studies, but we also demonstrate that present-day quasi-radially striking faults with non-vertical dips can be consistent with an original tangential

strike orientation. This does not mean that all radial faults in the dome formed at the same time and in the same way as the Kopjeskraal faults - it is inevitable that collapse of the crater wall would have occurred diachronously, younging radially outwards. Moreover, [Simpson \(1978\)](#) and [Manzi et al. \(2021\)](#) demonstrated faults cutting folds that formed, respectively, during formation of the rim syncline (Modification Stage 2) and collapse of the central uplift (Modification Stage 3).

Notwithstanding the arguments presented above, it is important to recognise that the Kopjeskraal faults display slips of ~400-500 m, which is 1-2 orders of magnitude smaller than the total displacement of the bulk Kopjeskraal rock volume during the cratering process. Modelling indicates that it achieved its farthest radial position of 33.9 km at $t = 59$ s (Figure 0.20A) – immediately prior to the formation of the faults, and that it experienced 15 km of centripetal, and 3 km of vertical displacement to reach its final position (Figure 0.20A: see also [Jahn and Riller, 2015](#)). There is, thus, potential for the development of low-angle faults with km-scale slip as inferred by [Grieve et al. \(2008\)](#); however, it must be recognised that only part of this displacement would be accommodated by faults, as it is primarily a bulk body motion directed towards a free surface.

Identifying faults with orientations close to bedding (as happens with the western end of the Western Fault; Figure 0.2) is particularly challenging, but one possible example in the Kopjeskraal area may be provided by the thin but persistent, ~2 km long, pseudotachylyte dyke in the Brixton Formation (Figure 0.2A). The dyke shows exceptional laminar flow banding and it exhibits a listric relationship with respect to bedding that, if subjected to the full 120° backrotation, generates a shallow centripetal dip. Although no kinematic indicators have been found, it would be a potential candidate for the listric normal faults portrayed by [Grieve et al. \(2008\)](#) and [Jahn and Riller \(2009\)](#). An alternative origin, related to flexural slip during central uplift rise (Modification Stage 2) (e.g., [Dressler and Reimold, 2004](#); [Manzi et al., 2021](#)) cannot be ruled out, but [Mashabela \(2016\)](#) proposed that it was truncated by the radial faults, which would favour a listric collapse origin during Modification Stage 1.

Within the context of the above kinematic interpretation, the large westward-tapering pseudotachylyte dyke in the gneisses south of the Western Fault, which locally appears to show evidence of a moderately steep NE dip, could be interpreted as filling a dilational Mode I megafault related to the Eastern Fault. Its present orientation does not correspond to either a tangential or radial orientation, such as proposed by [Lieger et al. \(2009\)](#) for other large dykes in the gneisses.

Unlike [Wieland et al.'s \(2005\)](#) interpretation of faults in the inner collar, based on iSALE modelling, the Kopjeskraal faults appear to predate the constrictional phase of central uplift formation. Surprisingly strong rotations accompanying the vertical uplift (Modification Stage 2) and collapse (Modification Stage 3) do not significantly disrupt the kinematic indicators. [Jahn and Riller \(2009\)](#) suggested concentric faults in the collar could be rotated listric transient crater fall extensional collapse faults now showing centripetal thrust slip. The Western Fault described here is approximately concentric, but it is noteworthy that our reconstruction shows that the hanging wall collapsed radially outwards.

6.5.2 Origin of Large-Volume Pseudotachylyte

As outlined in [Section 6.1](#), various mechanisms have been proposed to explain the origin of melt breccia dykes in the Vredefort Dome and other large impact structures. Recently, [Spray and Biren \(2021\)](#) have documented the differences between shock melt veins, intrusive dyke offshoots from the impact melt sheet, and smaller- volume fault-related pseudotachylyte in the crater basement rocks in the central parts of the Manicouagan impact structure. [Lambert \(1981\)](#), [Martini \(1991\)](#) and [Spray \(1992\)](#) proposed that impact structures contain two types of “pseudotachylyte” – one (A- or S-type) that crystallised under shock compression that is similar to the shock melt veins found in meteorites and that involves small melt volumes, and a second type (B- or E-type) that forms bodies with volumes that are up to several orders of magnitude greater. Because of the generally accepted link between pseudotachylyte and friction melting by the end of last century, [Reimold \(1995, 1998\)](#) cautioned against it being applied in impact structures for two main reasons: first, the attainment of shock pressures above 60 GPa will produce melt without the need for friction, so doubt was created about whether the formation of A/S-type veins involved friction melting, and convincing evidence was lacking of faults showing sufficient slip magnitude to account for B/E-type pseudotachylyte dykes in impact structures by friction melting. Recently, [Spray and Boonsue \(2018\)](#) have argued that the coesite-stishovite-bearing melt veins in the collar of the Vredefort Dome are friction melts that formed by mm-scale slip on fault planes during the rise of the shock pressure wave, and that crystallisation over the next 1-2 seconds happened under elevated shock pressure conditions. In Manicouagan, [Spray and Biren \(2021\)](#) have demonstrated the spatial, geometric and kinematic relationships between 1-10 cm wide pseudotachylyte dykes and faults showing slip amounts up to a few decimetres, consistent with the scaling laws described by [Melosh \(2005\)](#). These pseudotachylytes cut the earlier S-type pseudotachylyte veins. They also identified a few examples of E-type pseudotachylyte dykes up to 1 m wide associated with faults with slips of

10^1 - 10^2 m, again favouring friction melting. [Spray and Thompson \(1995\)](#) proposed that highly voluminous pseudotachylyte dykes in the Sudbury structure formed part of a large concentric fault related to the excavation phase of cratering; however, unequivocal spatial, geometric and kinematic links between large-volume pseudotachylyte dykes and large-slip faults has remained elusive.

Our observations differ from those of [Dressler and Reimold \(2004\)](#) and [Lieger et al. \(2009\)](#), which respectively propose either a random orientation or a combined tangential and radial preferred orientation pattern for dykes in the Vredefort Dome, with the consequence that the former favour a “flash melting” origin for the dykes while the latter proposes significant melt mobility that obscures its origin. The idea that the large dykes in the dome represent decompression melts formed during rapid uplift ([Mohr-Westheide and Reimold, 2011](#)) runs into difficulties from both a kinetic and timing perspective.

Despite the absence of exposure of the actual fault traces in the Kopjeskraal area, the offset of lithological contacts provides excellent constraints on both their location and magnitude of slip; furthermore, flanking outcrops, especially in the metabasalt and gneisses along the Western Fault and in the vicinity of the Brixton Formation quartzite along the Eastern Fault, preserve key spatial, geometric and kinematic evidence that ties pseudotachylyte formation to the respective faults. These include the characteristic geometries (generation plane and wedge-shaped vein offshoots, network breccia relay domains or linkage duplexes between offset or overlapping faults) described by [Sibson \(1977\)](#) from tectonic fault zones.

Although most pseudotachylyte dykes show evidence of a matrix that is metamorphically recrystallised owing to the elevated post-shock temperatures ([Ivanov, 2005](#); see [Chapter 2.2](#)), rare spherulitic-textured plagioclase microlites in a metabasalt-hosted dyke and flow-aligned microlites in the large dyke in the granitoid gneisses south of the Western Fault indicate a precursor melt phase. Geochemical analysis of the dykes confirms local melt derivation from wall rocks, as does the evidence of wall rock-dominated clast populations. The chemical compositions and mineral clast populations of the dykes are consistent with the preferential melting of mechanically weak ferromagnesian minerals (amphibole, biotite and/or cordierite) and feldspar relative to stronger minerals such as quartz. However, simple tie-ins to wall rocks are difficult owing to melt movement and hydrothermal overprint. The geochemical evidence thus fits the field, and petrographic evidence that shearing, cataclasis, and comminution in the wall rocks drove friction-induced selective mineral melting, as proposed by [Spray \(1992; 1997\)](#).

Scaling calculations (e.g., [Melosh, 2005](#)) indicate that faults with hundreds of metres of slip are capable of producing pseudotachylyte dykes with individual widths of metres to tens of metres and lengths of hundreds of metres. [Lieger et al. \(2009\)](#) suggested that pseudotachylyte dykes in the core of the dome are allochthonous and were emplaced late during cratering associated with central uplift collapse (corresponding to Modification Stage 3) into radial and tangential tensional/dilational zones. At 550 m long and up to 17 m wide, the largest dyke at Kopjeskraal is of comparable size to the largest dykes described by [Dressler and Reimold \(2004\)](#) and [Lieger et al. \(2011\)](#). Although its exact orientation and spatial relationship to the faults are not visible, it widens towards the Eastern Fault. It has a subvertical flow foliation that suggests a steep overall orientation. It does not fit the radial or tangential orientations that [Lieger et al. \(2011\)](#) observed, which underpin their late-stage emplacement model. This geometry is not inconsistent with that expected for a wedge-shaped offshoot from a sinistral strike-slip fault.

6.5.3 Implications

Complex crater modification and particularly central uplift formation are the least well-understood aspects of impact crater formation. While oversteepened transient crater walls are typically envisaged as slumping inwards along listric normal faults in 2D section, this does not account for crater basement heterogeneity or the tangential space problem (e.g. Radial transpression ridges, etc. ([Kenkmann et al., 2014](#)).

The superfaults from Sudbury described by [Spray \(1997\)](#) and faults described by [Killick \(1992\)](#) from the goldfields are all interpreted as involving centripetally-directed normal-slip and contain abundant pseudotachylyte. The Kopjeskraal Fault System conforms with large-volume pseudotachylyte genesis linked to crater wall collapse, but the fault pairs are quite symmetric. To understand this, we need to consider crater dimensions – large low-angle asymmetric listric faults with multi-kilometre displacements, as envisaged by [Grieve et al. \(2008\)](#) and [Jahn and Riller \(2015\)](#) may exist, but the Kopjeskraal Fault System involves faults with 100s m of displacement – these may thus be secondary faults within a larger unidirectional system. Evidence of a first-order listric fault may be in the Brixton Formation; unfortunately, it is impossible to determine kinematics on that fault, and the alternative possibility of it being related to flexural slip during the central uplift rise cannot be ruled out.

First-order scale hydrodynamic behaviour of the central uplift rock volume is envisaged to occur by more chaotic, independently moving, blocks separated by faults with frictional

resistance reduced by shock or seismic vibrations ([Ivanov and Kostuchenko, 1997](#); [Garde and Klausen, 2016](#)). [Senft and Stewart \(2009\)](#) propose that large-slip events favour dynamic weakening. However, there is little to suggest that the early-formed faults were reactivated by subsequent movements of the central uplift during the ensuing several minutes. One explanation could be that early melt expulsion and quenching led to faults locking up, and future slip occurs on new faults. [Reches \(1978\)](#) suggested 3 different orientations are needed to allow adjustments. This would explain faults associated with constrictional folds, and late-stage normal faults cutting collapse-related folds. The presence of significant pseudotachylyte volumes in these younger Modification Stage 3 faults may imply that pseudotachylyte was generated throughout the modification phase; alternatively, as extensional features, they may have been drainage sites for still-molten pseudotachylyte melts from older, larger-slip faults like the Kopjeskraal Fault System. This would be similar to the mechanism proposed to explain the impact-melt dykes, which also occupy late extensional sites. However, the latter melt must have travelled kilometres, and there is no evidence that the pseudotachylyte melts moved significant distances from the wall rocks from which they were generated. Future work on large boulders along fault traces may be useful to look for mixing.

What our study shows, is that more careful consideration of the 3D geometry of faults in the Vredefort Dome is essential. The study suggests that sense can be made of impact fault kinematics even in the highly complex central uplifts. The preservation of early-formed fault slip evidence despite significant subsequent body rotation and translation with further slip movements is surprising. The lack of reactivation may be a consequence of the deep levels below the crater surface. In contrast at shallower levels, proximity to the free surface might facilitate more chaotic block slips and rotations during multiple stress changes. The more constrained deep environment might “lock up” early formed faults.

If large volumes of pseudotachylyte in the collar and core are generated in Modification Stage 1 in association with faulting they could form an excellent lubricant to facilitate the ensuing block slip and rotations needed to form the central uplift in Modification Stage 2. Also, pseudotachylyte genesis in the collapsing crater may initiate diachronously, becoming younger (by a few 10s of seconds) outwards.

6.6 Conclusions

Although faulting and pseudotachylyte formation in impacts have for a long time been considered chaotic and unordered, this study suggests that the Kopjeskraal Fault System forms

at a discrete time interval during Modification Stage 1, between 65-69 seconds, of the Vredefort impact event. We show that stress fields, even in impact settings can be discerned from traditional structural approaches and that pseudotachylyte formation occurs as a result of friction along the fault planes. Although closer to the crater floor, the proximity to the free surface may allow far more unconstrained and complex block movements. In the Vredefort Dome at its deep erosional level, applying fundamental structural geology techniques to central uplifts may definitively constrain the stresses that lead to fault formation and their associated pseudotachylytes. Although our study can not completely rule out other mechanisms, it makes a compelling case for a non-shock, frictional, origin for voluminous and extensive pseudotachylyte dykes in central uplifts. In this study, it may, thus, be named pseudotachylyte, *sensu stricto*. Unlike [Wieland et al.'s \(2005\)](#) interpretation of faults in the inner collar, based on iSALE modelling, these faults appear to predate the constrictional phase of central uplift formation. Surprisingly, the strong rotations accompanying the vertical uplift (Modification Stage 2) and collapse (Modification Stage 3) do not appear to significantly overprint the kinematic indicators. [Jahn and Riller \(2009\)](#) suggested concentric faults in the collar could be rotated listric transient crater fall extensional collapse faults now showing centripetal thrust slip. The Western Fault described here is approximately concentric, but it is noteworthy that our reconstruction shows that the hanging wall collapsed radially outwards. Implications for models of pseudotachylyte genesis in the Vredefort Dome at the outcrop and microscale show all features as consistent with pseudotachylyte development by fault-slip-induced cataclasis, friction melting and emplacement in fault-controlled sites.

6.7 Acknowledgements

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Chapter 7

DISCUSSION AND CONCLUSIONS

7.1 Discussion and Conclusions

Numerical modelling successfully reconstructed the gross crater morphology and phases of impact cratering by partitioning large volumes of rocks into manageable coherent cells. It employs hydrodynamic codes, however, to achieve fluid movements in the rock volume when, in reality, the rocks behave by buckling and shattering. In order to get a coherent, realistic model that accurately represents the way that brittle rocks behave during impact, this study has attempted to compare and cross-validate field data and that provided by numerical modelling.

Many of the structural studies from the Vredefort Dome were carried out prior to the consensus that it was an impact event and were predominantly in the gold fields on the outskirts of the basin. The features were, therefore, not assessed with impact dynamics in mind. Numerical modelling has provided scientists with a fundamental understanding of how the stresses change with the evolution of the crater, richly enhancing our ability to categorise structures sequentially. This study aimed to reconstruct the sequence of impact-induced brittle and ductile structural changes in the lower WRG of the dome whilst investigating their geometric and spatial relationships, deformation mechanisms, stress and strains. Furthermore, both bedding and faults were used to assess (a)symmetry in the dome. More specifically, looking at links between shatter cones, and the shatter cleavage; folds, faults and pseudotachylite dykes.

Given that shock pressure and strain rate attenuate faster horizontally than vertically ([Rae et al., 2021](#)) perhaps the depth at which the Vredefort Dome is exposed offers unique insights into impact cratering. This study reported the first occurrence of tangential folds comprising disharmonic, open, asymmetric, horizontal to plunging-inclined anticline-syncline pairs with centripetally dipping axial planes and right-way-up intermediate limbs. They display centripetal-down vergence (anticline radially outward of the syncline) that is consistent with steep inward-directed shear of the overturned strata. They are attributed to the subvertical collapse of the Vredefort central uplift during Modification Stage 3 at $t = 200 - 230$ s. These folds differ from those previously reported by [Wieland et al. \(2005\)](#) which are subradially oriented, vertical to plunging-inclined, gentle folds, interpreted as the products of convergent flow (constriction) that would have developed during the initial stages of central uplift formation (Modification Stage 2) at $t = 72 - 87$ s.

The shatter cones, shatter cleavage, and rare horsetail fractures within the inner collar of the Vredefort Dome record the complexity of the shock wave and rarefaction wave interaction at $t = 3 - 9$ s. This study was able to confirm that a relationship between the shatter cleavage and horsetail fractures occurs; however, a more statistically significant study would make the association more robust if outcrops with better 3D exposure could be found. This study preliminarily proposes that the horsetail fractures branch off the CLA shatter cleavage and, thus, share a similar geometric relationship to the shatter cones. The shatter cones are preliminarily shown to be formed along the bisector of the shatter cleavages, although again, a dome-wide study is needed to strengthen the argument.

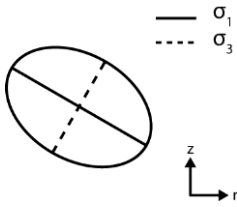
The dome-wide analysis of fault and bedding strikes as recorded on the 1:50,000 map of the Dome ([Bisschoff, 2000](#)) revealed no obvious bilateral symmetry that could indicate an oblique impact. An asymmetric Vredefort central uplift cannot be fully discounted as asymmetry likely decreases with increasing depth, and the Vredefort Dome is eroded by between 8 and 10 km. A further complicating factor might be the increased amounts of collapse associated with increased central uplift size, which may overprint or locally distort early-formed second-order features, and which this study shows to be the case (multiple ages of faults, late-stage folds) .

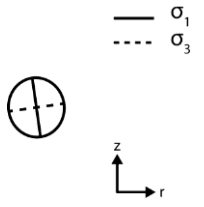
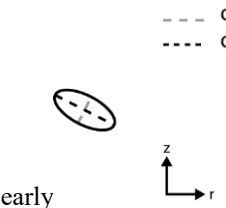
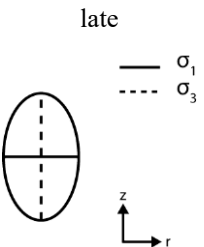
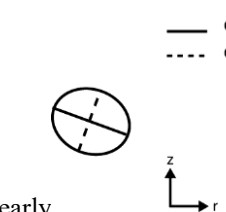
In contrast to pseudotachylytes in the granitic core of the Vredefort Dome, pseudotachylytes in the lower WRG in the inner collar are linked to faults displaying clear offsets of sufficiently large magnitude to account for the anomalously large melt volumes. A key finding of this study is the need for 3D, rather than strike-only, geometric analysis of faults, fractures, pseudotachylytes and bedding.

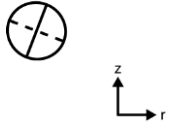
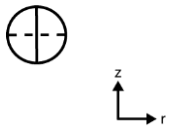
While acknowledging that strain is highly heterogenous on a variety of scales, this study has achieved success in placing what appear to be reasonable constraints on the relative timing of several macroscopic deformation features in the central uplift of a complex impact crater by comparing the predicted stress-strain history to that obtained from numerical modelling of relatively large unit cells. Part of this success may be attributed to deep levels at which this study area is preserved, investigations of less eroded craters may well experience difficulties due to the proximity of the rock volume to free surface. Further testing of this approach is recommended by investigating the structural histories of other regions of the Vredefort central uplift, located at different radial distances, and in different sectors of the Vredefort Dome.

Table 0.1 provides a summary of the sequence of impact-induced structural changes in the lower WRG in the Vredefort Dome.

Table 0.1: Summary of the features in the inner collar of the Vredefort Dome.* Indicates may also be formed at a later stage. A clockwise rotation is, in this case, centrifugal vergence.

Phase	Stress Ellipse at Time of Formation (r-z plane)	Initial Feature	Current Feature	Initial Stress Orientations	Association	Reported by
Pre-impact	-	Half graben structure, Listric faults. West-side down slip, NNE strike	Asymmetric listric fault system with up to 3 km of vertical displacement, east-west extension and north-south shortening	Vertical shortening and tangential east-west extension	Ventersdorp-age, lacks pseudotachylyte	Mashabela (2016)
Contact and Compression	-	Shatter cleavages/MSJS	Shatter cleavages/MSJS: rotated by 98° clockwise in the r-z plane	σ_1 52° inwards σ_2 38° outwards σ_3 is in the hoop direction	Shatter cones	Manton (1965) ; Nicolaysen and Reimold (1987) ; Albat and Mayer (1989)
		Shatter cones	Shatter cones: Rotated by 98° clockwise in the r-z plane		MSJS, Shatter cleavage	Dietz (1960) ; Hargraves (1961) ; Manton (1965)
		Shear Fractures*	Shear fractures: Rotated by 98° clockwise in the r-z plane		Pseudotachylyte	Wieland et al. (2005)
Excavation		-	-	σ_1 19° to 30° inwards σ_2 is in the hoop direction σ_3 84° to 60° outwards	-	-

Modification Stage 1		Low-angle, centripetally-dipping, concentrically striking, rotated normal crater wall collapse faults	Rotated to centrifugal to steep orientations during the rise of the central uplift. i.e. Brixton Formation Pseudotachylyte	σ_1 31° to 82° inwards σ_2 is in the hoop direction σ_3 59° to 8° outward	Pseudotachylyte	Grieve et al. (2008) ; Jahn and Riller (2009) ; Manzi (2025)
Modification Stage 2 Central peak entrainment	 early	F1: radial km-scale folds brittle-ductile	F1 folds: Rotated clockwise 85° in the r-z plane	σ_1 is in the hoop direction σ_2 87° to 63° outwards σ_3 3° to 27° inwards Constriction – tangential shortening	Brecciated hinges with pseudotachylytes	Wieland et al. (2005) ; Manzi et al. (2021)
	 late	Radially striking faults	Radially striking faults: rotated clockwise 46° in the r-z plane (reverse-steeply oblique-slip)	σ_1 61° to 0° outwards σ_2 is in the hoop direction σ_3 20° to 90° inwards Transpressional, constrictional stresses	km-scale folds	Lilly, 1978 ; Wieland et al. (2005) ; Jahn and Riller (2009)
Modification Stage 3 Central peak collapse	 early	F2: Tangentially-oriented asymmetric, disharmonic folds, anticline positioned radially outward of the syncline, inward-dipping axial planes and subhorizontal to moderately-plunging hinges	F2: Rotated 5° clockwise in the r-z plane	σ_1 0° to 61° to 20° inwards σ_2 is in the hoop direction σ_3 90° to 27° to 70° Outwards radial and tangential extension; oblate strain, vertical flattening	Tangential faults	Manzi et al. (2021)

	late — σ_1 - - - σ_3 	Subvertical faults with broadly tangential strike and centre-down displacement	Unchanged	σ_1 58° to 15° to 69° outwards σ_2 is in the hoop direction σ_3 32° to 85° to 21° inwards	Pseudotachylytes	Manzi et al. (2021)
Modification Stage 4 Relaxation	— σ_1 - - - σ_3 	Jointing	Unchanged	Radial and tangential extension	Faulting	Wieland et al. (2005)
		Faulting	Granophyre dykes: faults provided the opportunity for the downward intrusion of impact melt	oblate strain pattern involving a vertical σ_1 and subhorizontal $\sigma_2 \approx \sigma_3$. Mode I extension	Granophyre Dykes	Wieland et al. (2005) ; Huber et al., 2020

7.2 Recommendations for Future Work

Although the findings of this study have improved our understanding of impact mechanics, the following scientific questions remain:

- The mechanism of shatter cleavage formation.
- How shatter cones, shatter cleavage, and horsetail fractures vary along a radial transect of the dome.

In this regard, it is recommended that the shatter cleavage be assessed based on the latest numerical modelling data provided by [Rae et al. \(2021\)](#) that account for the change in orientation from radial based on the outcrop location from the point of impact. By doing a radial transect with the relevant rotations, this data can be compared and may yield better results.

REFERENCES

- Agangi, A., Hofmann, A., Hegner, E., Xie, H., Teschner, C., Slabunov, A. and Svetov, S.A., 2020, The Mesoarchaeon Dominion Group and the onset of intracontinental volcanism on the Kaapvaal craton—geological, geochemical and temporal constraints. *Gondwana Research*, 84, 131-150.
- Ahrens, T.J. and O'Keefe, J.D., 1972, Shock melting and vaporization of lunar rocks and minerals. *The Moon*, 4 (1), 214-249.
- Ahrens, T.J., O'Keefe, J.D. and Gibbons, R.V., 1973, Shock compression of a recrystallized anorthositic rock from Apollo 15. In *Lunar and Planetary Science Conference Proceedings*, Vol. 4, p. 2575.
- Al Barazi, S., Riller, U. and Hecht, L., 2009, Geochemistry of pseudotachylytes in target rocks of the Sudbury impact structure, Ontario, Canada: *Meteoritics and Planetary Science*, 44, A27.
- Albat, H.M. and Mayer, J.J., 1989, Megascopic planar shock fractures in the Vredefort Structure: a potential time marker? *Tectonophysics*, 162, 265-276.
- Alexandre, P., Andreoli, M.A.G., Jamison, A. and Gibson, R.L., 2006, $^{40}\text{Ar}/^{39}\text{Ar}$ age constraints on low-grade metamorphism and cleavage development in the Transvaal Supergroup (central Kaapvaal Craton, South Africa); implications for the tectonic setting of the Bushveld igneous complex. *South African Journal of Geology*, 109 (3), 393-410.
- Allen, N.H., Nakajima, M., Wünnemann, K., Helhoski, S. and Trail, D., 2022, A revision of the formation conditions of the Vredefort crater. *Journal of Geophysical Research: Planets*, 127(8), p.e2022JE007186.
- Alneasan, M., Behnia, M. and Bagherpour, R., 2020, The effect of Poisson's ratio on the creation of tensile branches around dynamic faults. *Journal of Structural Geology*, 131, p.103950.
- Amsden, A. A., Ruppel, H. M. and Hirt, C. W., 1980, SALE: A simplified ALE computer program for fluid flow at all speeds (LA-8095). New Mexico: Los Alamos National Laboratories.

- Anders, M.H. and Wiltschko, D.V., 1994, Microfracturing, paleostress and the growth of faults. *Journal of Structural Geology*, 16 (6), 795-815.
- Anderson, E.M., 1951, *The Dynamics of Faulting and Dyke Formation With Application to Britain*, 2nd ed., 206 pp., Oliver and Boyd, Edinburgh.
- Anderson, J.L. and Schultz, P.H., 2006, Flow-field centre migration during vertical and oblique impacts. *International Journal of Impact Engineering*, 33 (1-12), 35-44.
- Anderson, J.L., Schultz, P.H. and Heineck, J.T., 2003, Asymmetry of ejecta flow during oblique impacts using three-dimensional particle image velocimetry. *Journal of Geophysical Research: Planets*, 108 (E8).
- Anderson, J.L., Schultz, P.H. and Heineck, J.T., 2004, Experimental ejection angles for oblique impacts: Implications for the subsurface flow-field. *Meteoritics and Planetary Science*, 39 (2), 303-320.
- Andreoli, M.A.G. and Hart, R.J., 1987, Explosive KREEP-norite from Namaqualand, South Africa, with implications for the Sudbury Irruptive. In *Proceedings of the International Workshop on Cryptoexplosions and Catastrophes in the Geological Record*, Section A, Parys, South Africa, 1-8.
- Antoine, L.A.G., Nicolaysen, L.O. and Niccol, S.L., 1990, Processed and enhanced gravity and magnetic images over the Vredefort structure and their interpretation. *Tectonophysics*, 171 (1-4), 63-74.
- Armstrong, R.A., Compston, W., Retief, E.A., Williams, I.T. and Welke, H.J., 1991, Zircon ion microprobe studies bearing on the age and evolution of the Witwatersrand triad. *Precambrian Research*, 53, 243-266.
- Armstrong, R.A., Lana, C., Reimolds, W.U. and Gibson, R.L., 2006. *Geological Society of America Special Paper 405 2006. Processes on the Early Earth*, 405, p.233.
- Artemieva, S.N. and Ivanov, B.A., 2001, Numerical simulation of oblique impacts: Impact melt and transient cavity. *Lunar Planetary Science*, 32, 1288.
- Ashwal, L.D., 2021, Sub-lithospheric mantle sources for overlapping southern African Large Igneous Provinces. *South African Journal of Geology*, 124, 421-442. <https://doi.org/10.25131/sajg.124.0023>.

- Austin, M. G., Thomsen, J. M., Ruhl S. F., Orpha, D. L., and Schultz, P. H, 1980, Calculational investigation of impact cratering dynamics: Material motions during the crater growth period. Proceedings of the 11th Lunar Planetary Science Conference.
- Bahat, D., 1982, Extensional aspects of earthquake induced ruptures determined by an analysis of fracture bifurcation. *Tectonophysics*, 83 (3-4), 163-183.
- Bahat, D., 1991, Tectonofractography. In *Tectonofractography* (pp. 239-324). Springer, Berlin, Heidelberg.
- Bahat, D., Rabinovitch, A. and Frid, V., 2005, *Tensile fracturing in rocks* (p. 570). Springer-Verlag Berlin Heidelberg.
- Bai, T. and Pollard, D.D., 2000, Fracture spacing in layered rocks: a new explanation based on the stress transition. *Journal of Structural Geology*, 22, 43-57.
- Bai, T., Maerten, L., Gross, M.R. and Aydin, A., 2002, Orthogonal cross joints: do they imply a regional stress rotation? *Journal of structural geology*, 24 (1), pp.77-88.
- Bai, T., Pollard, D.D. and Gao, H., 2000, Explanation for fracture spacing in layered materials. *Nature*, 403, 753.
- Baratoux, D. and Melosh, H.J., 2003, The formation of shatter cones by shock wave interference during impacting. *Earth and Planetary Science Letters*, 216 (1-2), 43-54.
- Baratoux, D. and Reimold, W.U., 2016, The current state of knowledge about shatter cones: introduction to the special issue. *Meteoritics and Planetary Science*, 51 (8), 1389-1434.
- Barnicoat, A.C., Henderson, I.H.C., Knipe, R.J., Yardley, B.W.D., Napier, R.W., Fox, N.P.C., Kenyon, A.K., Muntingh, D.J., Strydom, D., Winklert, K.S., Lawrence, S.R. and Cornford, C., 1997, Hydrothermal gold mineralization in the Witwatersrand Basin. *Nature*, 386, 820-824.
- Bates, R.L., and Jackson, J.A. (Eds.), 1987, *Glossary of Geology*. American Geological Institute, Alexandria, VA, 788 pp.
- Beach, A. and Smith, R., 2007, Structural geometry and development of the Witwatersrand Basin, South Africa. In: Ries, A.C., Butler, R.W.H., Graham, R.H. (Eds.), *Deformation of the Continental Crust: The Legacy of Mike Coward*: Geological Society, London, Special Publications, 272, 533-542.

- Beaton K.J., Gibson R.L., Bybee G.M., Walker S., Diener J.F.A. and Baxter E.F., 2022, Amphibolite facies metamorphism in lower Witwatersrand Supergroup rocks exposed in the Vredefort Dome - a Ventersdorp LIP connection. *South African Journal of Geology*, 125 (2): 191-210. <https://doi.org/10.25131/sajg.125.0016>.
- Bhat, H.S., Rosakis, A.J. and Sammis, C.G., 2012, A micromechanics based constitutive model for brittle failure at high strain rates. *Journal of Applied Mechanics*, 79 (3).
- Bickle, M.J. and Eriksson, K.A., 1982, Evolution and subsidence of early Precambrian sedimentary basins. *Philosophical Transactions of the Royal Society of London Series A: Mathematical and Physical Sciences*, 305, 225-244.
- Bisschoff, A.A., 1982, Thermal metamorphism in the Vredefort Dome:(text of Presidential Address delivered at Geocongress' 81 in Pretoria, July 1981). *South African Journal of Geology*, 85 (1), 43-57.
- Bisschoff, A.A., 1988, The history and origin of the Vredefort Dome. *South African Journal of Science*, 84 (5), p.413.
- Bisschoff, A.A., 2000, The geology of the Vredefort Dome (Explanation of Geological Sheets 2627CA, CB, CC, CD, DA, DC and 2727 AA, AB, BA - 1:50 000 scale). Council for Geoscience, South Africa.
- Bjork, R.L., 1961, Analysis of the formation of Meteor Crater, Arizona: A preliminary report. *Journal of Geophysical Research*, 66(10), 3379-3387.
- Bjørnerud, M. and Magloughlin, J.F., 2004, Pressure-related feedback processes in the generation of pseudotachylytes. *Journal of Structural Geology*, 26(12), 2317-2323.
- Bottke Jr, W.F., Love, S.G., Tytell, D. and Glotch, T., 2000, Interpreting the elliptical crater populations on Mars, Venus, and the Moon. *Icarus*, 145 (1), 108-121.
- Bowen, N.L., 1913, The melting phenomena of the plagioclase feldspar. *American Journal of Science*. 35, 577–599. <https://doi.org/10.2475/ajs.s4-35.210.577>.
- Braile, L.W., 2001, Seismic Waves and the Slinky. IRIS Consortium.
- Brink, M.C., Waanders, F.B. and Bisschoff, A.A., 1997, Vredefort: a model for the anatomy of an astrobleme. *Tectonophysics*, 270, 83-114.
- Brink, M.C., Waanders, F.B., Bisschoff, A.A. and Gay, N.C., 2000, The Foch Thrust-Potchefstroom Fault structural system, Vredefort, South Africa: a model for impact-

- related tectonic movement over a pre-existing barrier. *Journal of African Earth Sciences*, 30, 99-117.
- Britt, D. T., Yeomans, D., Housen, K., and Consolmagno, G., 2002, Asteroid density, porosity, and structure. In *Asteroids III*, 485-500.
- Bumby, A.J., Eriksson, P.G. and Van der Merwe, R., 1998, Compressive deformation in the floor rocks to the Bushveld Complex (South Africa): evidence from the Rustenburg Fault Zone. *Journal of African Earth Sciences*, 27 (3-4), 307-330.
- Bunch, T.E. and Quaide, W.L., 1968, Shatter cones in the Danny Boy nuclear crater. *Shock metamorphism of natural materials*, p.285.
- Burke, K., Kidd, W.S.F. and Kusky, T.M., 1985, Is the Ventersdorp rift system of southern Africa related to a continental collision between the Kaapvaal and Zimbabwe Cratons at 2.64 Ga ago? *Tectonophysics*, 115, 1-24.
- Burke, K., Kidd, W.S.F. and Kusky, T.M., 1986, Archean foreland basin tectonics in the Witwatersrand, South Africa. *Tectonics*, 5, 439-456.
- Bush, R., 2014, Structural and metamorphic analysis of the White Waters Ridge area, northern collar of the Vredefort Dome, South Africa [Honours Thesis]: University of the Witwatersrand, Johannesburg, pp. 110.
- Camacho, A., Vernon, R.H. and Gerald, J.F., 1995, Large volumes of anhydrous pseudotachylyte in the Woodroffe Thrust, eastern Musgrave Ranges, Australia. *Journal of Structural Geology*, 17 (3), 371-383.
- Cardwell, R.K., Chinn, D.S., Moore, G.F. and Turcotte, D.L., 1978, Frictional heating on a fault zone with finite thickness *Geophysical Journal of the Royal Astronomical Society*, 52, 525-530.
- Chalivendra, V.B. and Rosakis, A.J., 2008, Interaction of dynamic mode-I cracks with inclined interfaces. *Engineering fracture mechanics*, 75 (8), 2385-2397.
- Clarke, D.B., Henry, A.S. and White, M.A., 1998, Exploding xenoliths and the absence of “elephants’ graveyards” in granite batholiths: *Journal of Structural Geology*, 20, 1325-1343, [https://doi.org/10.1016/S0191-8141\(98\)00082-0](https://doi.org/10.1016/S0191-8141(98)00082-0).

- Clendenin, C.W., Charlesworth, E.G. and Maske, S., 1988, Tectonic style and mechanism of early Proterozoic successor basin development, southern Africa. *Tectonophysics*, 156 (3-4), 275-291.
- Clough, C.T., Maufe, H.B. and Bailey, E.B., 1909, The cauldron- subsidence of Glen Coe, and the associated igneous phenomena. *Quarterly Journal of the Geological Society of London*, 65, 611-678.
- Collettini, C., Tesei, T., Scuderi, M.M., Carpenter, B.M. and Viti, C., 2019, Beyond Byerlee friction, weak faults and implications for slip behavior. *Earth and Planetary Science Letters*, 519, 245-263.
- Collins, G. S., 2014, Numerical simulations of impact crater formation with dilatancy. *Journal of Geophysical Research: Planets*, 119, 2600-2619. <https://doi.org/10.1002/2014JE004708>.
- Collins, G.S., Kenkmann, T., Osinski, G.R. and Wünnemann, K., 2008, Mid-sized complex crater formation in mixed crystalline-sedimentary targets: Insight from modeling and observation. *Meteoritics and Planetary Science*, 43, 1955-1977.
- Collins, G.S., Lynch, E., McAdam, R. and Davison, T.M., 2017, A numerical assessment of simple airblast models of impact airbursts. *Meteoritics and Planetary Science*, 52 (8), 1542-1560.
- Collins, G.S., Melosh, H.J. and Ivanov, B.A., 2004, Modelling damage and deformation in impact simulations. *Meteoritics and Planetary Science*, 35, 217-231.
- Collins, G.S., Melosh, H.J., Morgan, J.V. and Warner, M.R., 2002, Hydrocode simulations of Chicxulub crater collapse and peak-ring formation. *Icarus*, 157 (1), 24-33.
- Collins, G.S., Wünnemann, K., Artemieva, N. and Pierazzo, E., 2013, Numerical modelling of impact processes. *Impact cratering: Processes and products*, 254-270.
- Cornell, D.H., Meintjes, P.G., van der Westhuizen, W.A. and Kristoffersen, M., 2024. Correlations on the southern Kaapvaal Craton Margin, 2: Ventersdorp strata at T’Kuip date the end of Klipriviersberg Volcanism. *South African Journal of Geology*, 127(4), pp.705-730.
- Courant, R. and Friedrichs, K.O., 1999, *Supersonic flow and shock waves* (Vol. 21). Springer Science and Business Media.

- Coward, M.P., Spencer, R.M. and Spencer, C.E., 1995, Development of the Witwatersrand Basin, South Africa. In: Coward, M.P., Ries, A.C. (Eds.), *Early Precambrian Processes: Geological Society, London, Special Publications*, 95, 243-269.
- Cox, M.A., Cavosie, A.J., Poelchau, M.H., Kenkmann, T., Miljković, K. and Bland, P.A., 2021, Asymmetric shock deformation at the Spider impact structure, Western Australia. *Meteoritics and Planetary Science*, 56 (2), 331-351.
- Crow, C. and Condie, K.C., 1988, Geochemistry and origin of late Archean volcanics from the Ventersdorp Supergroup, South Africa. *Precambrian Research*, 42 (1-2), 19-37.
- Dahl, J.M. and Schultz, P.H., 2001, Measurement of stress wave asymmetries in hypervelocity projectile impact experiments. *International Journal of Impact Engineering*, 26 (1-10), 145-155.
- Dankert, B.T. and Hein, K.A.A., 2010, Evaluating the structural character and tectonic history of the Witwatersrand Basin. *Precambrian Research* 177, 1-22.
- Dawson E., 2009, Meteorite impact shatter cones—Adiabatic shear bands? DYMAT 2009: 9th International Conference on the Mechanical and Physical Behavior of Materials under Dynamic Loading, Brussels—Belgium, 7-11 September 2009, Vol. 2, 1471-1477.
- De Kock, M.O., Beukes, N.J. and Armstrong, R.A., 2012, New SHRIMP U-Pb ages from the Hartswater Group, South Africa: implications for correlations of the Neoarchean Ventersdorp Supergroup on the Kaapvaal craton and with the Fortescue Group on the Pilbara craton. *Precambrian Research*, 204-205, 66-74.
- Deseta, N., Boonsue, S., Gibson, R.L. and Spray, J.G., 2017, Shock-induced kelyphite formation in the core of a complex impact crater. *Contributions to Mineralogy and Petrology*, 172, pp.1-19.
- Di Toro, G., and Pennacchioni, G., 2005, Fault plane processes and mesoscopic structure of a strong-type seismogenic fault in tonalities (Adamello batholith, Southern Alps). *Tectonophysics*, 402, 55-80.
- Dietz, R.S., 1960, Meteorite Impact Suggested by Shatter Cones in Rock: Three cryptoexplosion structures yield new evidence of natural hypervelocity shocks. *Science*, 131 (3416), 1781-1784.

- Dietz, R.S., 1963, Astroblemes, ancient meteorite-impact structures on the Earth. In: Middlehurst B.M. and Kuiper G.P. (Eds.), *The Moon, Meteorites and Comets*. University of Chicago Press, Chicago, III, 285-300.
- Dietz, R.S., 1968, Shatter cones in cryptoexplosion structures. In: French B.M. and Short N.M., (Eds.), *Shock Metamorphism of Natural Materials*. M.D. Mono Book Corp., Baltimore, Md., 267-285.
- Dobson, D.P., Montheil, L., Paine, J.J. and Thomson, A.R., 2021, Peritectic melting of mica in fault-related pseudotachylite melts and potassium mass balance as an indicator of fluid-absent source conditions. *G-cubed* 22, e2020GC009217. <https://doi.org/10.1029/2020gc009217>.
- Dörfler, M.A. and Kenkmann, T., 2020, Central uplift collapse in acoustically fluidized granular targets: Insights from analog modeling. *Meteoritics and Planetary Science*, 55, 441-456. <https://doi.org/10.1111/maps.13442>
- Dressler, B.O. and Reimold, W.U., 2004, Order or chaos? Origin and mode of emplacement of breccias in floors of large impact structures. *Earth-Science Reviews*, 67, 1-54.
- Dreuse, R., Doman, D., Santimano, T. and Riller, U., 2010, Crater floor topography and impact melt sheet geometry of the Sudbury impact structure, Canada. *Terra Nova*, 22, 463-469.
- Ebert, M., Poelchau, M.H., Kenkmann, T. and Schuster, B., 2020, Tracing shock-wave propagation in the Chicxulub crater: Implications for the formation of peak rings. *Geology*, 48 (8), 814-818.
- Ellis, S. 2005. An investigation of Witwatersrand Basin structures related to the Vredefort impact event. Internal Research Report, Geological Research Unit, Gencor, 117, 80 pp.
- Eriksson, P.G., Altermann, W., Catuneanu, O., Van der Merwe, R. and Bumby, A.J., 2001, Major influences on the evolution of the 2.67-2.1 Ga Transvaal basin, Kaapvaal craton. *Sedimentary Geology*, 141-142, 205-231.
- Eriksson, P.G., Banjeree, S., Nelson, D.R., Rigby, M.J., Catuneanu, O., Sarkar, S., Roberst, R.J., Ruban, D., Mtinkulu, M.N. and Sundur Raju, P.V., 2009, A Kaapvaal craton debate: nucleus of an early small supercontinent or affected by an enhanced accretion event. *Gondwana Research*, 15, 354-372.

- Evans, M.A. and Dunne, W.M., 1991, Strain factorization and partitioning in the North Mountain thrust sheet, central Appalachians, USA. *Journal of Structural Geology*, 13(1), 21-35.
- Ferriere L. and Osinski G. R. 2010, Shatter cones and associated shock-induced microdeformations in minerals—New investigations and implications for their formation (abstract #1392). 41st Lunar and Planetary Science Conference. CD-ROM.
- Ferrill, D.A., and Dunne, W.M., 1989, Cover deformation above a blind duplex: An example from West Virginia, USA. *Journal of Structural Geology*, 11 (4), 421-431.
- Ferrill, D.A., Evans, M.A., McGinnis, R.N., Morris, A.P., Smart, K.J., Wigginton, S.S., Gulliver, K.D., Lehrmann, D., de Zoeten, E. and Sickmann, Z., 2017, Fault zone processes in mechanically layered mudrock and chalk. *Journal of Structural Geology*, 97, 118-143.
- Fineberg, J., Sharon, E. and Cohen, G., 2003, Crack front waves in dynamic fracture. *International Journal of Fracture*, 121 (1), 55-69.
- Fiske, P.S., Nellis, W.J., Lipp, M. and Lorenzana, H., 1995, Pseudotachylytes generated in shock experiments: Implications for impact cratering products and processes. *Science*, 270, 5234, 281-283
- Fossen, H., 2010, *Structural Geology*, Cambridge University Press, Cambridge, pp. 463. <https://doi.org/10.1017/CBO9780511777806>.
- French, B.M., 1998, *Traces of Catastrophe. Lunar and Planet. A Handbook of Shock-Metamorphic Effects in Terrestrial Meteorite Impact Structures*. Lunar and Planetary Institute Contribution, No. 954, Lunar planetary Institute, pp. 120.
- Fricke, A., Medenbach, O. and Schreyer, W., 1990, Fluid inclusions, planar elements and pseudotachylytes in the basement rocks of the Vredefort structure, South Africa. *Tectonophysics*, 171, 169-183.
- Friese A. E. W., Charlesworth E. G. and McCarthy T. S., 1995, Tectonic processes within the Kaapvaal Craton during the Kibaran (Grenville) Orogeny: Structural, geophysical and isotopic constraints from the Witwatersrand Basin and environs. *Economic Geology*. Research Unit, University of the Witwatersrand, Johannesburg, Information Circular. 292, 1-67.

- Frimmel, H.E. and Minter, W.E.L., 2002, Recent developments concerning the geological history of the Witwatersrand gold deposits, South Africa. In: Goldfarb, R., Nielsen, R.L. (Eds.), *Global Exploration 2002 — Integrated Methods for Discovery: Special Publication*, 9, Society of Economic Geologists, Littleton, 17-45.
- Frimmel, H.E., 2005, Archaean atmospheric evolution: evidence from the Witwatersrand gold fields, South Africa. *Earth-Science Reviews* 70, 1-46.
- Gangi, A.F., Min, K.D. and Logan, J.M., 1977, Experimental folding of rocks under confining pressure: Part IV—Theoretical analysis of faulted drape-folds. *Tectonophysics*, 42 (2-4), 227-260.
- Garde, A.A. and Klausen, M.B., 2016, A centennial reappraisal of the Vredefort pseudotachylytes: shaken, not stirred by meteorite impact. *Journal of the Geological Society*, 173 (6), 954-965.
- Gash, P.S., 1971, A study of surface features relating to brittle and semibrittle fracture. *Tectonophysics*, 12 (5), 349-391.
- Gault, D.E. and Wedekind, J.A., 1978, Experimental studies of oblique impact. In *Lunar and Planetary Science Conference Proceedings*, 9, 3843-3875.
- Gibson, H.M. and Spray, J.G., 1998, Shock-induced melting and vaporization of shatter cone surfaces: Evidence from the Sudbury impact structure. *Meteoritics and Planetary Science*, 33 (2), 329-336.
- Gibson, R.L. and Jones, M.Q.W., 2002, Late Archaean to Palaeoproterozoic geotherms in the Kaapvaal craton, South Africa: Constraints on the thermal evolution of the Witwatersrand basin. *Basin Research*, 14, 169-181.
- Gibson, R.L. and Reimold, W.U., 2005, Shock pressure distribution in the Vredefort impact structure, South Africa. *Large meteorite impacts III*, 384, 329-349.
- Gibson, R.L. and Reimold, W.U., 2008, *Geology of the Vredefort impact structure: A guide to sites of interest*, 97. Council for Geoscience, pp.181.
- Gibson, R.L. and Wallmach, T., 1995, Low pressure-high temperature metamorphism in the Vredefort Dome, South Africa: Anticlockwise pressure-temperature path followed by rapid decompression. *Geological Journal*, 30, 319-331.

- Gibson, R.L., Courtnage, P.M. and Charlesworth, E.G., 1999, Bedding-parallel shearing and related deformation in the lower Transvaal Supergroup north of the Johannesburg Dome, South Africa. *South African Journal of Geology*, 102, 99-108.
- Gibson, R.L., Reimold, W.U. and Stevens, G., 1998, Thermal-metamorphic signature of an impact event in the Vredefort dome, South Africa. *Geology*, 26, 787-790.
- Gibson, R.L., Reimold, W.U. and Wallmach, T., 1997, Origin of pseudotachylyte in the lower Witwatersrand Supergroup, Vredefort Dome (South Africa): constraints from metamorphic studies. *Tectonophysics*, 283, 241-262.
- Gibson, R.L., Reimold, W.U., Ashley, A.J. and Koeberl, C., 2002, Metamorphism on the Moon: A terrestrial analogue in the Vredefort dome, South Africa? *Geology*, 30 (5), 475-478.
- Gibson, R.L., Reimold, W.U., Phillips D. and Layer P.W., 2000, $^{40}\text{Ar}/^{39}\text{Ar}$ constraints on the age of metamorphism in the Witwatersrand Supergroup, Vredefort Dome (South Africa). *South African Journal of Geology*, 103, 175-190.
- Gilbert, G.K., 1893, *The Moon's Face: A Study of the Origin of Its Features*. Philosophical Society of Washington, XII, 3, 241-292.
- Grasemann, B., Exner, U. and Tschegg, C., 2011, Displacement-length scaling of brittle faults in ductile shear. *Journal of structural Geology*, 33 (11), 1650-1661.
- Grieve, R. and Therriault, A., 2000, Vredefort, Sudbury, Chicxulub: three of a kind? *Annual Review of Earth and Planetary Sciences*, 28, 1, 305-338.
- Grieve, R.A., 1987, Terrestrial impact structures. *Annual Review of Earth and Planetary Sciences*, 15.
- Grieve, R.A., Reimold, W.U., Morgan, J., Riller, U. and Pilkington, M., 2008, Observations and interpretations at Vredefort, Sudbury, and Chicxulub: Towards an empirical model of terrestrial impact basin formation. *Meteoritics and Planetary Science*, 43, 855-882.
- Grieve, R.A.F. and Robertson, P.B., 1979, The terrestrial cratering record: I. Current status of observations. *Icarus*, 38 (2), 212-229.
- Grieve, R.A.F., Coderre, J.M., Robertson, P.B. and Alexopoulos, J., 1990, Microscopic planar deformation features in quartz of the Vredefort structure: Anomalous but still suggestive of an impact origin. *Tectonophysics*, 171, 185-200.

- Griffith, A.A., 1920, Philosophical Transactions of the Royal Society of London. A221, 163.
- Griffith, W.A., Di Toro, G., Pennacchioni, G., Pollard, D.D. and Nielsen, S., 2009, Static stress drop associated with brittle slip events on exhumed faults. *Journal of Geophysical Research: Solid Earth*, 114 (B2).
- Grocott, J., 1981, Fracture geometry of pseudotachylyte generation zones: a study of shear fractures formed during seismic events. *Journal of Structural Geology*, 3, 2, 169-178.
- Gross, M.R., 1993, The origin and spacing of cross joints: examples from the Monterey Formation, Santa Barbara Coastline, California. *Journal of Structural Geology*, 15 (6), 737-751.
- Gumsley, A., Stamsnijder, J., Larsson, E., Söderlund, U., Naeraa, T., de Kock, M., Sałacińska, A., Gawęda, A., Humbert, F. and Ernst, R., 2019, Neoarchean large igneous provinces on the Kaapvaal Craton in southern Africa re-define the formation of the Ventersdorp Supergroup and its temporal equivalents: *GSA Bulletin*. <https://doi.org/10.1130/B35237.1>.
- Guy, B.M., Ono, S., Gutzmer, J., Kaufman, A.J., Lin, Y., Fogel, M.L. and Beukes, N.J., 2012, A multiple sulfur and organic carbon isotope record from non-conglomeratic sedimentary rocks of the Mesoarchean Witwatersrand Supergroup, South Africa. *Precambrian Research*, 216, 208-231.
- Hargraves, R.B. and White, J.C., 1996, Micro-shock deformation adjacent to the surface of shatter cones from the Beaverhead impact structure, Montana. *Journal of Geology*, 104 (2), 233-238.
- Hargraves, R.B., 1961, Shatter cones in the rocks of the Vredefort ring. *Transactions of the Geological Society of South Africa*, 64, 147-161.
- Harris, C., Fourie, D.S. and Fagereng, A., 2013, Stable isotope evidence for impact-related pseudotachylyte formation at Vredefort by local melting of dry rocks. *South African Journal of Geology*, 116 (1), 101-118.
- Hart, R.J., Andreoli, M.A., Tredoux, M. and De Wit, M.J., 1990, Geochemistry across an exposed section of Archaean crust at Vredefort, South Africa: With implications for mid-crustal discontinuities. *Chemical Geology*, 82, 21-50.

- Hasch, M., Reimold, W.U., Raschke, U. and Zaag, P.T., 2016, Shatter cones at the Keurusselkä impact structure and their relation to local jointing. *Meteoritics and Planetary Science*, 51 (8), 1534-1552.
- Heider, N. and Kenkmann, T., 2003, Numerical simulation of temperature effects at fissures due to shock loading. *Meteoritics and Planetary Science*, 38 (10), 1451-1460.
- Henkel, H. and Reimold, W.U., 1998, Integrated geophysical modelling of a giant, complex impact structure: anatomy of the Vredefort Structure, South Africa. *Tectonophysics*, 287, 1-20.
- Henkel, H. and Reimold, W.U., 2002, Magnetic model of the central uplift of the Vredefort impact structure, South Africa. *Journal of Applied Geophysics*, 51 (1), 43-62.
- Hofmann A., 2011, Transvaal Supergroup, South Africa. In: Gargaud M. et al. (Eds.) *Encyclopedia of Astrobiology*. Springer, Berlin, Heidelberg. https://doi.org/10.1007/978-3-642-11274-4_1609.
- Holland, T.H., 1900, The charnockite series, a group of Archean hypersthene rocks in peninsular India. *Memoirs of the Geological Survey of India*, 28, II, 119-249.
- Holsapple, K.A., 1987, The scaling of impact phenomena. *International Journal of Impact Engineering*, 5(1-4), 343-355.
- Hossain, S., 2015, Quantification of Impact-Induced Brittle Structures: An Analysis of Stress- and Scale-Dependence of Brittle Deformation Processes. [PhD Thesis]: Universität München, Germany, pp.135.
- Huber, M.S., Kovaleva, E. and Riller, U., 2020, Modeling the geochemical evolution of impact melts in terrestrial impact basins: Vredefort granophyre dikes and Sudbury offset dikes. *Meteoritics & Planetary Science*, 55(10), pp.2320-2337.
- Huffman, A.R. and Reimold, W.U., 1996, Experimental constraints on shock-induced microstructures in naturally deformed silicates. *Tectonophysics*, 236, 165-217.
- Hugoniot, H., 1887, M'emoire sur la propagation du mouvement dans les corps et sp'ecialement dans les gaz parfaits. *J. l'Ecole Polytech*, 57, 3-97.
- Hung, C.C., Kuo, L.W., Spagnuolo, E., Wang, C.C., Di Toro, G., Wu, W.J., Dong, J.J., Lin, W., Sheu, H.S., Yeh, E.C. and Hsieh, P.S., 2019, Grain fragmentation and frictional melting during initial experimental deformation and implications for seismic slip at

- shallow depths. *Journal of Geophysical Research: Solid Earth* 124 (11), 11150–11169.
<https://doi.org/10.1029/2019JB017905>.
- Ishii, E., 2016, Far-field stress dependency of the failure mode of damage-zone fractures in fault zones: Results from laboratory tests and field observations of siliceous mudstone. *Journal of Geophysical Research: Solid Earth*, 121 (1), 70-91.
- Ivanov, B. A. and Artemieva, N. A., 2002, In: *Catastrophic Events and Mass Extinctions: Impact and Beyond*. C. Koeberl, C. and MacLeod, K. G. (Eds.), Geological Society of America Special Paper, 356, 619-630.
- Ivanov, B. A., Deniem, D. and Neukum, G., 1997, Implementation of dynamic strength models into 2D hydrocodes: Applications for atmospheric breakup and impact cratering. *International Journal of Impact Engineering*, 17 (1-5), 375-386.
[https://doi.org/10.1016/S0734-743X\(97\)87511-2](https://doi.org/10.1016/S0734-743X(97)87511-2).
- Ivanov, B.A. and Deutsch, A., 1999, Sudbury impact event: Cratering mechanics and thermal history.
- Ivanov, B.A. and Kostuchenko, V.N., 1997, March. Block oscillation model for impact crater collapse. In *Conference Paper, 28th Annual Lunar and Planetary Science Conference*, p. 631. (Vol. 28, p. 631).
- Ivanov, B.A., 2005, Numerical modeling of the largest terrestrial meteorite craters. *Solar System Research*, 39, 381-409.
- Jahn, A. and Riller, U., 2009, A 3D model of first-order structural elements of the Vredefort Dome, South Africa - importance for understanding central uplift formation of large impact structures. *Tectonophysics*, 478, 221-229.
- Jahn, A. and Riller, U., 2015, Kinematics of large terrestrial impact crater formation inferred from structural analysis and three-dimensional block modeling of the Vredefort Dome, South Africa. *Special Papers Geological Society of America*, 518, 85-97.
- Johnson, B. C., Campbell, C. S. and Melosh, H. J., 2016, The reduction of friction in long runout landslides as an emergent phenomenon. *Journal of Geophysical Research*, 121, 881-889.
- Johnson, G.P. and Talbot, R.J., 1964, A theoretical study of the shock wave origin of shatter cones. Air Force Institute of Technology Wright-Patterson, AFB, Ohio.

- Johnson, G.W. and Violet, C.E., 1958, Phenomenology of contained nuclear explosions. University of California Radiation Laboratory.
- Jolley, S.J., Freeman, S.R., Barnicoat, A.C., Phillips, G.M., Knipe, R.J., Pather, A., Fox, N.P.C., Strydom, D., Birch, M.T.G., Henderson, I.H.C. and Rowland, T.W., 2004, Structural controls on Witwatersrand gold mineralisation. *Journal of Structural Geology*, 26, 1026-1086.
- Jolley, S.J., Stuart, G.W., Freeman, S.R., Knipe, R.J., Kershaw, D., McAllister, E., Barnicoat, A.C. and Tucker, R.F., 2007, Progressive deformation of a Late Orogenic Thrust System, from duplex development to extensional reactivation and disruption: Witwatersrand Basin, South Africa. Geological Society, London, Special Publications, 272, 543-569.
- Kamo, S.L., Reimold, W.U., Krogh, T.E. and Colliston, W.P., 1996, A 2.023 Ga age for the Vredefort impact event and a first report of shock metamorphosed zircons in pseudotachylitic breccias and granophyre. *Earth and Planetary Science Letters*, 144, 369-387.
- Kelly, A. and Tyson, W.R., 1965, Tensile properties of fiber-reinforced metals: copper/tungsten and copper/molybdenum. *Journal of Mechanics and Physics of Solids*, 13 (6), 329-350.
- Kenkmann, T. and Poelchau, M.H., 2008, March. Matt Wilson: an elliptical impact crater in Northern Territory, Australia. In 39th Annual Lunar and Planetary Science Conference (No. 1391, p. 1027).
- Kenkmann, T. and Poelchau, M.H., 2009, Low-angle collision with Earth: the elliptical impact crater Matt Wilson, NT, Australia. *Geology*, 37, 459-462.
- Kenkmann, T. and von Dalwigk, I., 2000, Radial transpression ridges: A new structural feature of complex impact craters. *Meteoritics and Planetary Science*, 35, 1189-1201.
- Kenkmann, T., 2002, Folding within seconds. *Geology*, 30, 231-234.
- Kenkmann, T., 2003, Dike formation, cataclastic flow, and rock fluidisation during impact cratering: an example from the Upheaval Dome structure, Utah. *Earth and Planetary Science Letters*, 214, 43-58.
- Kenkmann, T., Collins, G.S. and Wünnemann, K., 2012, The modification stage of crater formation. *Impact cratering: Processes and products*, 60-75.

- Kenkmann, T., Hergarten, S., Kuhn, T. and Wilk, J., 2016, Formation of shatter cones by symmetric fracture bifurcation: Phenomenological modeling and validation. *Meteoritics and Planetary Science*, 51 (8), 1519-1533.
- Kenkmann, T., Hornemann, U. and Stöffler, D., 2000, Experimental generation of shock-induced pseudotachylytes along lithological interfaces. *Meteoritics and Planetary Science*, 35, 1275-1290.
- Kenkmann, T., Jahn, A., Scherler, D. and Ivanov, B.A., 2005, Structure and formation of a central uplift: A case study at the Upheaval Dome impact crater, Utah. *Large meteorite impacts III*, 384, 85-115.
- Kenkmann, T., Poelchau, M.H. and Wulf, G., 2014, Structural geology of impact craters. *Journal of Structural Geology*, 62, 156-182.
- Kenkmann, T., Reimold, W.U., Khirfan, M., Salameh, E., Konsul, K. and Khoury, H., 2010, The complex impact crater Jebel Waqf as Suwwan in Jordan: effects of target heterogeneity and impact obliquity on central uplift formation. *Geological Society of America Special Paper 465*. In: Gibson, R.L. and Reimold, W.U. (Eds.), *Large Meteorite Impacts and Planetary Evolution IV*, 471-487.
- Kenkmann, T., Sturm, S., Krüger, T., Salameh, E., Al-Raggad, M. and Konsul, K., 2017, The structural inventory of a small complex impact crater: Jebel Waqf as Suwwan, Jordan. *Meteoritics and Planetary Science*, 52, 1351-1370.
- Kenkmann, T., Vasconcelos, M.A., Crosta, A.P. and Reimold, W.U., 2011, The complex impact structure Serra da Cangalha, Tocantins State, Brazil: *Meteoritics and Planetary Science*, 46, 875-889.
- Kennedy, G.C., Wasserburg, G.J., Heard, H.C. and Newton, R.C., 1962, The upper three-phase region in the system $\text{SiO}_2\text{-H}_2\text{O}$. *American Journal of Science*, 260 (7), 501-521.
- Kieffer, S.W. and Simonds, C.H., 1980, The role of volatiles and lithology in the impact cratering process. *Reviews of Geophysics*, 18 (1), 143-181.
- Killick, A.M. and Scheepers, R., 2005, Controls to hydrothermal gold mineralization in the Witwatersberg Goldfield; situated in the floor to the south of the Bushveld Complex, South Africa. *Journal of African Earth Sciences*, 41, 235-247.
- Killick, A.M., 1992, Pseudotachylytes of the West Rand Goldfield, Witwatersrand Basin, South Africa. [PhD. Thesis]: Rand Afrikaans University, Johannesburg, pp. 273.

- Killick, A.M., 1994, The geochemistry of pseudotachylyte and its host rocks from the West Rand Goldfield, Witwatersrand Basin, South Africa: implications for pseudotachylyte genesis. *Lithos*, 32 (3-4), 193-205.
- Killick, A.M., Thwaites, A.M., Germs, G.J.B. and Schoch, A.E., 1988, Pseudotachylyte associated with a bedding-parallel fault zone between the Witwatersrand and Ventersdorp Supergroups, South Africa. *Geologische Rundschau*, 77 (1), 329-344.
- Koeberl, C., Reimold, W.U. and Shirey, S.B., 1996, Re-Os isotope and geochemical study of the Vredefort Granophyre: Clues to the origin of the Vredefort structure, South Africa. *Geology*, 24, 913-916.
- Kositcin, N. and Krapež, B., 2004, Relationship between detrital zircon age-spectra and the tectonic evolution of the Late Archaean Witwatersrand Basin, South Africa. *Precambrian Research*, 129, 141-168.
- Kovaleva, E., Zamyatin, D. and Habler, G., 2019, Granular zircon from Vredefort granophyre (South Africa) confirms the deep injection model for impact melt in large impact structures. *Geology*, 47, 691-694.
- Kriens, B.J., Shoemaker, E.M. and Herkenhoff, K.E., 1999, Geology of the Upheaval Dome impact structure, southeast Utah. *Journal of Geophysical Research*, 104, 18867-18887.
- Kulander, B.R. and Dean, S.L., 1995, Observations on fractography with laboratory experiments for geologists. Geological Society, London, Special Publications, 92 (1), 59-82.
- Lambert, P., 1977. Les Effects des Ondes de Choc Naturelles et Artificielles et le Cratere d'Impacte de Rochechouart (Limousin, France). Ph.D. Thesis, Univ. Paris-Sud, (unpubl.).
- Lambert, P., 1981, Breccia dikes: Geological constraints on the formation of complex craters. In: P.H. Schultz and R.B. Merrill (Editors), Multi-Ring Basins. *Proc. Lunar Planet. Sci.*, 12A: 59-78.
- Lana, C., Gibson, R.L. and Reimold, W.U., 2003, Impact tectonics in the core of the Vredefort dome: implication for formation of central uplifts in very large impact structures. *Meteoritics and Planetary Science*, 38, 1093-1107.

- Lana, C., Reimold, W.U., Gibson, R.L., Koeberl, C. and Siegesmund, S., 2004, Nature of the archean midcrust in the core of the Vredefort dome, Central Kaapvaal Craton, South Africa. *Geochimica et Cosmochimica Acta*, 68(3), pp.623-642.
- Lana, C., Souza Filho, C.R., Marangoni, Y.R., Yokoyama, E., Trindade, R.I.F., Tovher, E. and Reimold, W.U., 2008, Structural evolution of the 40 km wide Araguainha impact structure, central Brazil. *Meteoritics and Planetary Science*, 43, 701-716.
- Langenhorst, F., Poirier, J.P., Deutsch, A. and Hornemann, U., 2002, Experimental approach to generate shock veins in single crystal olivine by shear melting. *Meteoritics and Planetary Science*, 37 (11), 1541-1553.
- Le Feuvre, M. and Wieczorek, M. A., 2011, Nonuniform cratering of the Moon and a revised crater chronology of the inner Solar System. *Icarus*, 214 (1), 1-20. <https://doi.org/10.1016/j.icarus.2011.03.010>.
- Lee, S.K., Han, R., Kim, E.J., Jeong, G.Y., Khim, H. and Hirose, T., 2017, Quasi-equilibrium melting of quartzite upon extreme friction. *Nature Geoscience*. 10 (6), 436–441. <https://doi.org/10.1038/ngeo2951>.
- Lee, Y.J. and Freund, L.B., 1990, Fracture initiation due to asymmetric impact loading of an edge cracked plate.
- Lieger, D., 2011, Structural and Geochemical Analyses of fragment-rich pseudotachylyte bodies and the Vredefort Granophyre of the Vredefort Impact Structure, South Africa [PhD. Thesis]. Free University of Berlin, Germany, pp. 212.
- Lieger, D., Riller, U. and Gibson, R.L., 2009, Generation of fragment-rich pseudotachylyte bodies during central uplift formation in the Vredefort Impact Structure, South Africa. *Earth and Planetary Science Letters*, 279, 53-64.
- Lieger, D., Riller, U. and Gibson, R.L., 2011, Petrographic and geochemical evidence for an allochthonous, possibly impact melt, origin of pseudotachylyte from the Vredefort Dome, South Africa. *Geochimica et Cosmochimica Acta*, 75 (16), 4490-4514.
- Lilly P. A. 1978, Deformation in the collar rocks of the Vredefort ring structure. Ph.D. thesis, University of the Witwatersrand, Johannesburg, South Africa. 430 p.
- Lilly, P.A., 1981, Shock metamorphism in the Vredefort collar: evidence for internal shock sources. *Journal of Geophysical Research*, 86, 10689-10700.

- Lin, A. and Shimamoto, T., 1998, Selective melting processes as inferred from experimentally generated pseudotachylytes. *Journal of Asian Earth Sciences*, 16 (5-6), 533-545.
- Lin, A., 1994, Glassy pseudotachylyte veins from the Fuyun fault zone, northwest China. *Journal of Structural Geology*, 16 (1), 71-83.
- Lin, A., 2008, *Fossil Earthquakes: The formation and preservation of pseudotachylytes*. Springer Berlin Heidelberg, pp. 348.
- Maddock, R.H., 1983, Melt origin of fault-generated pseudotachylytes demonstrated by textures. *Geology*, 11 (2), 105-108.
- Maddock, R.H., 1992, Effects of lithology, cataclasis and melting on the composition of fault-generated pseudotachylytes in Lewisian gneiss, Scotland. *Tectonophysics*, 204 (3-4), 261-278.
- Maddock, R.H., Grocott, J. and Van Nes, M., 1987, Vesicles, amygdales and similar structures in fault-generated pseudotachylytes. *Lithos*, 20 (5), 419-432.
- Magloughlin, J. F., and Spray, J. G., 1992, Frictional melting processes and products in geological materials: introduction and discussion. *Tectonophysics*, 204, 197-206.
- Mambane, P.W., Hein, K.A.A., Twemlow, S.G. and Manzi, M.S.D., 2011, Pseudotachylyte in the South Boundary Fault at the Cooke Shaft, Witwatersrand Basin, South Africa. *South African Journal of Geology* 114 (2), 109-120.
- Manton, W.I., 1965, The orientation and origin of shatter cones in the Vredefort Ring. *Annals of the New York Academy of Sciences*, 123 (2), 1017-1049.
- Manzi, M.S.D., Hein, K.A.A., King, N. and Durrheim, R.J., 2013, Neoarchean tectonic history of the Witwatersrand Basin and Ventersdorp Supergroup: new constraints from high resolution 3D seismic reflection data. *Tectonophysics* 590, 94-105.
- Manzi, S., Gibson, R.L., and Tshibubudze, A., 2021, Dynamics of collapse of an impact central uplift: Evidence from folds and faults in the collar of the Vredefort Dome, South Africa, *in* Reimold, W.U., and Koeberl, C., eds., *Large Meteorite Impacts and Planetary Evolution VI: Geological Society of America Special Paper 550*, 585–606, [https://doi.org/10.1130/2021.2550\(27\)](https://doi.org/10.1130/2021.2550(27)).

- Marchi, S., Mottola, S., Cremonese, G., Massironi, M., and Martellato, E., 2009, A new chronology for the moon and mercury. *The Astronomical Journal*, 137 (6), 4936–4948. <https://doi.org/10.1088/0004-6256/137/6/4936>.
- Martini, J.E.J., 1978, Coesite and stishovite in the Vredefort Dome, South Africa. *Nature*, 272, 715-717.
- Martini, J.E.J., 1991, The nature, distribution, and genesis of the coesite and stishovite associated with the pseudotachylyte of the Vredefort Dome, South Africa. *Earth and Planetary Science Letters*, 103, 285-300.
- Masaitis, V.L., Naumov, M.V. and Mashchak, M.S. 1999, Anatomy of the Popigai impact crater, Russia. *Special Papers, Geological Society of America*, 1-18.
- Mashabela, S., 2011, Structural Geology of the Panvlakte Anticline in the Cooke 3 Gold Mine. [Honours Thesis], University of the Witwatersrand Johannesburg, South Africa, pp.39.
- Mashabela, S., 2016, Analysis of pre-impact and impact-induced geological structures in the northern collar of the Vredefort Dome, South Africa. [MSc Thesis]: University of the Witwatersrand, Johannesburg, South Africa, pp. 212.
- Maxwell, D.E., 1977, Simple Z model for cratering, ejection, and the overturned flap. In *Impact and explosion cratering. Planetary and terrestrial implications*, 1003-1008.
- McCarthy, T.S., 2006, The Witwatersrand Supergroup. In: *The Geology of South Africa*, Johnson, M.R., Anhaeusser, C.R., and Thomas, R.J., (Eds.), Geological Society of South Africa, Council for Geoscience Pretoria, 155-186.
- McCarthy, T.S., Cadle, A.B. and Horrocks, P., 1982, Thrusting in Witwatersrand rocks at Langerman's Kop, Johannesburg. *Transactions of the Geological Society of South Africa*, 85, 189-201.
- McCarthy, T.S., Charlesworth, E.G. and Stanistreet, I.G., 1986, Post-Transvaal structural features of the northern portion of the Witwatersrand Basin. *Transactions of the Geological Society of South Africa*, 89, 311-323.
- McCarthy, T.S., Stanistreet, I.G. and Robb, L.J., 1990, Geological studies related to the origin of the Witwatersrand basin and its gold mineralization - an introduction and a strategy for research and exploration. *South African Journal of Geology*, 93, 1-4.

- McGinnis, R.N., Ferrill, D.A., Morris, A.P., Smart, K.J. and Lehrmann, D., 2017, Mechanical stratigraphic controls on natural fracture spacing and penetration. *Journal of Structural Geology*, 95, 160-170.
- McKenzie, D. and Brune, J.N., 1972, Melting on fault planes during large earthquakes. *Geophysical Journal of the Royal Astronomical Society*, 29, 65-78.
- Melosh, H. J., 1979, Acoustic fluidization: a new geological process? *Journal of Geophysical Research*, 84, 7513-7520.
- Melosh, H. J., 1996, Dynamical weakening of faults by acoustic fluidization. *Nature*. 379, 601-606.
- Melosh, H. J., Ryan, E. V. and Asphaug, E., 1992, Dynamic fragmentation in impacts: Hydrocode simulation of laboratory impacts. *Journal of Geophysical Research*, 97 (E9), 14735-14759. <https://doi.org/10.1029/92JE01632>
- Melosh, H.J., 1977, Crater modification by gravity: a mechanical analysis of slumping. In: Roddy, D.J., Pepin, R.O. and Merrill, R.B., (Eds.), *Impact and Explosion Cratering*. Pergamon Press, New York, 1245-1260.
- Melosh, H.J., 1989, *Impact Cratering: A Geological Process*. Oxford University Press, New York, pp. 245.
- Melosh, H.J., 2005, The mechanics of pseudotachylite formation in impact events. *Impact tectonics*. Springer, Berlin, Heidelberg, 55-80.
- Melosh, H.J., 2007, A hydrocode equation of state for SiO₂. *Meteoritics and Planetary Science*, 42 (12), 2079-2098.
- Melosh, H.J., and Ivanov, B.A., 1999, Impact crater collapse. *Annual Review of Earth and Planetary Science*, 27, 385-415.
- Meyers, M.A., 1994, *Dynamic behavior of materials*. John Wiley and Sons, pp. 86.
- Milton, D. J., Glikson, A.Y. and Brett, R., 1996, Gosses Bluff—A latest Jurassic impact structure, central Australia. Part 1: Geological structure, stratigraphy, and origin. *Journal of Australian Geology and Geophysics*, 16, 453-486.
- Milton, D.J., 1977, Shatter cones—an outstanding problem in shock mechanics. In: *Impact and Explosion Cratering: Planetary and Terrestrial Implications*, 703-714.

- Mohr-Westheide, T. and Reimold, W.U., 2011, Formation of pseudotachylitic breccias in the central uplifts of very large impact structures: Scaling the melt formation. *Meteoritics and Planetary Science*, 46 (4), 543-555.
- Mohr-Westheide, T., Reimold, W.U., Riller, U. and Gibson, R.L., 2009, Pseudotachylitic breccia and microfracture networks in Archean gneiss of the central uplift of the Vredefort Impact Structure, South Africa. *South African Journal of Geology*, 112 (1), 1-22.
- Molezzi, M., 2018, Modelling the Witwatersrand Basin: A window into Neoarchaeo-Palaeoproterozoic crustal-scale tectonics. [MSc dissertation], University of the Witwatersrand, South Africa, pp. 228.
- Molezzi, M.G., Hein, K.A. and Manzi, M.S., 2019, Mesoarchaeo-Palaeoproterozoic crustal-scale tectonics of the central Witwatersrand basin - Interpretation from 2D seismic data and 3D geological modelling: *Tectonophysics*, 761, 65-85.
- Morgan, J., Warner, M., Collins, G.S., Melosh, H.J. and Christeson, G.L., 2000, Peak-ring formation in large impact craters: geophysical constraints from Chicxulub. *Earth and Planetary Science Letters*, 183, 347-354.
- Morgan, J.V., Gulick, S.P., Bralower, T., Chenot, E., Christeson, G., Claeys, P., Cockell, C., Collins, G.S., Coolen, M.J., Ferrière, L. and Gebhardt, C., 2016, The formation of peak rings in large impact craters. *Science*, 354, 878-882.
- Mungall, J.E., Kamo, S.L. and McQuade, S., 2016, U-Pb geochronology documents out-of-sequence emplacement of ultramafic layers in the Bushveld Igneous Complex of South Africa. *Nature Communications*, 7, 13385.
- Nel, L.T., 1927, The geology of the country around Vredefort: An explanation of the geological map (No. 6). Government Printing and Stationery Office.
- Ngo, D., Huang, Y., Rosakis, A., Griffith, W.A. and Pollard, D., 2012, Off-fault tensile cracks: A link between geological fault observations, lab experiments, and dynamic rupture models. *Journal of Geophysical Research: Solid Earth*, 117 (B1).
- Ngo, D., Huang, Y., Rosakis, A., Griffith, W.A. and Pollard, D.D., 2009, Off-fault tensile cracks: a link between geological fault observations, experiments and earthquake rupture models. In AGU Fall Meeting, Abstracts T52A-06.

- Nicolaysen, L.O. and Reimold, W.U., 1987, Shatter cones revisited. In International. Workshop on Cryptoexplosion and Catastrophe in the Geological Record. Parys, South Africa, Vol. 2, p. 8.
- Nicolaysen, L.O. and Reimold, W.U., 1999, Vredefort shatter cones revisited. *Journal of Geophysical Research: Solid Earth*, 104 (B3), 4911-4930.
- O'Keefe, J.D., Stewart, S.T., Lainhart, M.E. and Ahrens, T.J., 2001, Damage and rock-volatile mixture effects on impact crater formation. *International Journal of Impact Engineering*, 26 (1-10), 543-553.
- Osinski G. and Spray J. G. 2006, Shatter cones of the Houghton impact structure, Canada. *Meteoritics and Planetary Science* 42:1945–1960.
- Osinski, G.R. and Ferrière, L., 2016, Shatter cones:(Mis) understood? *Science Advances*, 2 (8), e1600616.
- Osinski, G.R. and Spray, J.G., 2005, Tectonics of complex crater formation as revealed by the Houghton impact structure, Devon Island, Canadian High Arctic. *Meteoritics and Planetary Science*, 40, 1813-1834.
- Papa, S., Spagnuolo, E., Di Toro, G., Cavallo, A., Favero, M., Camacho, A. and Pennacchioni, G., 2021, Selective clast survival in an experimentally-produced pseudotachylyte. *Journal of Structural Geology*, 147, p.104328.
- Paprika, D., Hofmann, A., Agangi, A., Elburg, M., Xie, H. and Hartmann, S., 2021, Age of the Dominion-Nsuze igneous province, the first intracratonic igneous province of the Kaapvaal craton. *Precambrian Research*, 363, 106335.
- Pierazzo, E. and Melosh, H.J., 1999, Hydrocode modeling of Chicxulub as an oblique impact event. *Earth and Planetary Science Letters*, 165 (2), 163-176.
- Pierazzo, E. and Melosh, H.J., 2000, Understanding oblique impacts from experiments, observations, and modeling. *Annual Review of Earth and Planetary Sciences*, 28 (1), 141-167.
- Pike, R.J., 1985, Some morphologic systematics of complex impact structures. *Meteoritics*, 20, 49-68.
- Pike, R.J., 1988, *Geomorphology of impact craters on Mercury*. Mercury, University of Arizona Press, 165-273.

- Pittarello, L., Nestola, F., Viti, C., Crósta, A.P. and Koeberl, C., 2015, Melting and cataclastic features in shatter cones in basalt from the Vista Alegre impact structure, Brazil. *Meteoritics and Planetary Science*, 50 (7), 1228-1243.
- Poelchau, M.H. and Kenkmann, T., 2008, Asymmetric signatures in simple craters as an indicator for an oblique impact direction. *Meteoritics and Planetary Science*, 43 (12), 2059-2072.
- Pollard, D. D. and Fletcher, R. C. 2005, *Fundamentals of structural geology*. Cambridge University Press, Cambridge, pp. 500.
- Pollard, D.D. and Aydin, A., 1988, Progress in understanding jointing over the past century. *Geological Society of America Bulletin*, 100 (8), 1181-1204.
- Pollard, D.D., and Segall, P., 1987, Theoretical displacements and stress near fractures in rocks, with applications of faults, joints, veins, dikes, and solution surfaces. In: Atkinson, B.K., (Eds.), *Fracture mechanics of rocks*: Academic Press, London, 277-349.
- Price, N.J., 1966, *Fault and Joint Development in Brittle and Semi-brittle Rocks*. Oxford: Pergamon Press, 15-31.
- Rae, A.S., Collins, G.S., Poelchau, M., Riller, U., Davison, T.M., Grieve, R.A., Osinski, G.R., Morgan, J.V. and Expedition, I.I., 2019, Stress-Strain Evolution During Peak-Ring Formation: A Case Study of the Chicxulub Impact Structure. *Journal of Geophysical Research: Planets*, 124, 396-417.
- Rae, A.S., Poelchau, M.H. and Kenkmann, T., 2021, Stress and strain during shock metamorphism. *Icarus*, 370, p.114687.
- Rae, A.S.P., Collins, G.S., Grieve, R.A.F., Osinski, G.R. and Morgan, J.V., 2017, Complex crater formation: Insights from combining observations of shock pressure distribution with numerical models at the West Clearwater Lake impact structure. *Meteoritics and Planetary Science*, 52 (7), 1330-1350.
- Ramsay, J.G., Huber, M.I. and Lisle, R.J., 1987, *The techniques of modern structural geology: Folds and fractures* (Vol. 2). Academic press.
- Reches, Z.E. and Dewers, T.A., 2005, Gouge formation by dynamic pulverization during earthquake rupture. *Earth and Planetary Science Letters*, 235 (1-2), 361-374.

- Reimold, W.U. and Colliston, W.P., 1992, Pseudotachylites of the Vredefort Dome and the surrounding Witwatersrand basin, South Africa. Geological Society of America Special Papers, 293, 177-196.
- Reimold, W.U. and Gibson, R.L., 2005, "Pseudotachylites" in large impact structures. In Impact tectonics (pp. 1-53). Springer, Berlin, Heidelberg.
- Reimold, W.U. and Gibson, R.L., 2006, The melt rocks of the Vredefort impact structure-Vredefort Granophyre and pseudotachylitic breccias: Implications for impact cratering and the evolution of the Witwatersrand Basin. *Geochemistry*, 66 (1), 1-35.
- Reimold, W.U. and Koeberl, C. 2021, (Eds.), Large Meteorite Impacts and Planetary Evolution VI (Vol. 550). Geological Society of America, p. 644.
- Reimold, W.U., 1991, The geochemistry of pseudotachylites from the Vredefort Dome, South Africa. *Neues Jahrbuch für Mineralogie. Abhandlungen*, 162 (2), 151-184.
- Reimold, W.U., 1995, Pseudotachylite in impact structures—generation by friction melting and shock brecciation?: A review and discussion. *Earth-Science Reviews*, 39(3-4), 247-265.
- Reimold, W.U., 1998, Exogenic and endogenic breccias: a discussion of major problematics. *Earth-Science Reviews*, 43(1-2), 25-47.
- Reimold, W.U., Andreoli, M. and Hart, R., 1985, March. Pseudotachylite from the Vredefort dome. In Lunar and Planetary Science Conference Proceedings, 16, 691-692.
- Reimold, W.U., Colliston, W.P., Dressler, B.O., Grieve, R.A.F. and Sharpton, V.L., 1994, Pseudotachylites of the Vredefort Dome and the surrounding Witwatersrand basin, South Africa. *Special papers-geological Society of America*, 177-177.
- Reimold, W.U., Hauser, N., Hansen, B.T., Thirlwall, M. and Hoffmann, M., 2017, The impact pseudotachylitic breccia controversy: Insights from first isotope analysis of Vredefort impact-generated melt rocks. *Geochimica et Cosmochimica Acta*, 214, 266-281.
- Reimold, W.U., Horsch, H. and Durrheim, R.J., 1990, The 'bronzite'-granophyre from the Vredefort structure - a detailed analytical study and reflections on the genesis of one of Vredefort's enigmas. In Lunar and Planetary Science Conference Proceedings, 20, 433-450.
- Reimold, W.U., Pybus, G.Q.J., Kruger, F.J., Layer, P.W. and Koeberl, C., 2000, The Anna's Rust Sheet and related gabbroic intrusions in the Vredefort Dome-Kibaran magmatic

- event on the Kaapvaal Craton and beyond? *Journal of African Earth Sciences*, 31 (3-4), 499-521.
- Reimold, W.U., Schulz, T., König, S., Koeberl, C., Hauser, N., Wannek, D. and Schmitt, R.T., 2021, Genesis of the mafic granophyre of the Vredefort impact structure (South Africa): Implications of new geochemical and Se and Re-Os isotope data.
- Riemann, B., 1860, Über die Fortpflanzung ebener Luftwellen von endlicher Schwingungsweite (Vol. 8). Verlag der Dieterichschen Buchhandlung.
- Riller, U., Lieger, D., Gibson, R.L., Grieve, R.A. and Stöffler, D., 2010, Origin of large-volume pseudotachylite in terrestrial impact structures. *Geology*, 38 (7), 619-622.
- Riller, U., Poelchau, M.H., Rae, A.S., Schulte, F.M., Collins, G.S., Melosh, H.J., Grieve, R.A., Morgan, J.V., Gulick, S.P., Lofi, J. and Diaw, A., 2018, Rock fluidisation during peak-ring formation of large impact structures. *Nature*, 562 (7728), 511-518. <https://doi.org/10.1038/s41586-018-0607-z>.
- Roddy, D.J. and Davis, L.K., 1977, Shatter cones formed in large-scale experimental explosion craters. In: Roddy, D.J., Pepin R.G. and Merrill R.B., (Eds.), *Impact and Explosion Cratering*. Pergamon, New York, N.Y., 715-750.
- Roering, C., Berlenbach, J. and Schweitzer, J.K., 1991, Guidelines for the classification of fault rocks. COMRO. User Guide, 21, pp. 22.
- Rosakis, A.J. and Ravichandran, G., 2000, Dynamic failure mechanics. *International journal of solids and structures*, 37 (1-2), 331-348.
- Rowe, C.D. and Griffith, W.A., 2015, Do faults preserve a record of seismic slip: A second opinion. *Journal of Structural Geology*, 78, 1-26.
- Ruf, J.C., Rust, K.A. and Engelder, T., 1998, Investigating the effect of mechanical discontinuities on joint spacing. *Tectonophysics*, 295 (1-2), 245-257.
- Sagy, A., Cohen, G., Reches, Z.E. and Fineberg, J., 2006, Dynamic fracture of granular material under quasi-static loading. *Journal of Geophysical Research: Solid Earth*, 111 (B4).
- Sagy, A., Fineberg, J. and Reches, Z.E., 2004, Shatter cones: Branched, rapid fractures formed by shock impact. *Journal of Geophysical Research: Solid Earth*, 109 (B10).
- Sagy, A., Reches, Z.E. and Fineberg, J., 2002, Dynamic fracture by large extraterrestrial impacts as the origin of shatter cones. *Nature*, 418 (6895), 310-313.

- Sagy, A., Reches, Z.E. and Roman, I., 2001, Dynamic fracturing: Field and experimental observations. *Journal of Structural Geology*, 23 (8), 1223-1239.
- Scherler, D., Kenkmann, T. and Jahn, A., 2006, Structural record of an oblique impact. *Earth and Planetary Science Letters*, 248 (1-2), 43-53.
- Schneider, E. and Wagner, G.A., 1976, Shatter cones produced experimentally by impacts in limestone targets. *Earth and Planetary Science Letters*, 32 (1), 40-44.
- Schultz, P.H., Eberhardy, C.A., Ernst, C.M., A'Hearn, M.F., Sunshine, J.M. and Lisse, C.M., 2007, The Deep Impact oblique impact cratering experiment. *Icarus*, 191 (2), 84-122.
- Schwarzman, E.C., Meyer, C.E. and Wilshire, H.G., 1983, Pseudotachylyte from the Vredefort Ring, South Africa, and the origins of some lunar breccias. *Geological Society of America Bulletin*, 94 (7), 926-935.
- Scoates, J.S., Wall, C.J., Friedman, R.M., Weis, D., Mathez, E.A. and VanTongeren, J.A., 2021, Dating the Bushveld Complex: Timing of Crystallization, Duration of Magmatism, and Cooling of the World's Largest Layered Intrusion and Related Rocks. *Journal of Petrology*, 62 (2), egaa107. <https://doi.org/10.1093/petrology/egaa107>.
- Shand, S.J., 1916, The pseudotachylyte of Parijs (Orange Free State). *Quarterly Journal of the Geological Society of London*, 72, 198-221.
- Sharon E., Cohen G., and Fineberg J., 2001, Propagating solitary waves along a rapidly moving crack front. *Nature*, 410, 68–71.
- Sharon E., Cohen G., and Fineberg J., 2002, Crack front waves and the dynamics of a rapidly moving crack. *Physical Review Letters*, 88, 085503.
- Sharon, E., Gross, S.P. and Fineberg, J., 1995, Local crack branching as a mechanism for instability in dynamic fracture. *Physical Review Letters*, 74 (25), 5096.
- Shaw, C.S., and Walton, E., 2013, Thermal modeling of shock melts in Martian meteorites: Implications for preserving Martian atmospheric signatures and crystallization of high-pressure minerals from shock melts. *Meteoritics and Planetary Science*, 48 (5), 758-770.
- Shoemaker, E. M., 1962, Interpretation of lunar craters, *Physics and astronomy of the Moon*, 283-359. New York: Elsevier.

- Sibson, R.H., 1975, Generation of pseudotachylyte by ancient seismic faulting. *Geophysical Journal of the Royal Astronomical Society*, 43, 775-794.
- Sibson, R.H., 1977, Fault rocks and fault mechanisms. *Journal of the Geological Society of London*, 133, 191-213.
- Simpson, C., 1978, The structure of the rim synclinorium of the Vredefort Dome. *Transactions of the Geological Society of South Africa*, 81, 115-121.
- Simpson, C., 1981, Occurrence and orientation of shatter cones in Pretoria Group quartzites in the collar of the Vredefort "Dome": Impact origin precluded. *Journal of Geophysical Resources*, 86, 10701-10706.
- Speers, E.C., 1957, The age relation and origin of common Sudbury Breccia. *The Journal of Geology*, 65 (5), 497-514.
- Spray, J. G., 1997, Superfaults, *Geology*, 25, 579-582.
- Spray, J.G. and Biren, M.B., 2021, Distinguishing friction-from shock-generated melt products in hypervelocity impact structures. *Large Meteorite Impacts and Planetary Evolution VI*, 550, 147-70.
- Spray, J.G. and Boonsue, S., 2018, Quartz–coesite–stishovite relations in shocked metaquartzites from the Vredefort impact structure, South Africa. *Meteoritics & Planetary Science*, 53(1), 93-109.
- Spray, J.G. and Thompson, L.M., 1995, Friction melt distribution in a multi-ring impact basin. *Nature*, 373, 6510, 130-132.
- Spray, J.G. and Thompson, L.M., 2008, Constraints on central uplift structure from the Manicouagan impact crater. *Meteoritics and Planetary Science*, 43, 2049-2057.
- Spray, J.G., 1987, Artificial generation of pseudotachylyte using friction welding apparatus: simulation of melting on a fault plane. *Journal of Structural Geology*, 9 (1), 49-60.
- Spray, J.G., 1992, A physical basis for the frictional melting of some rock-forming minerals. *Tectonophysics*, 204 (3-4), 205-221.
- Spray, J.G., 1993, Viscosity determinations of some frictionally generated silicate melts: Implications for fault zone rheology at high strain rates. *Journal of Geophysical Research: Solid Earth*, 98 (B5), 8053-8068.
- Spray, J.G., 1995, Pseudotachylyte controversy: Fact or friction? *Geology*, 23 (12), 1119-1122.

- Spray, J.G., 1998, Localized shock-and friction-induced melting in response to hypervelocity impact. Geological Society, London, Special Publications, 140 (1), 195-204.
- Spray, J.G., 2010, Frictional melting processes in planetary materials: From hypervelocity impact to earthquakes. Annual Review of Earth and Planetary Sciences, 38, 221-254.
- Spray, J.G., 2016, August. Experimental vibration melting without offset: Post-shock materials behaviour. In 79th Annual Meeting of the Meteoritical Society (Vol. 79, No. 1921, 6118).
- Stanistreet, I.G., McCarthy, T.S., Charlesworth, E.G., Myers, R.E. and Armstrong, R.A., 1986, Pre-Transvaal wrench tectonics along the northern margin of the Witwatersrand Basin. Tectonophysics, 131, 53-74.
- Stöffler, D. and Langenhorst, F., 1994, Shock metamorphism of quartz in nature and experiment: I. Basic observation and theory. Meteoritics, 29 (2), 155-181.
- Stöffler, D., 1971, Progressive metamorphism and classification of shocked and brecciated crystalline rocks at impact craters. Journal of Geophysical Research, 76 (23), 5541-5551.
- Stöffler, D., 1984, Glasses formed by hypervelocity impact. Journal of Non-Crystalline Solids, 67 (1-3), 465-502.
- Sundaram, B.M. and Tippur, H.V., 2016, Dynamics of crack penetration vs. branching at a weak interface: an experimental study. Journal of the Mechanics and Physics of Solids, 96, 312-332.
- Therriault, A.M., Grieve, R.A.F. and Reimold, W.U., 1995, How big is Vredefort? Meteoritics, 30, 586-587.
- Therriault, A.M., Grieve, R.A.F. and Reimold, W.U., 1997, Original size of the Vredefort Structure: Implications for the geological evolution of the Witwatersrand Basin. Meteoritics and Planetary Science, 32 (1), 71-77.
- Therriault, A.M., Reimold, W.U. and Reid, A.M., 1996, The Vredefort granophyre: Part 1. Field studies. South African Journal of Geology, 99, 1-21.
- Thompson, L.M. and Spray, J.G., 1996, Pseudotachylyte petrogenesis: constraints from the Sudbury impact structure. Contributions to Mineralogy and Petrology, 125 (4), pp.359-374.

- Thompson, S. L. and Lauson, H. S., 1974, Improvements in the CHART D radiation-hydrodynamic code III: Revised analytic equations of state (SC-RR-710714). Albuquerque, New Mexico: Sandia National Laboratory.
- Tinker, J., De Wit, M.J. and Grotzinger, J., 2002, Seismic stratigraphic constraints on Neoproterozoic—Paleoproterozoic evolution of the Western Margin of the Kaapvaal Craton, South Africa. *South African Journal of Geology*, 105, 107-134.
- Toyoshima, T., 1990, Pseudotachylite from the Main Zone of the Hidaka metamorphic belt, Hokkaido, northern Japan. *Journal of Metamorphic Geology*, 8 (5), 507-523.
- Tsikalas, F., 2005, Mjølnir Crater as a result of oblique impact: asymmetry evidence constrains impact direction and angle. In *Impact tectonics*. pp. 285-306. Springer, Berlin, Heidelberg.
- Turtle, E.P. and Pierazzo, E., 1998, Constraints on the size of the Vredefort impact crater from numerical modeling. *Meteoritics and Planetary Science*, 33 (3), 483-490.
- Turtle, E.P., Pierazzo, E. and O'Brien, D.P., 2003, Numerical modeling of impact heating and cooling of the Vredefort impact structure. *Meteoritics and Planetary Science*, 38 (2), 293-303.
- Tuttle, O.F., and Bowen, N.L., 1958, Origin of granite in the light of experimental studies in the system NaAlSi₃O₈-KAlSi₃O₈-SiO₂-H₂O. *Geological Society of America, Mem.* 74, 153.
- Twiss, R.J. and Moores, E.M., 1992, *Structural geology*. Macmillan.
- Violay, M., Nielsen, S., Gibert, B., Spagnuolo, E., Cavallo, A., Azais, P., Vinciguerra, S. and Di Toro, G., 2014, Effect of water on the frictional behavior of cohesive rocks during earthquakes. *Geology* 42 (1), 27–30. <https://doi.org/10.1130/G34916.1>.
- Wang Y., Shan W., Zhou X and Wang P., 2021, Exact solutions and bifurcation for the resonant nonlinear Schrödinger equation with competing weakly nonlocal nonlinearity and fractional temporal evolution, *Waves in Random and Complex Media*, 31(6), 1859-1878, DOI: 10.1080/17455030.2019.1706013.
- Wenk, H.R., 1978, Are pseudotachylites products of fracture or fusion? *Geology*, 6, 507-511.

- Wieland, F., 2006, Structural analysis of impact-related deformation in the collar rocks of the Vredefort Dome, South Africa [PhD. Thesis]: University of the Witwatersrand, Johannesburg, South Africa, pp. 343.
- Wieland, F., Gibson, R.L. and Reimold, W.U., 2005, Structural analysis of the collar of the Vredefort Dome, South Africa—Significance for impact-related deformation and central uplift formation. *Meteoritics and Planetary Science*, 40, 1537-1554.
- Wieland, F., Reimold, W.U. and Gibson, R.L., 2006, New observations on shatter cones in the Vredefort impact structure, South Africa, and evaluation of current hypotheses for shatter cone formation. *Meteoritics and Planetary Science*, 41(11), pp.1737-1759.
- Wilk, J. and Kenkmann, T., 2015a, Shatter Cones from the MEMIN Impact Experiments. Bridging the Gap III: Impact Cratering. In *Nature, Experiments, and Modeling*, 1861, p.1066.
- Wilk, J. and Kenkmann, T., 2015b, The surface structure of shatter cones in experimental impact craters. In *46th Annual Lunar and Planetary Science Conference*, 1832, p. 2637.
- Wilk, J. and Kenkmann, T., 2016, Formation of shatter cones in MEMIN impact experiments. *Meteoritics and Planetary Science*, 51(8), 1477-1496.
- Wilk, J., Hamann, C., Fazio, A., Luther, R., Hecht, L., Langenhorst, F. and Kenkmann, T., 2018, Petrographic investigation of shatter cone melt films recovered from MEMIN impact experiments in sandstone and iSALE modeling of their formation boundary conditions. *Meteoritics & Planetary Science*, 53(8), pp.1569-1593.
- Willemse, J., 1937, On the Old Granite of the Vredefort region and some of its related rocks. *Transactions of the Geological Society of South Africa*, 40, 43-119.
- Wilshire, H. G. and Howard, K. A., 1968, Structural patterns in central uplifts of cryptoexplosion structures as typified by Sierra Madera. *Science*, 162, 258-261.
- Wilshire, H.G., 1971, Pseudotachylyte from the Vredefort Ring, South Africa. *The Journal of Geology*, 79 (2), 195-206.
- Winkler, R., Luther, R., Poelchau, M.H., Wünnemann, K. and Kenkmann, T., 2018, Subsurface deformation of experimental hypervelocity impacts in quartzite and marble targets. *Meteoritics and Planetary Science*, 53 (8), 1733-1755.

- Wright, T.W., 2002, The physics and mathematics of adiabatic shear bands. Cambridge University Press.
- Wünnemann, K., Collins, G. S. and Melosh, H. J., 2006, A strain-based porosity model for use in hydrocode simulations of impacts and implications for transient crater growth in porous targets. *Icarus*, 180 (2), 514-527. <https://doi.org/10.1016/j.icarus.2005.10.013>.
- Xu, L.R., Huang, Y.Y. and Rosakis, A.J., 2003, Dynamic crack deflection and penetration at interfaces in homogeneous materials: experimental studies and model predictions. *Journal of the Mechanics and Physics of Solids*, 51 (3), 461-486.

APPENDIX A

Pseudotachylytes

Appendix A

Table A1: Mineral phase data used for TIMA analysis

Mineral Name	O	Si	Al	Ca	K	Fe	P	Mg	Na	H	Mn	Ti	F	Hf	Zr	REE	Density [g/cm³]
Albite	48,7	31,5	10,8	0,76					8,3								2,62
Aluminosilicate mixture																	
Anorthite	38,5	30,2	18,5	12,3	0,53												2,74
Apatite	38,1			39,7			18,4						3,8				3,15
Augite																	
Augite Fe rich																	
Biotite	43,4	19,4	6,2		9	6,4		14		0,41			1,1				3,09
Heamatite/magnetite	30					70											4,72
Ilmenite	31,6					36,8						31,6					4,72
Kaolinite	55,8	21,8	20,9							1,6							2,6
Muscovite	47,4	21,1	20,3		9,8					0,46			0,95				2,77
O rich Quartz	78,5	21,5															2,6
Orthoclase	46	30,3	9,7		14												2,9
Plagioclase	47,3	31,1	10	7,4					4,3								2,68
Quartz	53,3	46,7															2,62
Titanite	36,7	12,3	0,94	23,3		1,6		0,05			0,04	25,1					3,48
Zircon	33,6	14,8												4,7	43,1	3,8	4,65

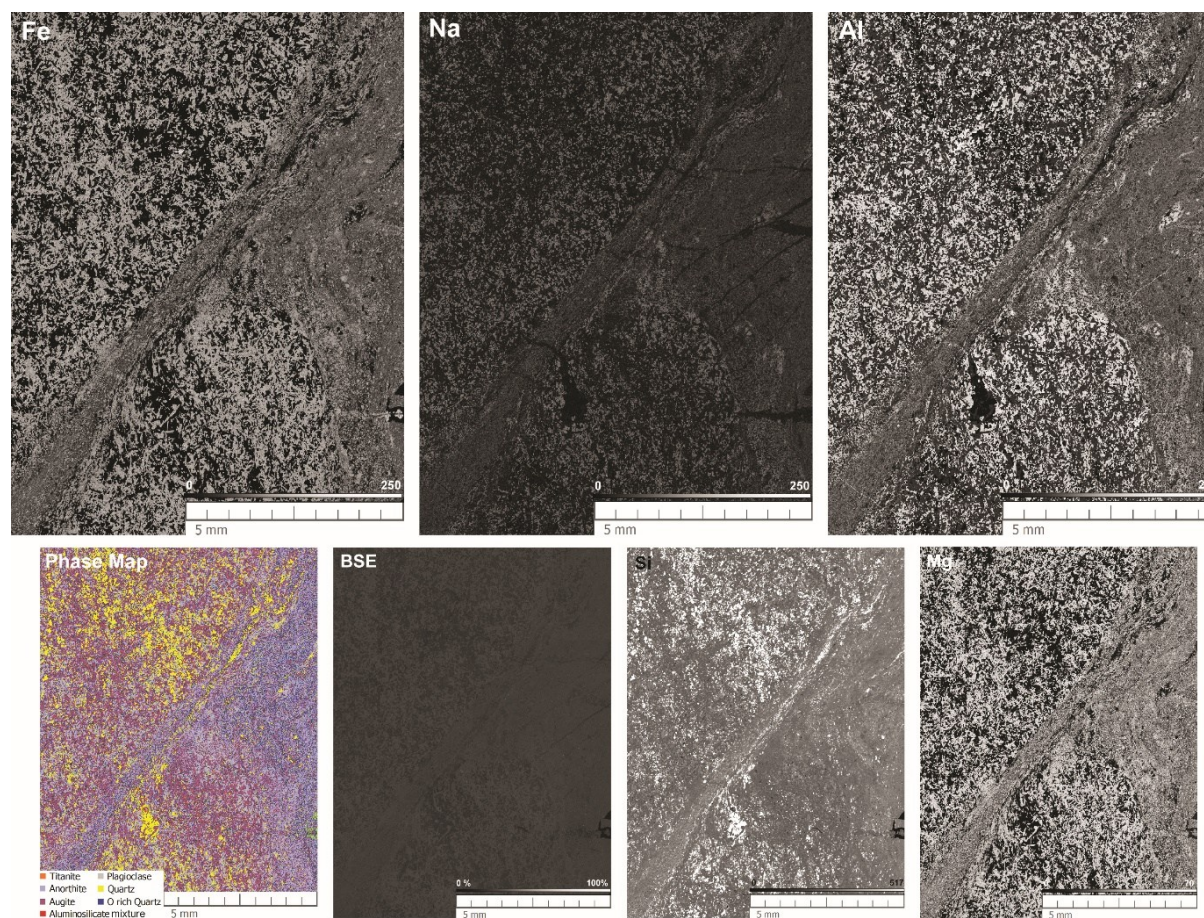


Figure A1: TIMA images of metabasalt pseudotachylyte showing well-defined margins, the absence of clasts of wall rock within the pseudotachylyte. Flow banding is visible. Clinopyroxene (augite) makes euhedral needles up to 1.5 mm. The pseudotachylyte is made up preferentially of plagioclase and anorthite. The pseudotachylyte composition is intermediate between that of the host mineralogy. Sample RAS-105 location UTM 35J 0532750 E 7026750 S.

Appendix A

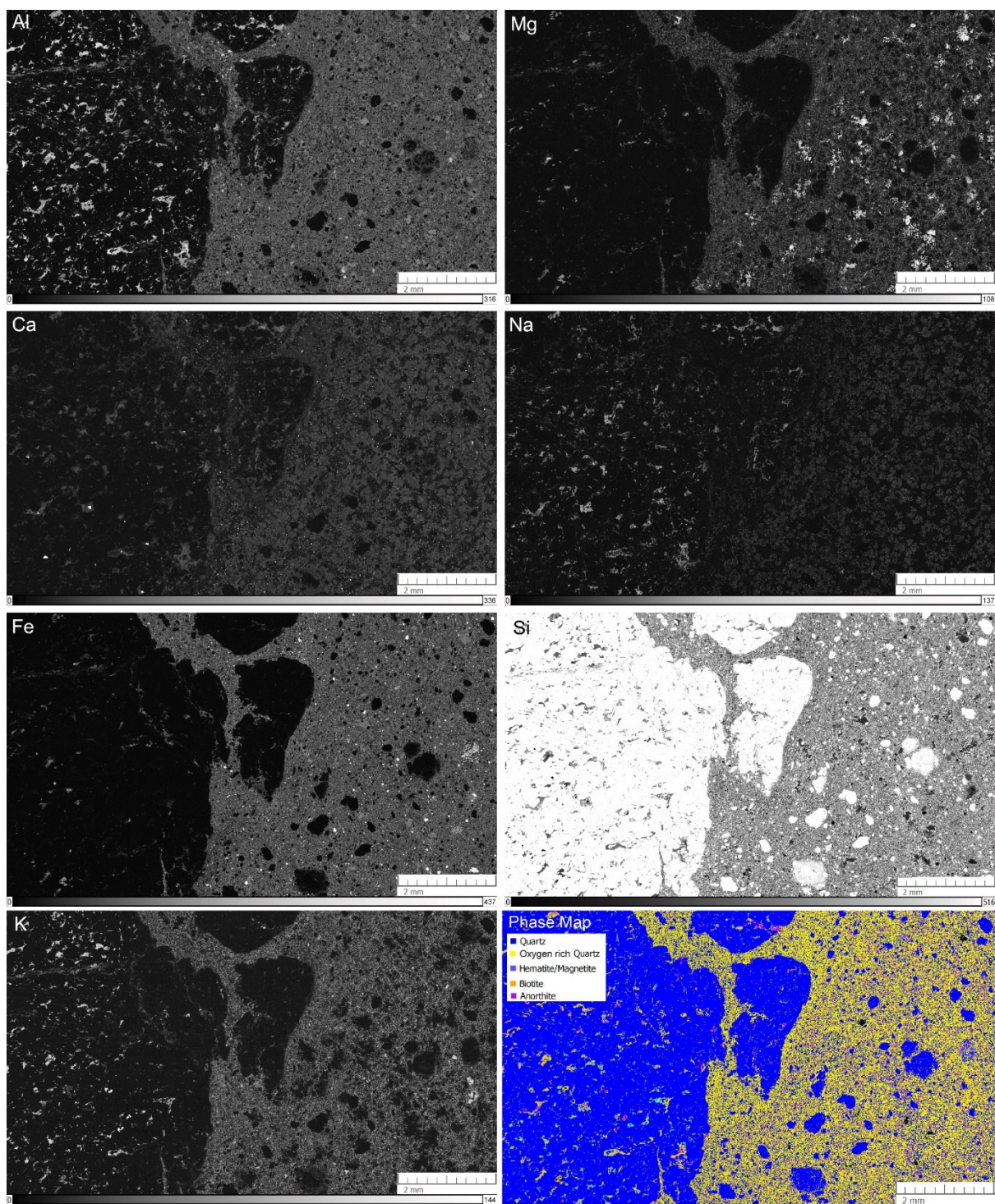


Figure A2: TIMA scan results of RAS-104. Sample location UTM 35J 0537574 E 7028690 S.

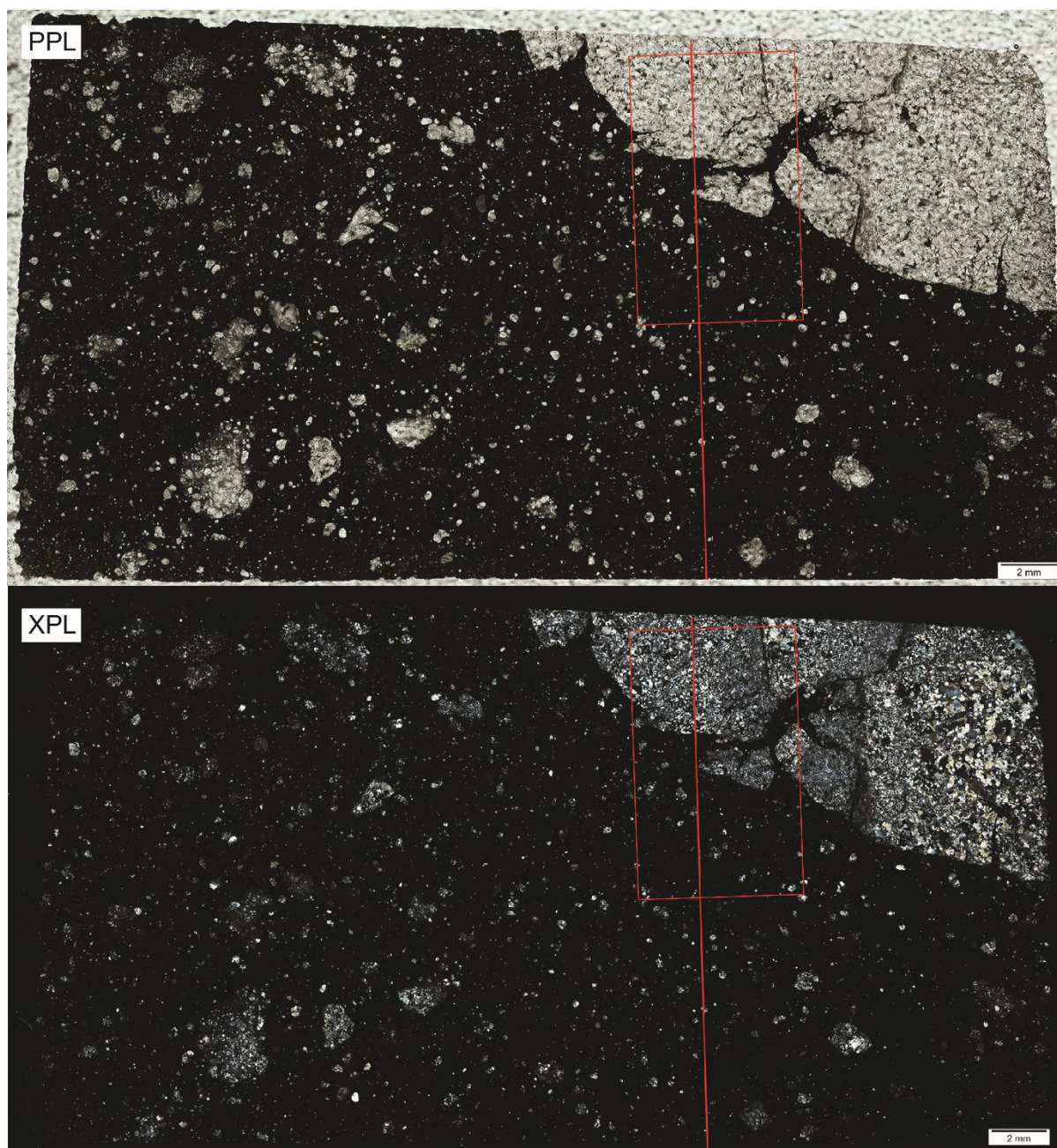


Figure A3: Thin section overview in PPL and XPL of RAS-104 a meta-ironstone pseudotachylyte. The red rectangle shows the TIMA scan area (Figure 6.34), and the red line indicates the EPMA scan line (Figure 6.35). The pseudotachylyte has abundant quartzite clasts within it. Sample location UTM 35J 0537574 E 7028690 S.

Appendix A

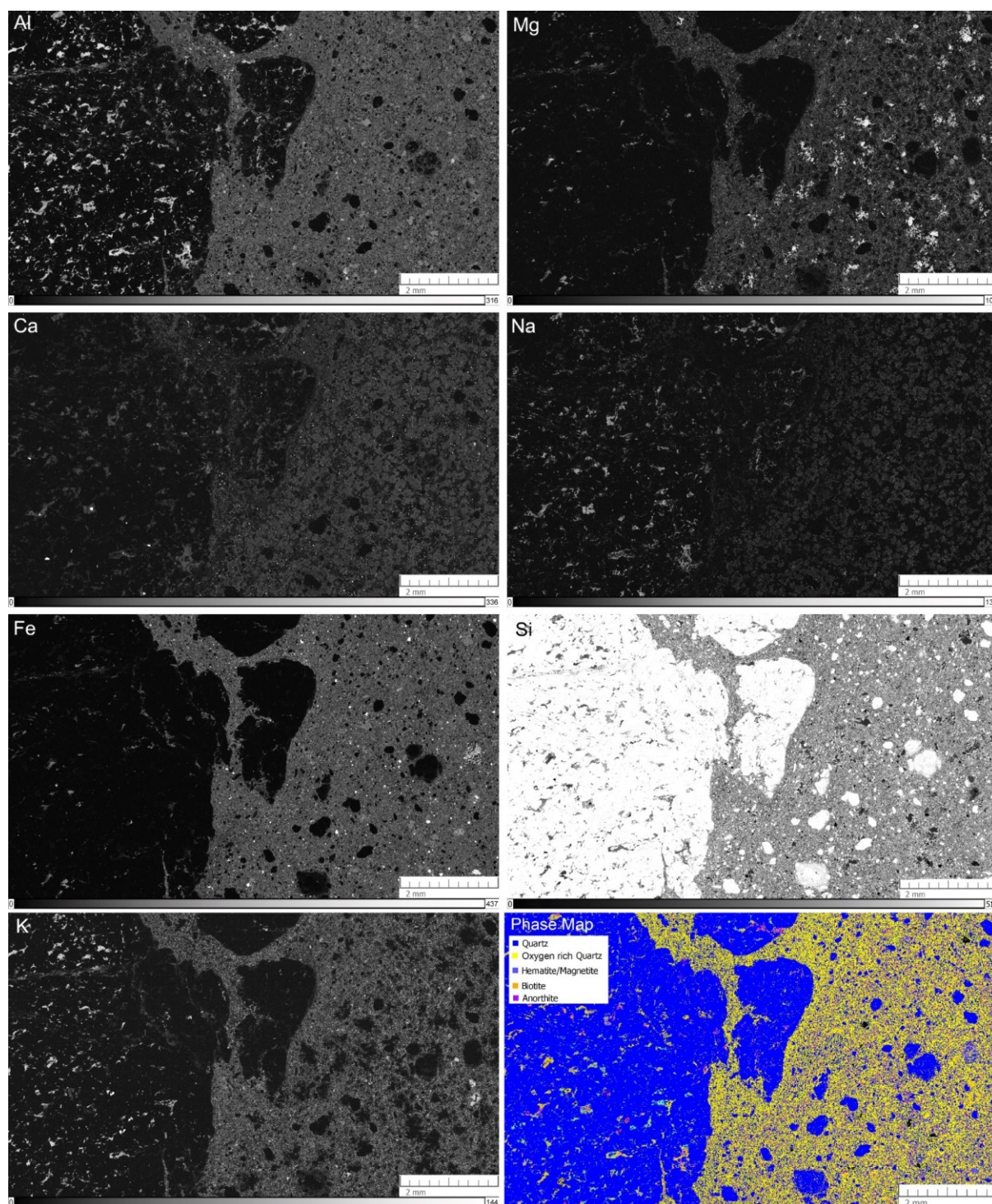


Figure A4: From left to right: Aluminium, magnesium, calcium, sodium, iron, silica, potassium elemental maps and phase map of quartzite/metaironstone pseudotachylyte RAS-104. The pseudotachylyte has abundant quartzite clasts within it. Sample location UTM 35J 0537574 E 7028690 S.

Appendix A

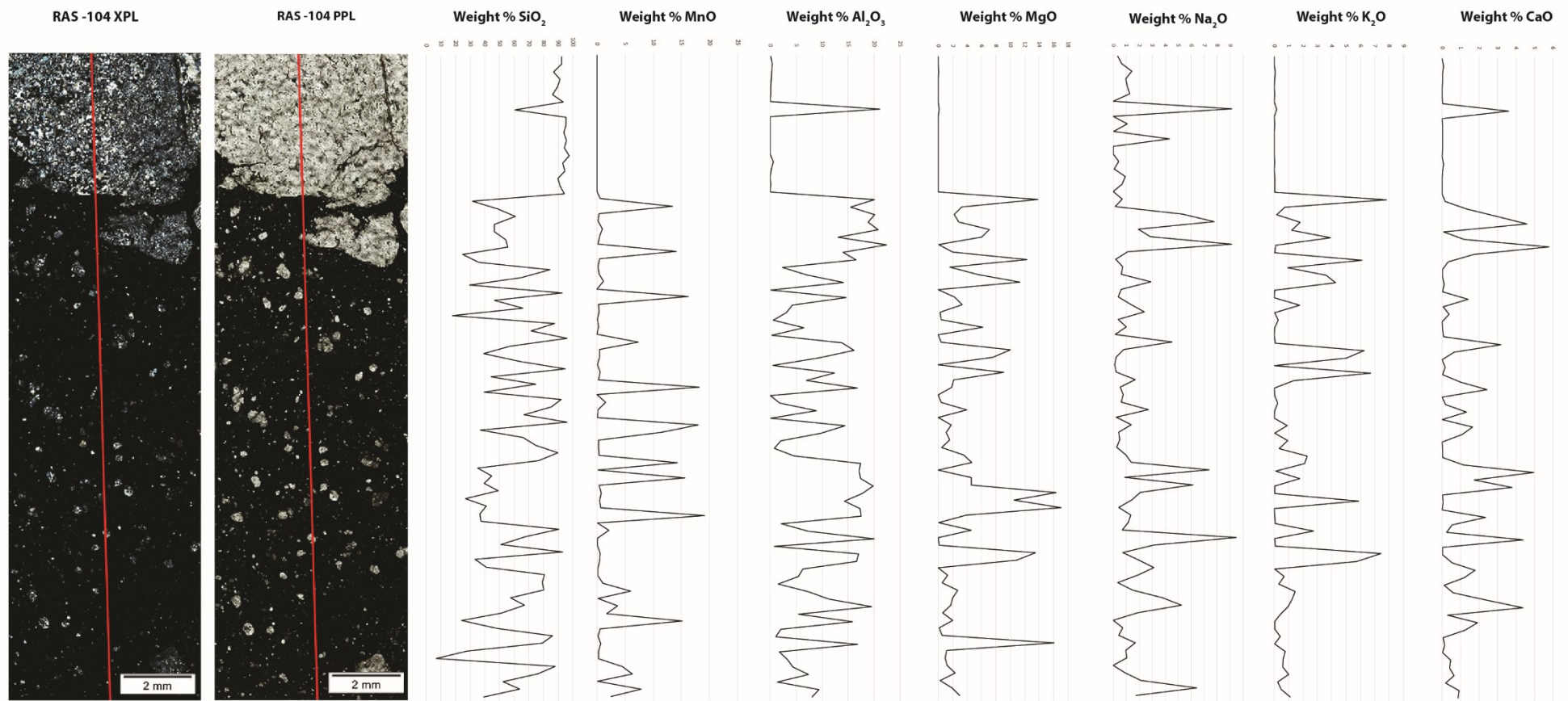


Figure A5: EPMA Scan line of RAS-104. Sample location UTM 35J 0537574 E 7028690

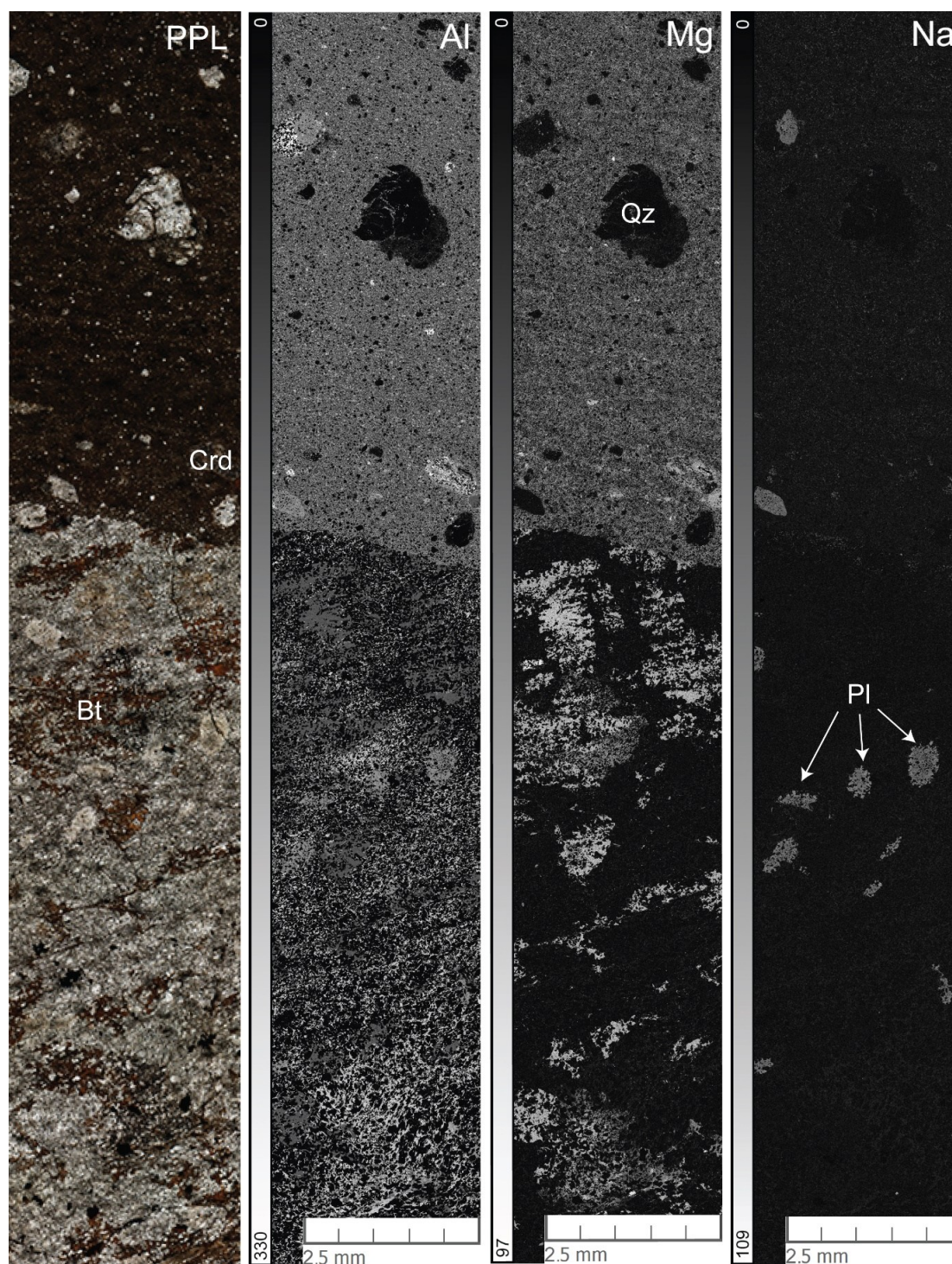


Figure A6: From left to right: Plane polarised light microphotograph, aluminium, magnesium and sodium elemental maps of Sample RAS-108, a metapelite pseudotachylite and the host showing the sharp, smooth contact. Clasts are less abundant in the matrix as compared to the quartzite samples. They show less internal grading with a very homogeneous aluminium and magnesium-rich matrix with small clasts of quartzite and, to a lesser extent, rounded plagioclase clast best seen in the Na elemental map. The quartzite clasts demonstrate that the melt had an allochthonous component. Sample location UTM 35J 537600 E 7028700 S

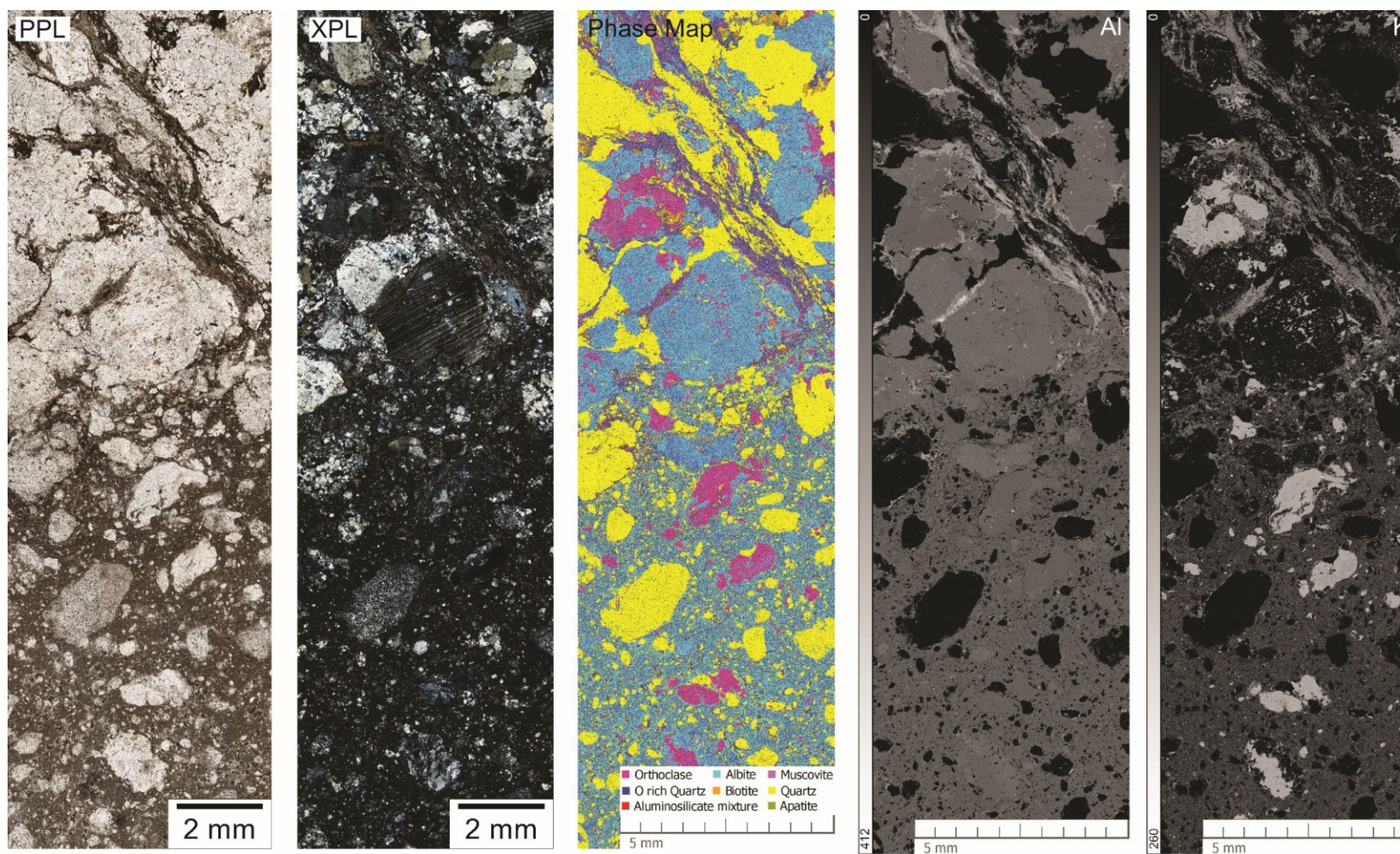


Figure A7: From left to right: Plane polarised light microphotograph, Cross polarised microphotograph, phase map, aluminium, and potassium elemental maps of Sample RAS-109, a granite pseudotachylyte. Sample location UTM 35J 538300 E 7026544 S.

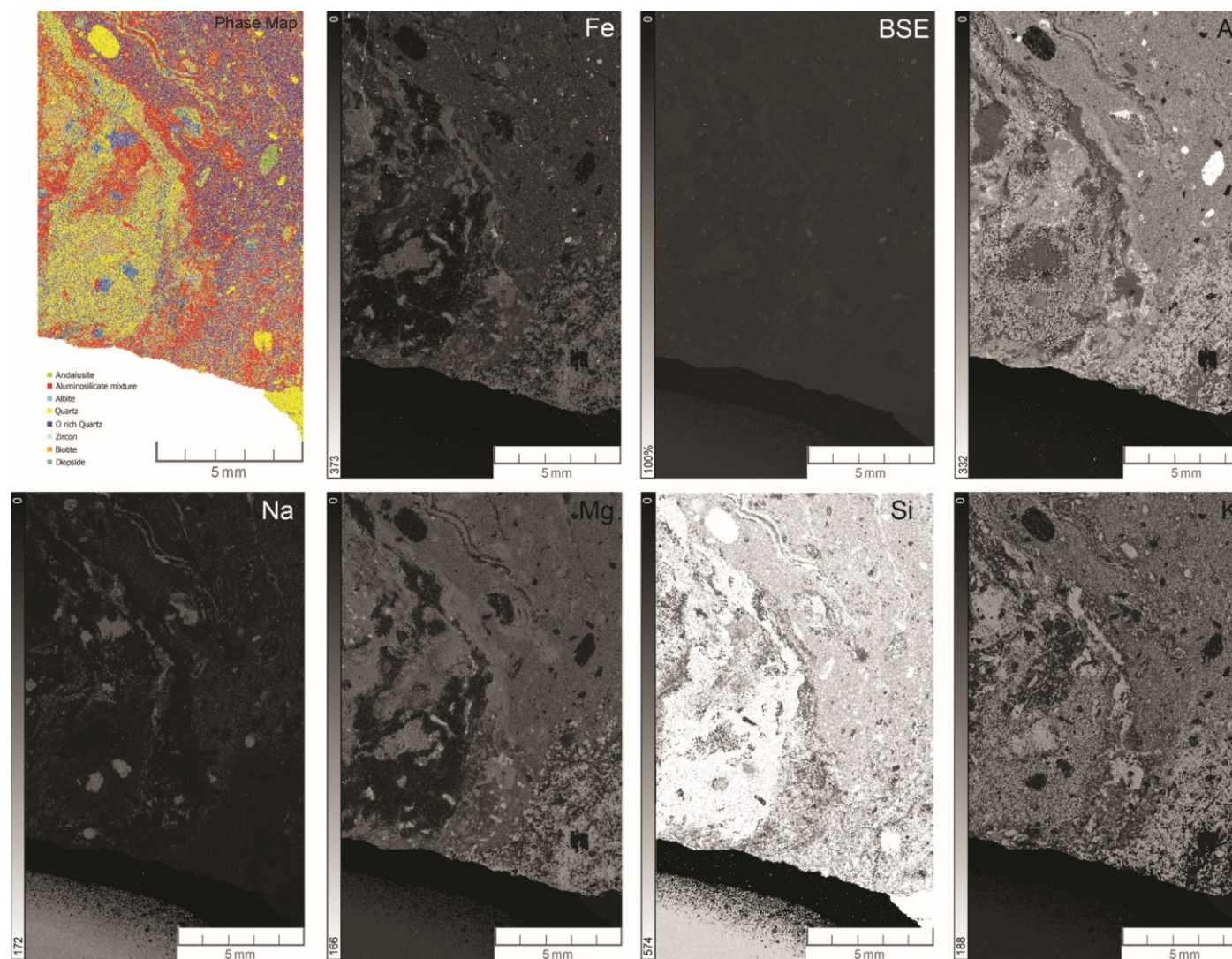


Figure A8: From left to right: Phase map, iron, BSE, aluminium, sodium, magnesium, silica, and potassium elemental maps of Sample RAS-100, a metapelite pseudotachylyte. Sample location UTM 35J 537952 E 7028599 S.

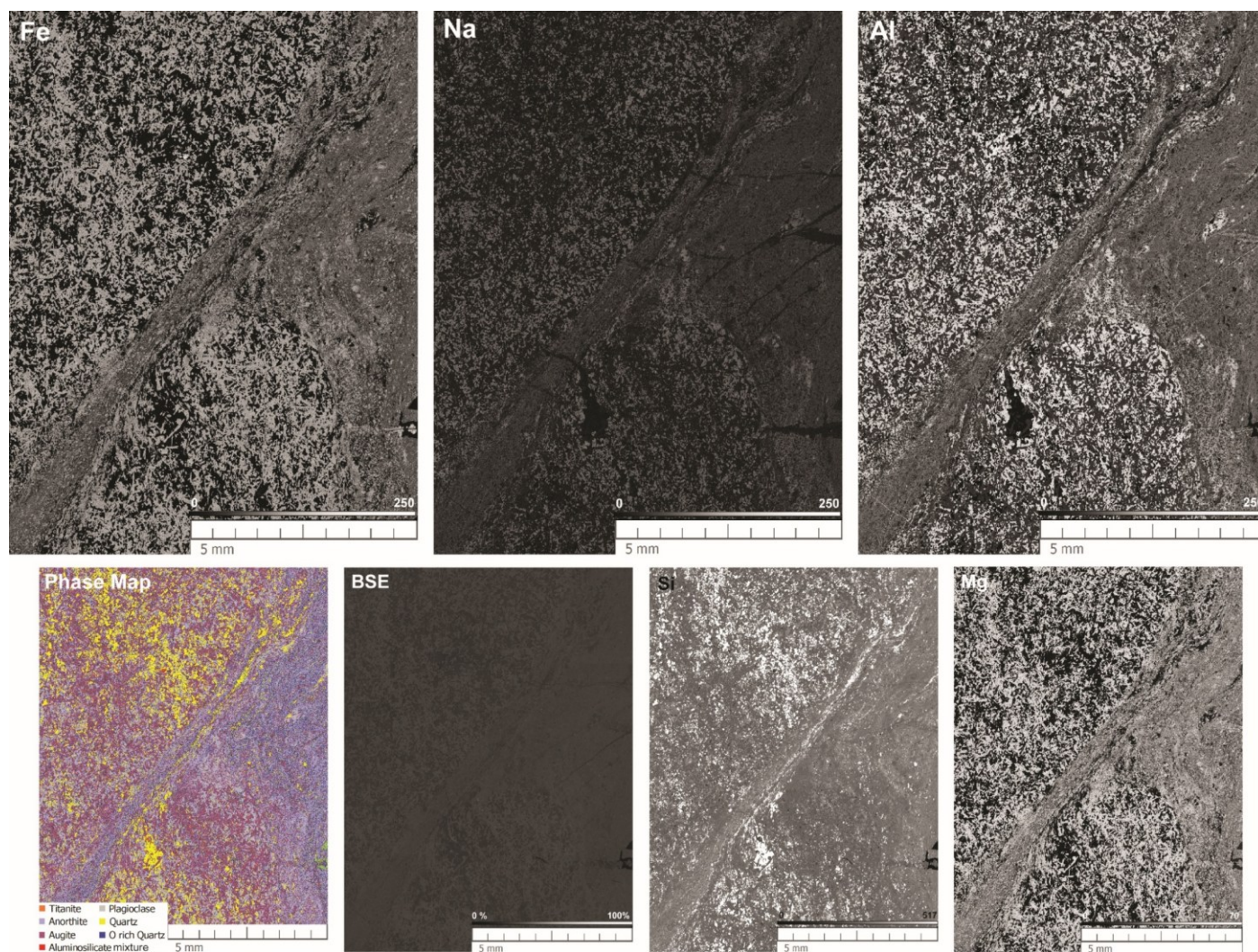


Figure A9: From left to right: Iron, sodium, aluminium, phase map, BSE, silica, and magnesium elemental maps of Sample RAS-107, a metabasalt pseudotachylyte. Sample location UTM 35J 532750 E 7026750 S

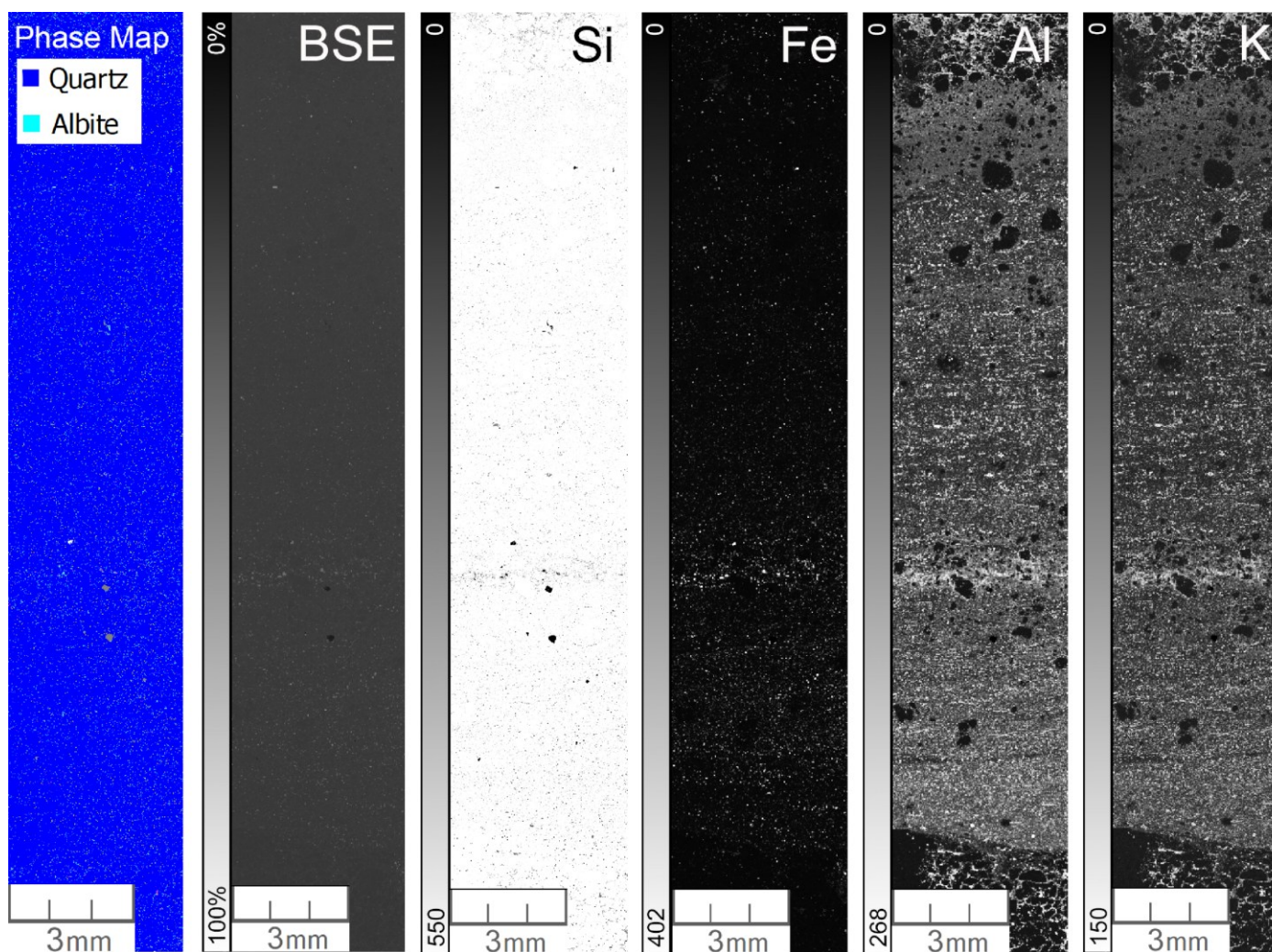


Figure A10: From left to right: Phase map, BSE, silica, iron, aluminium, and potassium elemental maps of Sample RAS-102, a quartzite pseudotachylyte. Sample location UTM 35J 0537659 E 7028591 S

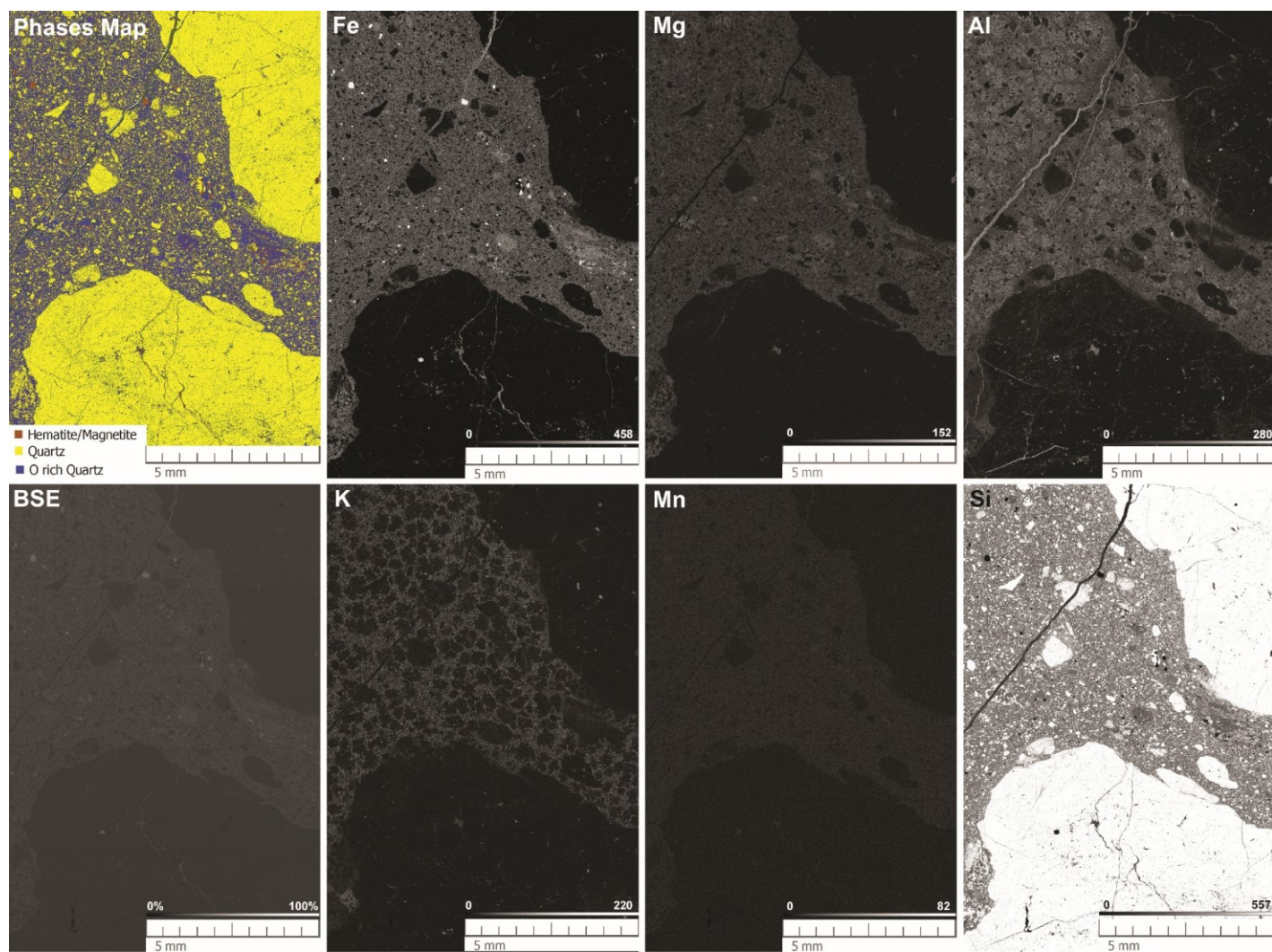


Figure A11: From left to right: Phase map, iron, magnesium, aluminium, BSE, potassium, manganese, and silica elemental maps of Sample RAS-110, a quartzite/ironstone pseudotachylyte. Sample location UTM 35J 0538587 E 7028007 S

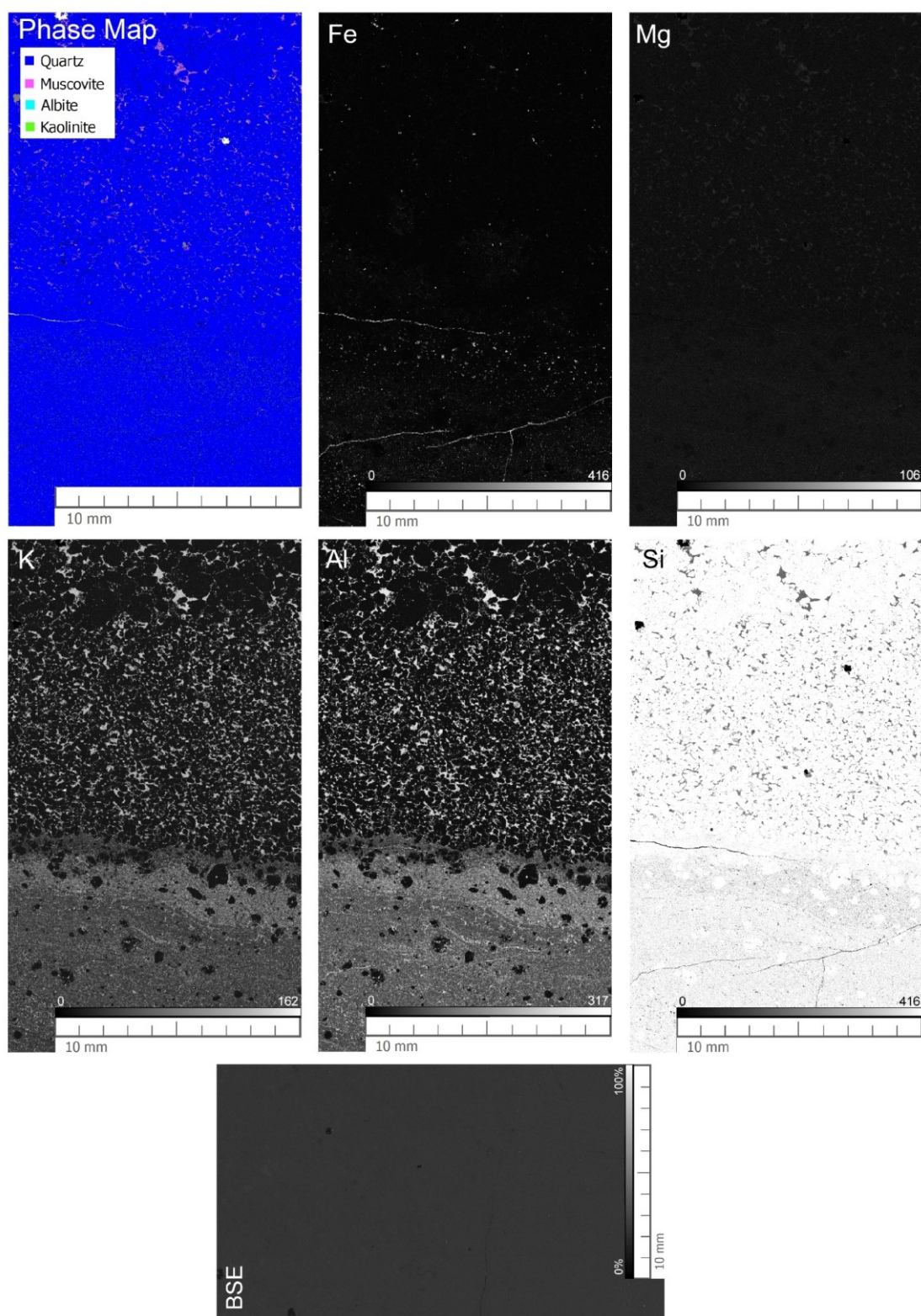


Figure A12: From left to right: Phase map, iron, magnesium, potassium, aluminium, silica, and BSE maps of Sample RAS-101, a metabasalt pseudotachylyte. Sample location location 35 J 0537618 E 7028608 S.

APPENDIX B

Bedding

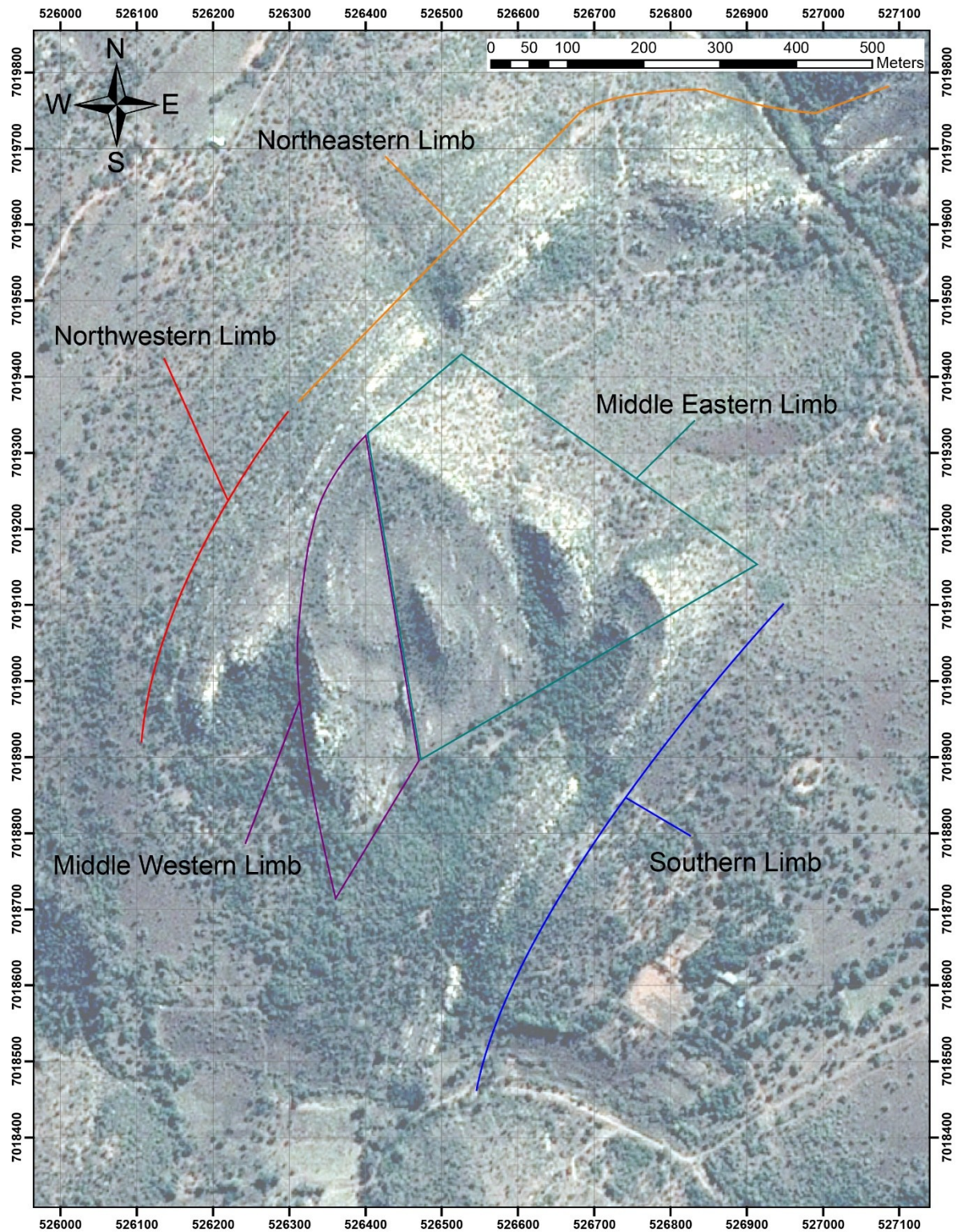


Figure B1: Google Earth Pro[®] map (imagery date: 06-11-2017) showing the various limbs of the Nauwpoort Folds.

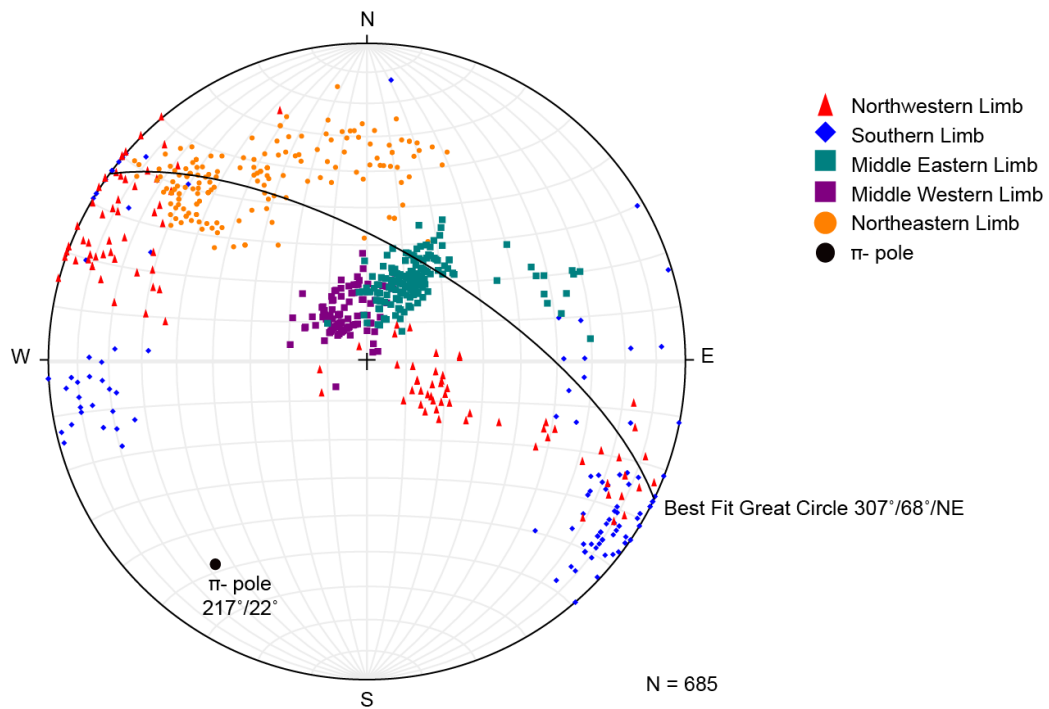


Figure B2: Equal area stereonet showing poles to all the bedding data and the best fit great circle 307°/68°/NE, π -pole 217°/22°. The bedding is inconsistent with uniform strata or that of a cylindrically folded sequence.

Google Earth Pro map ® showing all the collected bedding data

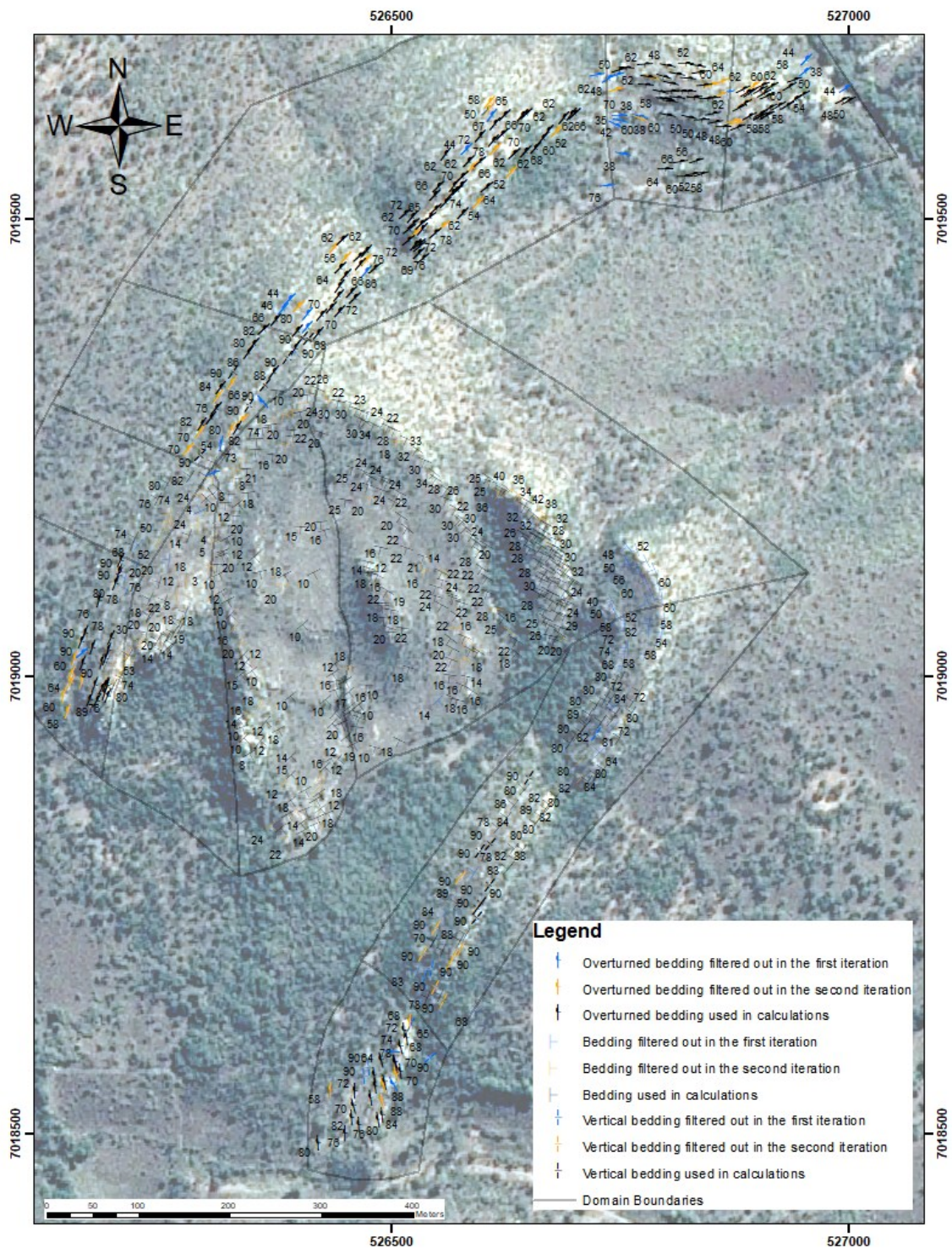


Figure B3: Google Earth Pro® map (imagery date: 06-11-2017) showing all the collected bedding measurements from the Nauwpoort locality.

Appendix B

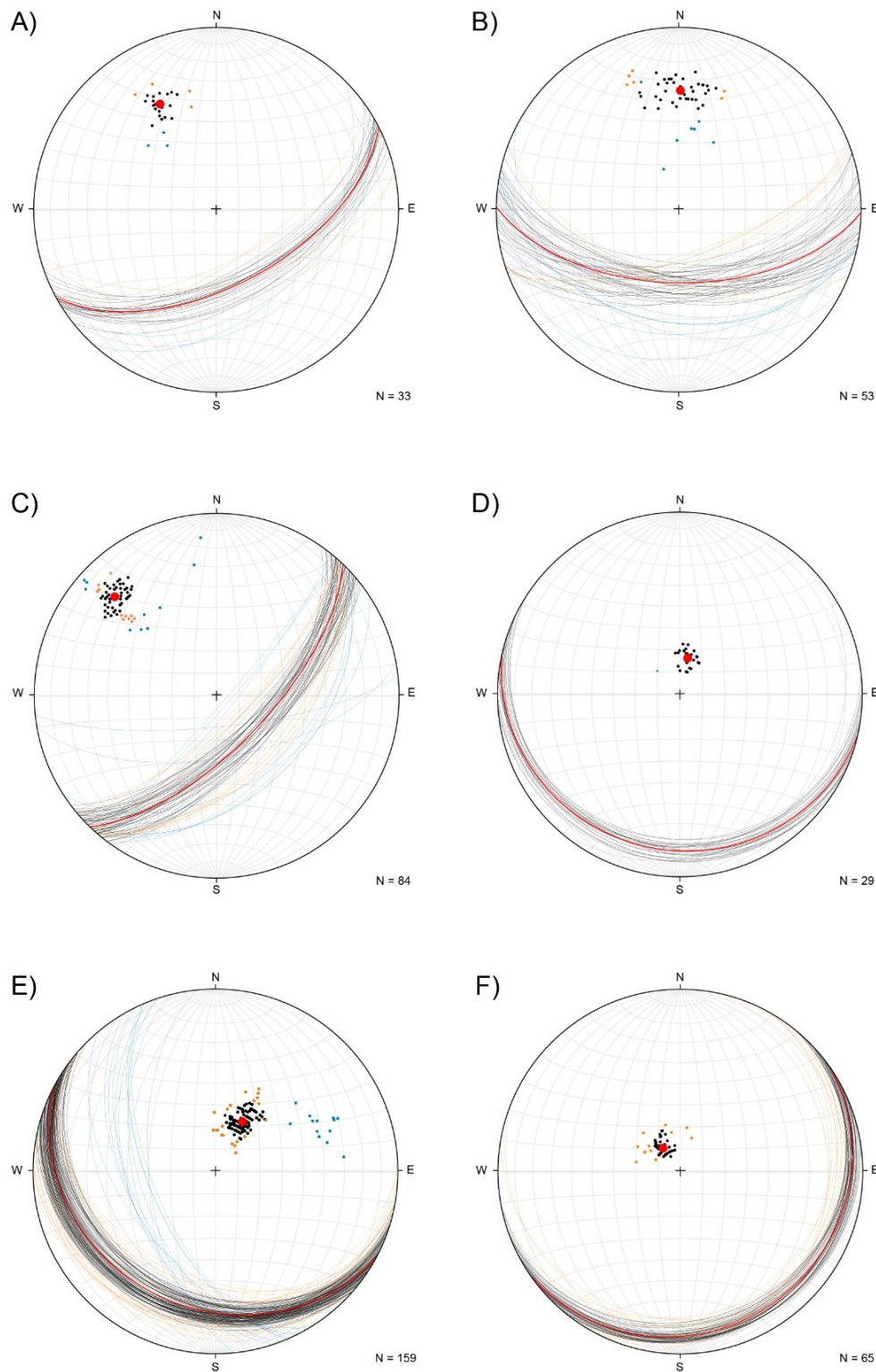


Figure B4: Equal area stereonets showing bedding measurements for domains A-F. The measurements in blue and orange were filtered out in the first and second iterations of filtration, respectively. The bold red plane corresponds to the mean bedding plane after filtration, recorded below in Table B1.

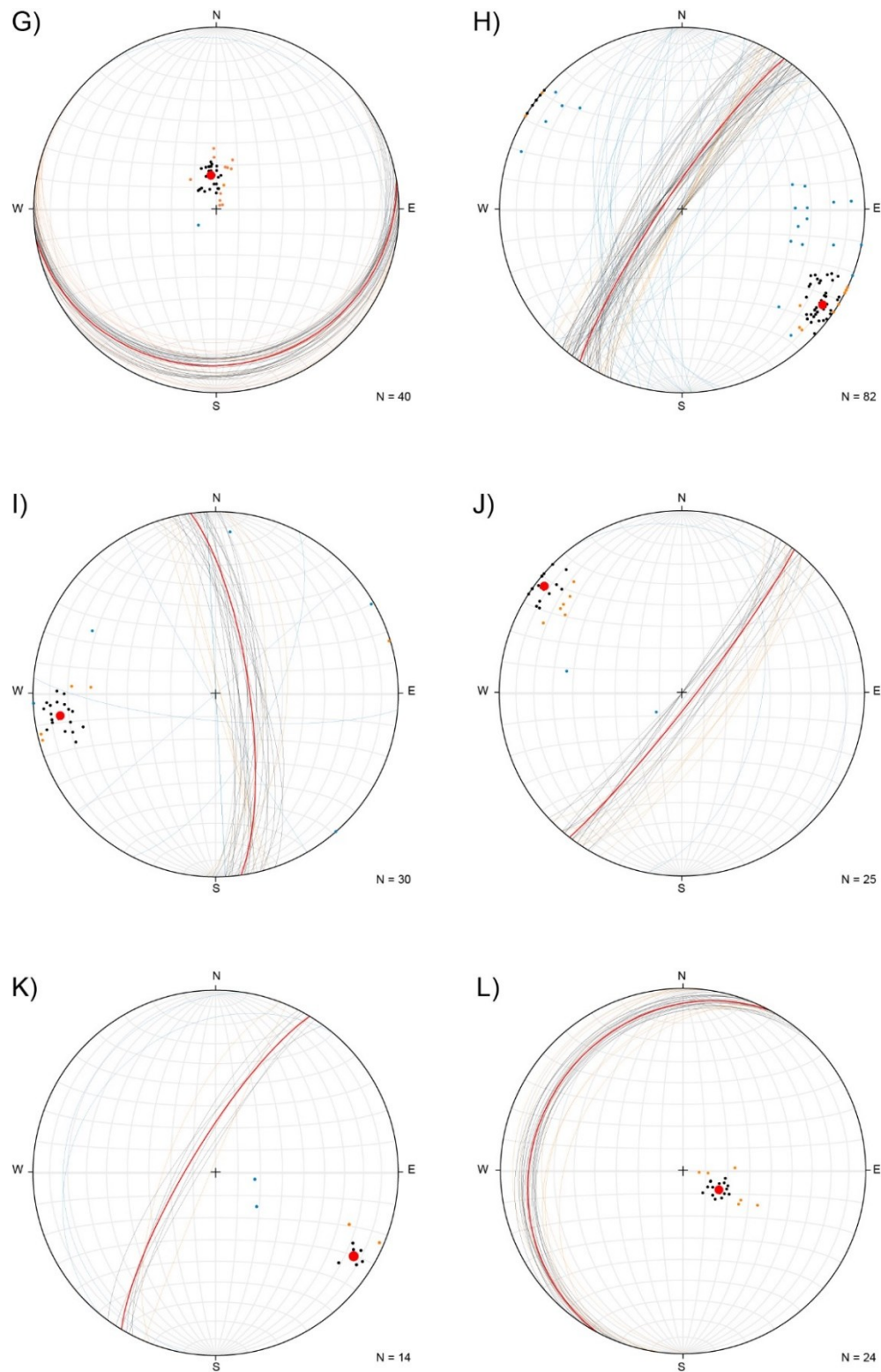


Figure B5: Equal area stereonets showing bedding measurements for domains G-L. The measurements in blue and orange were filtered out in the first and second iterations of filtration, respectively. The bold red plane corresponds to the mean bedding plane after filtration, recorded above below table B1.

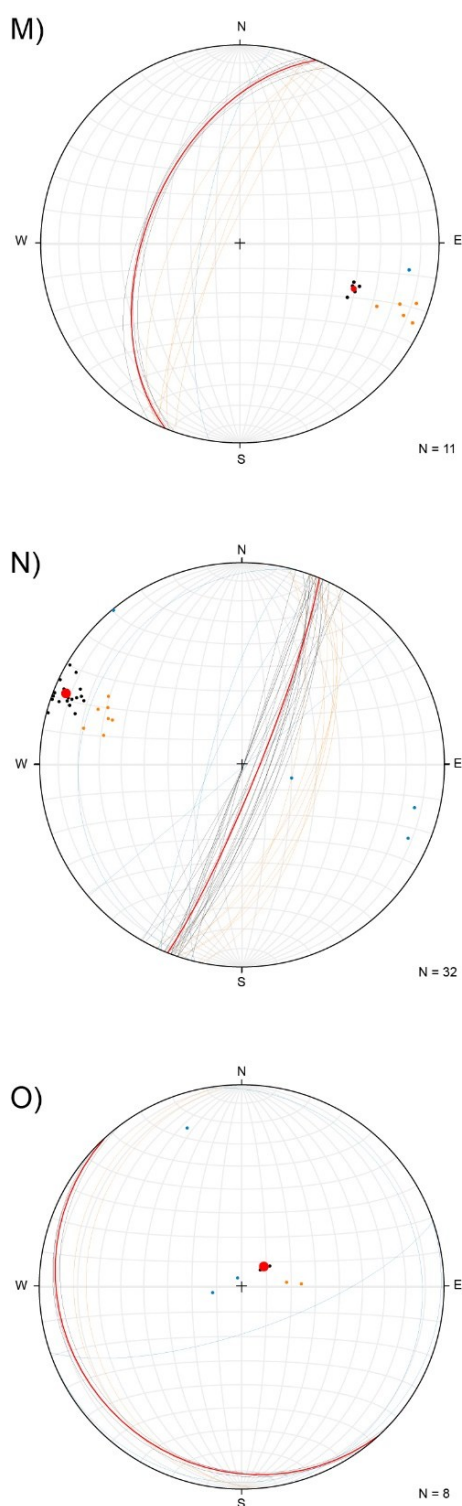


Figure B6: Equal area stereonets showing bedding measurements for domains M-O. The measurements in blue and orange were filtered out in the first and second iterations of filtration, respectively. The bold red plane corresponds to the mean bedding plane after filtration, recorded below in Table B1.

Table B1: Bedding results for each domain, showing the number of measurements taken (N), standard deviation (STD), dip direction (DD) and the resultant values for each successive iteration of filtration.

		Average bedding orientation before filtration					Average bedding orientation after first filtration					Average bedding orientation after second filtration							
Domain	Limb	Strike	Dip	DD	STD	N	Strike	Dip	DD	STD	N	Strike	Dip	DD	STD	N			
A	Northeastern Limb	061	56	SSE	6	7	33	063	57	SSE	5	5	30	062	57	SSE	4	4	26
B	Northeastern Limb	090	55	S	12	10	53	090	58	S	12	6	45	091	57	SSW	10	5	38
C	Northeastern Limb	045	66	ESE	7	9	84	044	67	ESE	4	6	73	044	67	ESE	4	4	60
D	Middle Eastern Limb	101	16	SSW	15	3	29	103	16	SSW	10	3	28						
E	Middle Eastern Limb	121	28	SSW	14	9	159	118	25	SSW	9	5	146	118	26	SSW	7	4	125
F	Middle Western Limb	055	14	SSE	16	4	65						055	13	SSE	10	3	53	
G	Middle Western Limb	094	16	SSW	41	6	40	088	16	SSE	19	6	39	081	16	SSE	13	4	27
H	Southern Limb	210	79	WNW	13	10	82	214	83	WNW	6	6	61	214	82	WNW	5	6	50
I	Southern Limb	346	77	ENE	28	9	30	352	76	ENE	6	8	24	352	75	ENE	5	6	19
J	Northwestern Limb	041	78	ESE	23	16	25	037	81	ESE	5	8	23	038	85	ESE	5	5	17
K	Northwestern Limb	208	67	WNW	9	25	10	209	78	WNW	5	7	8	211	78	WNW	3	4	6
L	Northwestern Limb	205	20	WNW	11	6	24						208	19	WNW	9	3	18	
M	Northwestern Limb	201	65	WNW	5	14	11	202	64	WNW	3	14	10	202	51	WNW	3	1	5
N	Northwestern Limb	038	77	ESE	52	14	32	021	78	ESE	4	10	28	022	84	ESE	4	5	21
O	Northwestern Limb	168	13	WSW	38	16	8	152	16	WSW	22	6	5	137	12	WSW	6	2	3

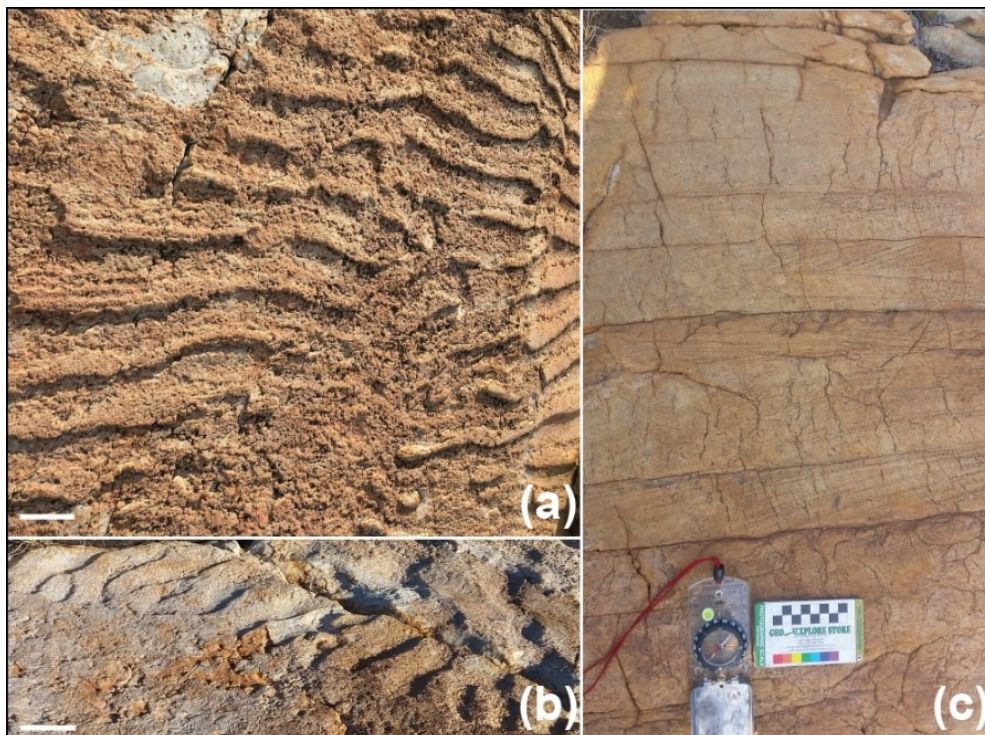


Figure B7: Sedimentary structures preserved in Rhq1 quartzite on Nauwpoort locality. The white line in each of the above photographs represents 10 cm. **a)** Asymmetrical ripples with a ripple index of 2.4 (35J 0526664 E 7019151 S); **b)** Asymmetric current ripples with a ripple

Appendix B

index of 3.34 (35J 0526643 E 7019184 S); **c)** Herringbone cross-bedding (35J 0526190 E 7019026 S). Cross bedding is used to determine right-way-up of bedding.

Domain A rotated to Domain B

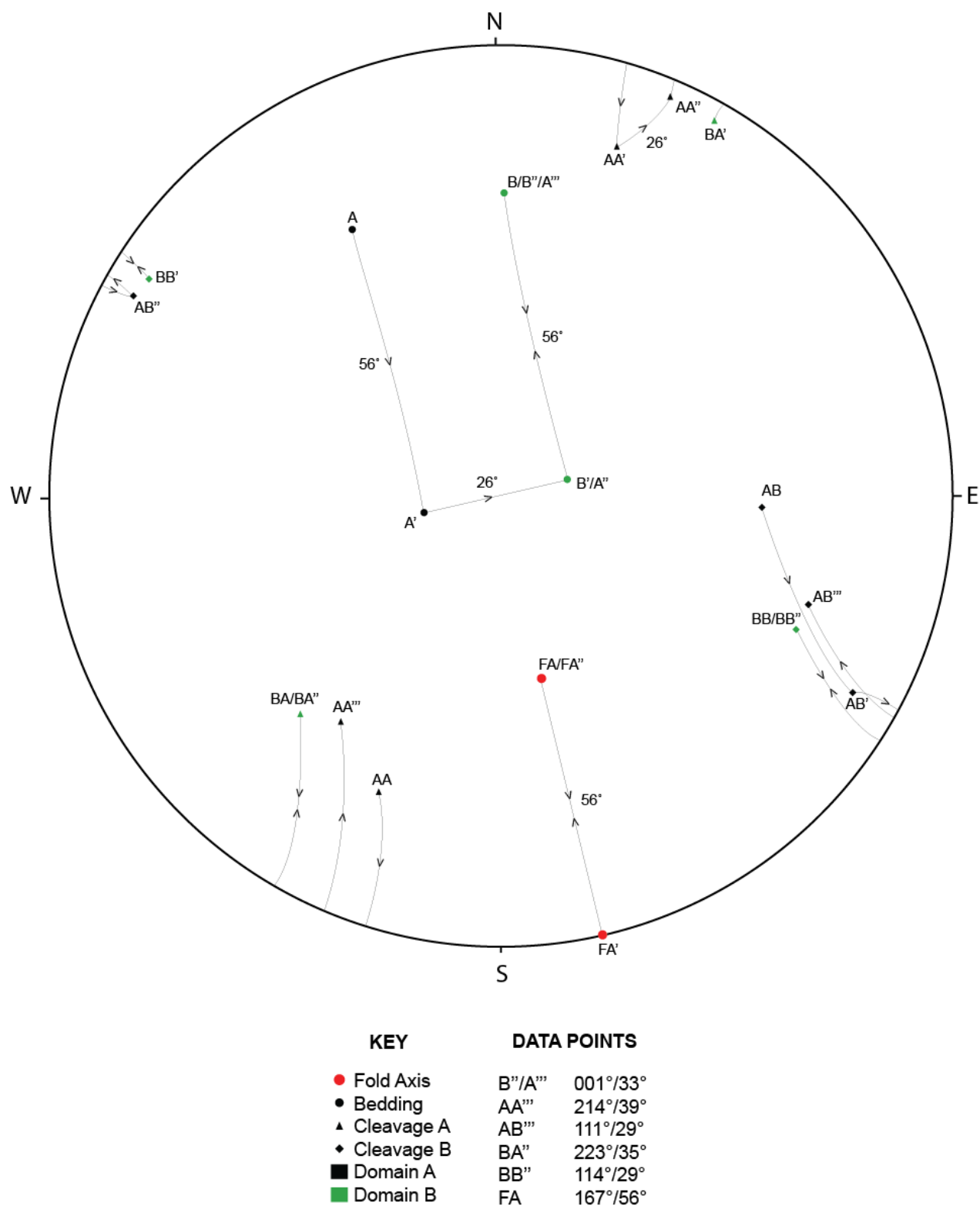


Figure B8: Equal area stereonet showing the rotation of Domain A to Domain B.

Domains AB rotated to Domain C

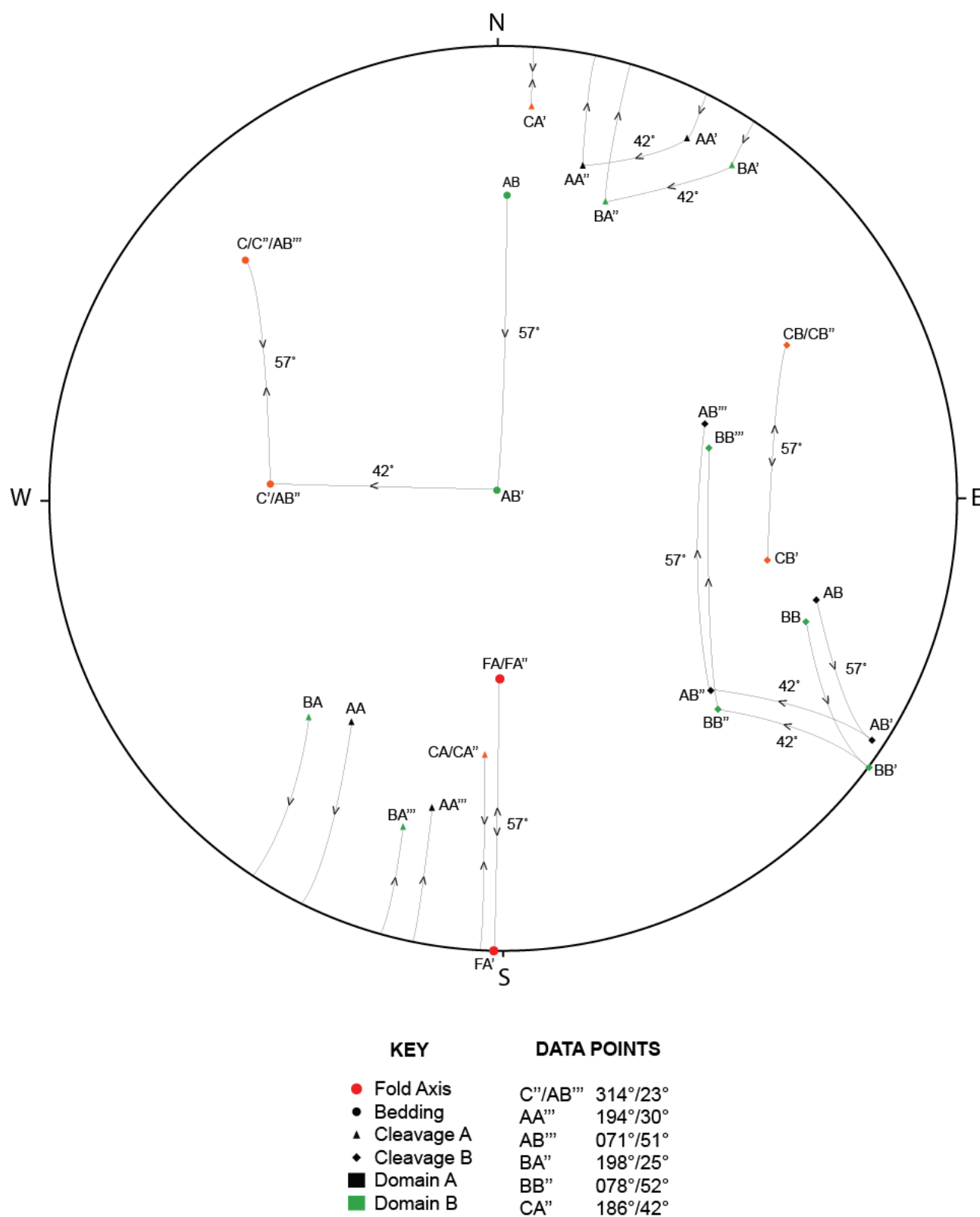


Figure B9: Equal area stereonet showing the rotation of Domain AB to Domain C.

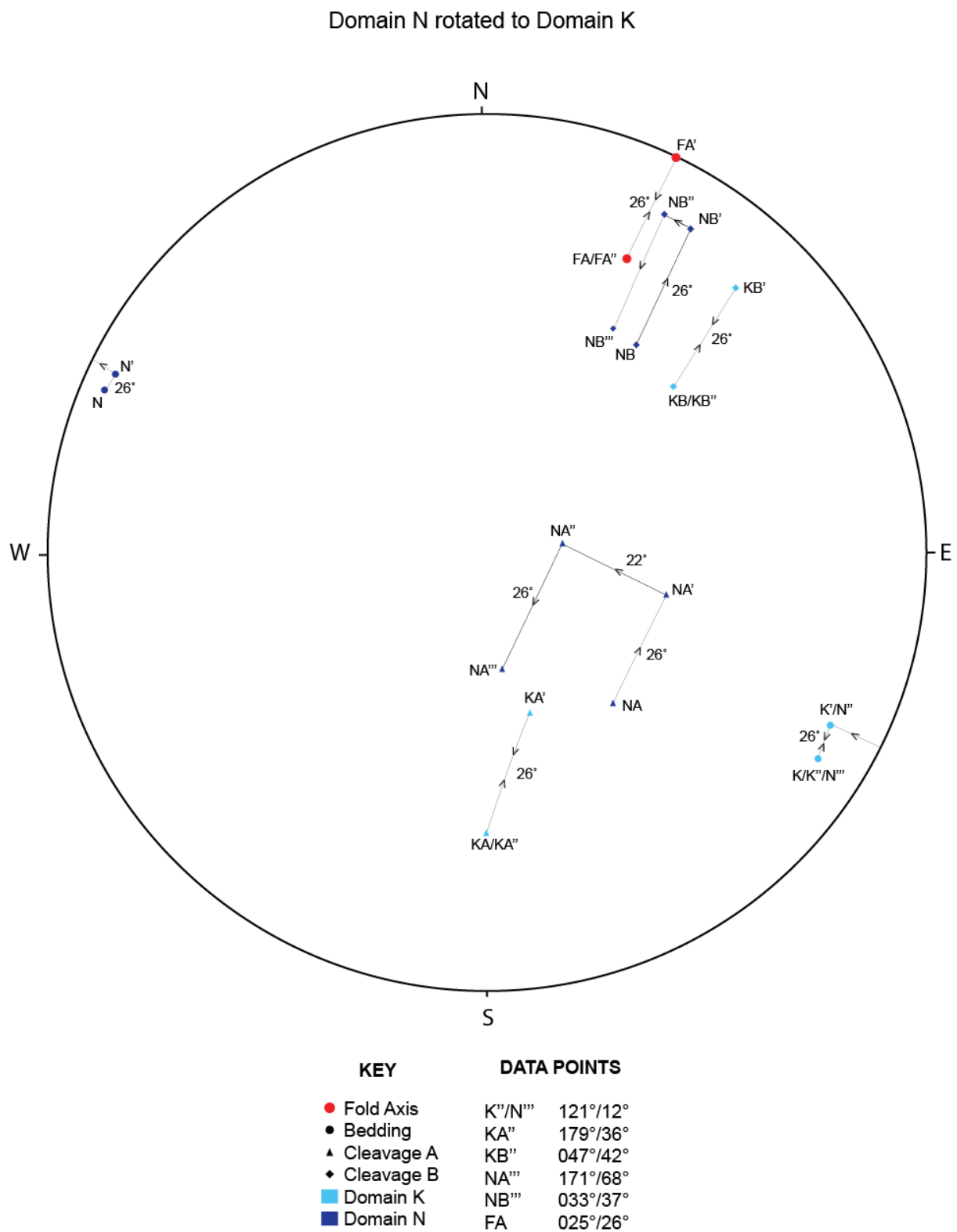


Figure B10: Equal area stereonet showing the rotation of Domain N to Domain K.

Domain KN rotated to Domain J

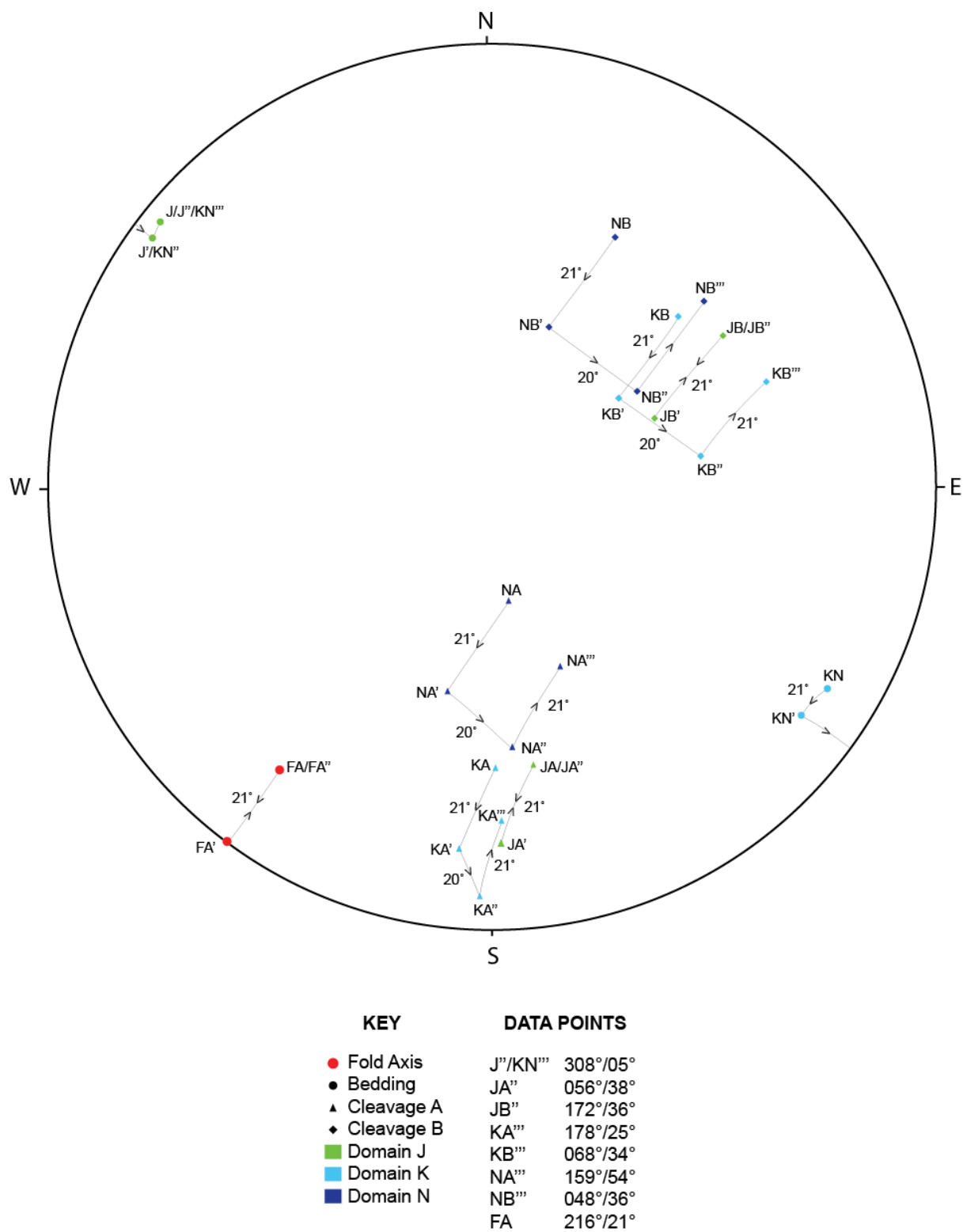


Figure B11: Equal area stereonet showing the rotation of Domain KN to Domain J.

Domain ABC rotated to Domain JKN

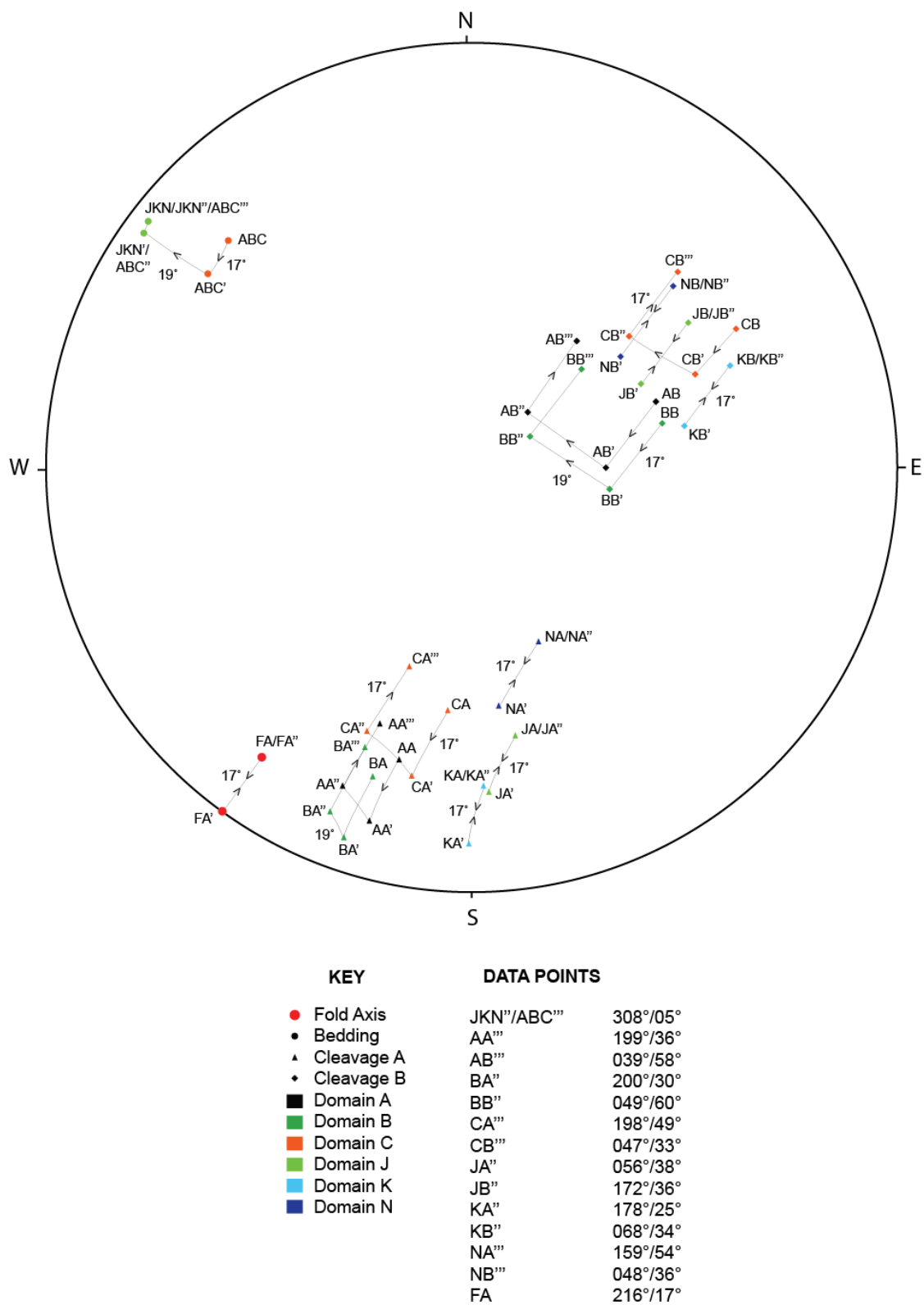


Figure B12: Equal area stereonet showing the rotation of Domain ABC to Domain JKN.

Domain DE and FG unfolded about an inclined fold axis

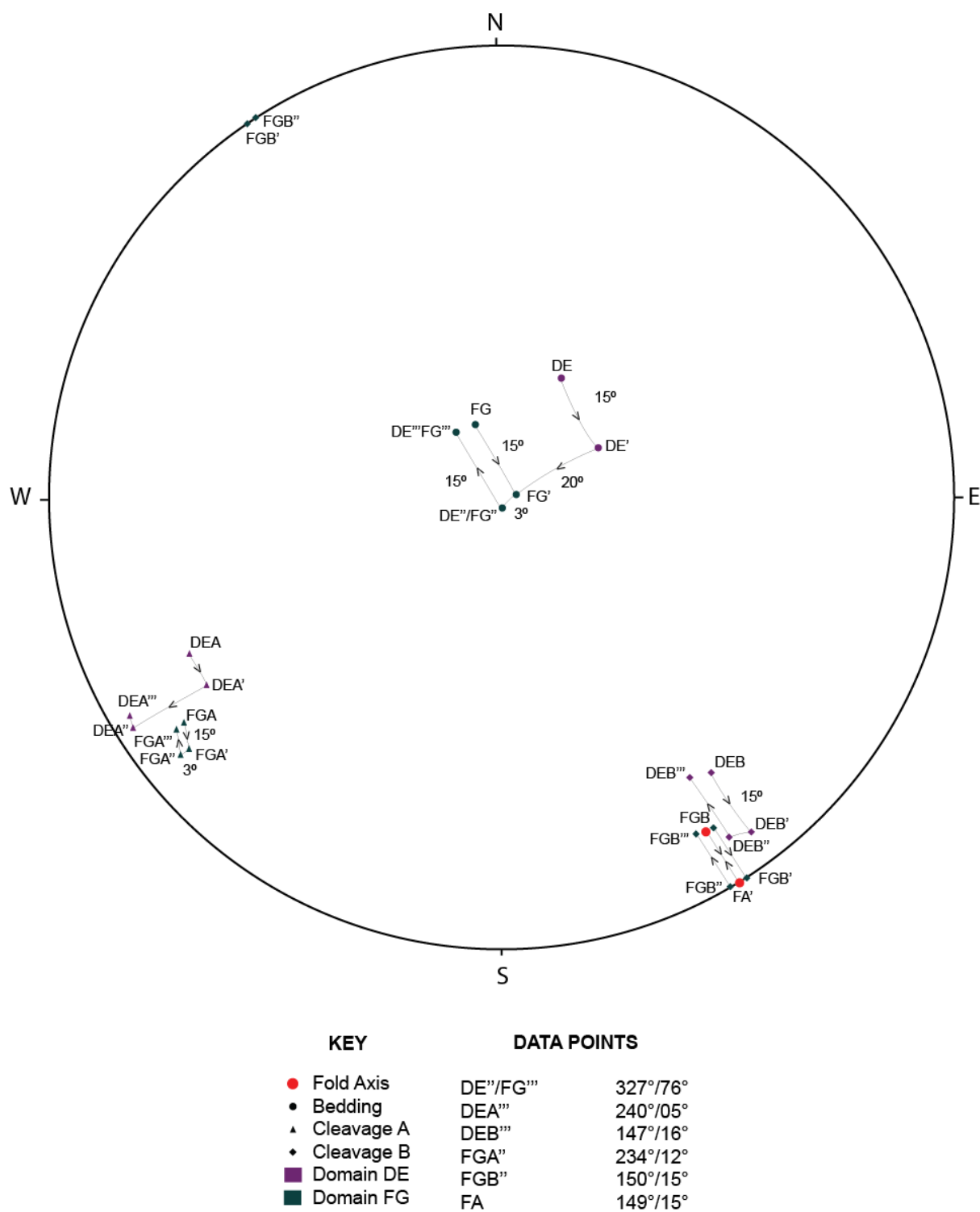


Figure B13: Equal area stereonet showing the rotation of Domain DE and Domain FG and about an inclined fold axis.

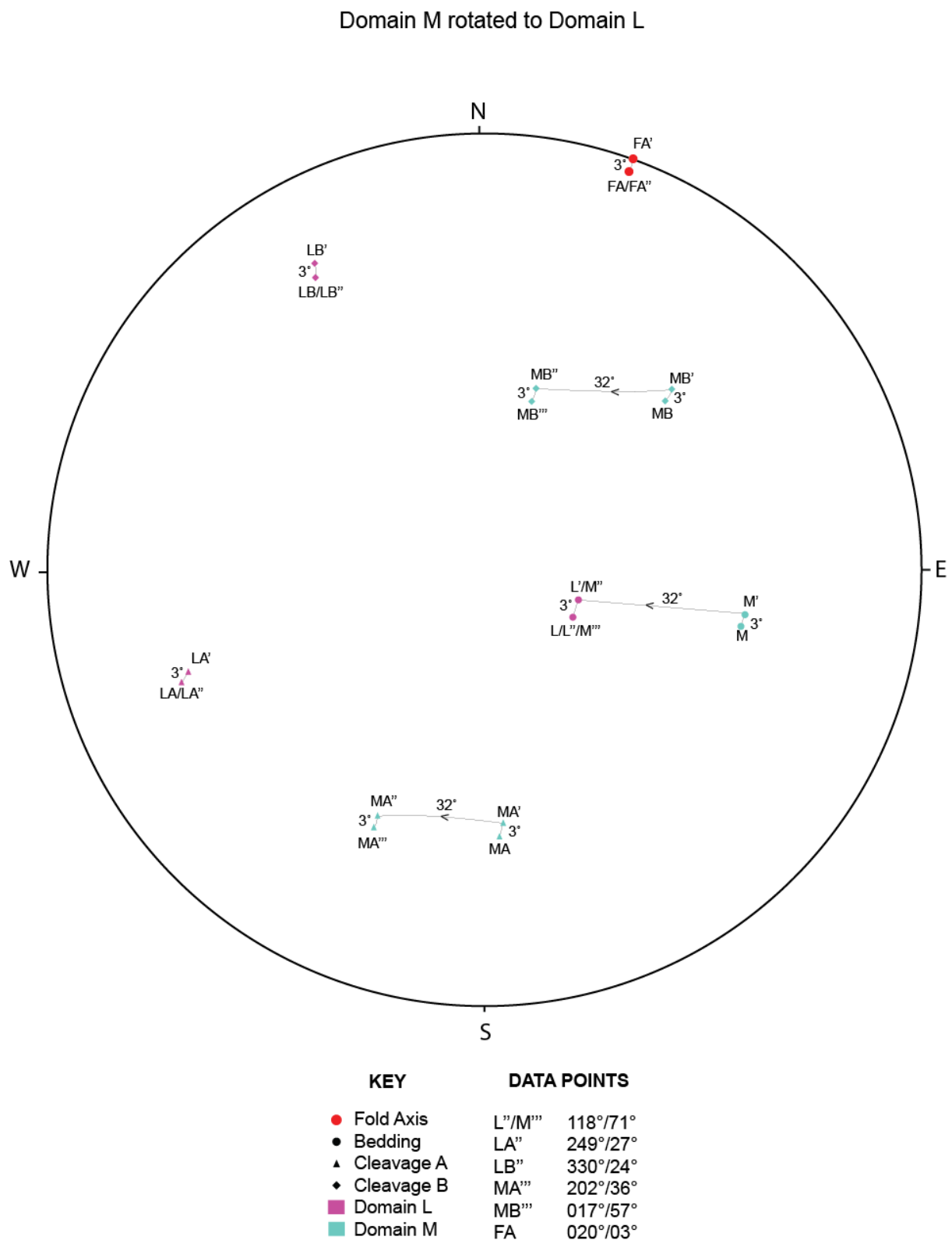


Figure B14: Equal area stereonet showing the rotation of Domain M to Domain L.

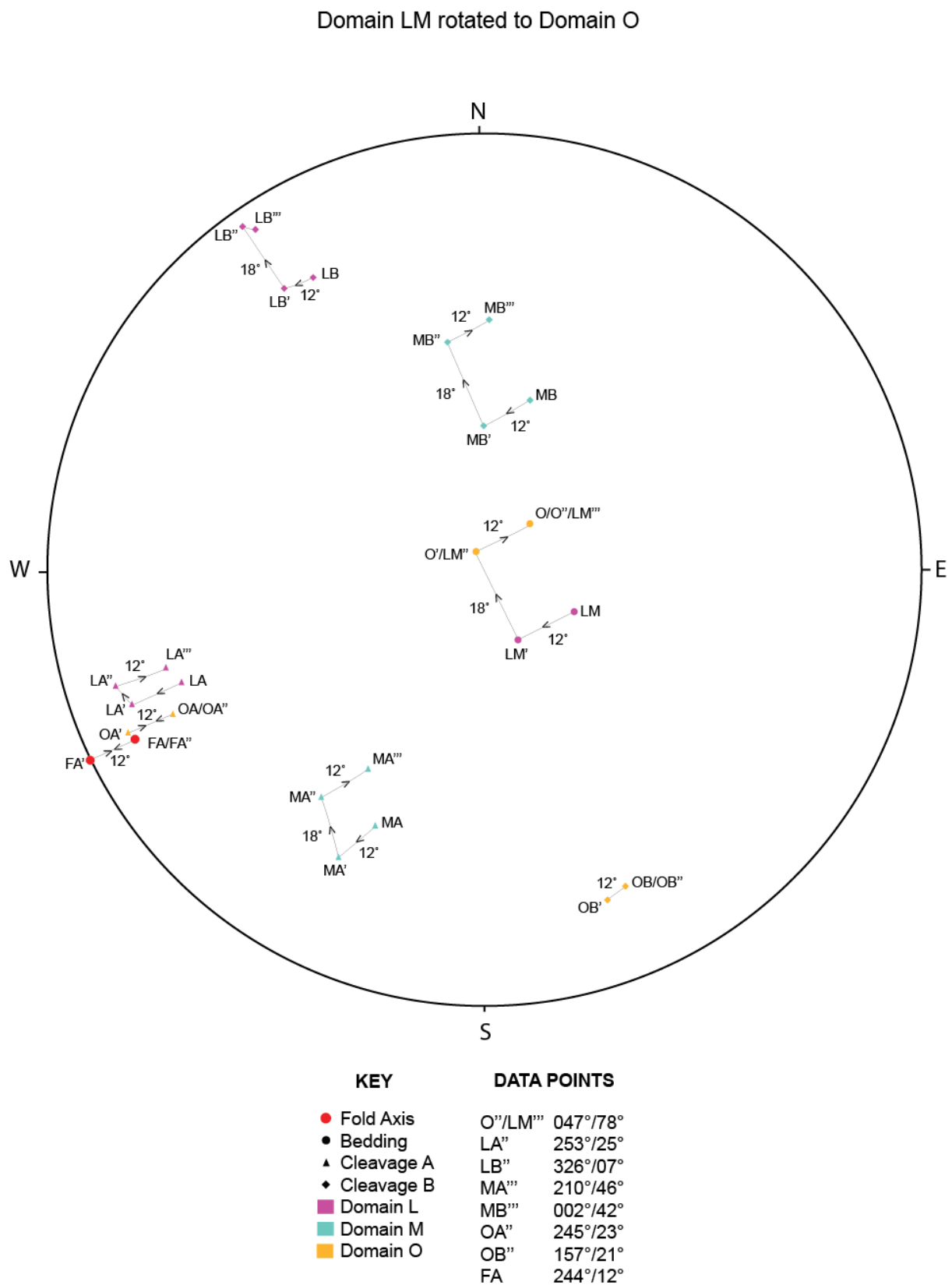


Figure B15: Equal area stereonet showing the rotation of Domain LM to Domain O.

Domain LMO rotated to Domain DEFG

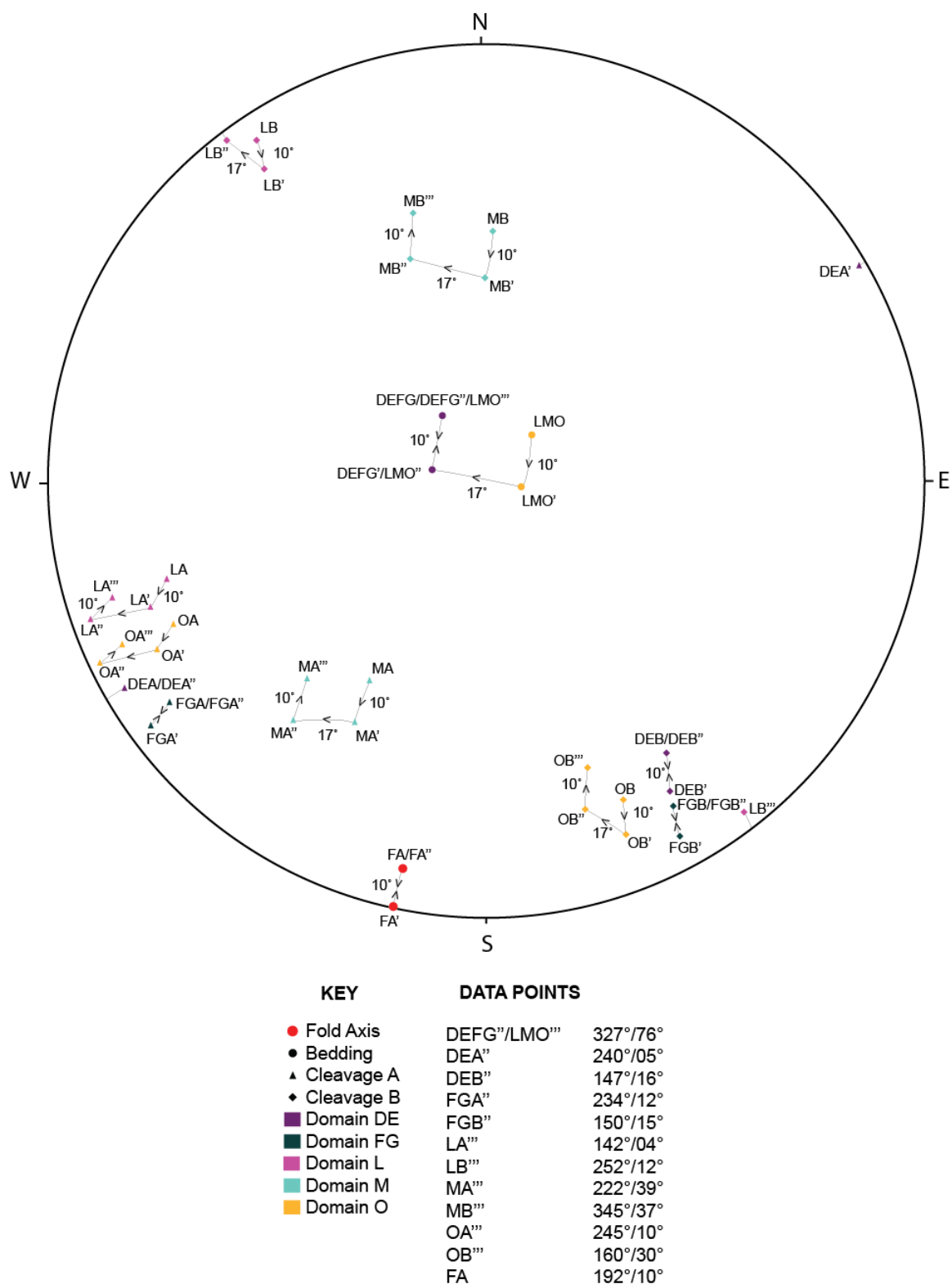


Figure B16: Equal area stereonet showing the rotation of Domain LMO to Domain DEFG.

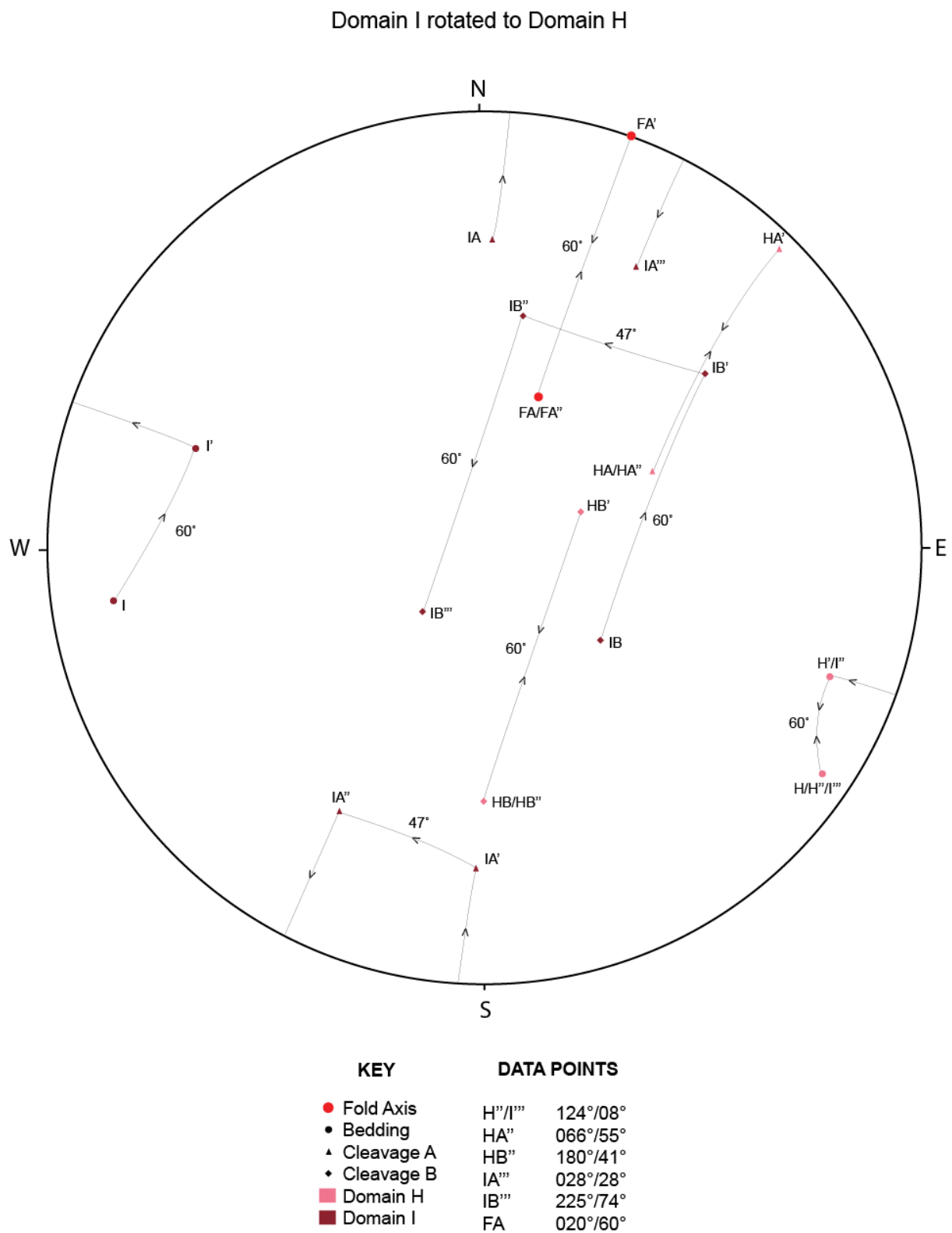


Figure B17: Equal area stereonet showing the rotation of Domain I to Domain H.

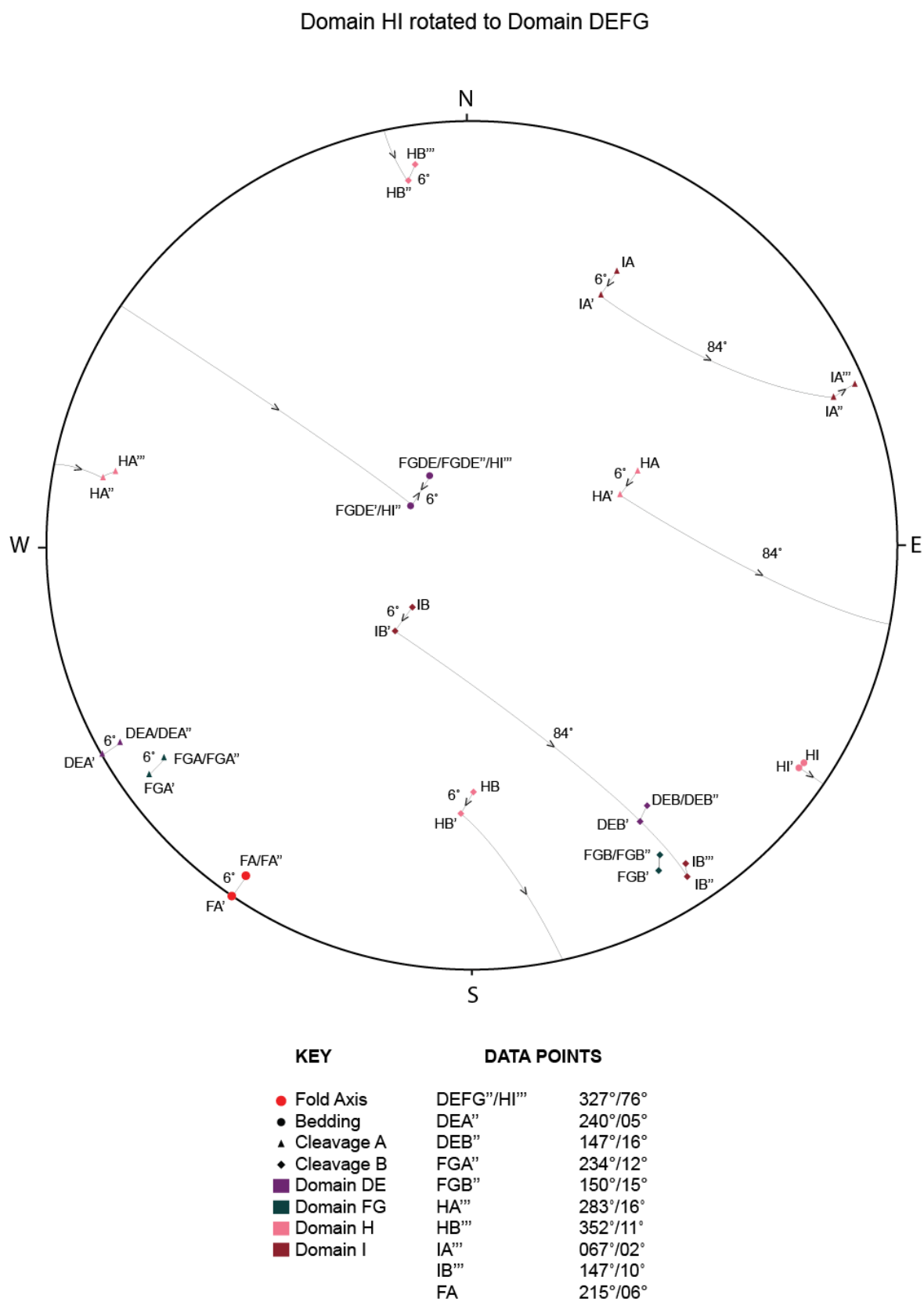


Figure B18: Equal area stereonet showing the rotation of Domain HI to Domain DEFG.

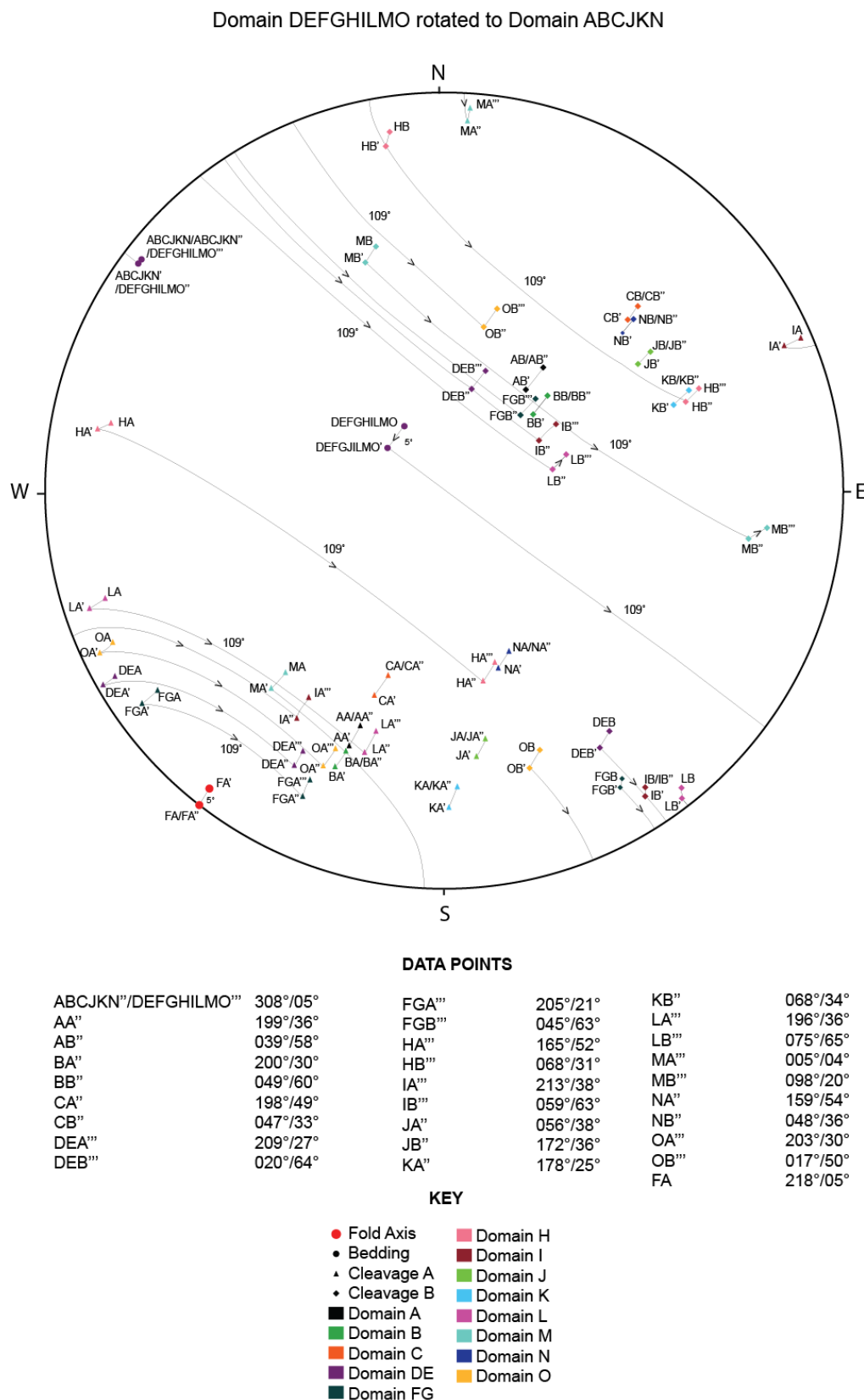


Figure B19: Equal area stereonet showing the rotation of Domain DEFGHILMO to Domain ABCJKN.

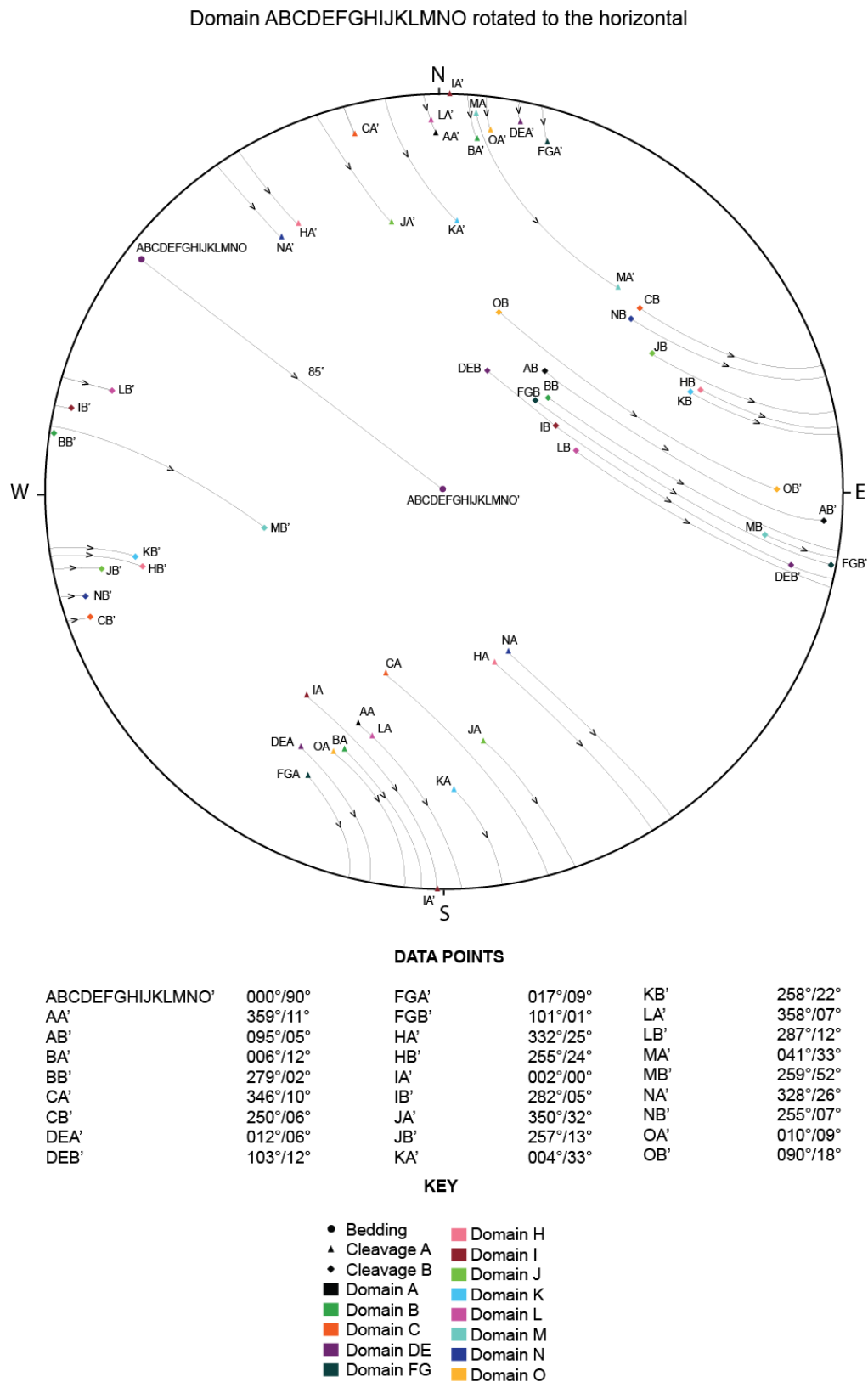
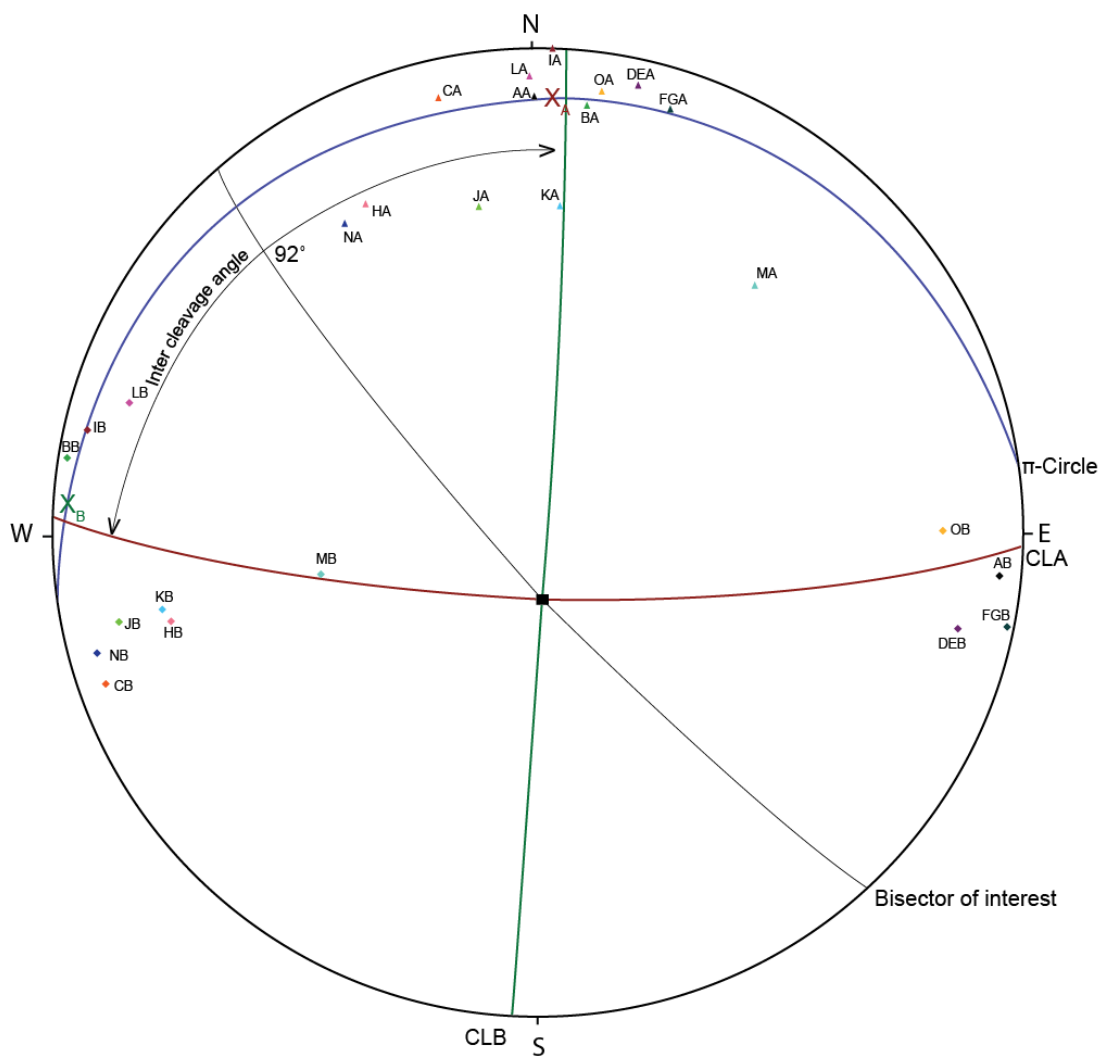


Figure B20: Equal area stereonet showing the rotation of Domain ABCDEFGHIJKLMNO to the horizontal.

Appendix B

Summary stereonet presenting the results of the final rotation



DATA POINTS

AA	359°/11°	FGA	017°/09°	KB	258°/22°
AB	095°/05°	FGB	101°/01°	LA	358°/07°
BA	006°/12°	HA	332°/25°	LB	287°/12°
BB	279°/02°	HB	255°/24°	MA	041°/33°
CA	346°/10°	IA	002°/00°	MB	259°/52°
CB	250°/06°	IB	282°/05°	NA	328°/26°
DEA	012°/06°	JA	350°/32°	NB	255°/07°
DEB	103°/12°	JB	257°/13°	OA	010°/09°
		KA	004°/33°	OB	090°/18°

KEY

● Bedding	Domain H	X_A	002°/11°
▲ Cleavage A	Domain I	X_B	273°/01°
◆ Cleavage B	Domain J	CLA Weighted Mean Plane of Cleavage A	092°/79°/S
■ Domain A	Domain K	CLB Weighted Mean Plane of Cleavage B	003°/89°/E
■ Domain B	Domain L	Bisector of interest	138°/83°/SW
■ Domain C	Domain M	π-Circle	262°/12°/NW
■ Domain DE	Domain N	■ Line of intersection between CLA and CLB	178°/79°
■ Domain FG	Domain O		

Figure B21: Equal area stereonet showing the final poles with the weighted mean pole.

Appendix B

Table B2: Table showing the final weightings for each cleavage on the Nauwpoort locality.

Domain	CLA				CLB			
	No. of data points	Azimuth	Plunge	Weight %	No. of data points	Azimuth	Plunge	Weight %
A	26	359	11	6,2	10	095	05	4,1
B	35	006	12	8,4	12	279	02	5,0
C	57	346	10	13,6	29	250	06	12,0
DE	132	12	06	31,5	81	103	12	33,6
FG	69	17	09	16,5	40	101	01	16,6
H	41	332	25	9,8	22	255	24	9,1
I	19	002	00	4,5	9	282	05	3,7
J	11	350	32	2,6	6	257	13	2,5
K	2	004	33	0,5	4	258	22	1,7
L	15	358	07	3,6	8	287	12	3,3
M	5	41	33	1,2	5	259	52	2,1
N	4	328	26	1,0	12	255	07	5,0
O	3	10	09	0,7	3	090	18	1,2
Total or Weighted Mean	419	002	11	100	241	273	01	100

Table B3: Geluksoord locality bedding results for each domain, showing the number of measurements taken (N), standard deviation (STD), dip direction (DD) and the resultant values after filtration.

Domain	Limb	Average bedding orientation before filtration						Average bedding orientation after filtration					
		Strike	Dip	DD	STD	N		Strike	Dip	DD	STD	N	
A	Eastern Limb	288	50	NNE	13	6	6	291	50	NNE	11	7	5
B	Eastern Limb	331	62	NE	29	23	8	309	67	NNE	13	12	4
C	Eastern Limb	331	68	NE	21	21	6	330	81	NE	14	8	3
D	Southwestern Limb	293	44	NNE	9	6	5	290	41	NNE	5	3	4
E	Middle Limb	256	41	NNW	0	16	2						
F	Northwestern Limb	177	63	WSW	8	8	6	180	64	W	3	9	5
G	Northwestern Limb	182	71	W	11	12	3	188	64	WNW	3	6	2
H	Southwestern Limb	317	52	NNE	18	11	7	322	47	NE	3	9	4
I	Eastern Limb	018	85	ESE	14	7	2						

APPENDIX C

Cleavages

Appendix C

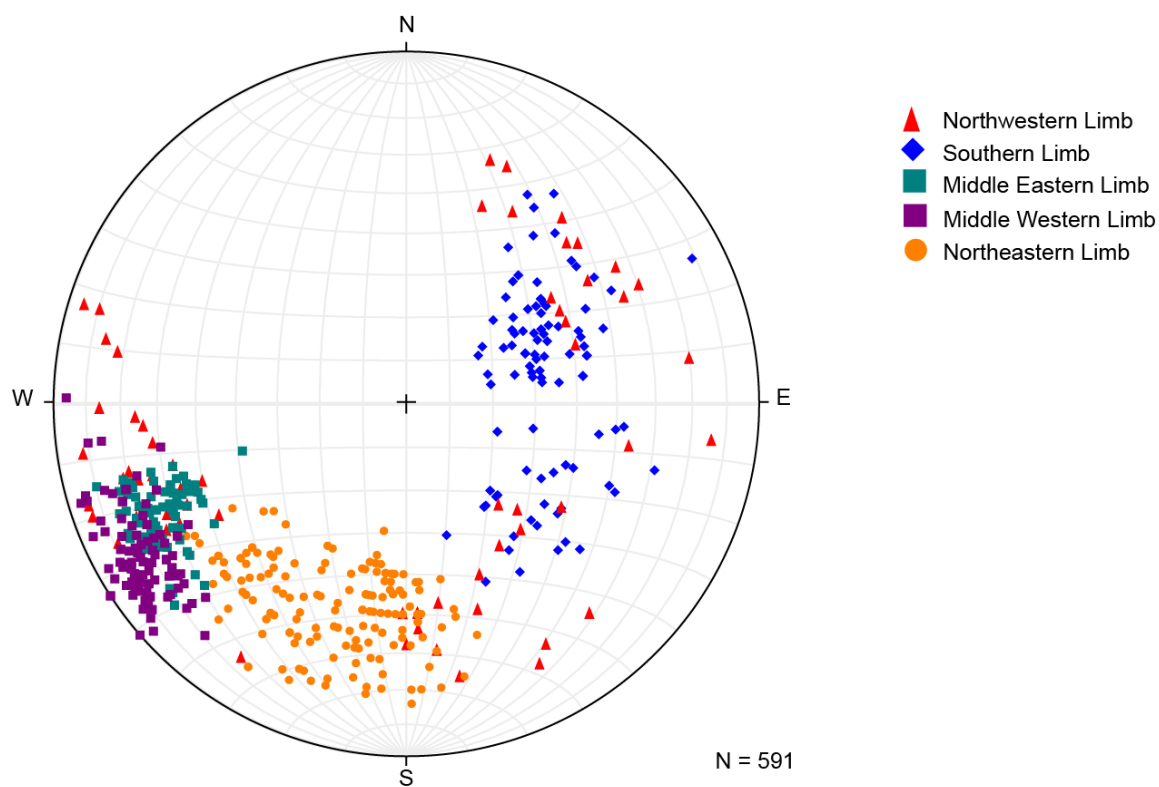


Figure C1: Equal area stereonet showing poles to all the CLA data on the Nauwpoort locality.

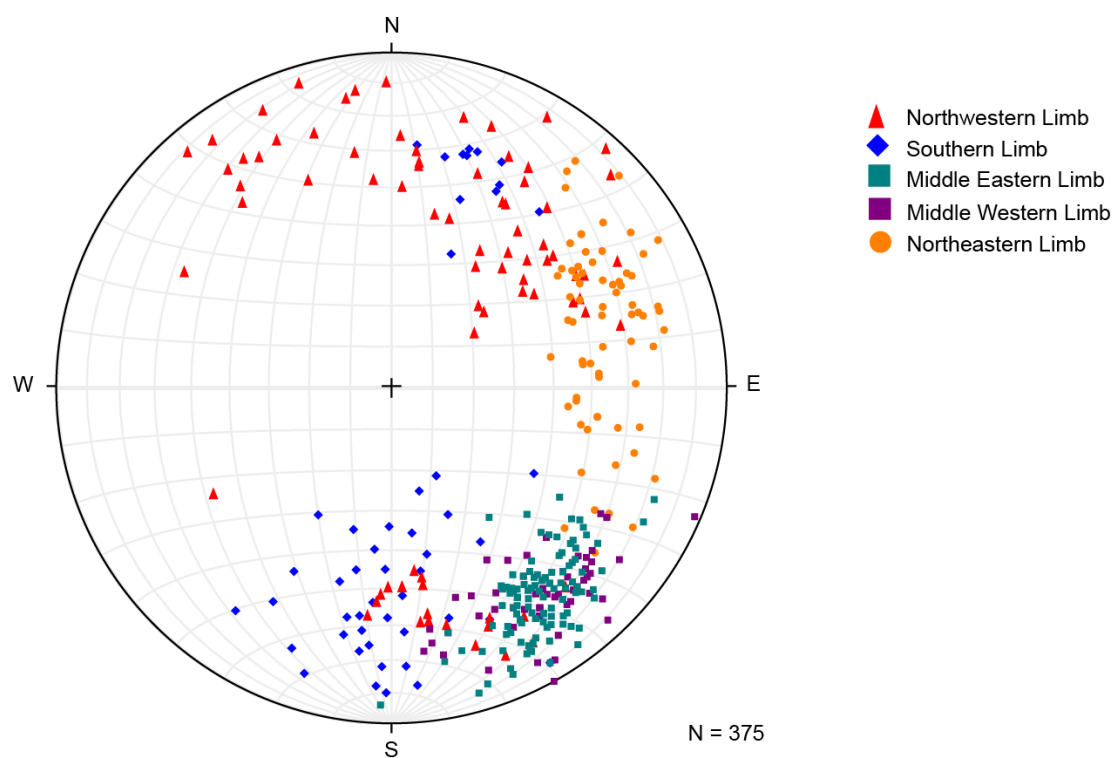


Figure C2: Equal area stereonet showing poles to all the CLB data.

Google Earth Pro map ® showing all the collected CLA data



Figure C3: Google Earth Pro® map showing CLA in the study area. Blue measurements were removed in the first iteration of filtration to show rotation along faults or in hinge zones. The second iteration of filtration (orange) was done visually on the stereonet to provide the least standard deviation and the tightest cluster of data points from which to obtain a robust average value for each domain.

Google Earth Pro map ® showing all the collected CLB data

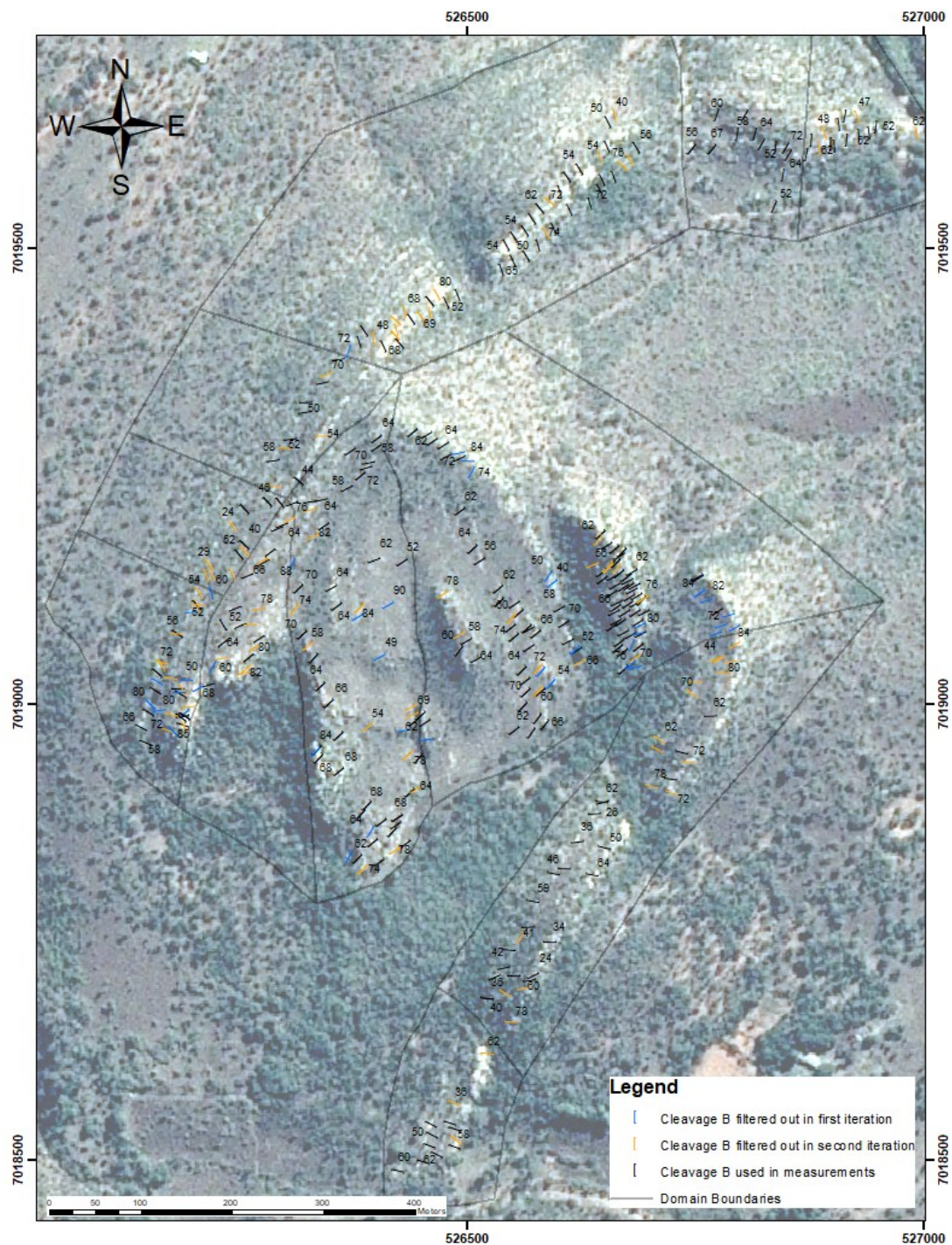
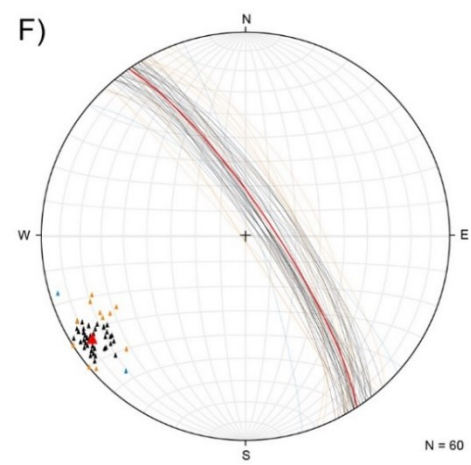
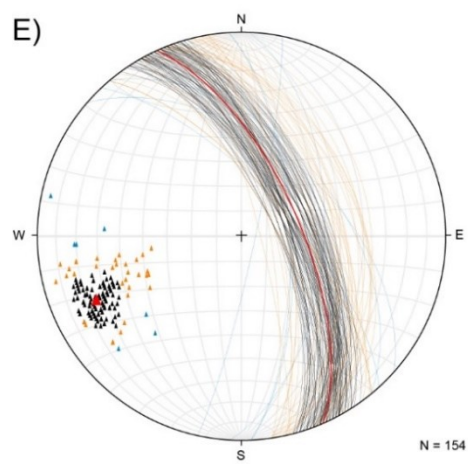
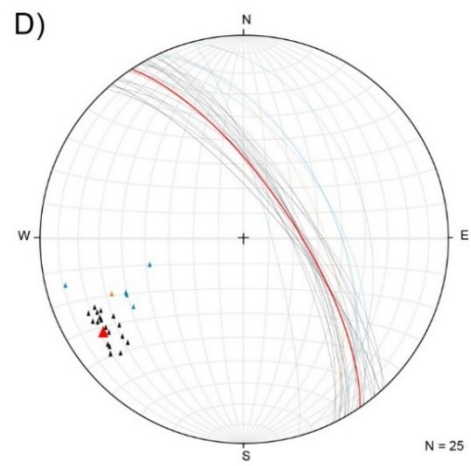
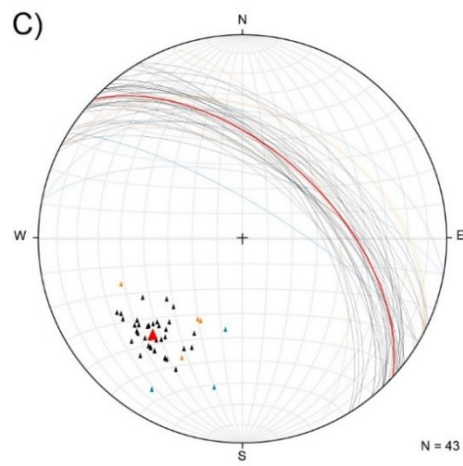
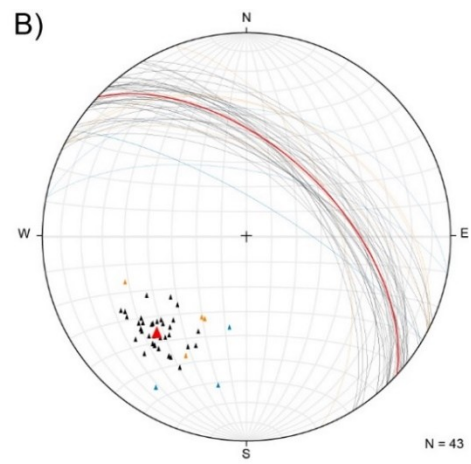
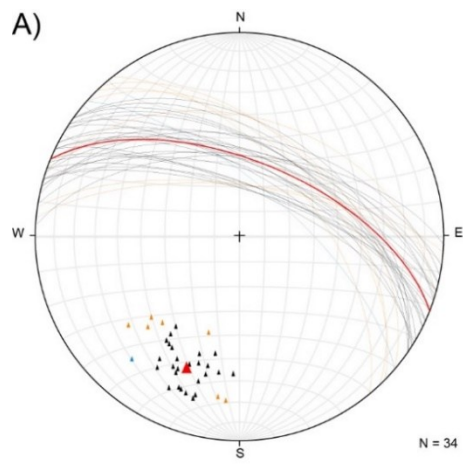


Figure C4: Google Earth Pro® map showing CLB in Study Area 1. Blue measurements were removed in the first iteration of filtration to show rotation along faults or in hinge zones. The second iteration of filtration (orange) was done visually on the stereonet to provide the least standard deviation and the tightest cluster of data points from which to obtain a robust average value for each domain.

Appendix C



Appendix C

Figure C5: Equal area stereonet showing CLA measurements for domains A-F. The measurements in blue and orange were filtered out in the first and second iterations of filtration, respectively. The bold red plane corresponds to the mean CLA plane after filtration.

Appendix C

Appendix C

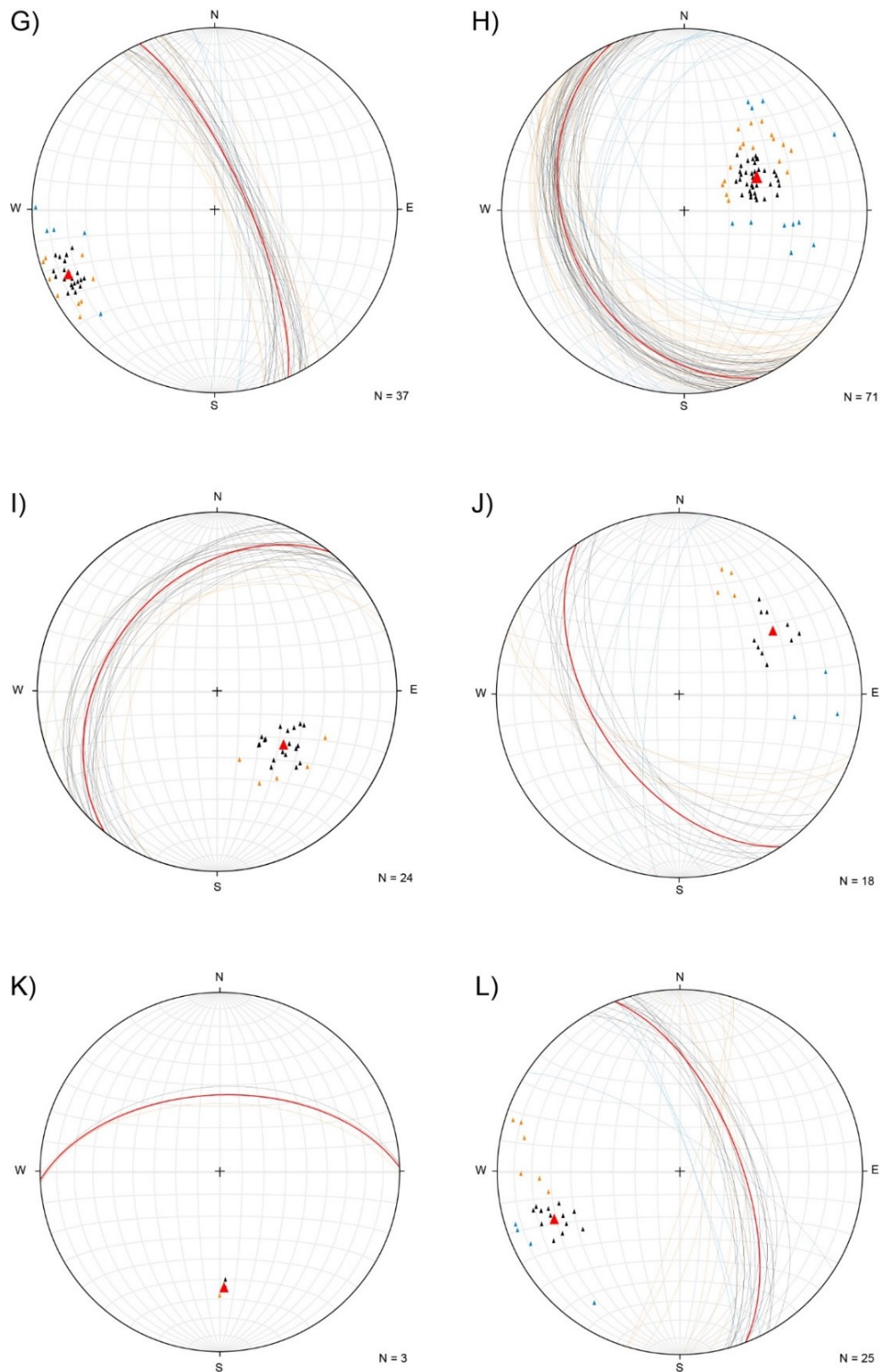


Figure C6: Equal area stereonet showing CLA measurements for domains G-L. The measurements in blue and orange were filtered out in the first and second iterations of filtration, respectively. The bold red plane corresponds to the mean CLA plane after filtration.

Appendix C

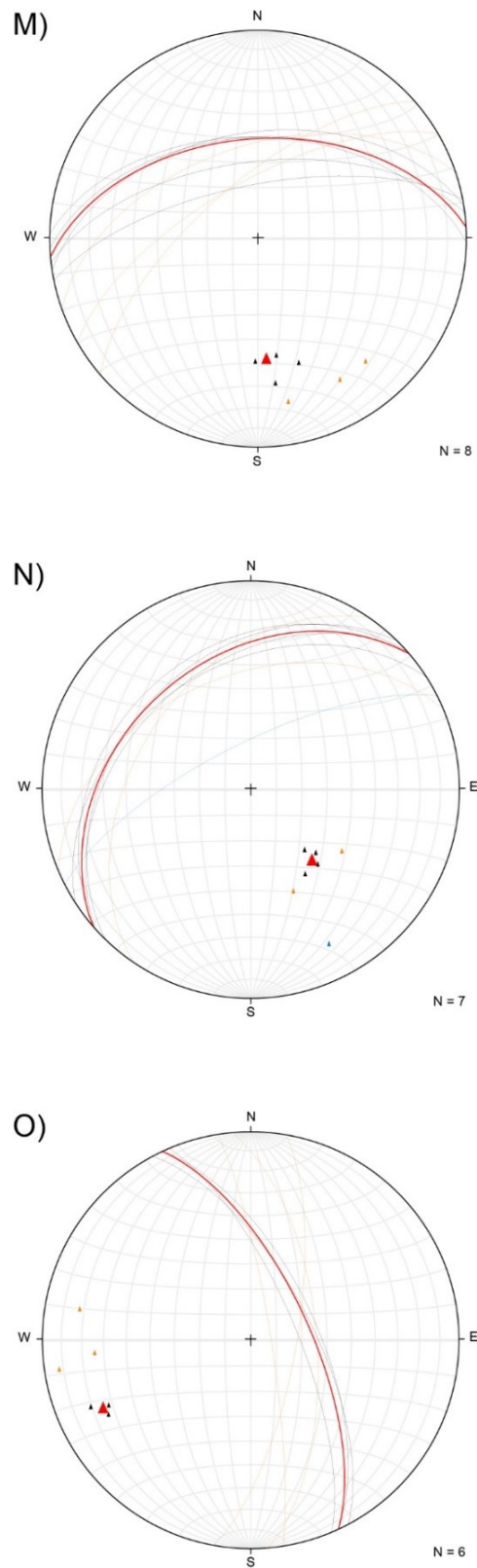


Figure C7: Equal area stereonets showing CLA measurements for domains M-O. The measurements in blue and orange were filtered out in the first and second iterations of filtration, respectively. The bold red plane corresponds to the mean CLA plane after filtration.

Appendix C

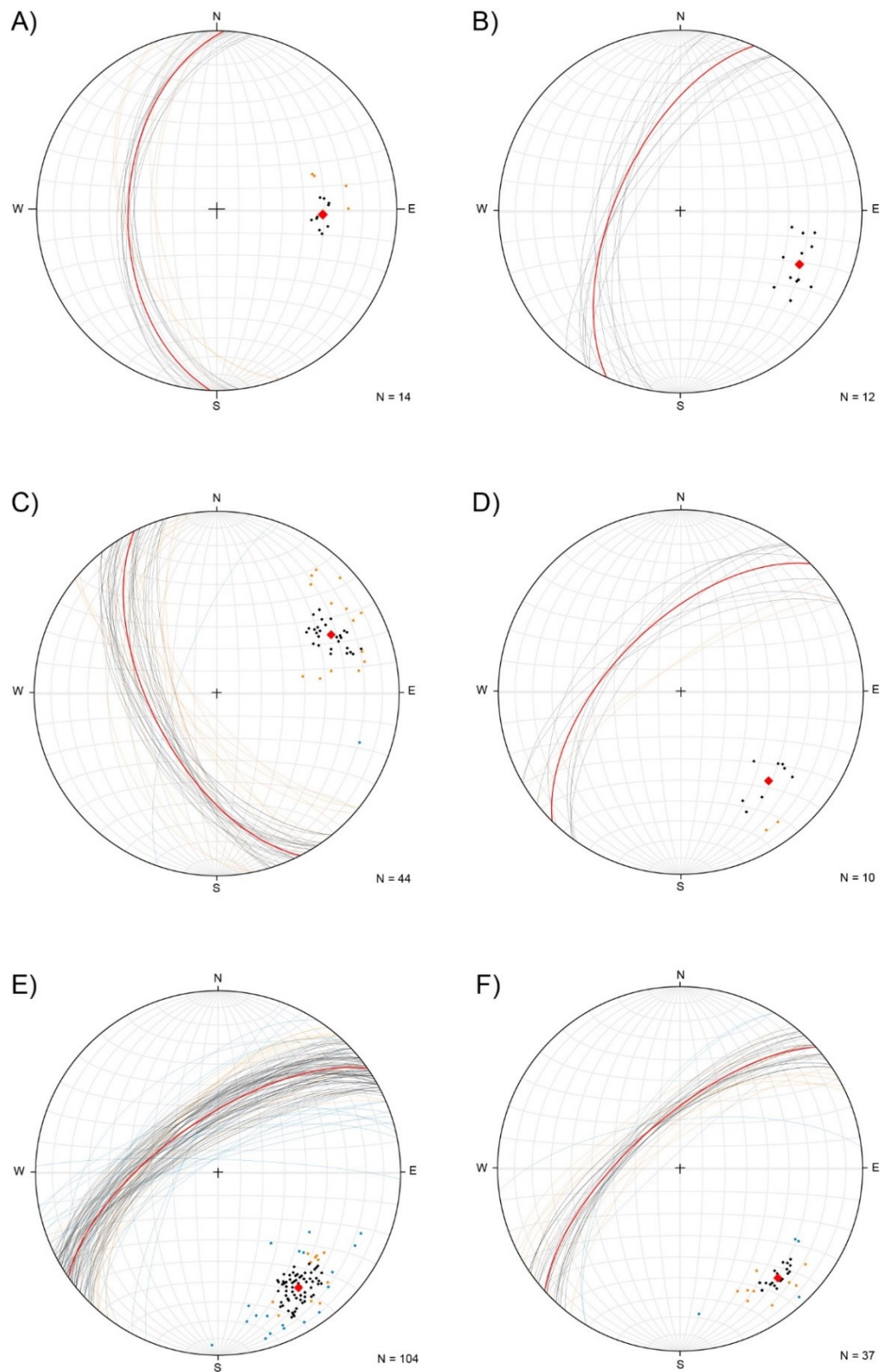


Figure C8: Equal area stereonet showing CLB measurements for domains A-F. The measurements in blue and orange were filtered out in the first and second iterations of filtration, respectively. The bold red plane corresponds to the mean CLB plane after filtration.

Appendix C

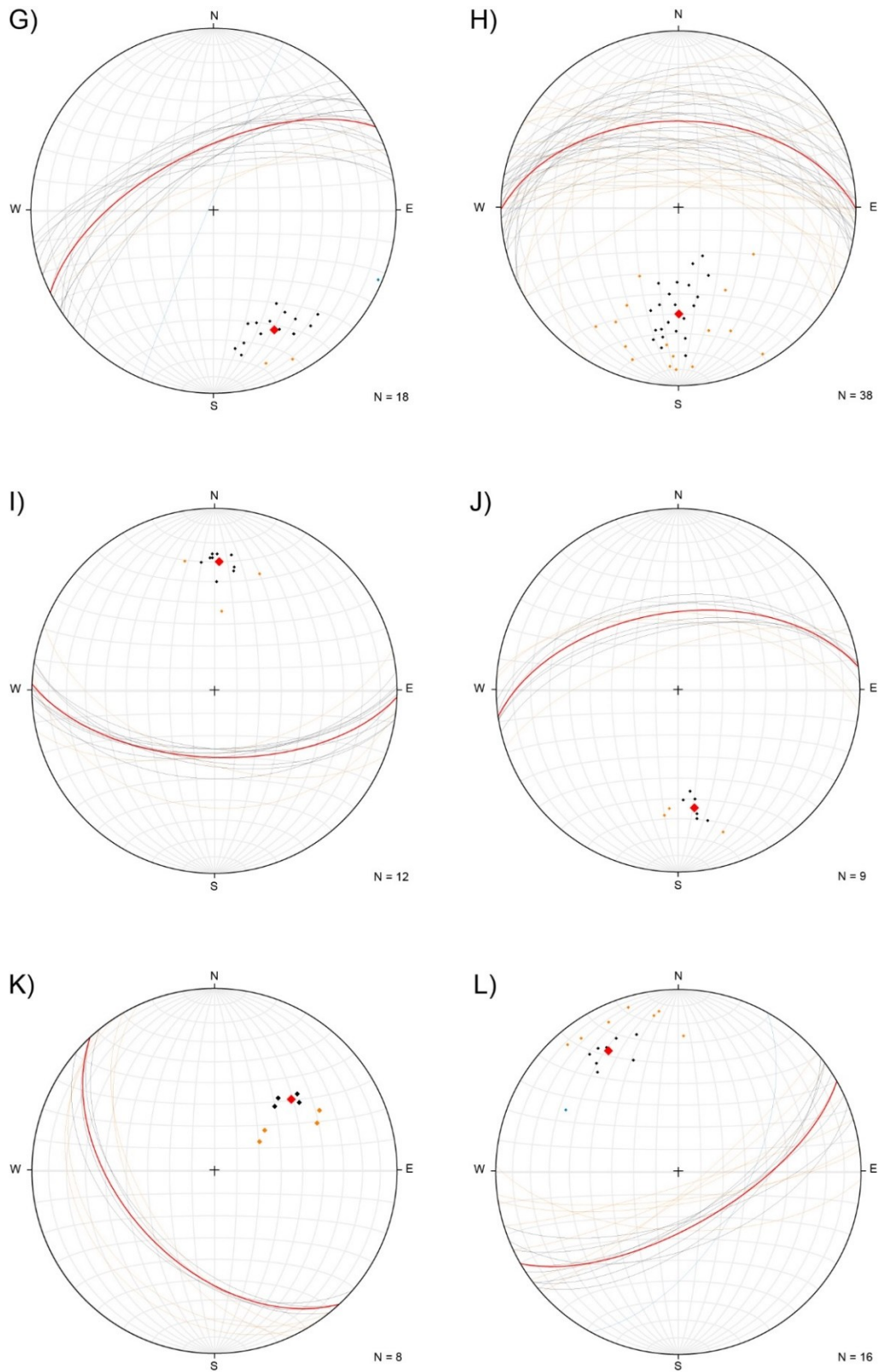


Figure C9: Equal area stereonet showing CLB measurements for domains G-L. The measurements in blue and orange were filtered out in the first and second iterations of filtration, respectively. The bold red plane corresponds to the mean CLB plane after filtration.

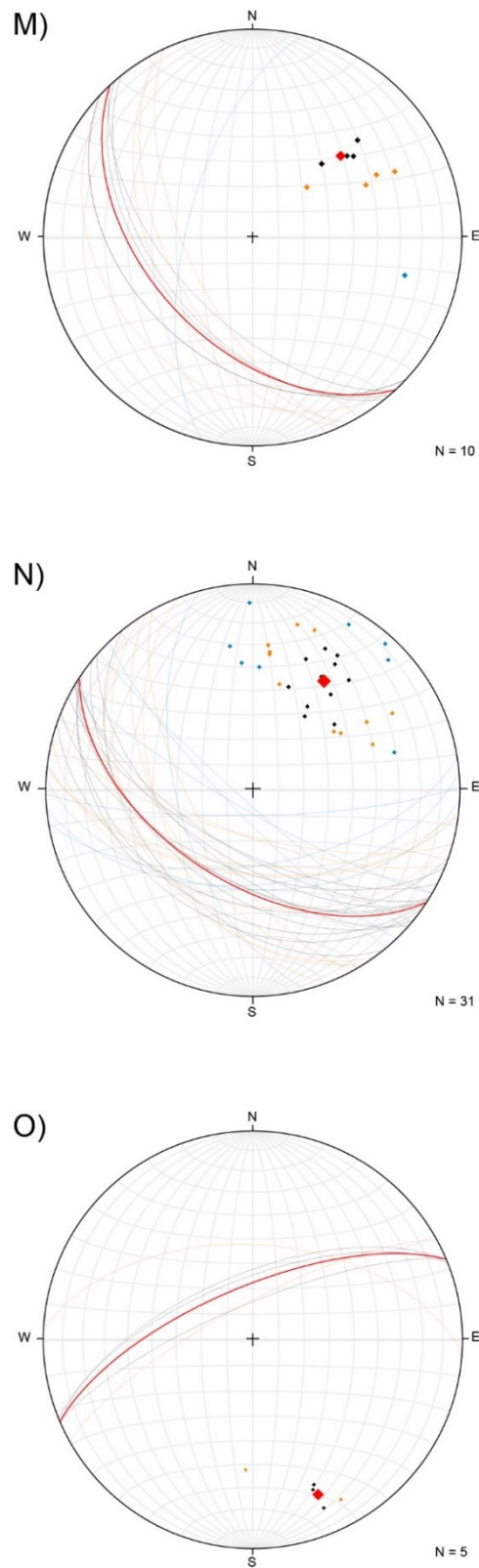


Figure C10: Equal area stereonets showing CLB measurements for domains M-O. The measurements in blue and orange were filtered out in the first and second iterations of filtration, respectively. The bold red plane corresponds to the mean CLA plane after filtration.

Appendix C

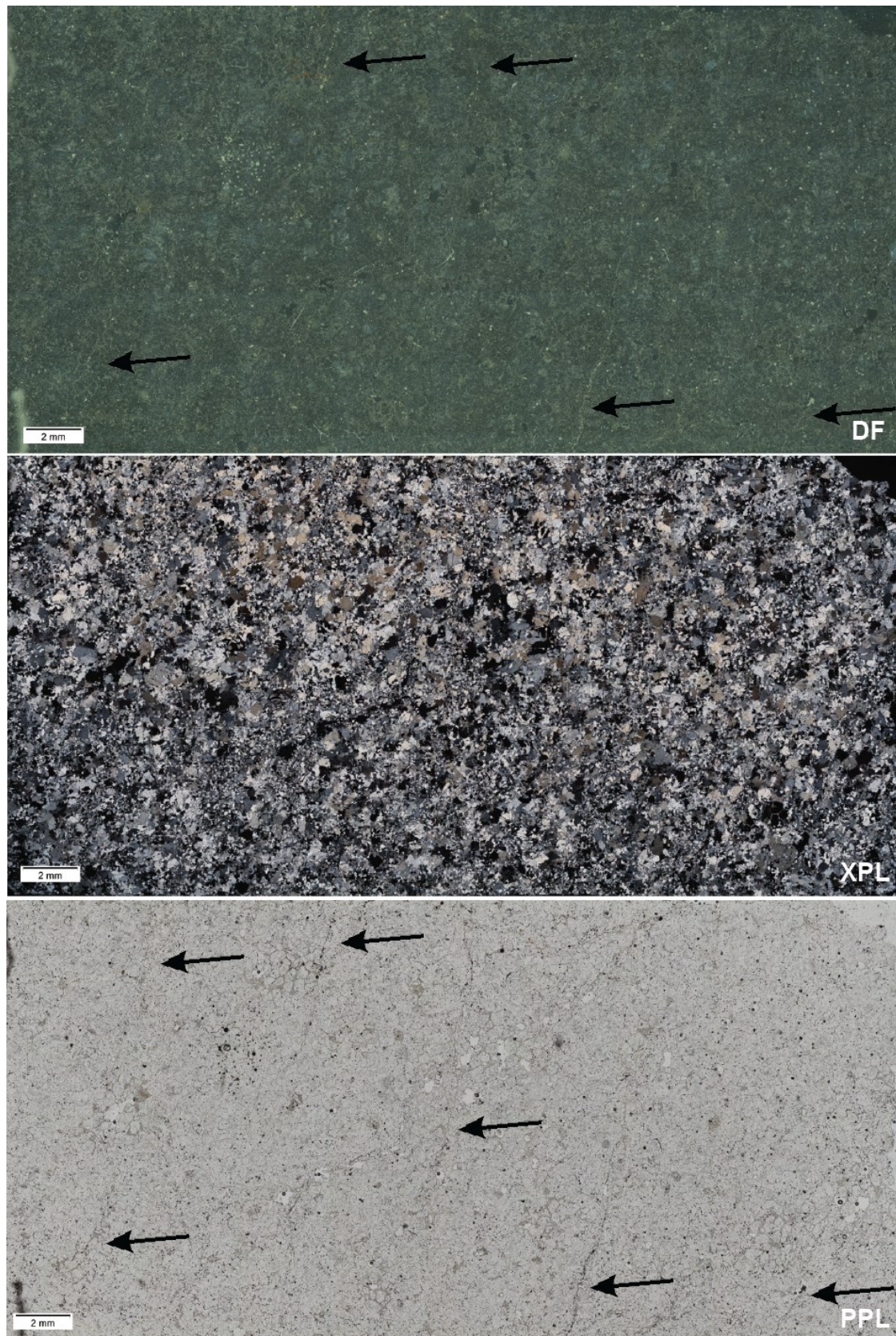


Figure C11: Thin section SM-3 showing shatter cleavage presented by shear fractures with an approximate spacing of 1 cm in a highly recrystallised medium-grained quartzite. Sample location (Nauwpoort, GPS location unknown – collected by Sagwadi Mabasa).

Appendix C

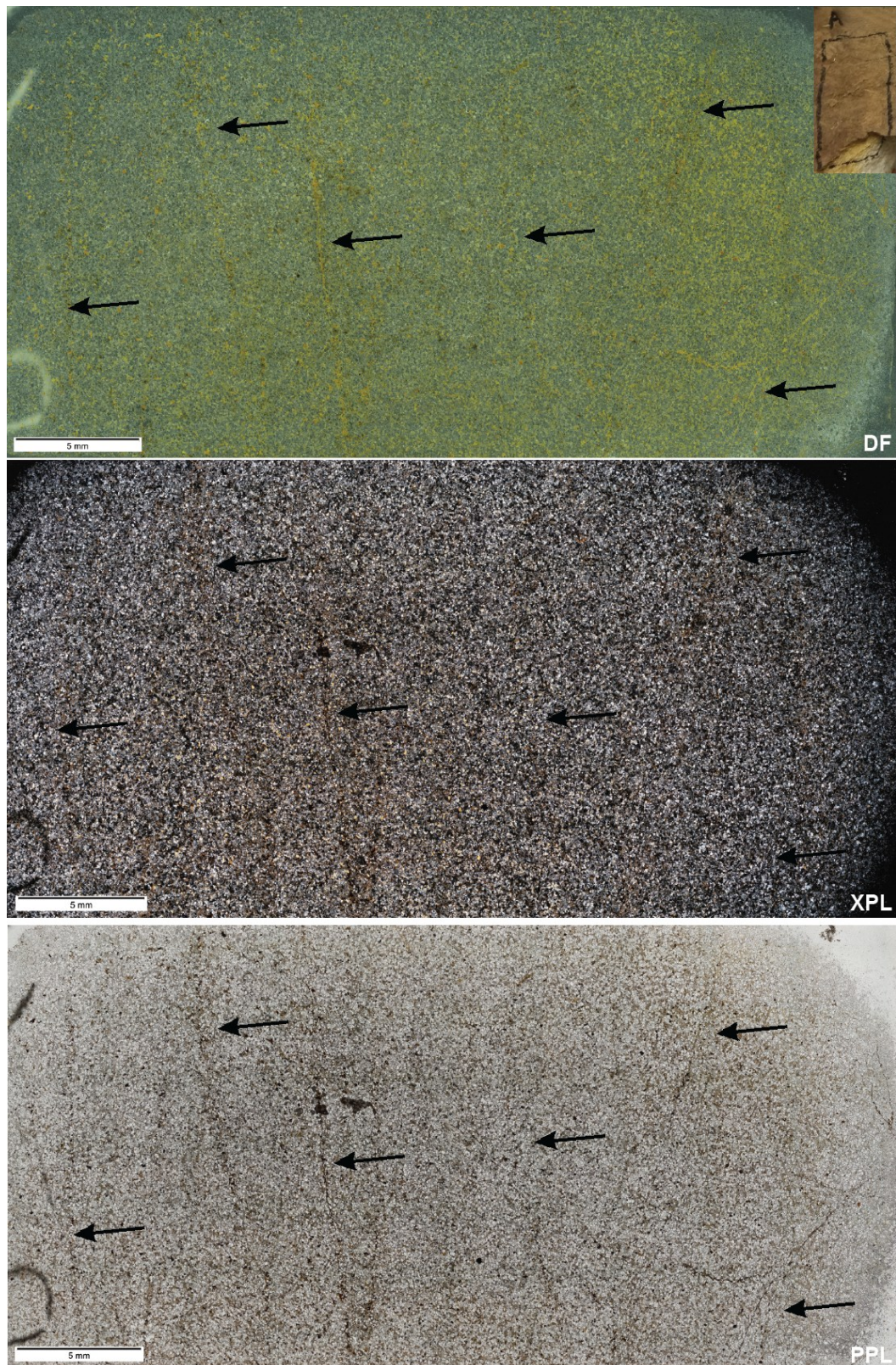


Figure C12: Thin section RAS-6A showing shatter cleavage presented by shear fractures in a fine-grained sandstone. The shear fractures are evident by preferential iron staining in DF. The inset in the top right corner shows the position from where the thin section was taken, which is parallel to sedimentary layering. Sample location (35J 0526700 E 7018898 S).

Appendix C

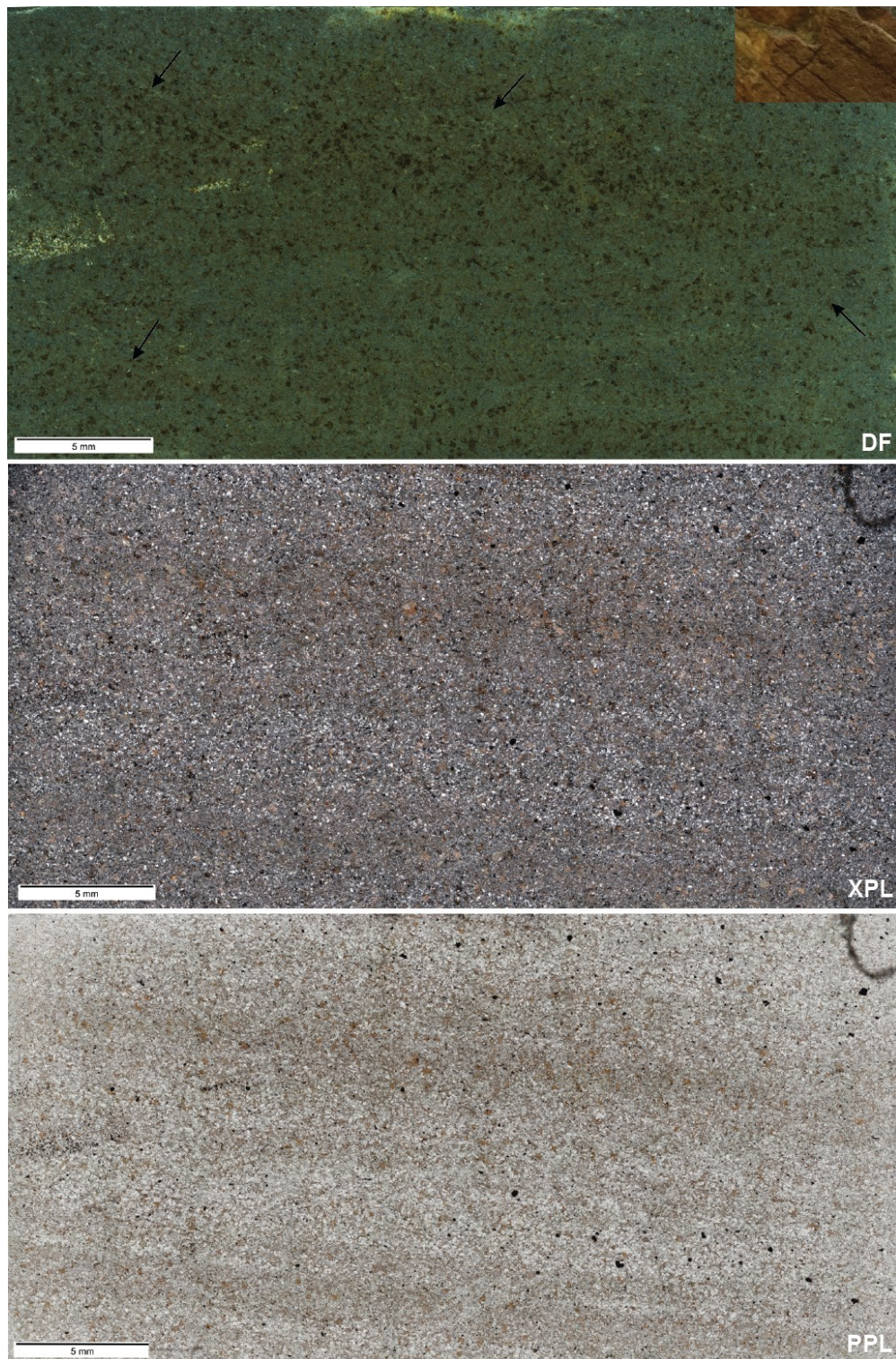


Figure C13: Thin section RAS-4 showing that although visibly present on the weathered surface outcrop, the shatter cleavage is not well-formed or marked by a significant feature in thin section. The inset in the top right corner shows the position from where the thin section was taken relative to the shatter cleavage. Sample location (35J 0526682 E 7019237 S).

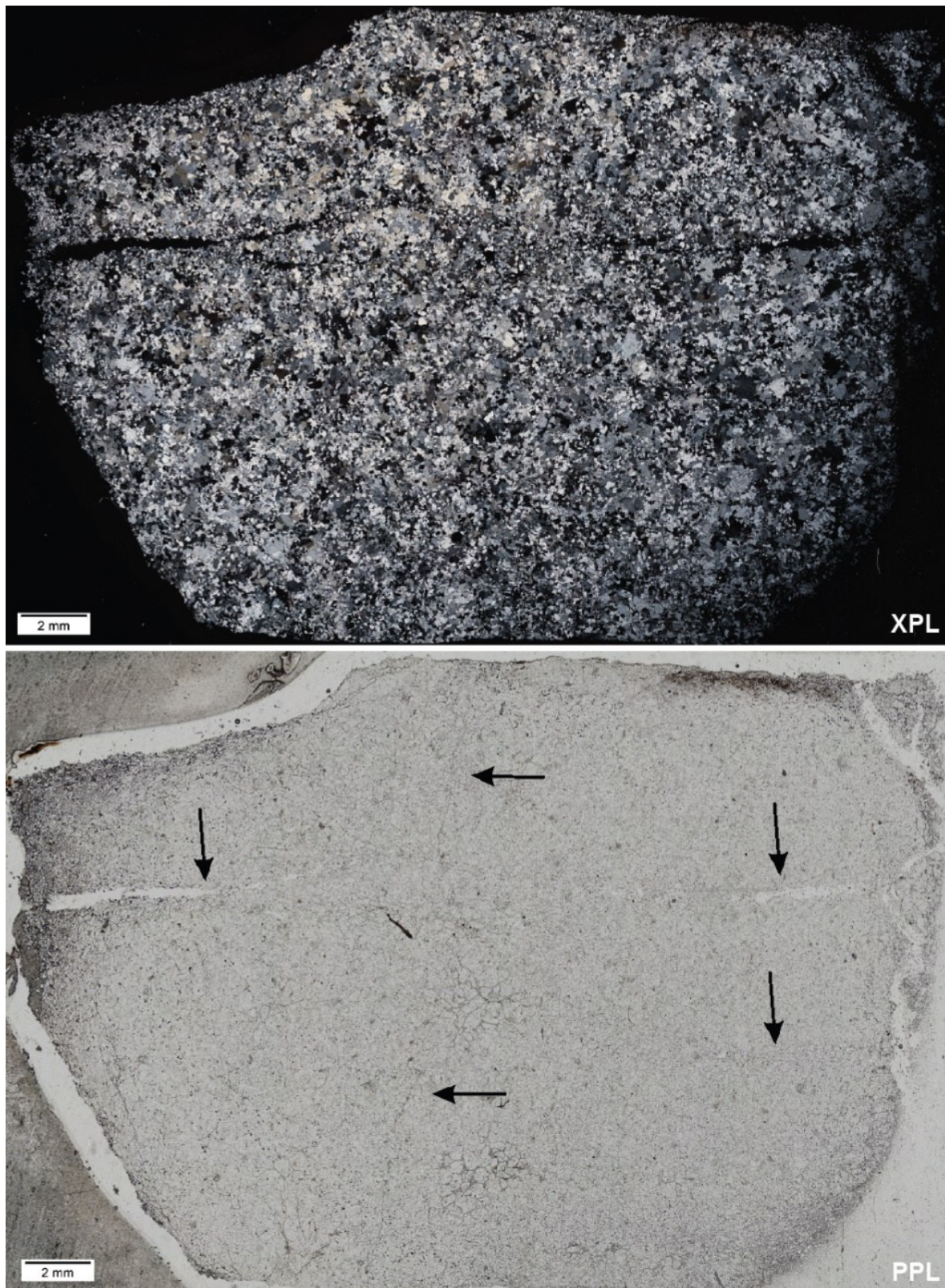
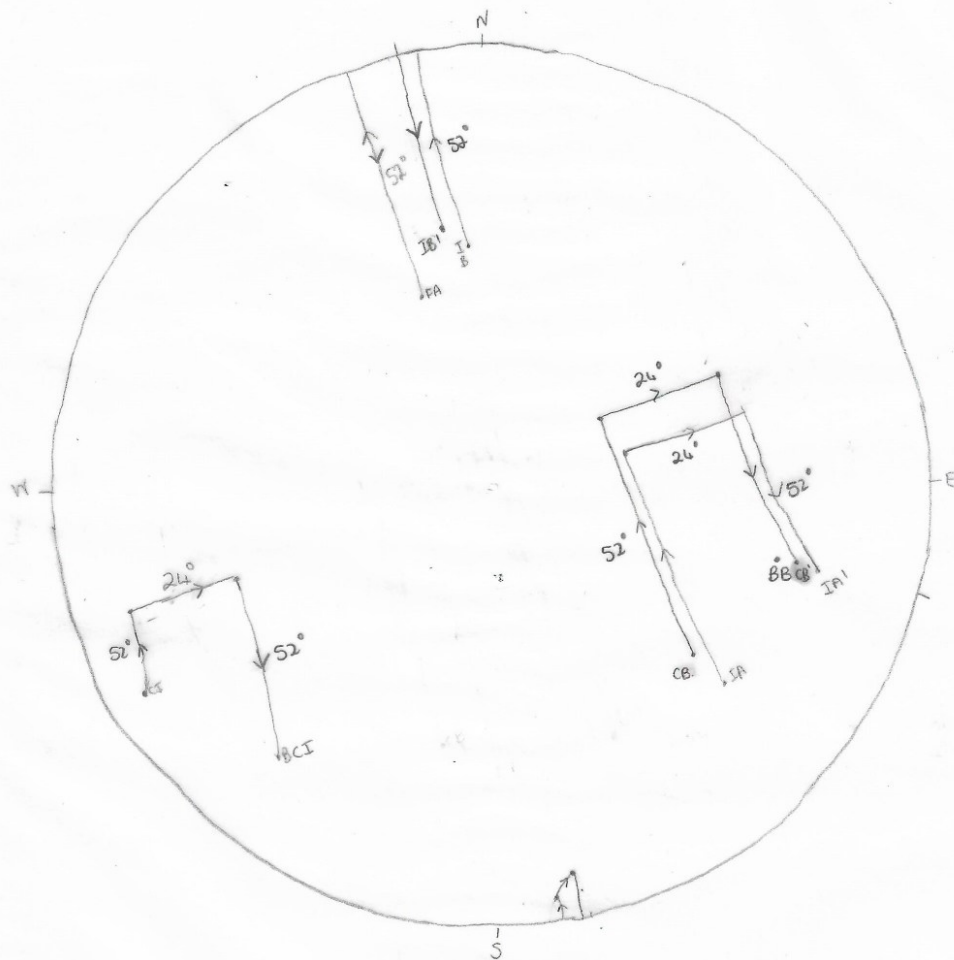


Figure C14: Thin section SM6-2 showing orthogonal shatter cleavage presented by fine-grained recrystallised quartz and fractures in a medium-grained quartzite. The shatter cleavage is seen to bifurcate around a large quartz grain. Sample location (Nauwpoort, GPS location unknown – collected by Sagwadi Mabasa).

Figure C15: Rotation of Domain I to Domain C.

②



FA $342^{\circ}/52^{\circ}$

Figure C16: Rotation of Domains CI to Domain B.

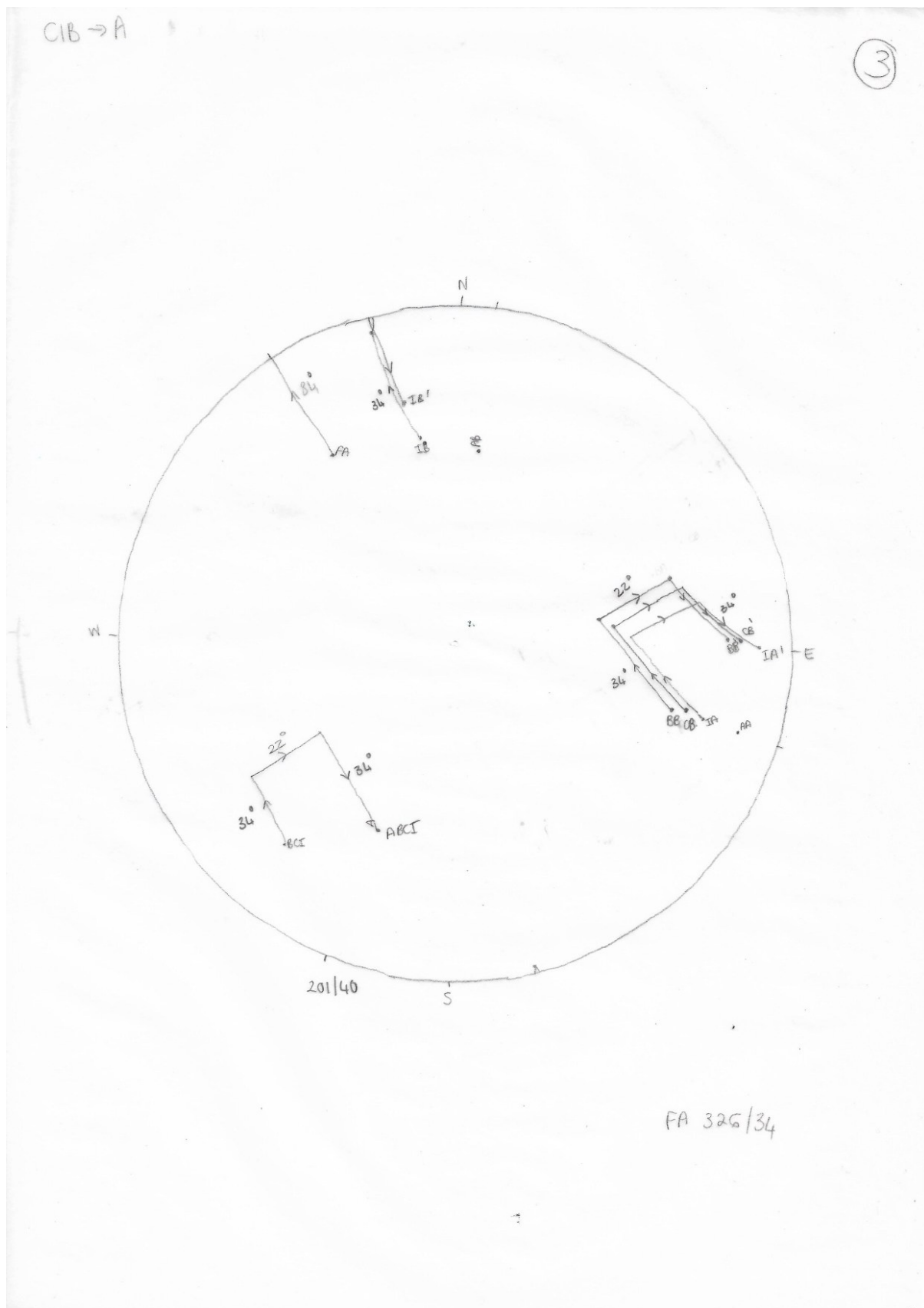


Figure C17: Rotation of Domains CIB to Domain A.

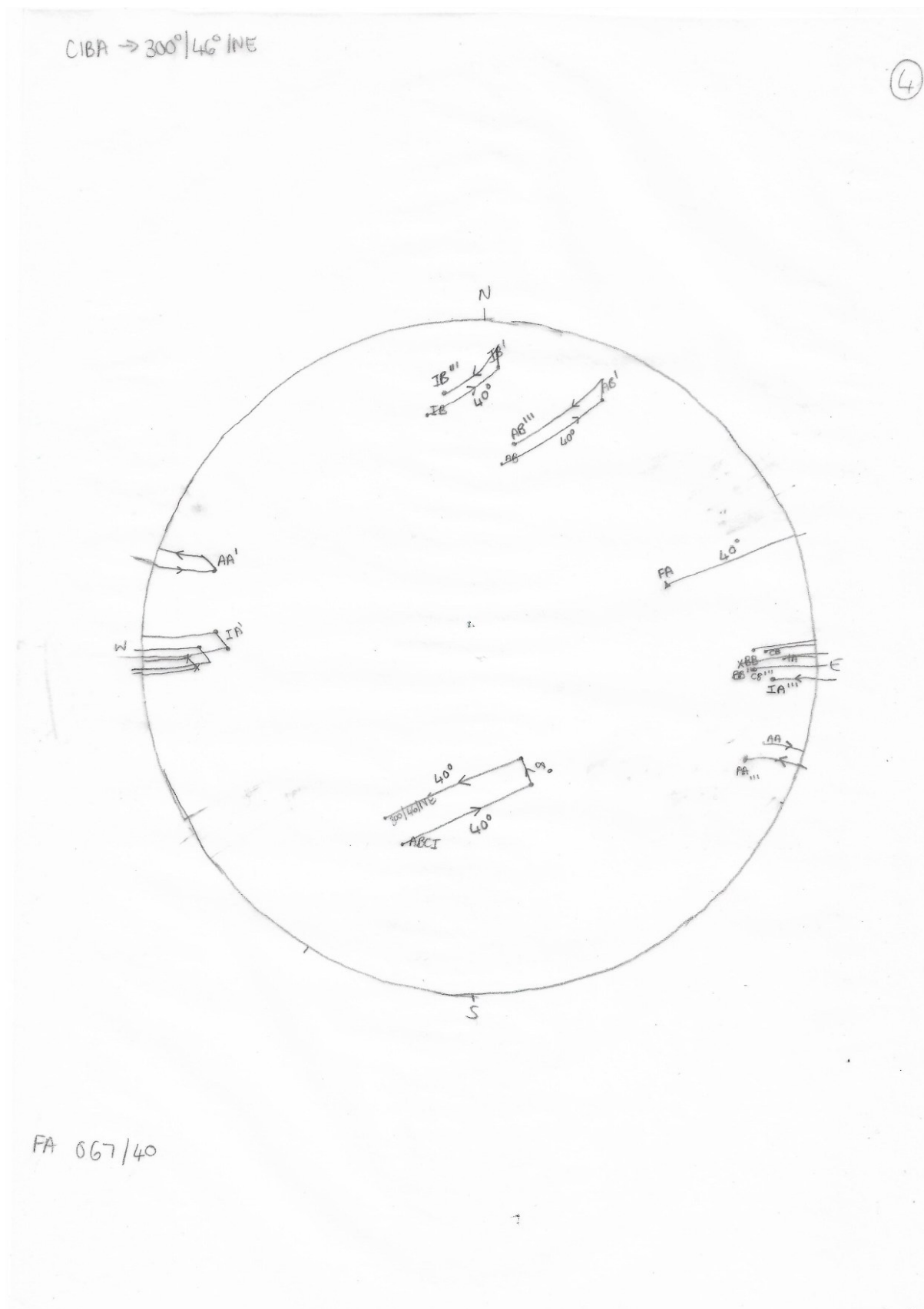


Figure C18: Rotation of Domains CIBA to $300^\circ/46^\circ$ NE.

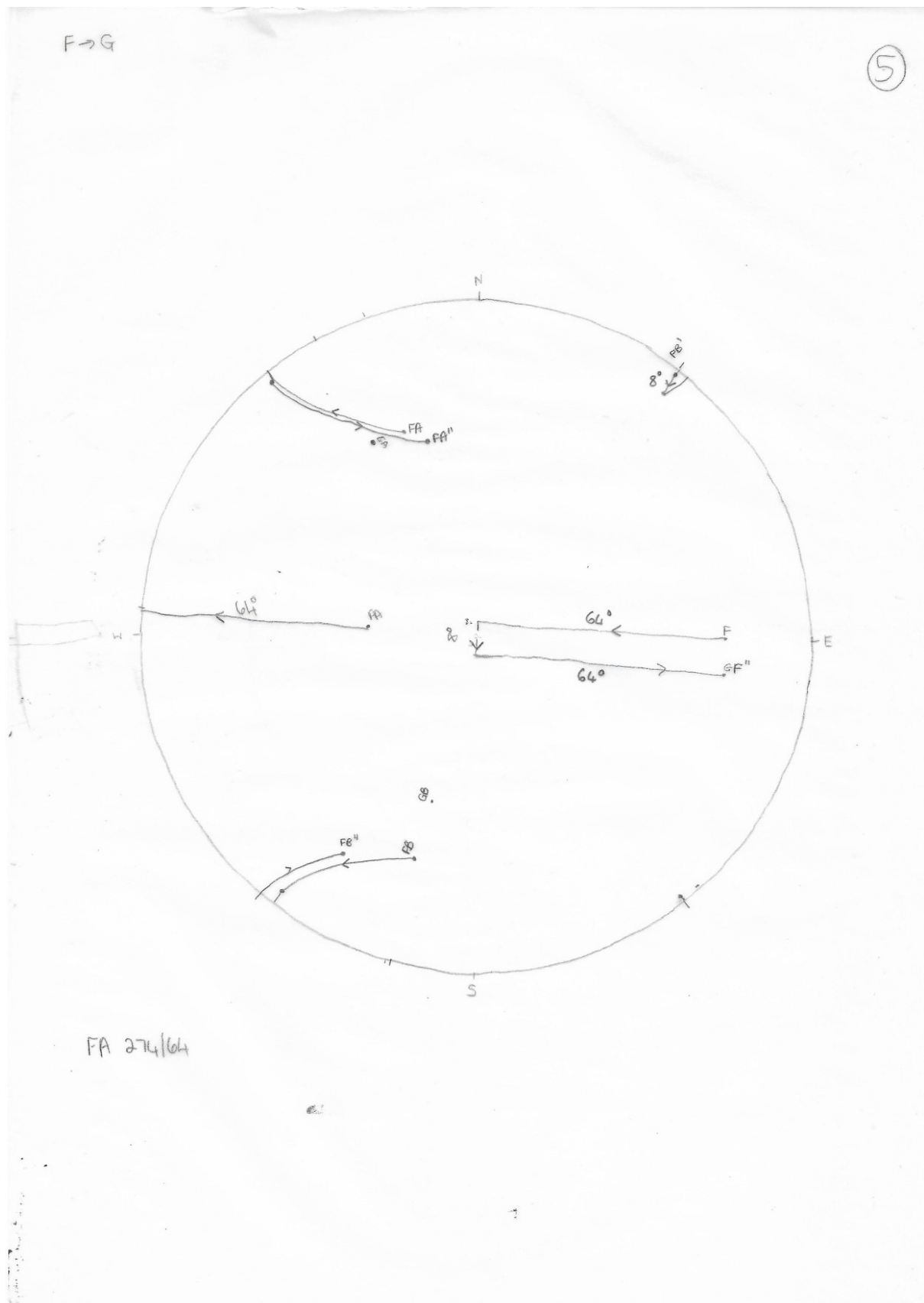


Figure C19: Rotation of Domain F to Domain G.

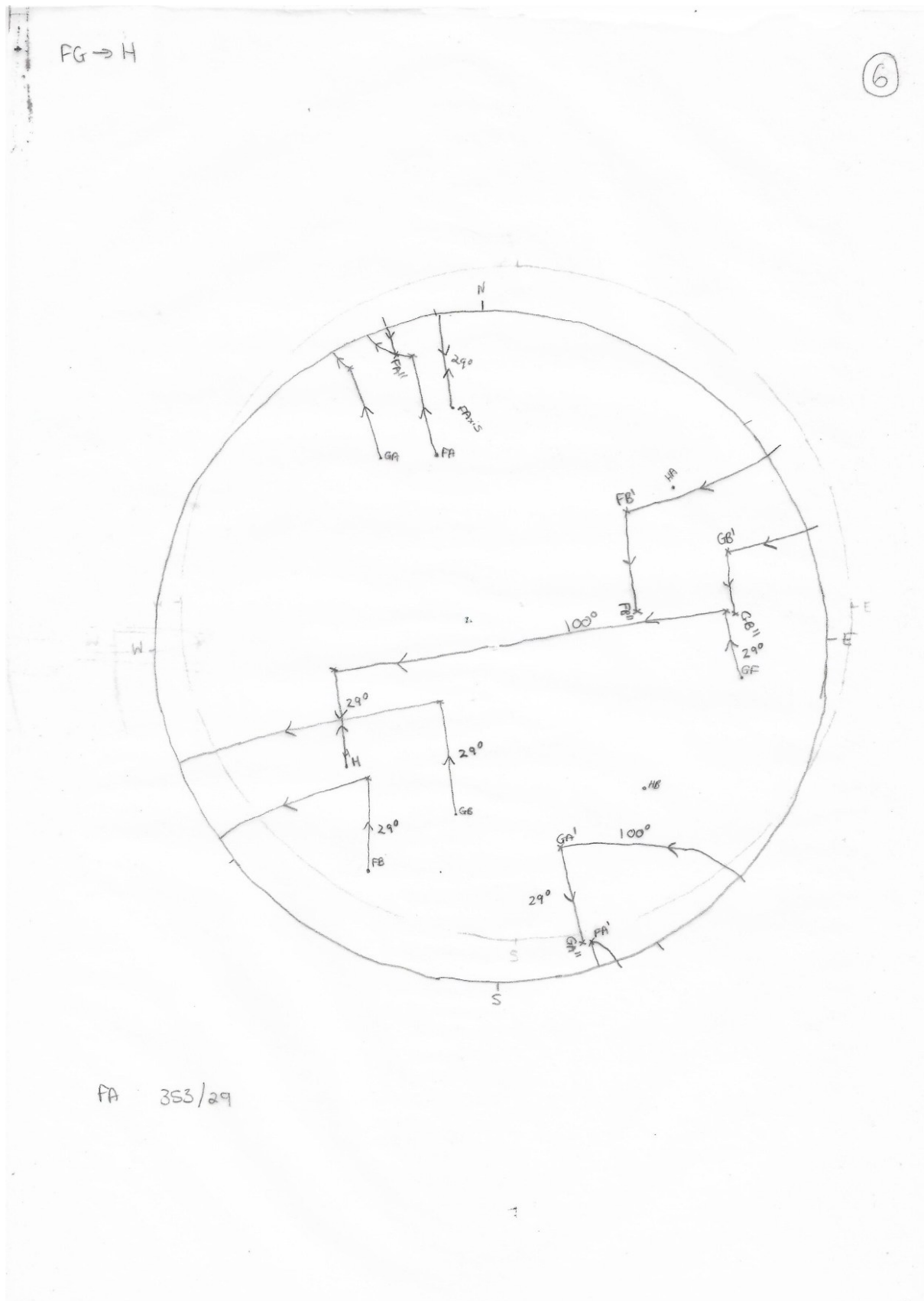


Figure C20: Rotation of Domain FG to Domain H.

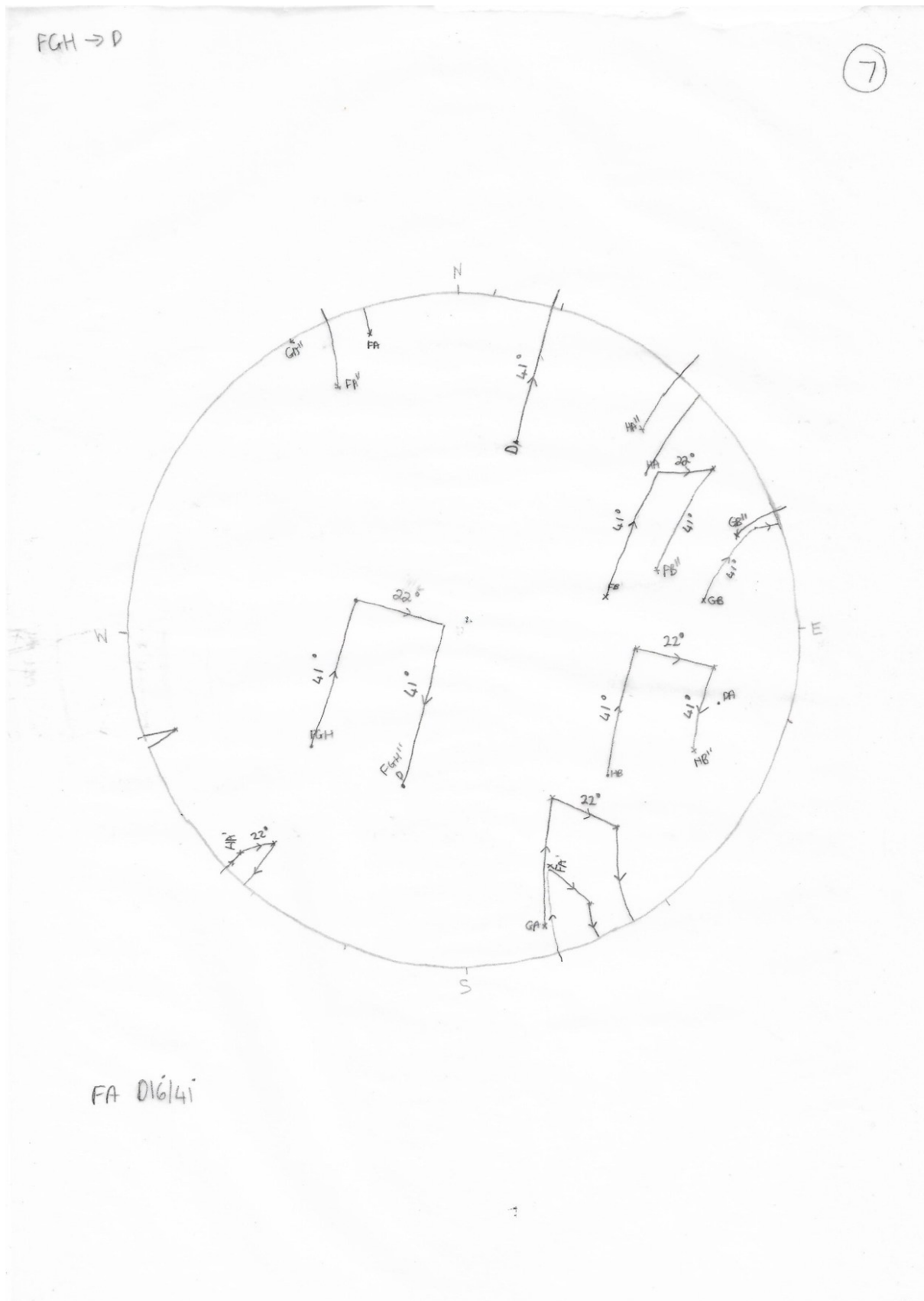


Figure C21: Rotation of Domain FGH to Domain G.

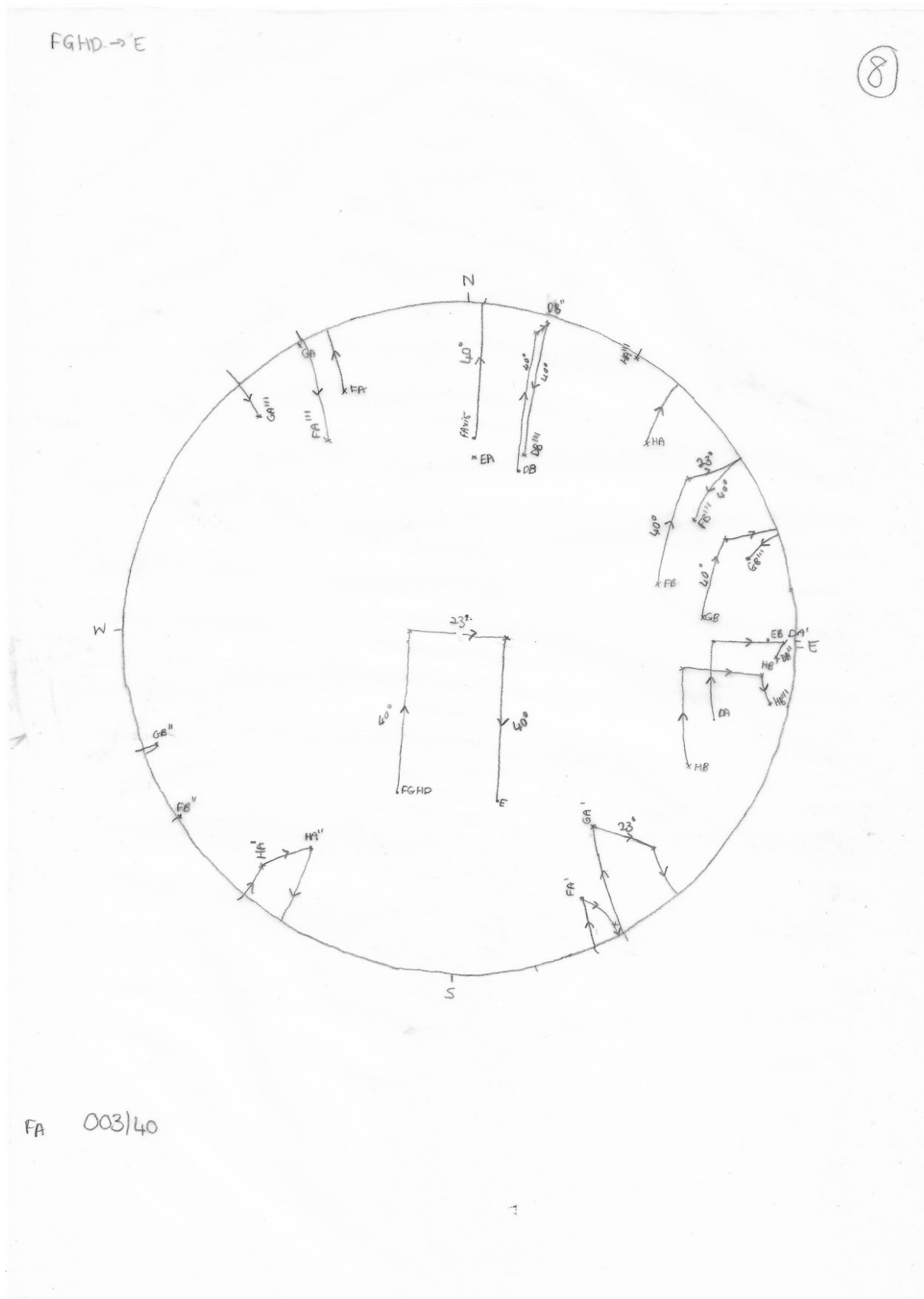


Figure C22: Rotation of Domain FGHD to Domain E.

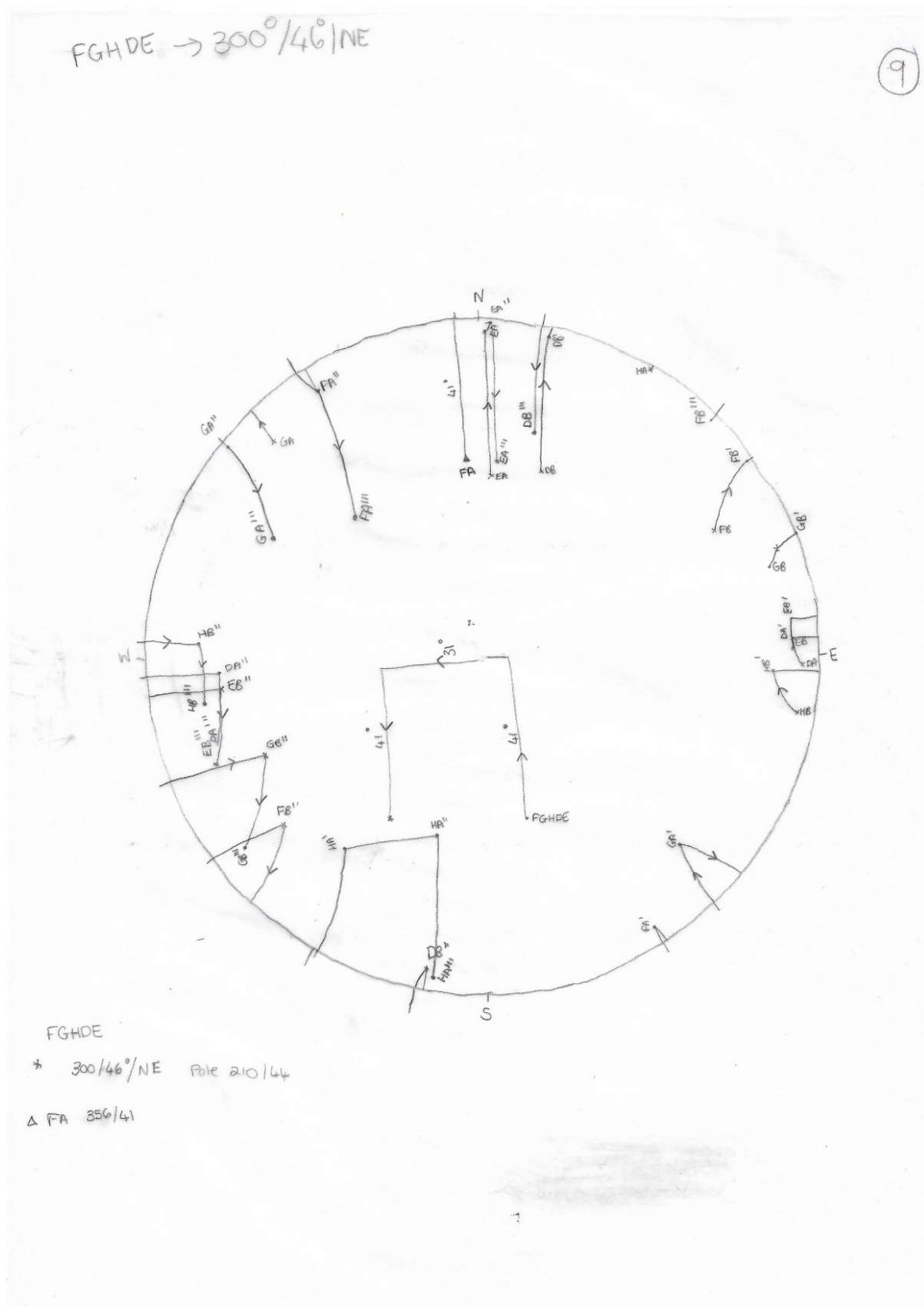
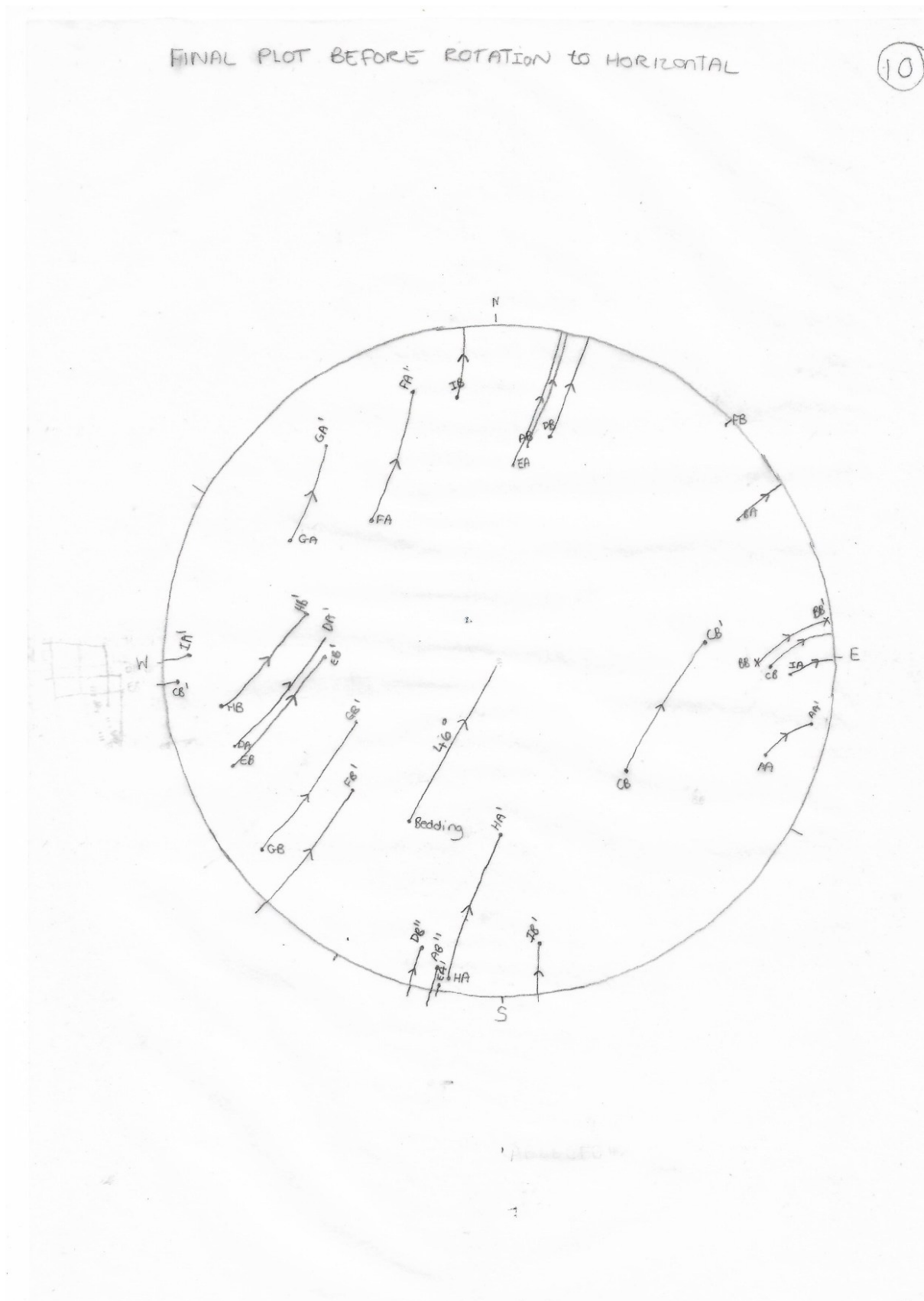


Figure C23: Rotation of FGHDE to $300^\circ/46^\circ/\text{NE}$.



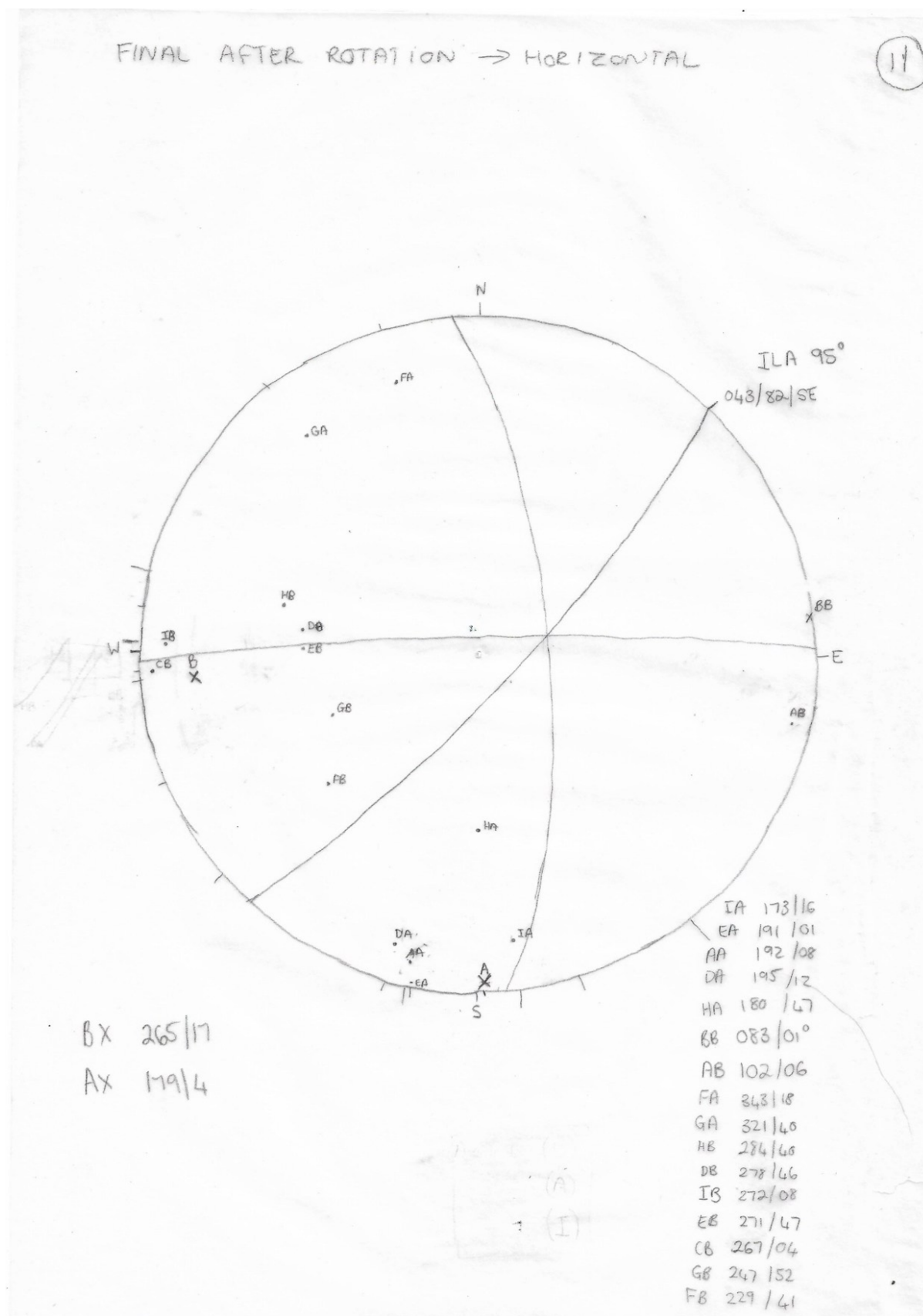


Figure C25: Result after rotation of all domains to the horizontal.

Appendix S

[Albat and Mayer \(1989\)](#) cleavage data (pairs only) rotated to the horizontal. The numbers indicated correlate to the location numbers used in the paper. These rotations were used to establish the bisectors used to create Figure 0.25.



Appendix S

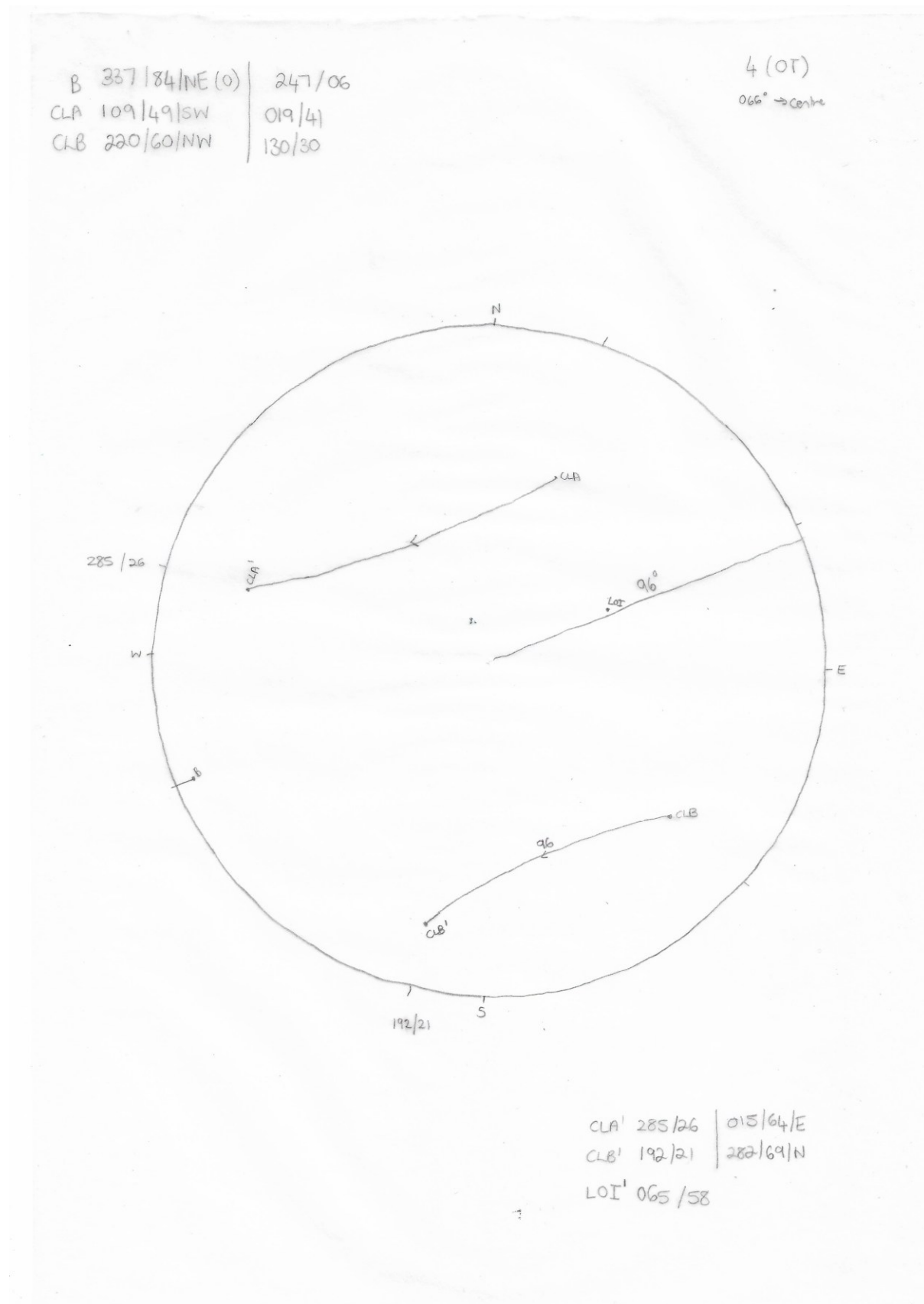


Figure S2: Rotation of Albat and Mayer (1989) location 4 to the horizontal.

Appendix S

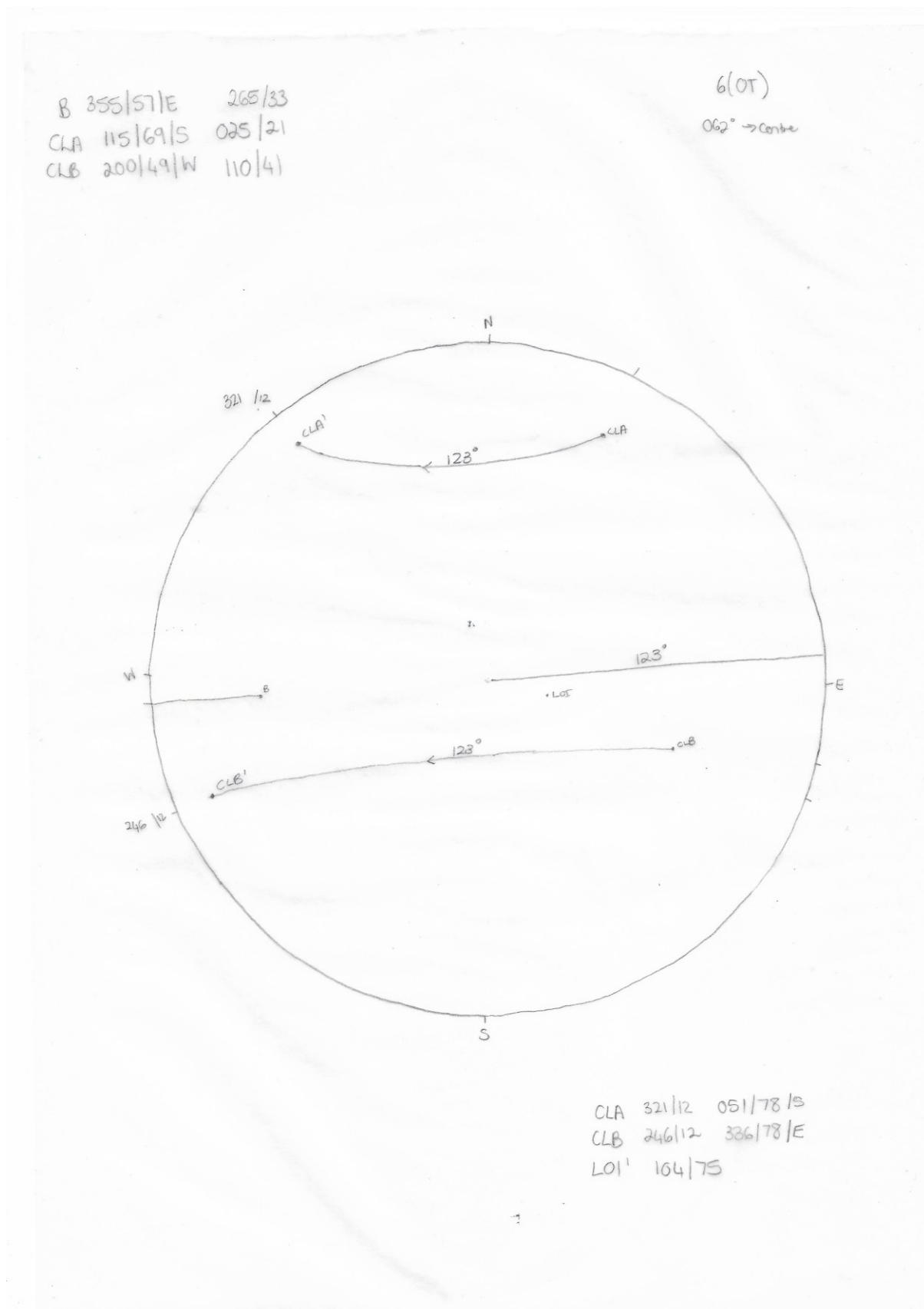


Figure S3: Rotation of Albat and Mayer (1989) location 6 to the horizontal.

Appendix S

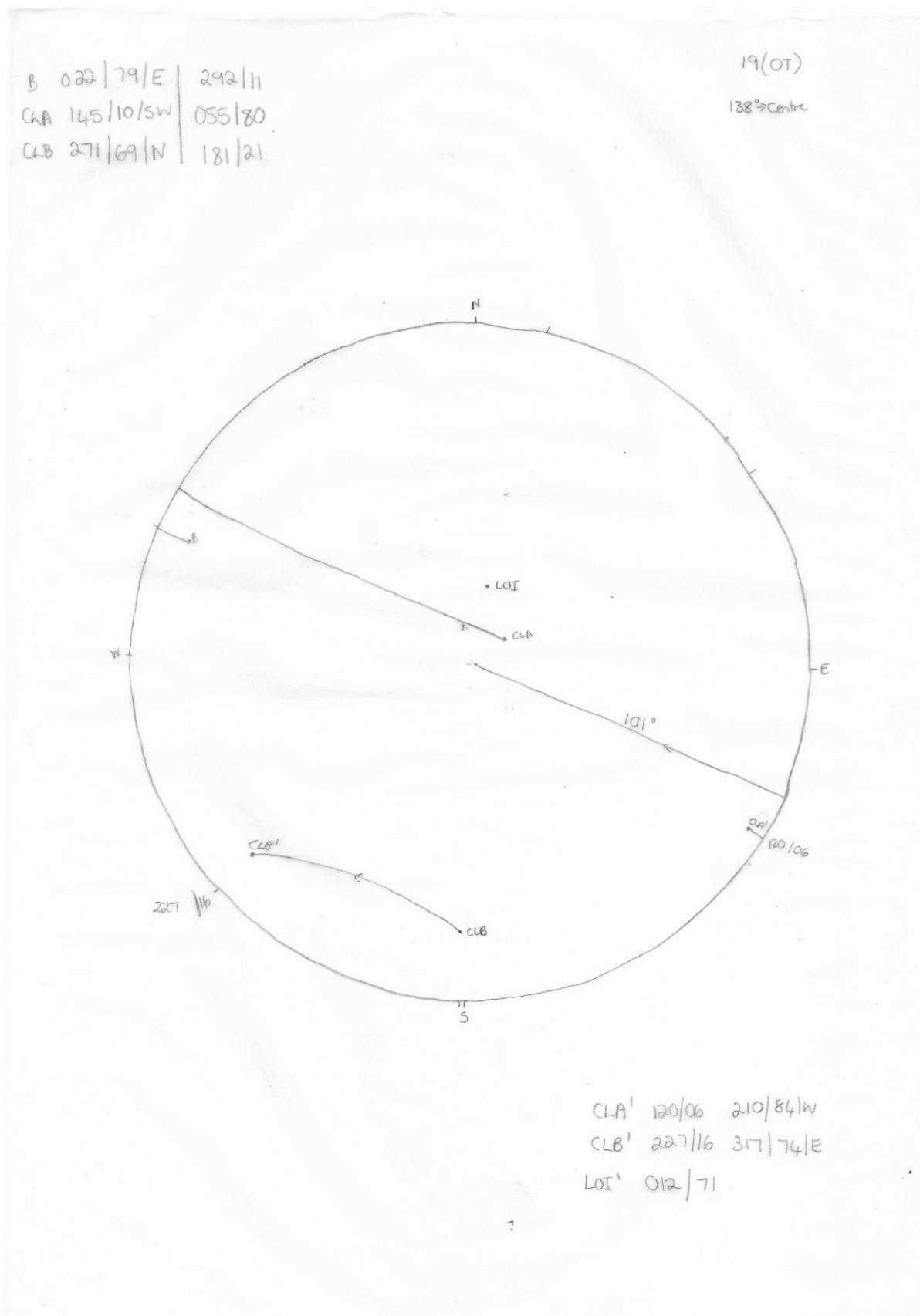


Figure S5: Rotation of Albat and Mayer (1989) location 19 to the horizontal.

Appendix S

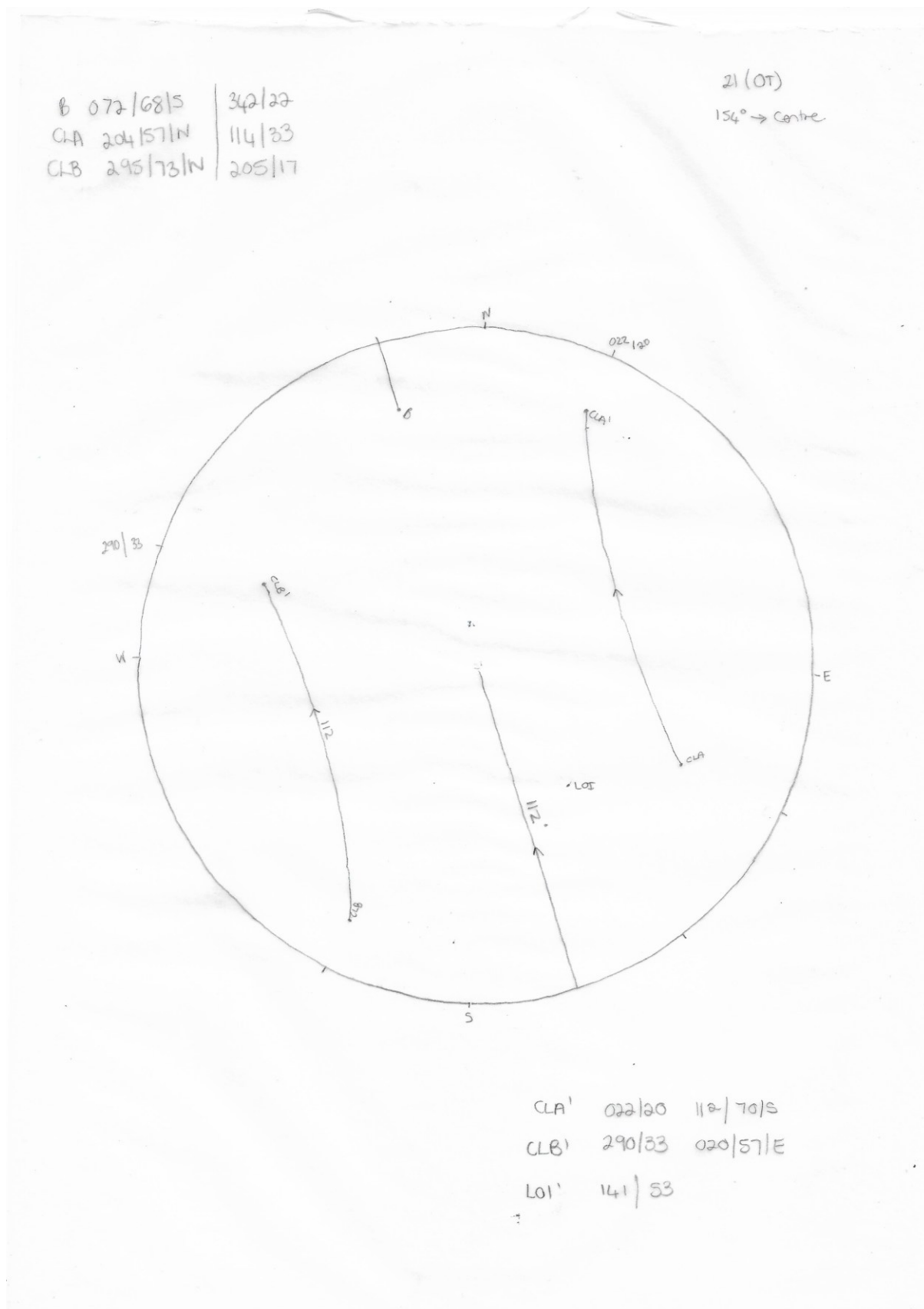


Figure S6: Rotation of Albat and Mayer (1989) location 21 to the horizontal.

Appendix S

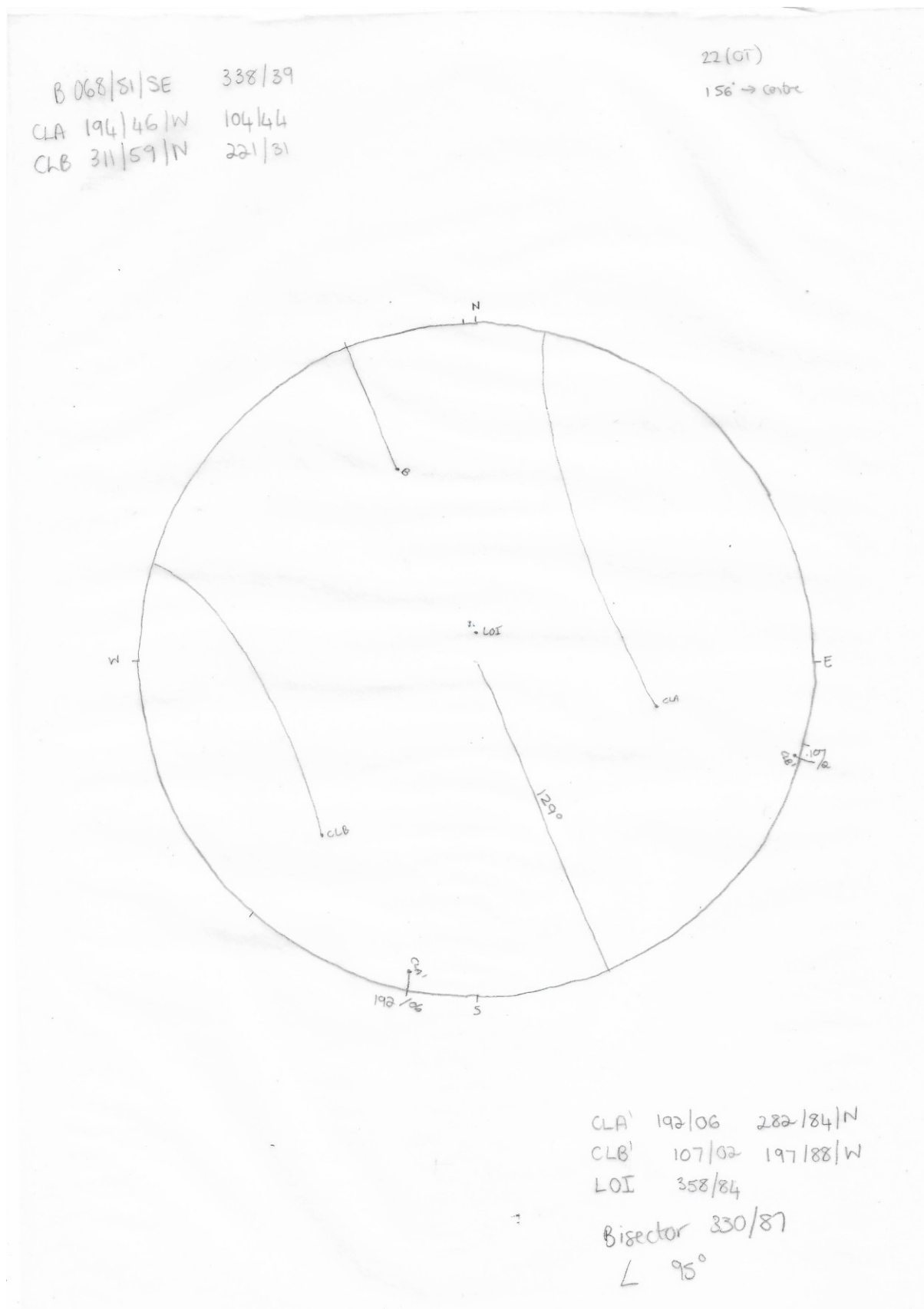


Figure S7: Rotation of Albat and Mayer (1989) location 22 to the horizontal.

Appendix S

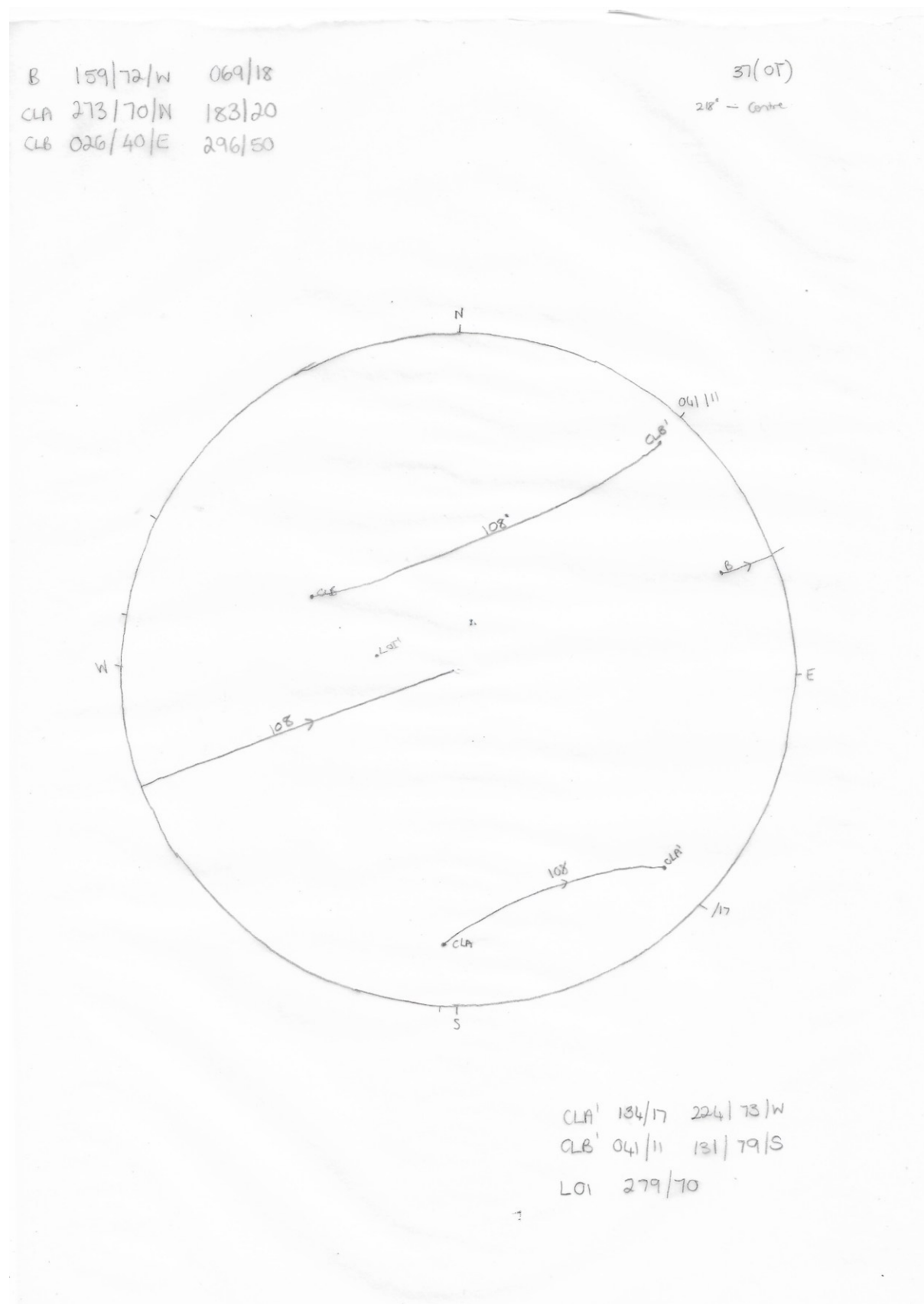


Figure S8: Rotation of Albat and Mayer (1989) location 37 to the horizontal.

Figure S9: Rotation of Albat and Mayer (1989) location 38 to the horizontal.

Appendix S

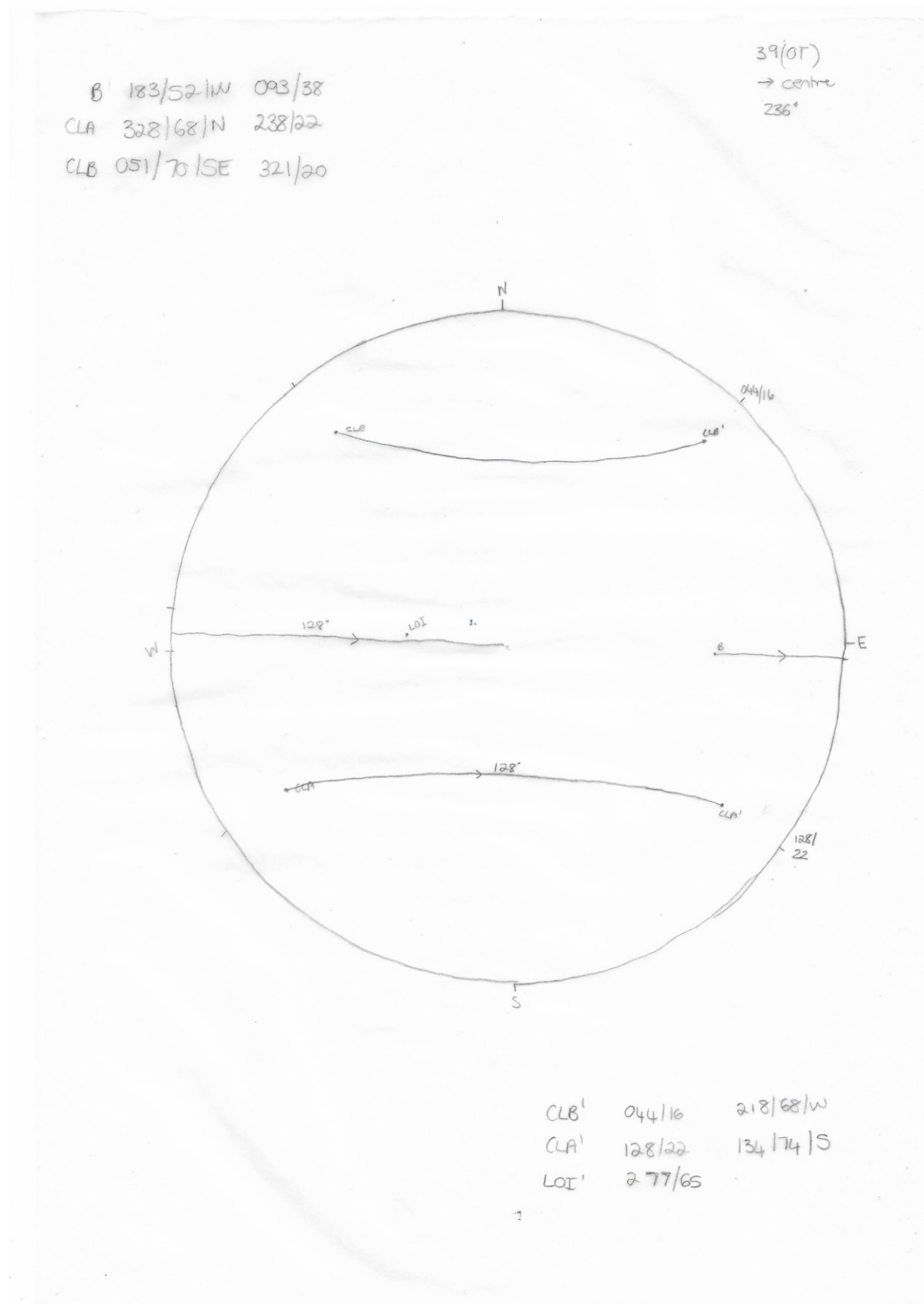


Figure S10: Rotation of Albat and Mayer (1989) location 39 to the horizontal.

Appendix S

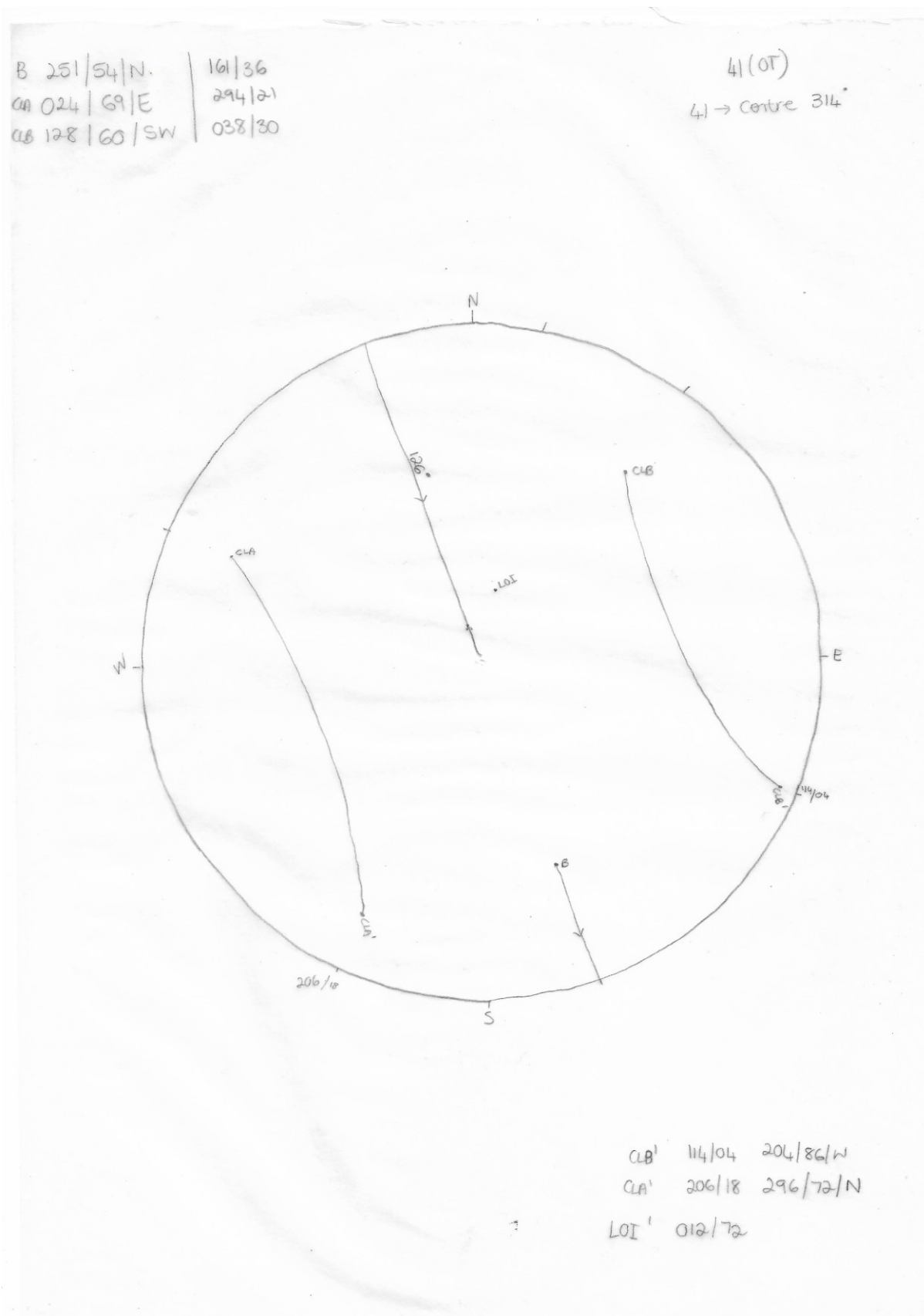


Figure S11: Rotation of Albat and Mayer (1989) location 41 to the horizontal.

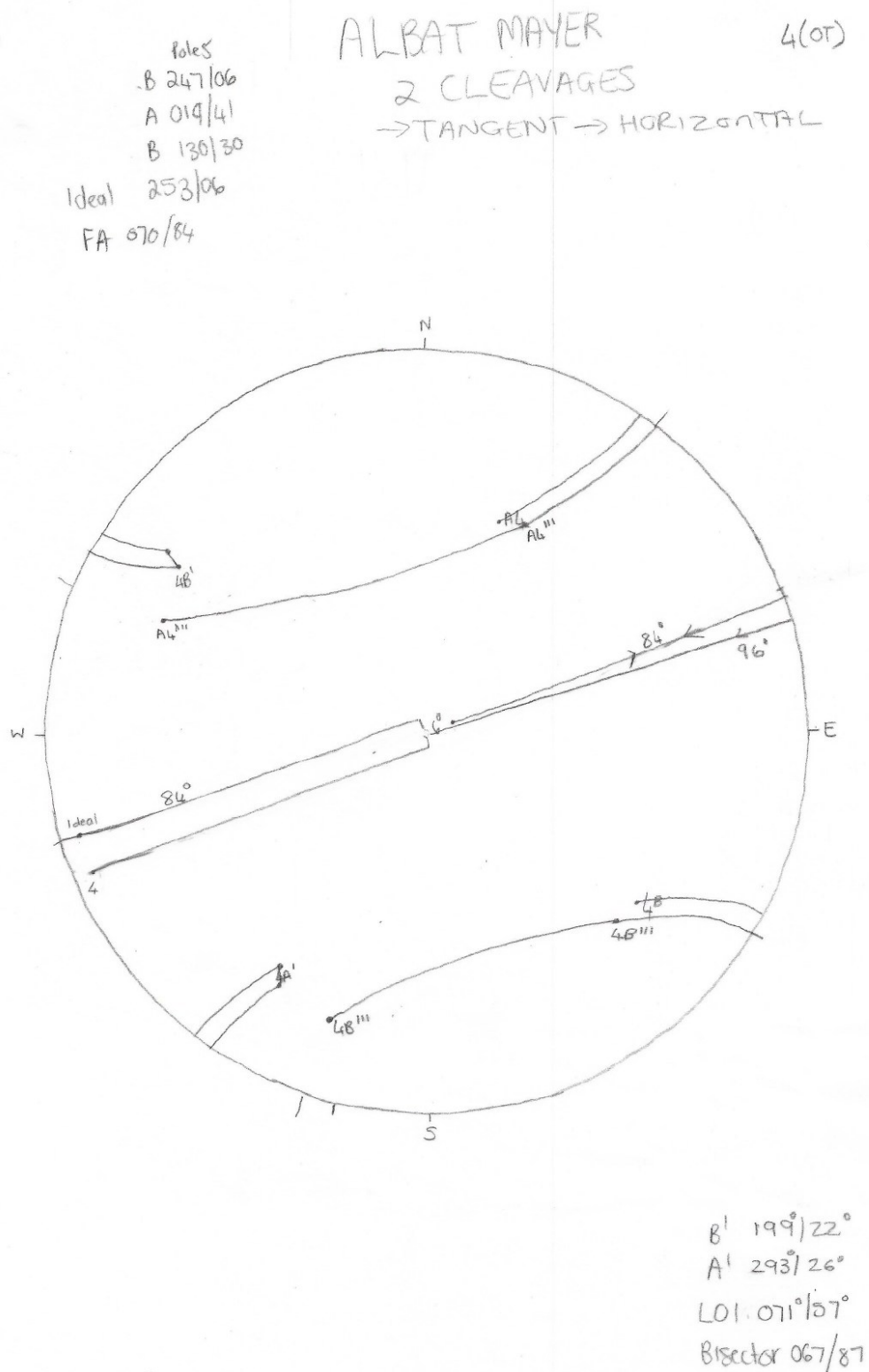
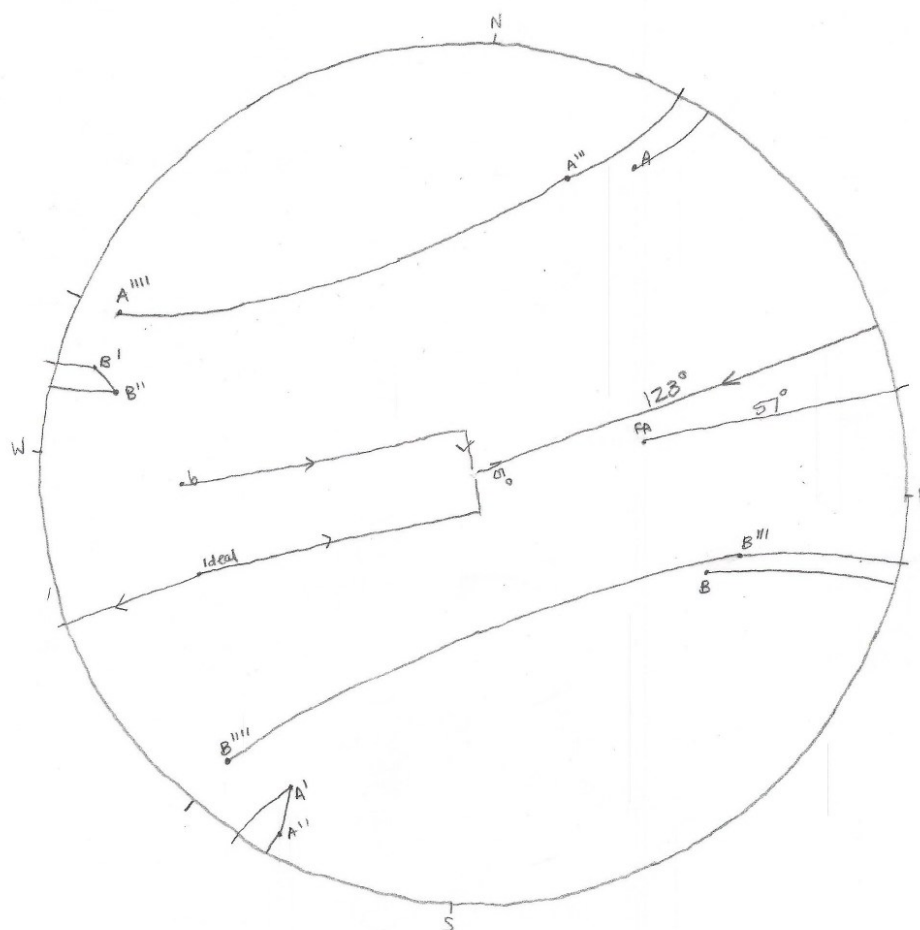


Figure S12: Rotation of Albat and Mayer (1989) location 4 to the tangential orientation and then to the horizontal.

Pole
247/33 Ideal

6(OT)



FA 076/57

B' 218/14

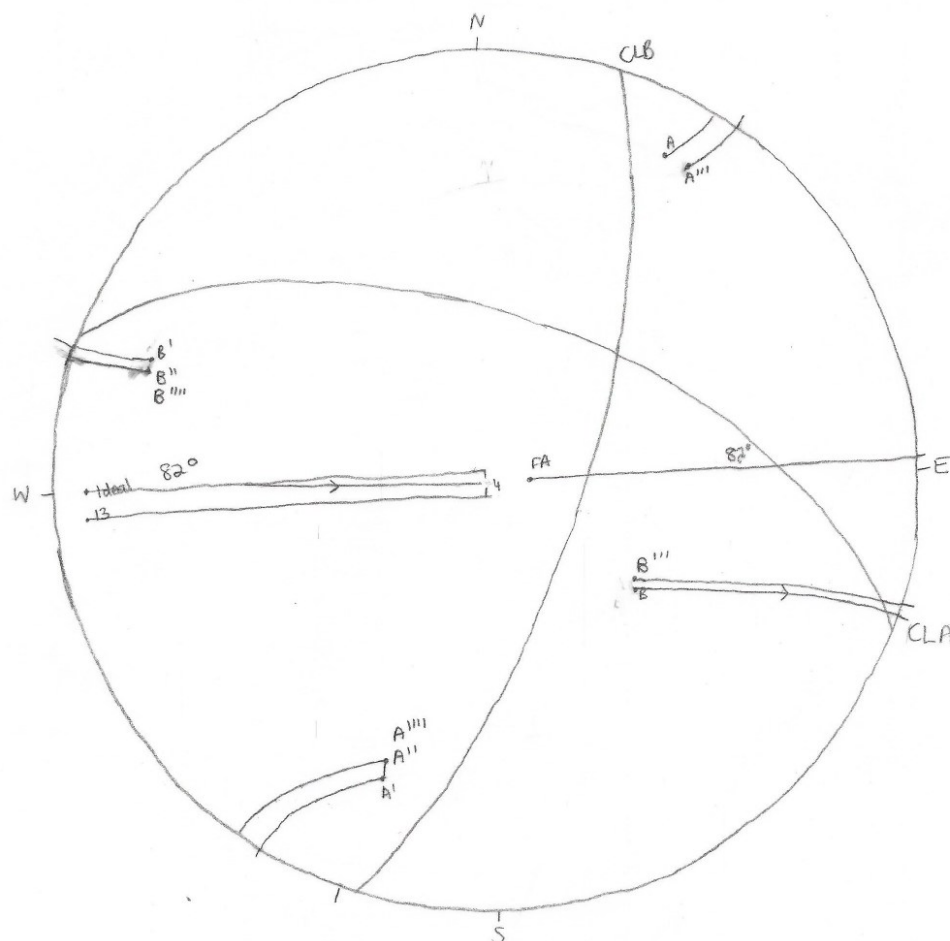
A' 292/10

L01 062/75

Bisector 254/87

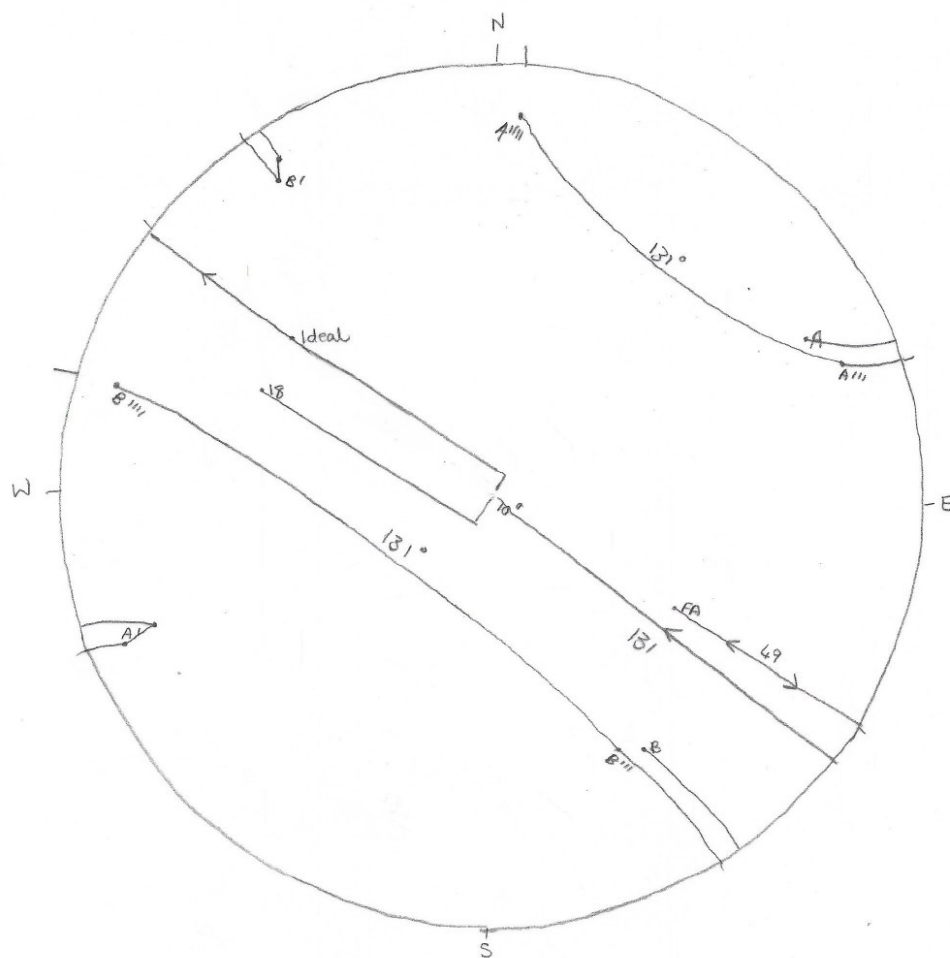
LC 108°/72°

Figure S13: Rotation of Albat and Mayer (1989) location 6 to the tangential orientation and then to the horizontal.



FA 088° / 82°
A' 201° / 32°
B' 290° / 18°
LO 048° / 55°
bisector 242° / 80°
LC 100°

Figure S14: Rotation of Albat and Mayer (1989) location 13 to the tangential orientation and then to the horizontal.



FA	$122^{\circ}49'$
A'	$004^{\circ}13'$
B'	$286^{\circ}10'$
LOF	$155/75$
Bisector	$145/88$
L	104°

Figure S15: Rotation of Albat and Mayer (1989) location 18 to the tangential orientation and then to the horizontal.

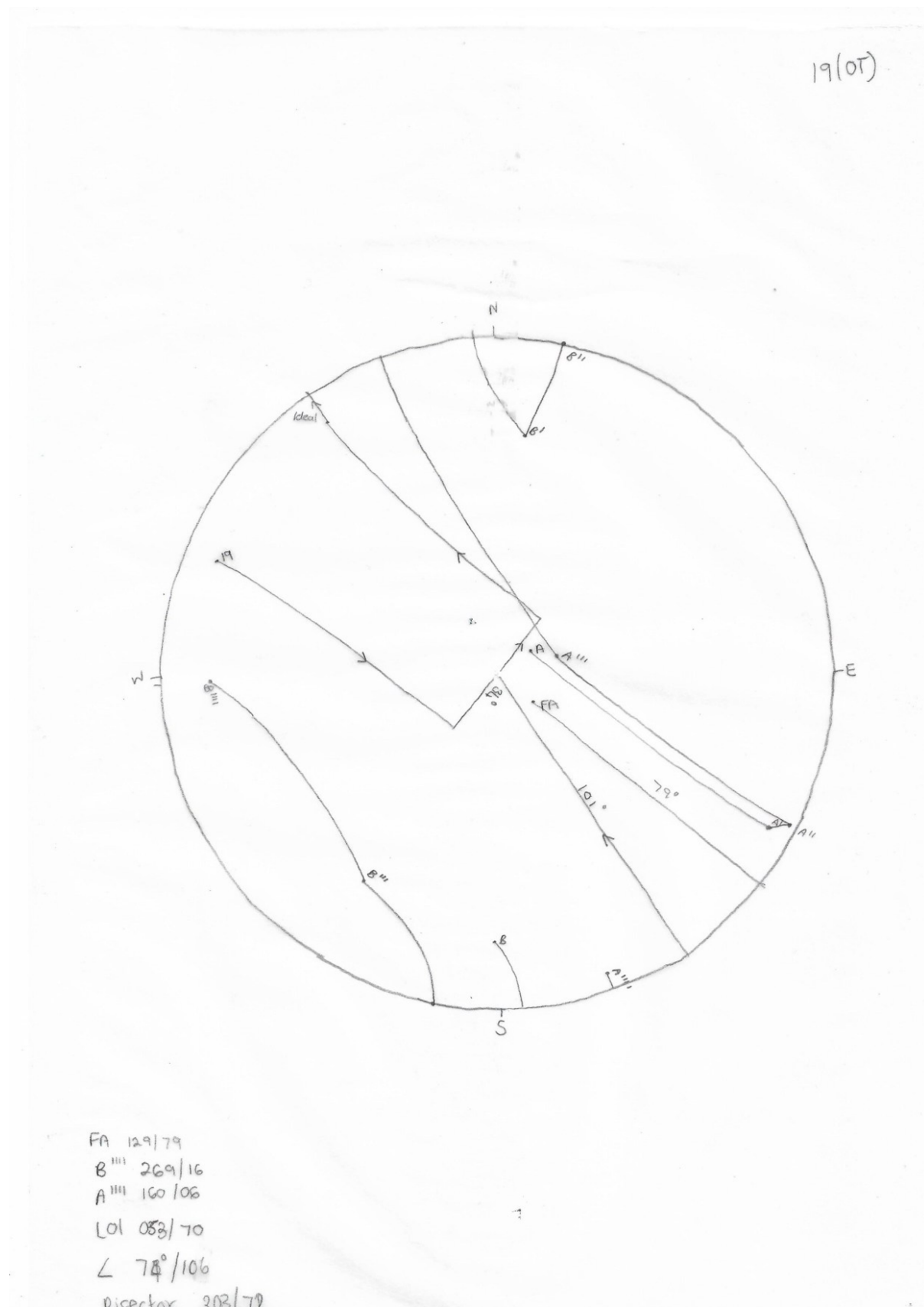
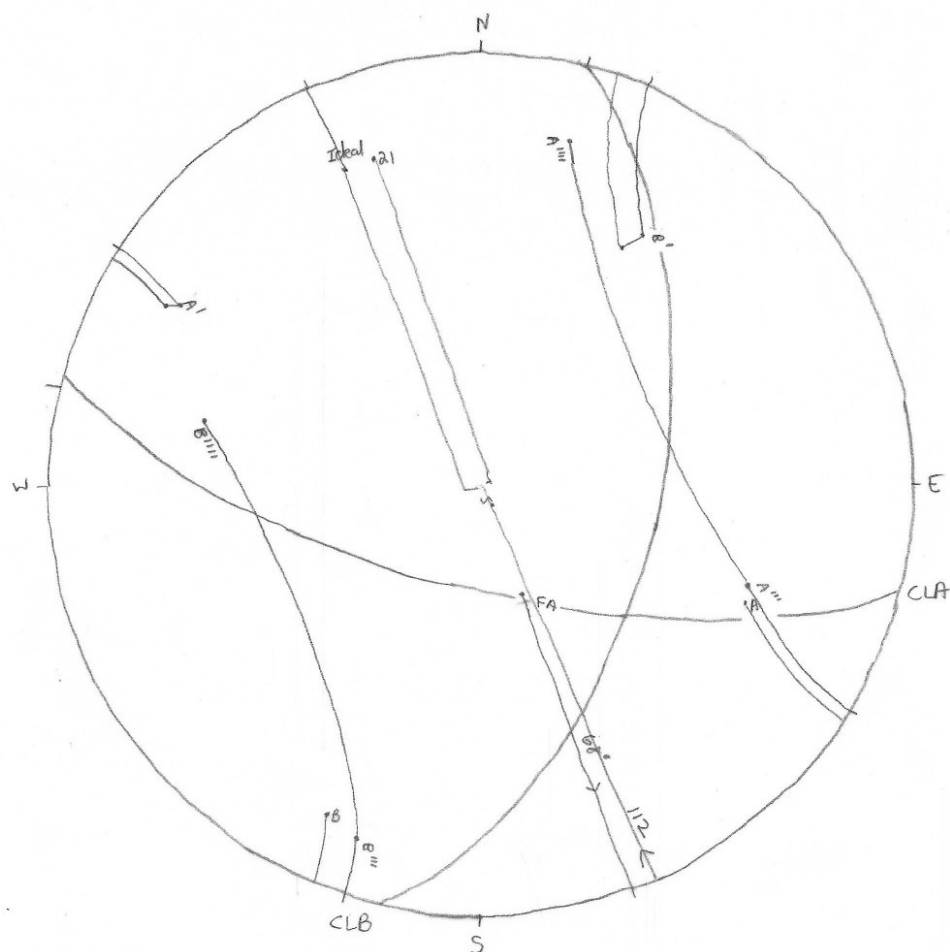


Figure S16: Rotation of Albat and Mayer (1989) location 19 to the tangential orientation and then to the horizontal.

21 (OT)



FA 160/68
 B' 283/34°
 A' 014/19
 LOT 131/53
 L 100/80°
 Bisector 325°/80

Figure S17: Rotation of Albat and Mayer (1989) location 21 to the tangential orientation and then to the horizontal.

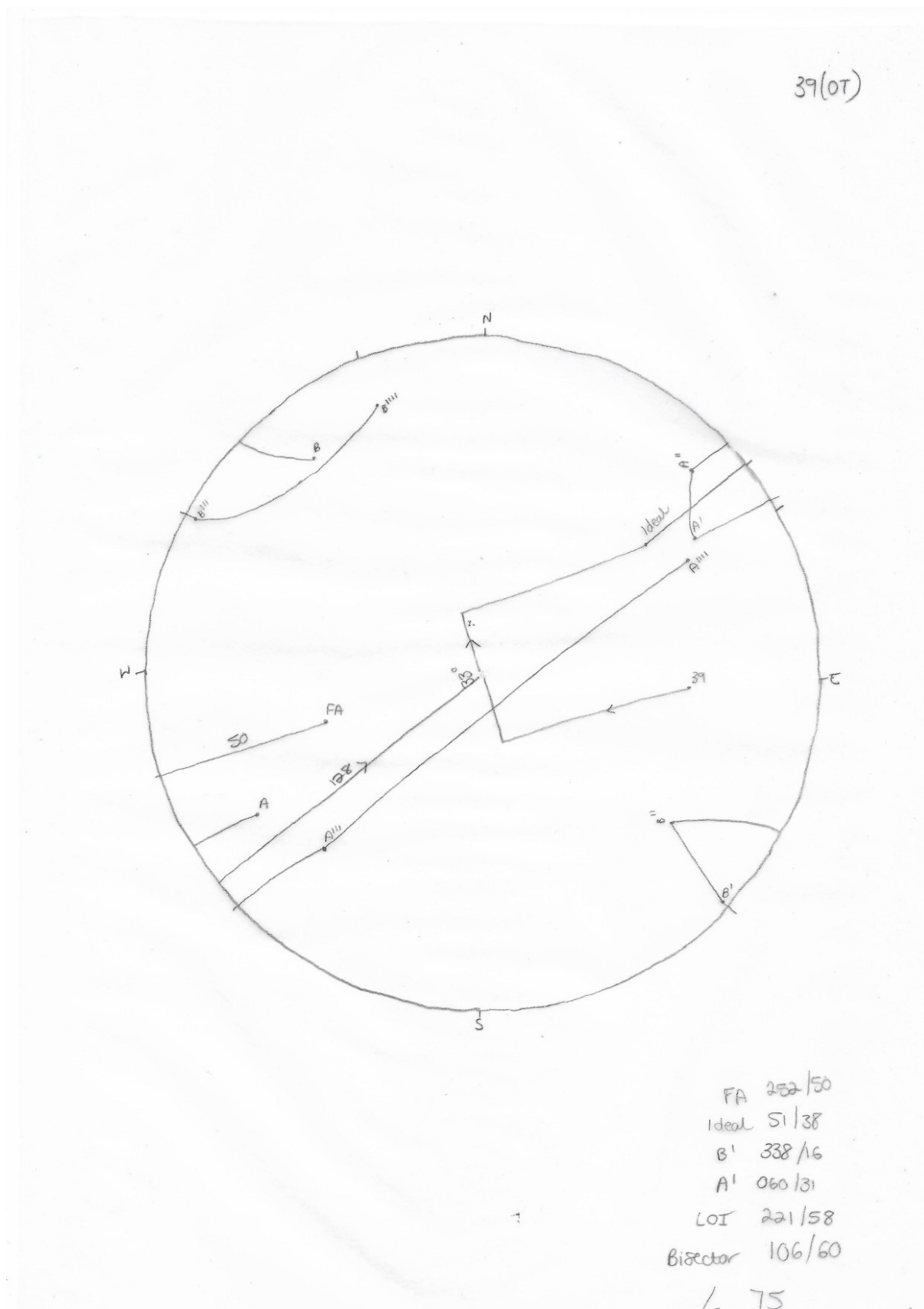


Figure S21: Rotation of Albat and Mayer (1989) location 39 to the tangential orientation and then to the horizontal.

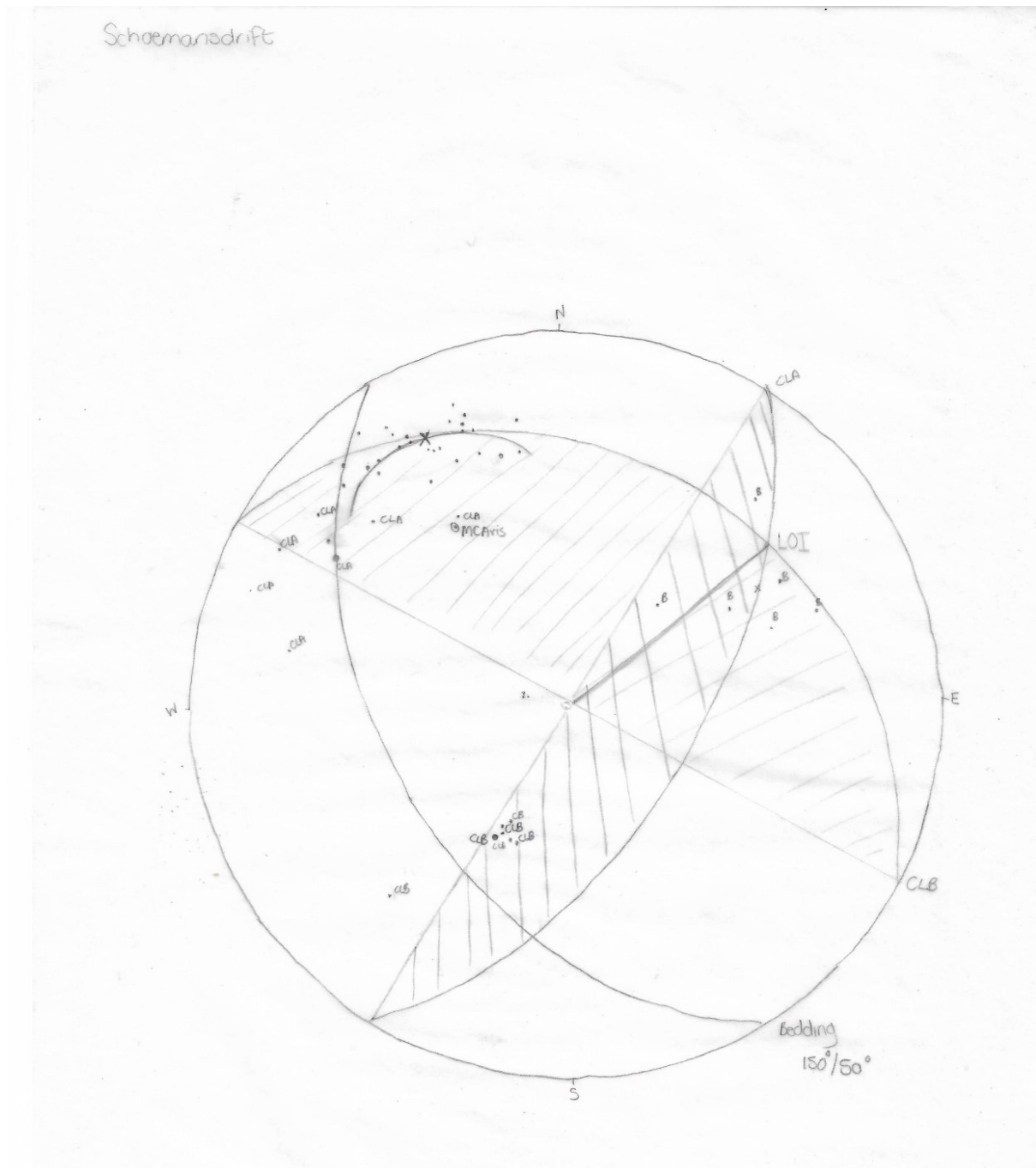


Figure S23: Schoemansdrift shatter cone data, showing the shatter cleavage orientation and the master shatter cone axis at the current outcrop. Bedding $n = 7$; CLB $n = 7$; CLA $n = 7$; Shatter cones $n = 25$.

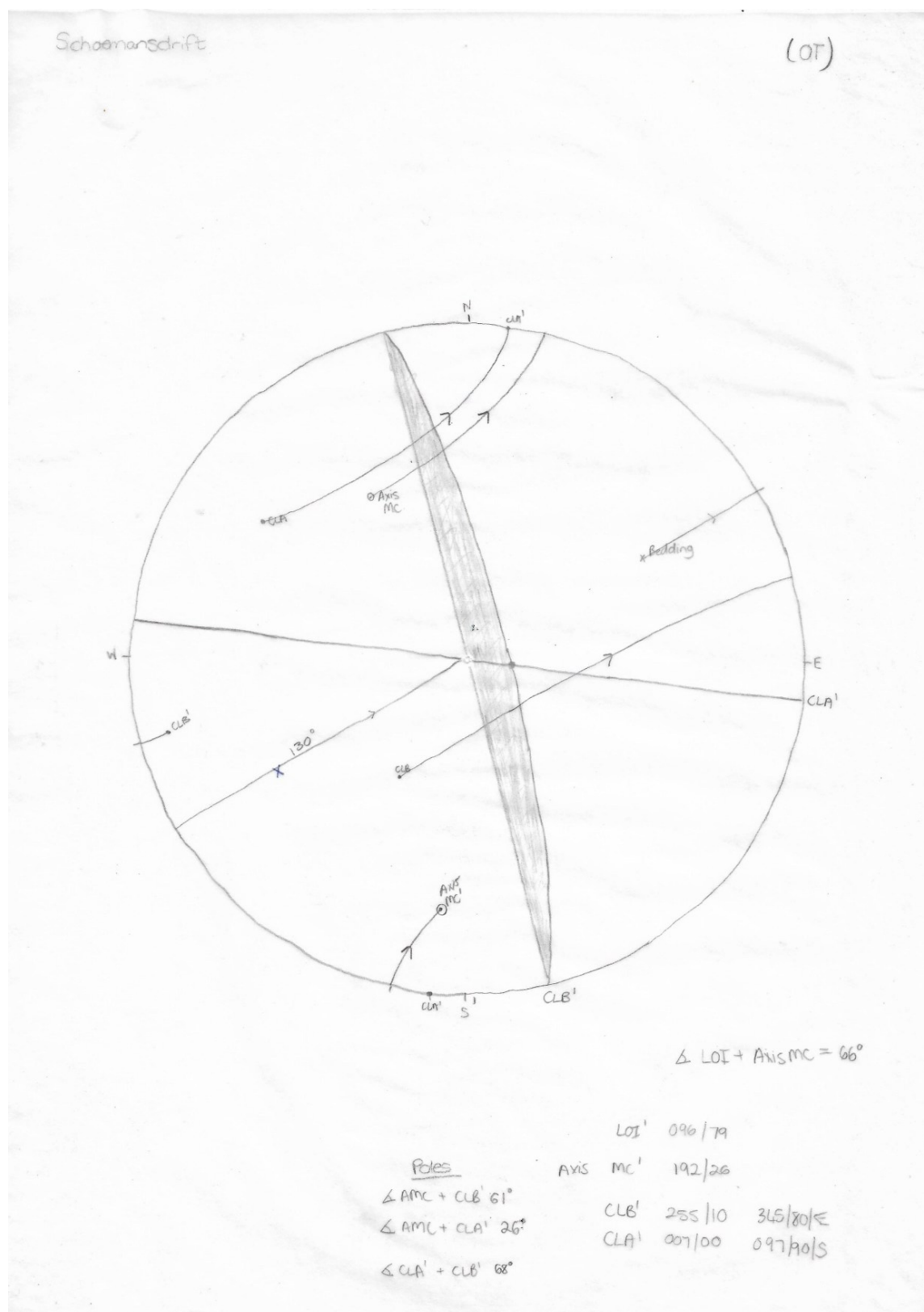


Figure S24: Schoemansdrift shatter cone data, showing the shatter cleavage orientation and the master shatter cone axis after rotation to the horizontal given an overturned dip.

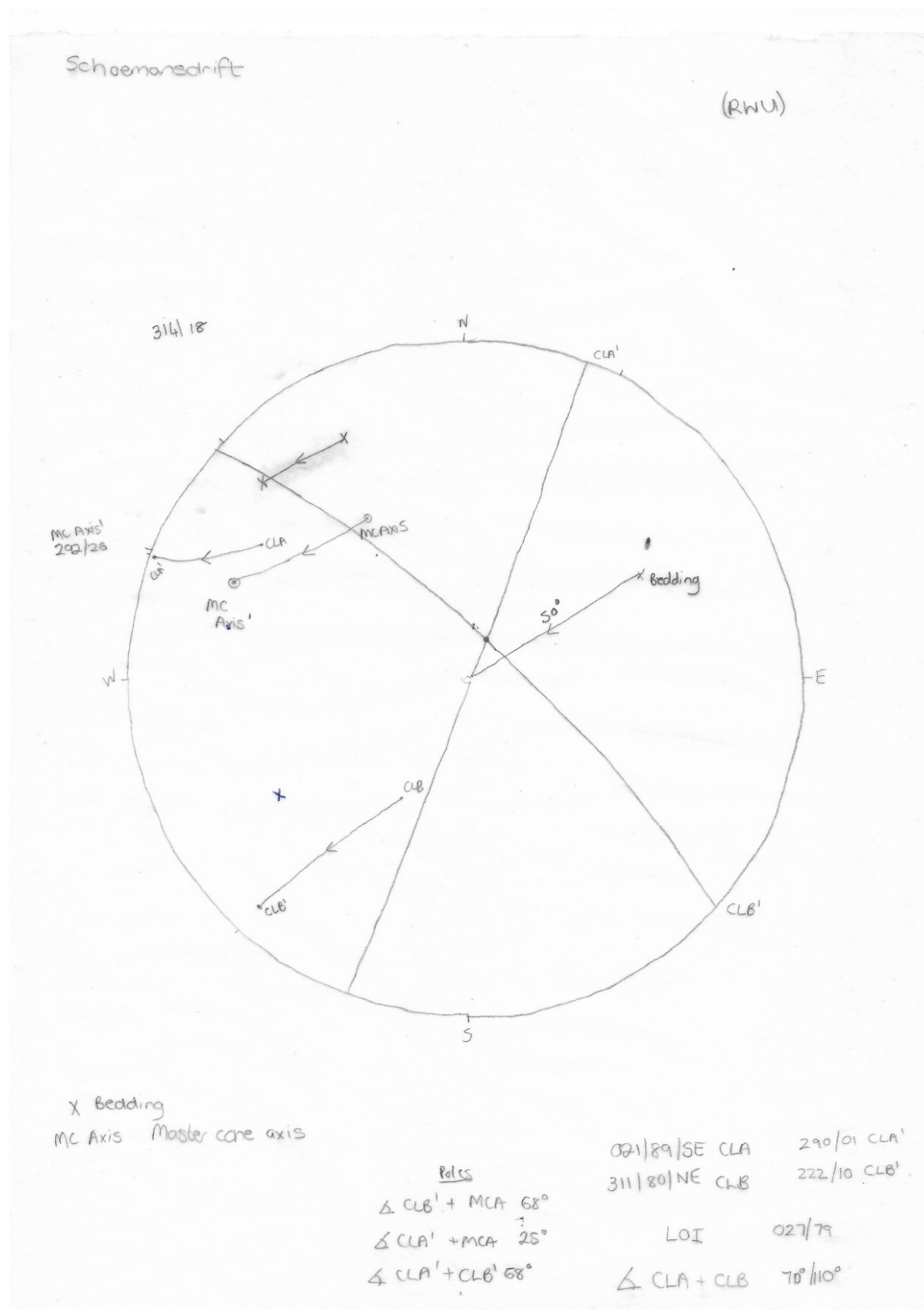


Figure S25: Schoemansdrift shatter cone data, showing the shatter cleavage orientation and the master shatter cone axis after rotation to the horizontal given a right-way-up dip.