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METAMORPHISM IN THE WITWATERSRAND BASIN: A PERSPECTIVE FROM THE VREDEFORT DOME

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METAMORPHISM IN THE WITWATERSRAND BASIN: A PERSPECTIVE FROM THE VREDEFORT DOME

by

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ABSTRACT

A model to explain the regional greenschist facies metamorphism observed in the rocks of the Witwatersrand goldfields is proposed based on new evidence from metapelitic units of the lower Witwatersrand Supergroup exposed in the collar of the Vredefort Dome. The P-T path derived for the rocks in the collar of the Dome indicates that they were subjected to a high geothermal gradient (~ 40 °C/km) mid-amphibolite facies metamorphic event concomitant with thickening of the overlying upper crust. Peak temperatures of 570-600 °C were reached at depths of 14-16 km during this event (denoted M1a here). After initial cooling following the M1a event, these rocks experienced rapid exhumation associated with the formation of the Dome, which led to a younger (M1b) paragenesis overprinting the M1a assemblages. The M1b paragenesis is consistent with a syn-doming geothermal gradient of ~ 30 °C/km for these rocks, indicating that they still retained high residual temperatures from the M1a event at the time of doming. This conclusion is supported by published evidence from the upper amphibolite to granulite facies rocks exposed in the basement core of the Dome, which also display high-temperature decompression textures. It is concluded that this evidence indicates that the M1a event affected the entire crustal section exposed in the Vredefort Dome and, thus, that it was of regional extent. The abnormal geothermal gradient attained during the M1a event, the anticlockwise P-T path and the partial overlap of the metamorphism with the Vredefort doming event, together provide strong support for a syn-Bushveld Complex timing for the metamorphism related to a major infusion of mafic magma below, and into, the lower crust. It is proposed that the greenschist facies metamorphic assemblages observed in the goldfields along the margins of the Witwatersrand Basin developed during this event. The lower grades in these rocks compared with the Vredefort Dome exposures reflects the shallower levels of burial of the rocks in the goldfields at the time of metamorphism.

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INTRODUCTION

During the past decade an increasing number of studies have established that the gold deposits of the Witwatersrand Basin contain widespread evidence of fluid-related transport and precipitation of gold, thus challenging the purely placer model (e.g., Hallbauer, 1986) for these deposits (see Frimmel (1994) and Phillips & Law (1994) and references therein). The resulting modified-placer (Frimmel et al., 1993; Frimmel, 1994) and epigenetic (Phillips & Myers, 1989) models recognise that metamorphism and hydrothermal fluid movement have played a major role in the evolution of these deposits. Consequently, if the genesis of the gold mineralization in the Witwatersrand Basin is to be fully understood, it is necessary to consider the post-depositional thermal history of the Basin.

Evidence that the rocks of the Witwatersrand goldfields experienced significant grades of metamorphism at some stage during their post-depositional history was first recorded by Young (1917). Since then, numerous occurrences of greenschist facies metamorphic assemblages have been described (e.g., Ramdohr, 1958; Viljoen, 1963; Schreyer & Bisschoff, 1982; Tweedie, 1986); however, it was only in the mid-1980s that a systematic regional study was undertaken throughout all the goldfields (Phillips, 1987). Phillips found a remarkable consistency in the metamorphic assemblages developed in pelitic horizons in all the goldfields, suggestive of relatively uniform lower greenschist facies metamorphic conditions. From this, he concluded that the rocks in the goldfields were affected by a regional-scale metamorphic event. Subsequent studies have confirmed Phillips' findings (e.g. Wallmach & Meyer, 1990; Frimmel, 1994; Zhou et al., 1994). Despite this agreement, however, the cause of the metamorphism and its timing has remained problematic.

Phillips & Law (1994) and Frimmel (1994) have recently attempted to synthesize these data into coherent thermal-hydrothermal models for the Basin. Unfortunately, while significant progress has been made in the study of fluids in the Basin (e.g., Phillips, 1988; Frimmel, 1994 and references therein), much of the essential data necessary to constrain the environment and causes of the metamorphism and to develop a thermal model is lacking. This is primarily because low-grade assemblages such as those observed in the goldfields are too fine-grained to preserve diagnostic reaction textures and inclusion relationships that are necessary for the estimation of P-T conditions and the construction of P-T paths and, until recently, to obtain meaningful geochronological results. An added problem in low-grade assemblages is that it may be difficult to distinguish separate metamorphic events and even prograde and retrograde parts of the same event. This information is best obtained in coarser-grained, higher grade, assemblages where mineral reaction and inclusion relationships, and mineral compositional zoning, can be used to establish P-T conditions and the P-T path during metamorphism.

With the exception of a few areas where slightly higher-grade assemblages involving kyanite, andalusite, cordierite and/or biotite have been reported in the Witwatersrand Supergroup (e.g., Schreyer & Bisschoff, 1982; Tweedie, 1986; Phillips, 1987), the only area in the Basin in which higher-grade metamorphism is observed is in the vicinity of the

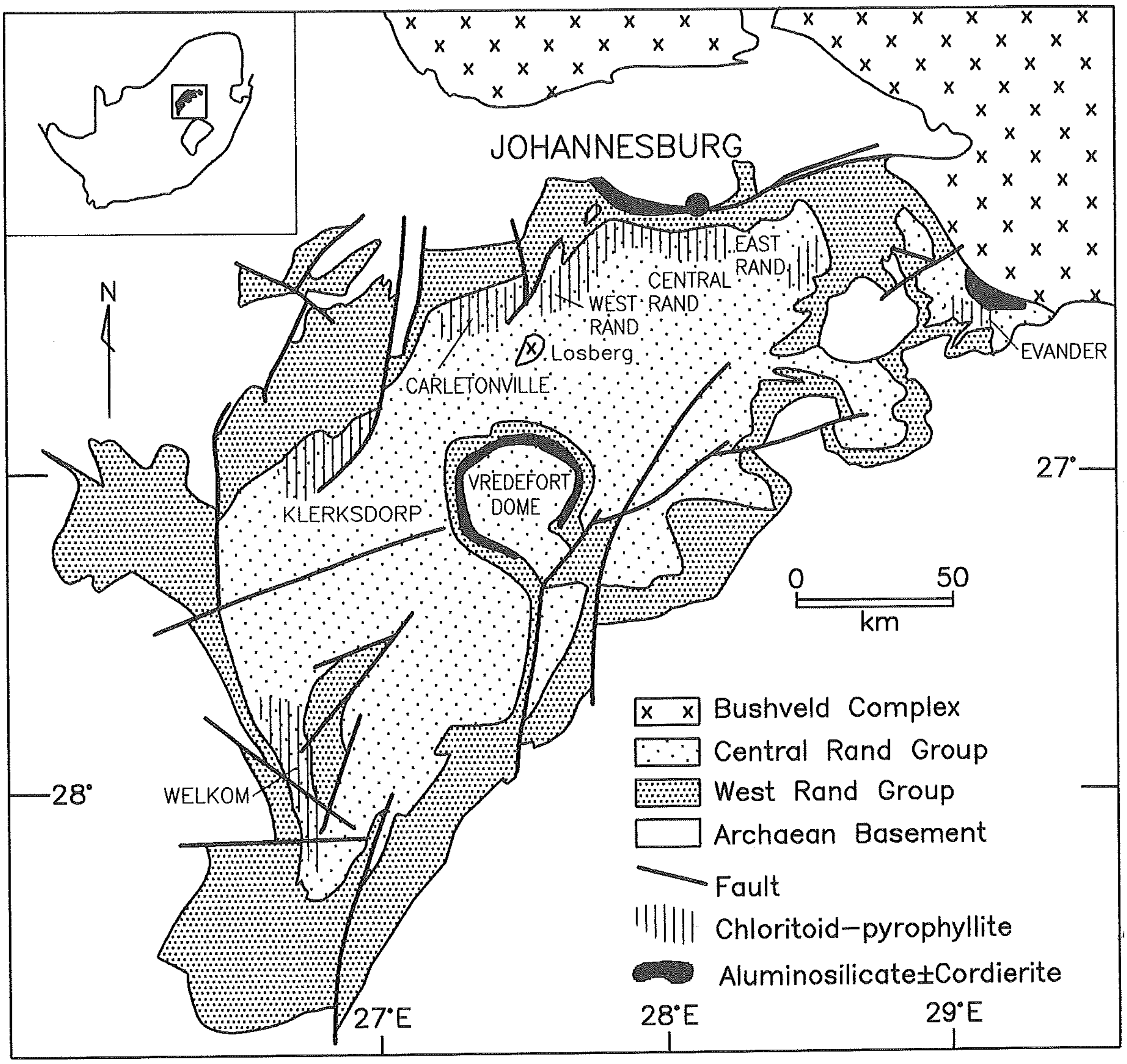


Figure 1: Approximate distribution of rocks of the Witwatersrand Supergroup and known metamorphic grade in the Witwatersrand Basin (modified from Frimmel, 1994, and Phillips & Law, 1994). The Vredefort Dome occupies a central position in the Basin. The position of the Losberg Complex, a correlative of the Bushveld Complex, is also indicated.

Vredefort Dome, situated in the geographic centre of the Basin (Fig. 1). Here rocks of the lower Witwatersrand Supergroup contain mid-amphibolite facies assemblages (Hall & Molengraaff, 1925; Nel, 1927; Bisschoff, 1982); however, this metamorphism has generally been interpreted as a localised effect associated with the formation of the Dome and, as such, has been regarded as unrelated to the metamorphism observed in the wider Basin (see, e.g., Phillips & Law, 1994; Frimmel, 1994).

In this paper, evidence is presented that suggests that the medium-grade metamorphism observed in the rocks of the Witwatersrand Supergroup in the vicinity of the

Dome formed part of a regional event that could account for the metamorphism observed in the goldfields. Additional constraints are discussed which enable the establishment of both the timing and the likely environment in which the Basin-wide metamorphism was effected.

REGIONAL GEOLOGICAL SETTING

The rocks of the Witwatersrand Supergroup form part of a late Archaean to Proterozoic supracrustal sequence (Fig. 2) deposited in a succession of basins on the Archaean granite-greenstone basement of the Kaapvaal Craton. Deposition commenced with the extrusion of the subaerial mafic volcanics of the Dominion Group at 3.07 Ga (Armstrong et al., 1991). This was followed by the deposition of up to 7-8 km of siliciclastic sediments belonging to the Witwatersrand Supergroup - a lower argillaceous-arenaceous sub-tidal marine sequence (West Rand Group), and an upper, arenaceous-rudaceous sequence (Central Rand Group) of fluviatile/alluvial fan origin. These rocks are overlain by up to 2-3 km of subaerial mafic lavas and subsidiary rift-related sediments of the 2.7 Ga (Armstrong et al., 1991) Ventersdorp Supergroup that are overlain, in turn, by the 2.6-2.06 Ga (F. Walraven, pers. comm., 1995) Transvaal Supergroup, comprising a 1-2 km thick lower dolomitic sequence (Chuniespoort Group) and an upper 3-4 km thick argillaceous-arenaceous sequence (Pretoria Group). The Rooiberg Group, a 2.06 Ga (Walraven, in press) felsitic volcanic sequence, was extruded shortly before the intrusion of up to 6 km of mafic-ultramafic magmas and younger granites of the Bushveld Complex into the upper levels of the Transvaal Sequence at 2.065-2.054 Ga (Walraven et al., 1990; Walraven & Hattingh, 1993). The Bushveld Complex crops out mainly to the north of the Witwatersrand Basin, but appears to have originally extended southwards over the Basin at least as far as the Losberg Complex (Coetzee & Kruger, 1989) (Fig. 1).

Following the formation of the Bushveld Complex, the region was subjected to widespread deformation associated with the formation of the Vredefort Dome (McCarthy et al., 1986). Among the features developed are a series of large-scale folds arranged tangentially around the Dome. The innermost of these, the Potchefstroom Synclinorium, is, to a large extent, responsible for the present-day preservation of the rocks of the Witwatersrand Supergroup and the geometry of the Basin (McCarthy et al., 1990). These authors have also suggested that regional tilting of the craton occurred at some time after the formation of the Dome, leading to variable amounts of exhumation of the rocks in the Witwatersrand Basin (up to 5-7 km in the vicinity of the Vredefort Dome, decreasing northwards). Phanerozoic sediments, lavas and intrusions of the Karoo Supergroup unconformably overlie the Basin.

METAMORPHISM IN THE WITWATERSRAND GOLDFIELDS

The cause of the metamorphism affecting the Witwatersrand goldfields has been the subject of considerable speculation in recent years; however, modelling has been hampered by the lack of information concerning the P-T-t path followed by the rocks during the metamorphism. This lack of P-T-t path information is to be expected in such low-grade

AGE (Ga)			THICKNESS MAXIMUM	VREDEFORT
	Bushveld Complex	Pretoria Group	3–4km	<4.5km
2.43	Transvaal Supergroup	Chuniespoort Group	1–2km	
2.65	Ventersdorp Supergroup	Platberg Group	4km	<2km
		Klipriviersberg Group		
2.70	Witwatersrand Supergroup	Central Rand Group	3km	5–6km
2.90		West Rand Group	4km	
	Dominion Group		2.5km	0.2km
3.07	Archaean Basement			

Figure 2: Generalized stratigraphic column of the Witwatersrand Basin, showing maximum thicknesses and estimates of stratigraphic thicknesses in the vicinity of the Vredefort Dome (see text for details). Sources of geochronological data referred to in text.

terranes where mineral assemblages are too fine-grained to preserve diagnostic reaction textures and inclusion relationships, and where the problem of distinguishing prograde and retrograde textures is particularly acute. In the face of these problems, authors have attempted to place the metamorphism within the context of known post-depositional events in the region. This has been done either directly, using geochronological data, or indirectly, by correlation with the known postdepositional history of the Basin.

Although P constraints in the goldfields are poor, there is general consensus that the metamorphism in the goldfields occurred under pressure conditions of 2-3 kbar (see discussion in Phillips & Law, 1994; Frimmel, 1994; Zhou et al., 1994). These estimates, corresponding to 7-11 km of overburden, require that the bulk, if not all, of the Transvaal Sequence sedimentation was complete prior to the metamorphic event. According to Phillips et al. (1989), the gold mineralisation in the Black Reef Formation, which they link to the metamorphic event, supports this timing. The greenschist facies metamorphism observed in the Ventersdorp Supergroup (Frimmel, 1994) provides further support for a post-Black Reef timing for this event.

Phillips & Myers (1989) suggested that the metamorphism in the goldfields might be related to the formation of the Vredefort Dome, based upon correlation of the syn-metamorphic cleavages that they observed in the goldfields with a regional cleavage attributed by McCarthy et al. (1986) to the doming event. However, given the complex post-depositional history of the Basin (e.g. Roering et al., 1990; Myers et al., 1990; Stanistreet & McCarthy, 1991), such a correlation appears premature, especially as several workers ascribe the growth of metamorphic minerals to a pre-cleavage event (e.g., Schreyer & Bisschoff; 1982; Frimmel, 1994) and the porphyroblasts described by McCarthy et al. (1986) appear to have undergone retrogression prior to, or during, formation of the regional cleavage (Frimmel, 1994; Courtnage et al., 1995).

Frimmel (1994) suggested that the regional metamorphism occurred at between 2.5 Ga and 2.6 Ga, based on radiogenic isotopic data from authigenic rutile (Robb et al., 1990) and zircon (R. Armstrong, unpublished data). Frimmel (1994) concluded that the metamorphism occurred during deposition of the Transvaal Supergroup and was followed by a Vredefort- or Bushveld- related retrogressive event. Phillips et al. (1989) argued that the cluster of ages around 2.3 Ga obtained from a variety of rocks in and around the Witwatersrand Basin may reflect the age of metamorphism, although they acknowledged that this is by no means a certain age.

Hart et al. (1995) suggested that the remanent magnetism in the rocks in the Vredefort Dome was reset during a thermal event that must have postdated doming. Layer et al. (1988) established that resetting of the magnetism occurred throughout the Witwatersrand Basin. They obtained a K-Ar age of 1.95 Ga from Witwatersrand shales in support of this and concluded that the rocks were affected by a regional thermal event that was related either to the Bushveld Complex or to the formation of the Vredefort Dome. Their inability to distinguish between the two reflects the close temporal overlap between the two events, a fact which has been noted by several authors who have attempted to reconcile both to a single event (e.g., Hamilton, 1970; Master, 1990; Elston, 1995). The issue of the relative timing of the Vredefort and Bushveld Complex events, and the implications which this has for the metamorphic history of the Witwatersrand Basin, are addressed in this paper.

GEOLOGY OF THE VREDEFORT DOME

The Vredefort Dome is located in the geographic centre of the structurally preserved remnants of the Witwatersrand Basin (Fig. 1). It comprises a 45-km wide core of predominantly granitic Archaean basement gneisses that are surrounded by a collar of steeply dipping to overturned strata belonging to the full stratigraphic sequence observed in the region (Dominion Group, Witwatersrand, Ventersdorp and Transvaal Supergroups) (Fig. 3). The collar is between 15 and 20 km wide, but estimates of the thickness of the supracrustal sequence suggest that deformation, probably related to the formation of the Vredefort Dome, thickened the stratigraphic sequence. Estimated pre-doming thicknesses for the different units are: 5-6 km for the Witwatersrand Supergroup, < 2 km for the Ventersdorp Supergroup (P. Linton, pers. comm., 1995) and < 4.5 km for the Transvaal Supergroup (P. Erikson, pers. comm., 1995). The rocks of the core and collar are intruded by numerous mafic sills and dykes, the bulk of which are of Ventersdorp age (Pybus et al., 1995) and the remainder of Bushveld Complex age (Bisschoff, 1972). Several small alkali granite and dioritic complexes, dated at approximately 2.2 Ga (Walraven & Elsenbroek, 1991), are also found in the vicinity of the Dome (Fig. 3). The southern part of the structure is unconformably overlain by Karoo sediments and volcanics.

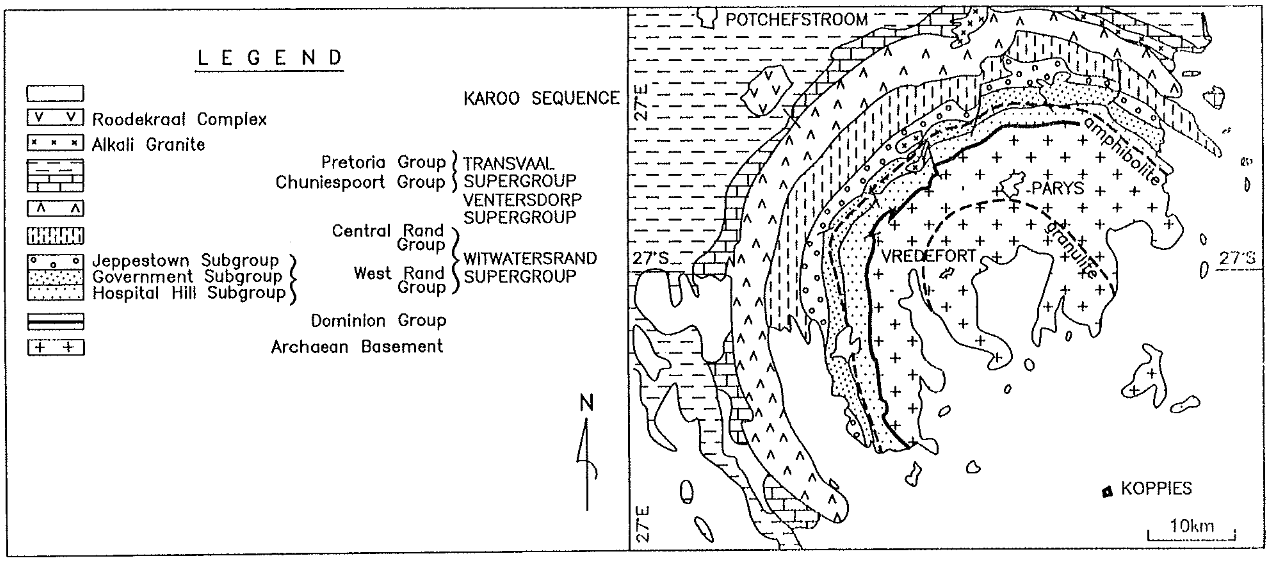


Figure 3: Geological map of the Vredefort Dome, showing the distribution of metamorphic grade.

The metamorphic grade of rocks exposed in the Vredefort Dome increases radially inwards. In the outer parts of the collar, the grade is reminiscent of the greenschist facies grades developed in the Witwatersrand Supergroup in the goldfields. The grade increases to mid-amphibolite facies in the lower parts of the Witwatersrand Supergroup (Hospital Hill and Government Reef Subgroups) and the Dominion Group (Hall & Molengraaff, 1925; Nel, 1927; Bisschoff, 1982; Jackson, 1994). The rocks in the outer parts of the basement core, comprising granitic to granodioritic migmatitic gneisses (the Outer Granite Gneiss; Stepto, 1990) contain upper amphibolite facies assemblages (Nel, 1927; Bisschoff, 1982), while those in the centre of the Dome, comprising leucocratic granofelses and subsidiary paragneisses and mafic orthogneisses (Inlandsee Leucogranofels Terrane and Steynskraal Metamorphic Zone; Stepto, 1990) contain granulite facies assemblages (e.g., Schreyer & Abraham, 1978; Schreyer et al., 1978; Stepto, 1990). The amphibolite-granulite transition occurs approximately 8-10 km from the core-collar contact (Bisschoff, 1982; Stepto, 1990). The limited metamorphic work that has been done on the basement rocks suggests that the basement contains a complex metamorphic history with at least three separate events having been recognised (e.g. Stepto, 1990). Geochronological studies (e.g. Hart et al., 1981; Allsopp et al., 1991) suggest that the basement was affected by events at approximately 3.1 Ga and 2.8 Ga; however, three lines of evidence confirm a third event:

1. Hart et al. (1981) obtained a Rb-Sr whole-rock age of 1.94 Ga for an undeformed anatectic granite in the basement;

2. Ar-Ar mineral stepheating ages from the basement core record a circa 2.0 Ga event (Allsopp et al., 1991); and

3. shock deformation features linked to the formation of the Dome have been variably recrystallised (e.g. Schreyer, 1983; Fricke et al., 1990; Grieve et al., 1990; Stepto, 1990).

The medium-grade metamorphism in the Witwatersrand Supergroup collar strata has been linked to this last event (e.g., Schreyer, 1983; Bisschoff, 1982; Stepto, 1990) which is perceived by most authors as being related to a localised igneous heat source that is centred on the Dome (e.g., Nel, 1927; Schreyer, 1983; Bisschoff, 1982; Stepto, 1990). This contact-metamorphic scenario for the Vredefort Dome metamorphism has led to workers in the wider Witwatersrand Basin discounting a possible link with the basin-wide metamorphism (e.g. Frimmel, 1994; Phillips & Law, 1994).

The formation of the Dome, at approximately 2.0 Ga, has been linked to high strain rate shock deformation processes that produced voluminous pseudotachylite breccias, shatter cones, planar microdeformation phenomena in quartz and the high-pressure quartz polymorphs (coesite and stishovite) (see review in Reimold, 1993). While there is general agreement that many of these features are consistent with the rocks having experienced shock deformation, two schools of thought exist on the cause of this shock event. The more popular hypothesis is that the Vredefort Dome formed in response to a large asteroid impact (e.g. Daly, 1947; Dietz, 1961). A second hypothesis relates the formation of the Dome to endogenous processes during which internal pressures built up sufficiently to cause shock deformation (e.g. Nicolaysen & Ferguson, 1990). A feature of the endogenous hypothesis is that it provides an explanation for the temporal and spatial overlap of the high-grade metamorphism with the doming event. It is this perception, that the metamorphic effects observed in the rocks of the Dome are a consequence of a localised heat source, that has led to the view that this metamorphism is unrelated to that seen in the goldfields (e.g. Frimmel, 1994; Phillips & Law, 1994).

In the past two years, an increasing body of evidence has accumulated in support of an impact origin for the Vredefort Dome. This includes:

1. new geochronological results that show that the voluminous pseudotachylite occurrences can be related to a single event at approximately 2.025 Ga (Trieloff et al., 1994; Spray et al., 1995; Kamo et al., 1995);

2. the confirmation that planar microdeformation features observed in quartz from the Dome developed in response to shock processes (Leroux et al., 1994) and the recognition of similar features in zircon (Kamo et al., 1995); and

3. the detection of an extraterrestrial Re-Os component in the Vredefort Granophyre, which has long been regarded as an impact-melt rock (Koeberl et al., in prep.).

This additional evidence in favour of an impact origin for the Vredefort Dome raises questions about the origin of the metamorphism and its relation to the doming event. In the remainder of this paper, evidence is discussed that resolves this issue and also provides clues to both the origin and the timing of the metamorphism in the Witwatersrand goldfields.

METAMORPHIC ROCKS OF THE LOWER WITWATERSRAND SUPERGROUP IN THE COLLAR OF THE VREDEFORT DOME

Previous Work

The mid-amphibolite facies metamorphic rocks of the lower Witwatersrand Supergroup in the collar of the Vredefort Dome have been intensively studied by Hall & Molengraaff (1925), Nel (1927) and Bisschoff (1969, 1982). They concluded that the rocks are hornfelses and that the metamorphism was thus a static or thermal event related to a hidden igneous pluton. Bisschoff (1982) described an apparent increase in grade towards the alkali granite intrusions. This, together with a noticeable bulge in the isograd pattern around the small alkali granite bodies in the collar, led him to speculate that they may represent the surface expression of the hidden pluton. Several authors have, however, pointed out that the low crystallization temperature of the alkali granite (650 °C, Bisschoff, 1982) cannot be reconciled with either the extent, or the high grade, of the metamorphism. Bearing these considerations in mind, Schreyer (1983) and Stepto (1990) speculated that the heat source would have to be a mafic intrusion. There is, however, no compelling evidence, either from field studies or from geophysical studies (Durrheim et al., 1991), that such a body exists (see discussion in Gibson, 1993). The idea of localised plutonic activity centred on the Dome and concomitant with the doming event (see above) has formed an integral part of the endogenous model for the formation of the Vredefort Dome.

Field Relations

Nel (1927) and Bisschoff (1982) mapped an "outer limit of contact metamorphism" in the collar of the Vredefort Dome based upon the first appearance of macroscopic porphyroblasts in the rocks. According to them, this limit is elliptical and slightly eccentric with respect to the stratigraphy, intersecting progressively higher stratigraphic levels towards the alkali granite intrusions in the northwestern sector of the collar. In contrast, Bisschoff (1982, Fig. 20) obtained a broadly concentric, bedding-parallel distribution of isograds from metamorphosed mafic sills in the collar rocks. During the course of this study, a sharp change was found between coarse-grained porphyroblastic assemblages in pelitic units of the Hospital Hill and lower Government Subgroups and pelitic units from stratigraphically higher levels that are fine-grained, highly weathered slates. Microscopically, these latter rocks are characterised by a chlorite-white mica paragenesis. In some samples, ghost outlines of rod-like porphyroblasts (possibly chloritoid?) suggest that a higher-grade paragenesis may have existed prior to an intense retrogression, possibly linked to a cleavage event. This may explain the presence of upper greenschist - lower amphibolite facies assemblages in the outer parts of the collar described by Nel (1927) and Bisschoff (1982). The poor outcrop in the northeastern and southwestern parts of the collar make it difficult to evaluate whether the discordance described by Nel and Bisschoff does, in fact, exist. It was also not possible to discern a 'staurolite isograd', as described by Bisschoff (1982, Fig. 20), around the alkali granites - although the coarsest-grained staurolite occurs in this area, and staurolite is also found throughout the exposed section of the collar.

The metapelitic rocks of the Hospital Hill and lower Government Subgroups are intercalated with orthoquartzites and banded iron formations, together with numerous metabasic sills ('epidiorites', Hall & Molengraaff, 1925). The pelites form rounded outcrops in which sedimentary layering and delicate sedimentary features such as graded bedding, cross-bedding and soft-sediment deformation structures are well-preserved. Coarsely porphyroblastic cordierite, andalusite, garnet, biotite and, occasionally, staurolite can be identified in outcrop.

At least two syn-metamorphic cleavages occur in the rocks, indicating that the metamorphism was dynamothermal rather than static (Gibson, 1993). The rocks are cut by spaced brittle fracture cleavages, microfaults, pseudotachylite, multiply striated joint surfaces and cataclasite breccias. The spaced fracture cleavages, in particular, involve small offsets of the peak metamorphic porphyroblasts, syn-metamorphic fabrics and bedding (Fig. 4).

Figure 4: Spaced brittle fracture cleavage displacing bedding, syn-metamorphic S1 cleavage and M1a cordierite porphyroblast (arrowed). Hospital Hill Subgroup, Kommandonek.

METAMORPHIC ASSEMBLAGES

The pelitic units contain a variety of mineral assemblages involving biotite, cordierite, andalusite, garnet and/or staurolite. Rare kyanite and fibrolite are also occasionally found in association with andalusite. The pelitic assemblages are frequently muscovite-poor or muscovite-deficient. Variation in the mineral parageneses appears to be predominantly a function of variation in bulk rock composition (see below).

Bulk Rock and Mineral Chemistry

The wide variety of mineral assemblages observed in the pelitic rocks reflects the wide range of bulk rock compositions encountered (Fig. 5; see also Bisschoff, 1982). The bulk compositions were obtained by XRF using the equipment in the Department of Geology, University of the Witwatersrand and the mineral compositions from the electron microprobe in the Department of Geology at the University of Pretoria. X_{Fe} values (= $Fe^{2+}/(Mg + Fe^{2+})$) have been calculated assuming that the bulk of the Ti^{4+} and Fe^{3+} occurs in biotite. This conforms with petrographic observations that show only limited magnetite and ilmenite in the pelites. X_{Fe} ranges from 0.43 to 0.71 and X_{Al} [= $(Al^{3+} - 3K^{+} - Na^{+})/(Al^{3+} - 3K^{+} - Na^{+} + Fe^{2+} + Mg^{2+})$] from 0.15 to 0.44. K_2O values are generally low, and are reflected in the paucity or absence of muscovite in many assemblages (Fig. 5b).

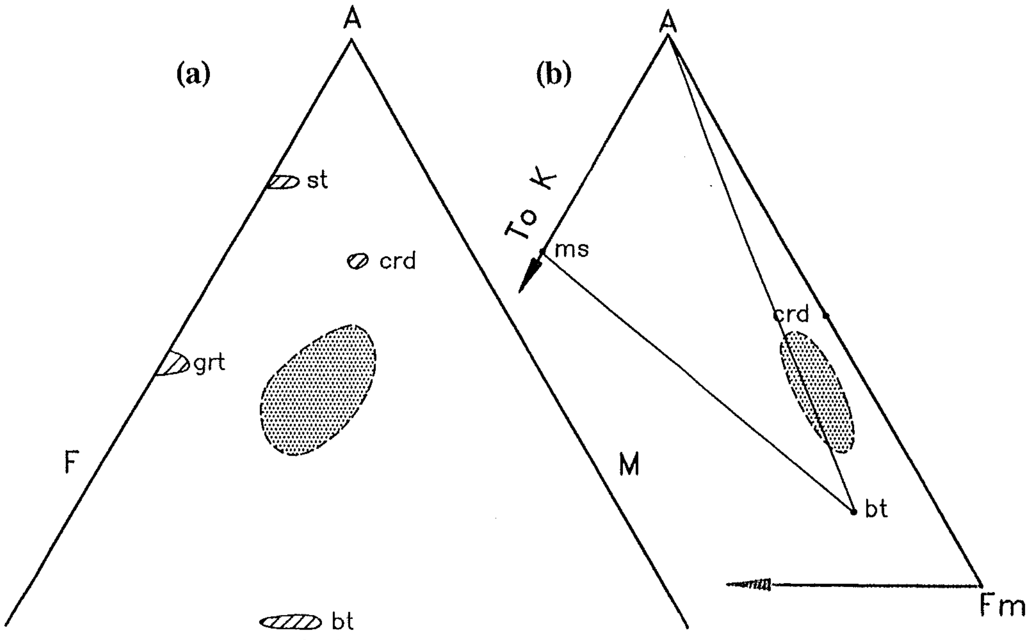


Figure 5: (a) AFM projection from MS+qtz+H_2O showing bulk rock compositions for metapelites from the lower West Rand Group, and representative mineral compositional data.
(b) AK(FM) projection showing bulk rock compositions for metapelites from the lower West Rand Group.

Mineral compositions show less variation than the bulk rock compositions, with biotite X_{Fe} values typically between 0.55-0.65, cordierite between 0.45-0.50 and staurolite ranging from the pure Fe end-member to rare, more Mg-rich compositions of X_{Fe} = 0.85. Little, if any, evidence was found of compositional zoning, despite clear textural evidence of at least two growth zones in garnet, cordierite and, in some cases, andalusite (Bisschoff, 1982; Gibson, 1993). In rare cases, some evidence of zoning was found in garnet, involving increasing Mg and Fe, and decreasing Mn and X_{Fe} from core to rim. Mn/(Fe+Mg+Mn+Ca) ranges up to 0.3 and X_{Fe} between 0.93-0.97.

(a)

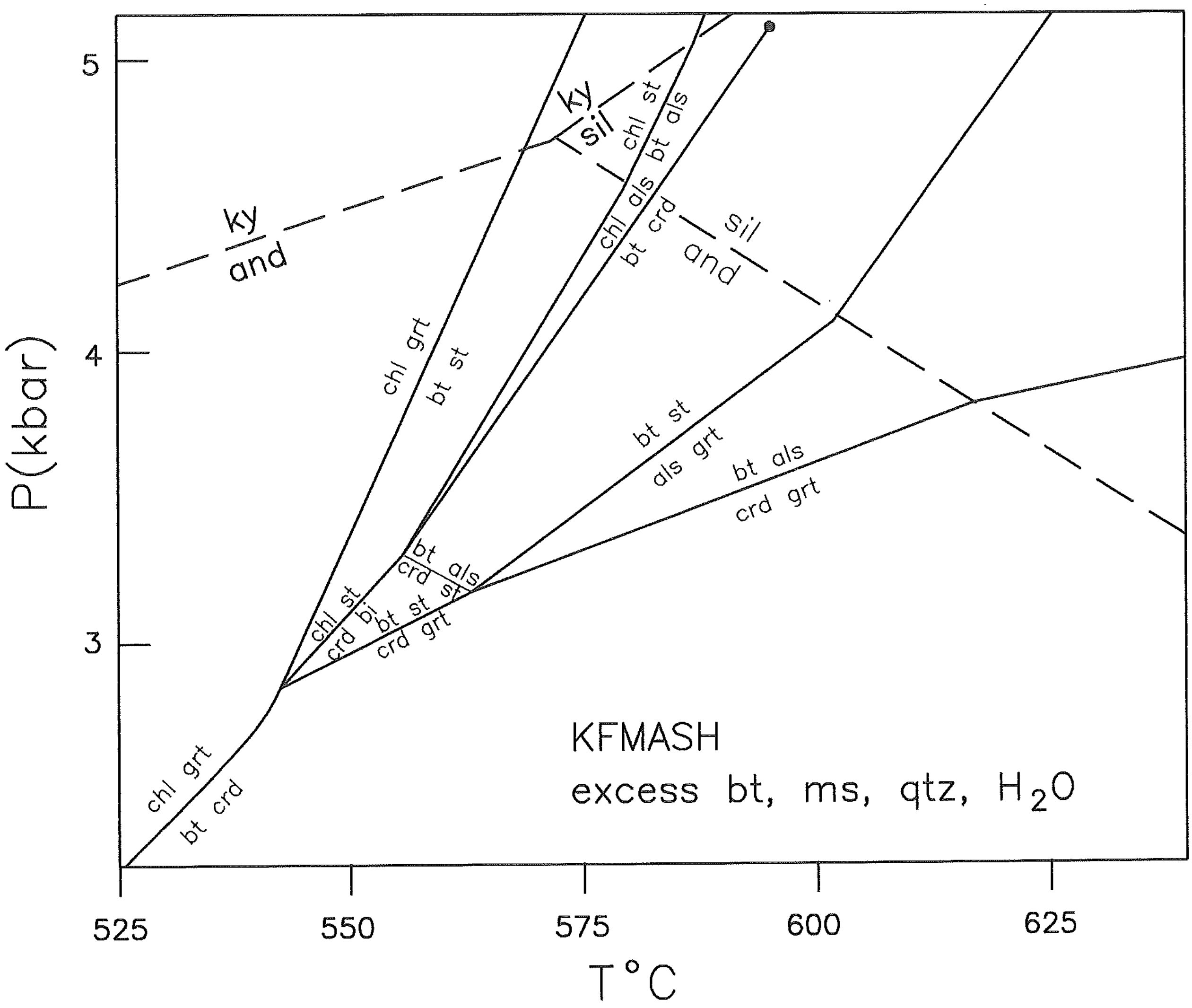
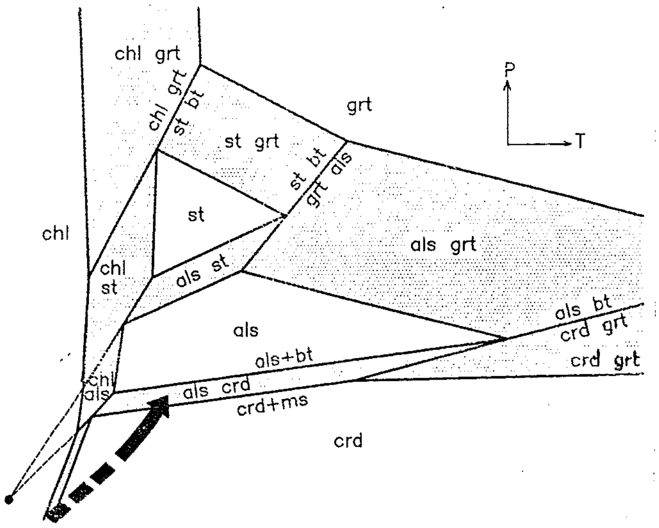


Figure 6: *(a) Calculated KFMASH P-T grid for reactions for garnet, staurolite, aluminosilicate, cordierite and chlorite (excess biotite, muscovite, quartz and H_2O) (from Dymoke & Sandiford, 1992).*
(b)-(d) Schematic P-T pseudosections for (b) low-X_{Fe}, (c) high-X_{Fe} and (d) intermediate-X_{Fe} bulk compositions, showing P-T paths derived from reaction textures and inclusion relations.

(b)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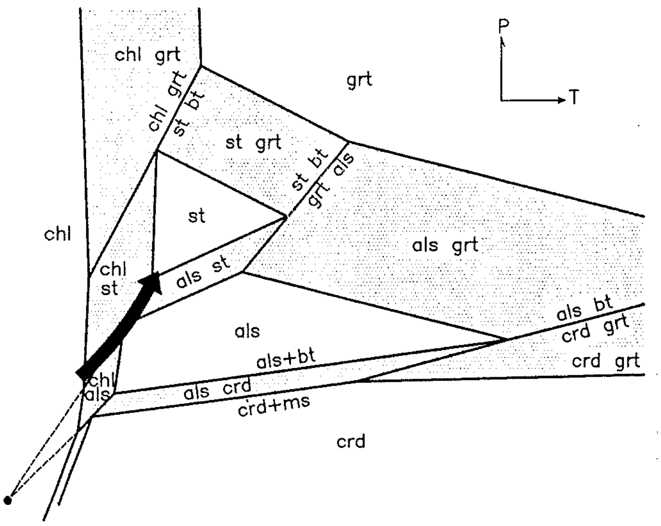
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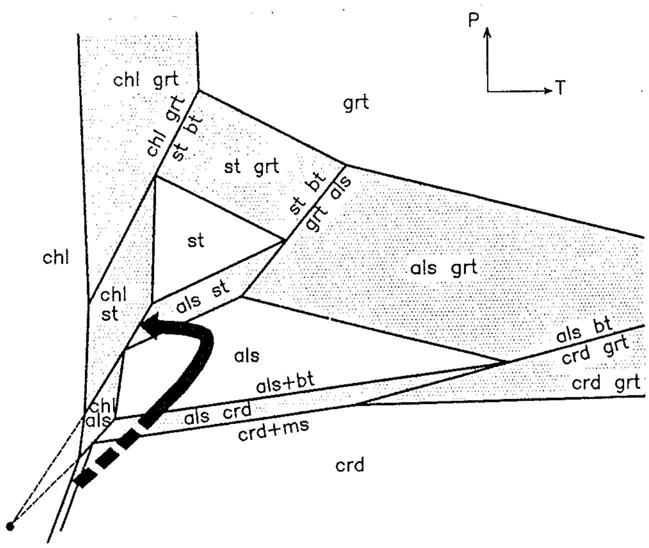
(d)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KFMASH Petrogenetic Grid, Reaction Textures and the P-T Path

The bulk composition of most pelitic rocks approximates the simplified chemical system KFMASH (K_2O-FeO-MgO-Al_2O_3-SiO_2-H_2O), for which well-constrained P-T petrogenetic grids are available (e.g. Harte & Hudson, 1979; Holland & Powell, 1990). The KFMASH petrogenetic grid for pelites, derived by Dymoke & Sandiford (1992), using the THERMOCALC programme of Holland & Powell (1990) (Fig. 6a). This thermodynamic grid is in good agreement with the petrologically derived grid of Harte & Hudson (1979), with the exception of the cordierite-andalusite reaction which has a positive slope in Figure 6. Although several of the observed assemblages from the Vredefort Dome are muscovite-poor, they strongly resemble the muscovite-present assemblages developed in intercalated lithologies. Hudson and Harte (1985) demonstrated a similar relationship between muscovite-poor and muscovite-present assemblages in the Buchan region and showed that muscovite can disappear and reappear in an assemblage as successive reactions are crossed along the P-T path. The authors have thus adopted the AFM projection and KFMASH grid for the bulk compositions observed in the Witwatersrand Supergroup.

The actual position of the divariant reactions used to derive the P-T path in P-T space varies as a function of bulk rock composition. For this reason, rocks with different bulk compositions are considered separately. During this study it was possible to identify assemblages related to two metamorphic "events" within individual samples - peak metamorphic assemblages were associated with the M1a event which was later overprinted by a lower-grade M1b event. For simplicity, the two events have been kept separate below.

The M1a Event

Pelites with the most magnesian and aluminous bulk compositions (X_{Fe} ~ 0.4-0.5; Fig. 5a) typically contain an M1a paragenesis comprising andalusite + cordierite + biotite ± muscovite. Mn-rich garnet is often present as an additional phase. Textural evidence of the timing between andalusite and cordierite is usually equivocal. However, cordierite is frequently included in andalusite (the reverse relationship is not seen) and, occasionally, andalusite occurs at the contact between cordierite and biotite-muscovite aggregates. The latter suggests a reaction such as

cordierite + muscovite = andalusite + biotite,

which implies a positive P-T path slope in Figure 6b.

In the pelites with Fe-rich bulk compositions ($X_{Fe} > 0.65$), no evidence of prograde cordierite is found and the assemblages comprise either andalusite + staurolite + biotite ± garnet or garnet + biotite + chlorite. There is some evidence to suggest that staurolite and andalusite grew after garnet, but this is not conclusive and the relationship between andalusite and staurolite could not be determined (Fig. 6c).

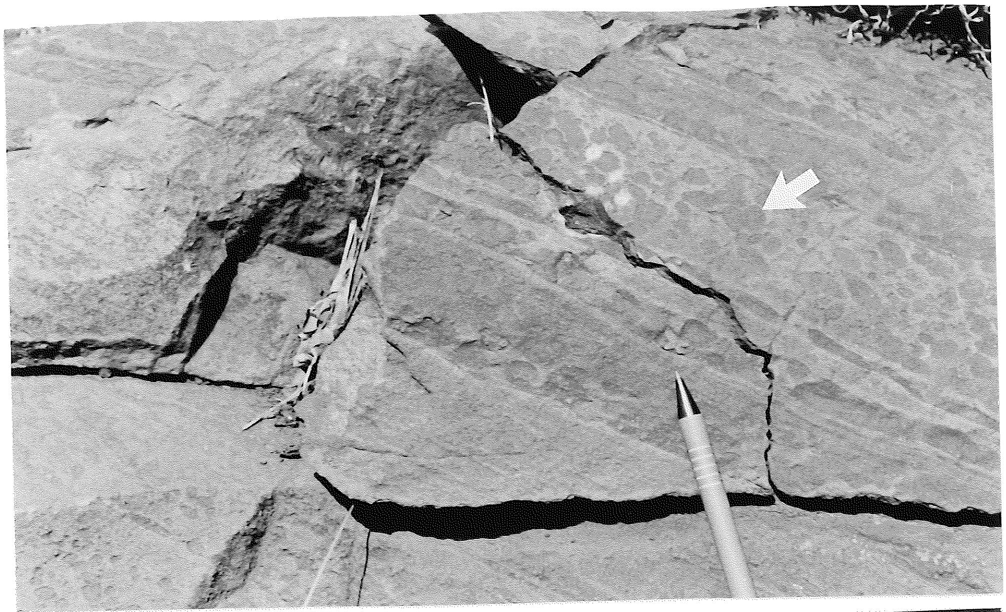
In intermediate bulk compositions (X_{Fe} ~ 0.55-0.65) lacking muscovite, the rocks contain pseudomorphed cordierite porphyroblasts in addition to andalusite and biotite (± garnet) (Fig. 7a,b). The andalusite is variably replaced by staurolite (Fig. 7c), with more

Figure 7: (a) Outcrop of intermediate-X_{Fe} pelite showing large M1a cordierite porphyroblasts (dark ovals, arrowed) and butterfly-twinned andalusite (indicated by pencil).

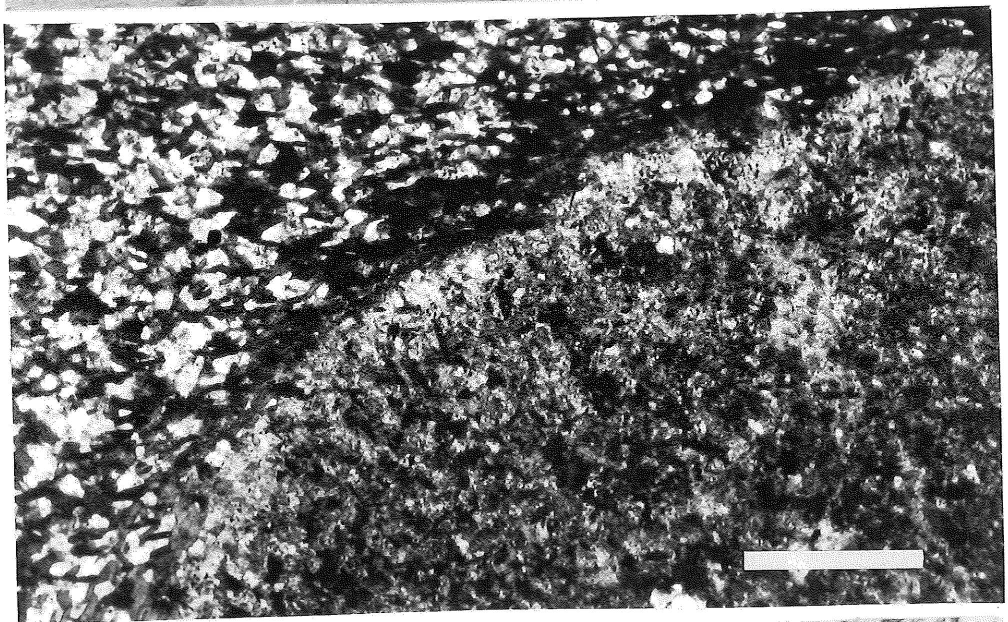
(b) Part of a cordierite porphyroblast from (a) showing pseudomorphous replacement by staurolite + biotite + quartz Scale bar 200 μm.

(c) Staurolite (arrowed) replacing andalusite porphyroblast. Scale bar 100 μm.

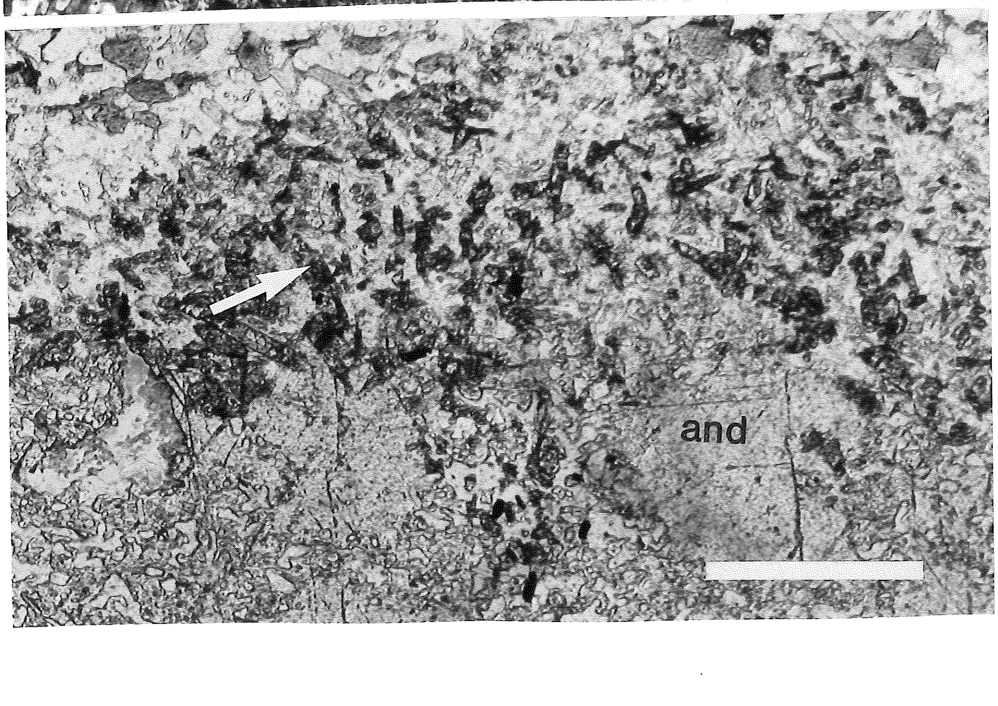
(a)



(b)



(c)



advanced pseudomorphing occurring in more biotite-rich areas. This indicates the reaction

andalusite + biotite = staurolite + muscovite

The absence of muscovite may be explained by the pseudomorphous replacement of cordierite by staurolite, biotite and quartz. The reaction:

cordierite + muscovite = staurolite + biotite + quartz

is not stable on conventional KFMASH grids with muscovite in excess. However, in muscovite-poor assemblages, if all the muscovite was consumed during prograde growth of cordierite and andalusite, cordierite could be stabilized beyond its normal limits to higher P. The production of muscovite by the andalusite breakdown reaction would, however, provide the catalyst for the cordierite to also break down to a staurolite-bearing paragenesis, seen in the pseudomorphs. The combined evidence from reaction textures thus suggests an anticlockwise P-T path during metamorphism, with staurolite developing during or slightly after the peak of M1a metamorphism (Fig. 6d).

The M1b Event

Evidence of the M1b event is most obvious in the more Fe-rich bulk compositions. It comprises the development of highly poikilitic cordierite porphyroblasts that overgrow the peak M1a paragenesis and that appear to be associated with fine-grained biotite aggregates. These cordierite porphyroblasts are characterised by extremely poikilitic, clouded cores surrounded by a clear, inclusion-free, rim zone (Fig. 8a). In biotite-rich areas, the rim may display a distinctive digitated appearance (Fig. 8a). This cordierite also occurs as coronae between M1a staurolite porphyroblasts and matrix biotite (Fig. 8b), indicating a reaction such as:

staurolite + biotite = cordierite + muscovite

The timing of the M1b event is constrained by the deformation event that produced the pseudotachylite-bearing brittle fracture cleavage. The cleavage cross-cuts and offsets the M1a porphyroblasts (Figs. 4, 8c) and the syn-metamorphic fabrics, and comminuted M1a porphyroblasts occur as inclusions in the pseudotachylite. However, the cleavage is overgrown by the M1b cordierite (Fig. 9a,b) and the pseudotachylite is recrystallised to a M1b cordierite + biotite paragenesis (Fig. 9c). This timing suggests that M1b occurred after the shock event that was responsible for the formation of the Vredefort Dome.

CONSTRAINTS ON P-T CONDITIONS

The petrogenetic grid in Figure 6a places semi-quantitative constraints on the M1a P-T conditions. According to Figure 6a the muscovite-bearing M1a parageneses are stable on the high-P side of the univariant reactions:

(a)

(b)

(c)

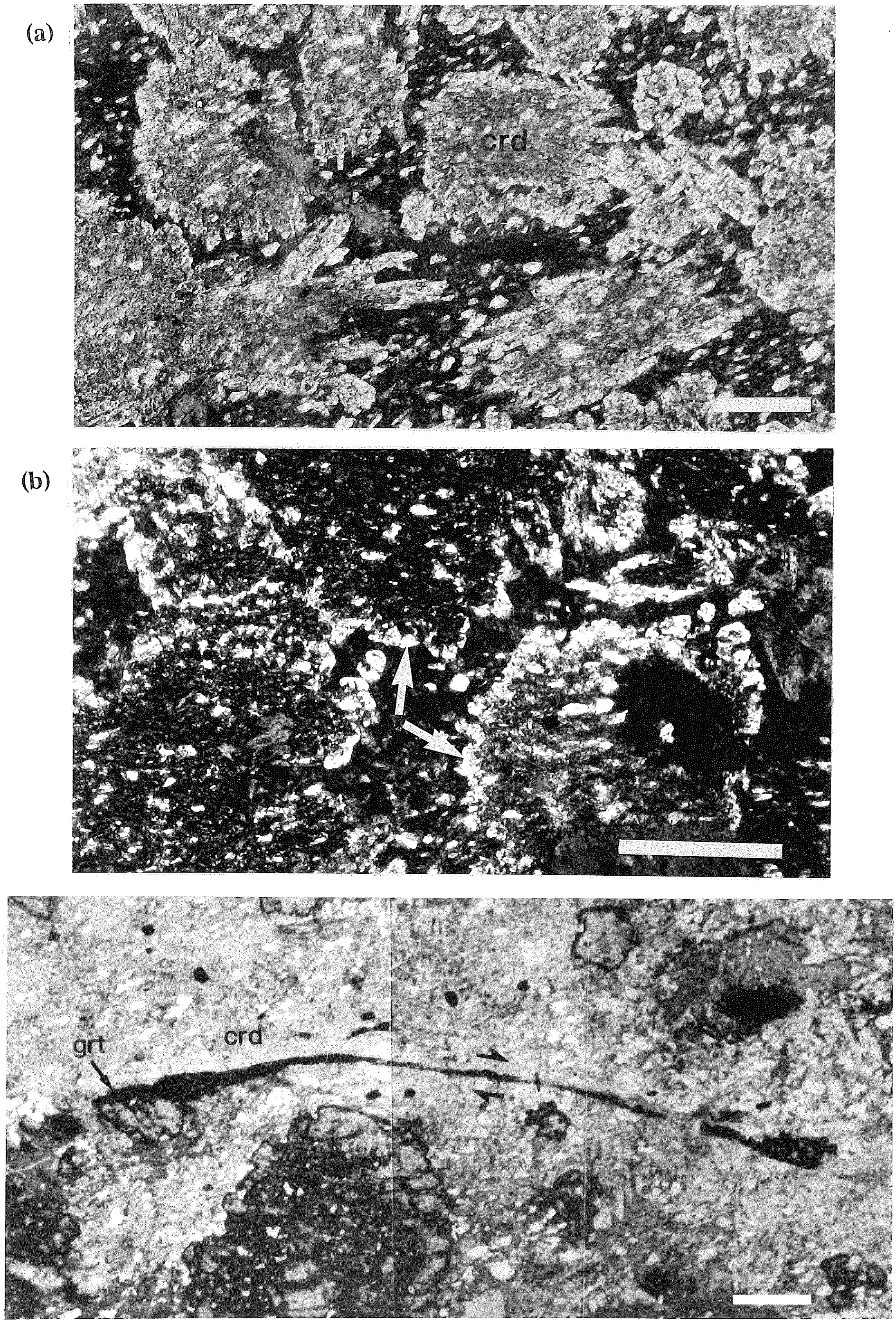


Figure 8: *(a) M1b cordierite in biotite matrix showing distinctive dentate margins. Scale bar 100 μm.*

(b) M1b cordierite coronae (arrowed) between M1a staurolite and biotite. Scale bar 100 μm.

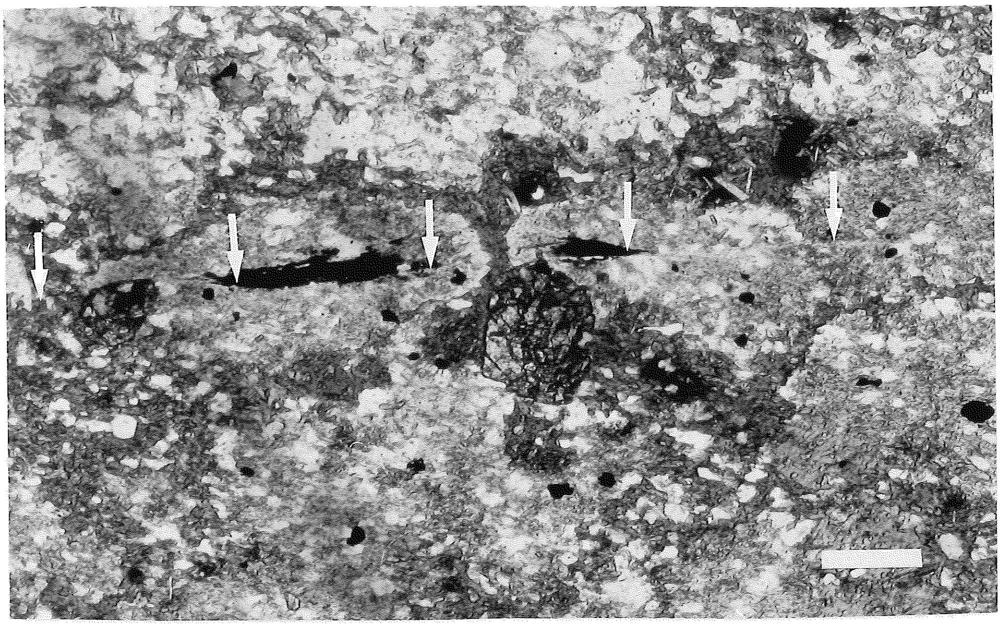
(c) M1a garnet sheared and comminuted along brittle fracture cleavage. Scale bar 200 μm.

Figure 9: (a) Brittle fracture cleavage (arrowed) overgrown by M1b cordierite-biotite paragenesis. The cleavage is defined in places by fine-grained opaques (black). Scale bar 200 μm.

(b) Enlarged SEM backscatter image of (a). Cordierite is medium grey, biotite light grey. Arrow indicates quartz grain truncated by cleavage.

(c) Pseudotachylite matrix recrystallized to M1b cordierite-biotite paragenesis (Compare with Fig. 8a). Arrow indicates biotite-chlorite clast. Scale bar 100 μm.

(a)



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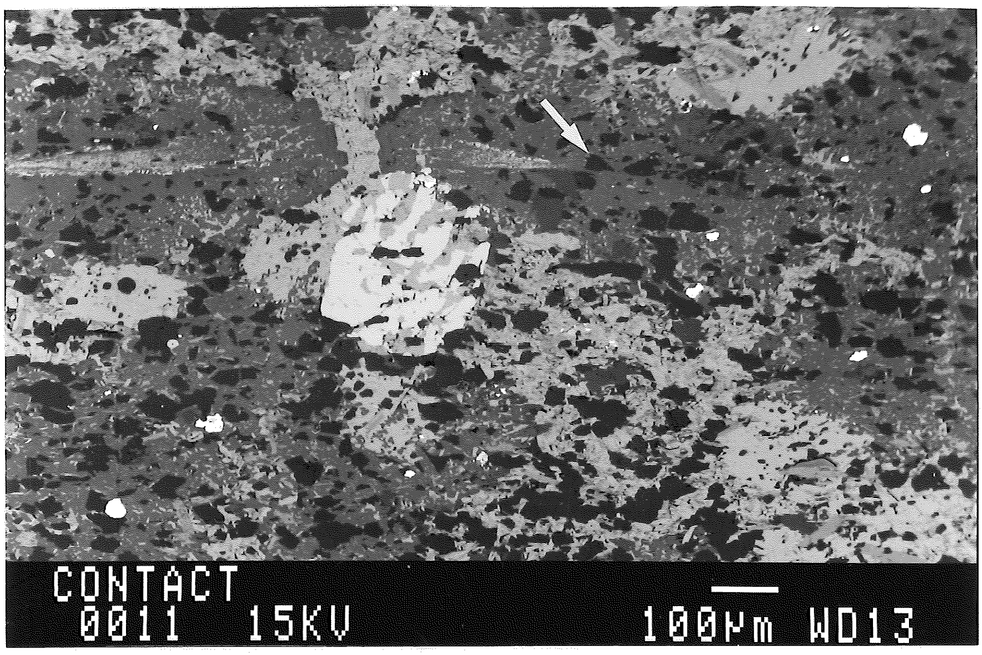
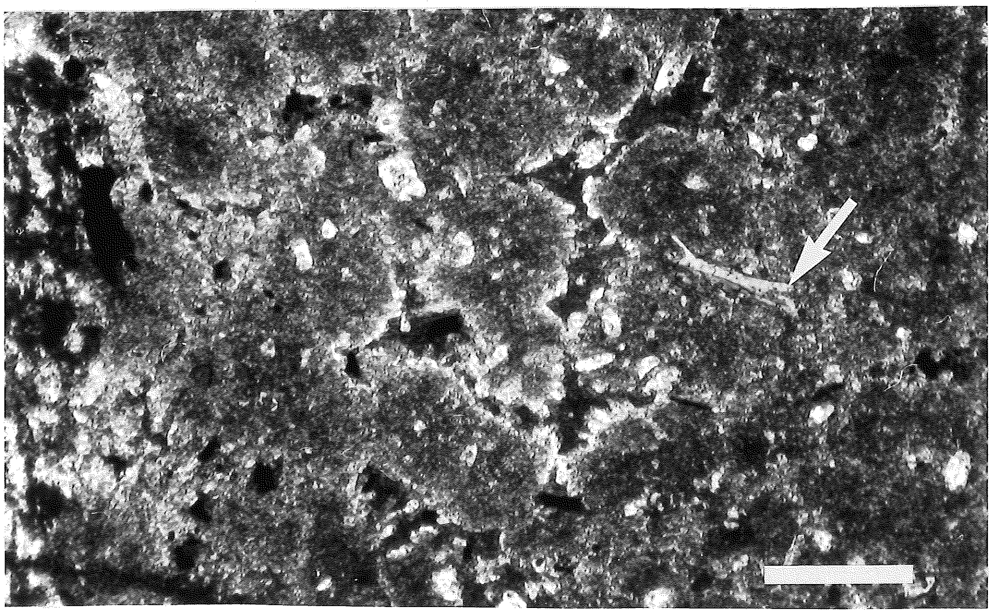


(c)



cordierite + garnet + muscovite = aluminium silicate + biotite + quartz; and

cordierite + staurolite + muscovite = aluminium silicate + biotite + quartz.

These equilibria define minimum P-T conditions of approximately 3.5 kbar, 550 °C. An upper P constraint is provided by the stability of andalusite in all assemblages. Using the aluminosilicate phase boundaries of Holland & Powell (1990), this implies a maximum P of approximately 4.5 kbar. The P-T pseudosection of Dymoke & Sandiford (1992) (Fig. 10a), which was constructed for bulk compositions closely resembling the Fe-rich pelites under discussion, suggests temperatures in the range 570-600 °C and pressures closer to 4.5 kbar, indicating an overburden of 14-16 km at the metamorphic peak and a peak geothermal gradient of approximately 40 °C km^{-1}. The P estimate is consistent with the estimated pre-doming thickness of 14 km obtained by Manton (1965).

The reaction indicated by the M1b cordierite coronae around staurolite is only possible over a very restricted range of P-T conditions (see discussion in Dymoke & Sandiford, 1992), however, stability is likely to be extended in a H_2O-free system. Restricted availability of H_2O, which is typical of rocks that have undergone prograde metamorphism, means that chlorite-forming retrograde reactions are inhibited, with the consequence that staurolite and cordierite stability fields may be extended to lower T. The authors have shown this schematically in Figure 10. The shallow dP/dT slope of the reaction suggests a strong component of isothermal decompression for the P-T path. This is consistent with the evidence that the deformation preceding the late cordierite growth was produced by high strain rates associated with the formation of the Dome, i.e. that exhumation occurred rapidly.

Maximum T estimates obtained from garnet-biotite and garnet-cordierite thermometry (unpublished data) are in the range 500-525 °C for the Thompson (1976) and Ferry & Spear (1978) calibrations. This discrepancy with respect to the peak T estimates obtained from the P-T pseudosection may reflect the high Mn content of the garnets. Alternatively, given the homogeneous nature of the garnet compositions, despite clear textural zoning (Gibson, 1993), it could reflect chemical re-equilibration associated with the M1b event. The topology of the grid of Dymoke & Sandiford (1992) suggests that the staurolite breakdown reaction to cordierite is stable at $T < 530$ °C and $P < 3.5$ kbar for $X_{Fe} = 0.7$ in the H_2O-depleted system, consistent with the geothermometry results. The coarse grain size of the M1b cordierite indicates that it is unlikely that the T during M1b could have been below ~ 500 °C.

Given the present-day preservation of the near-complete overburden from the time of metamorphism in the vicinity of the Dome, and evidence that at least approximately 14 km of overburden existed immediately prior to doming (Manton, 1965) an initial isobaric cooling path is inferred for the metamorphic rocks after the M1a peak prior to the decompression event. The M1b textures suggest that the upper crustal geothermal gradient at the time of exhumation must have been in the order of 30 °Ckm^{-1}.

The post-pseudotachylite (and post-cleavage) timing of the low-P cordierite paragenesis indicates that exhumation of these rocks from depths of 14-16 km was associated with a high strain-rate event. Furthermore, this exhumation occurred after the attainment of

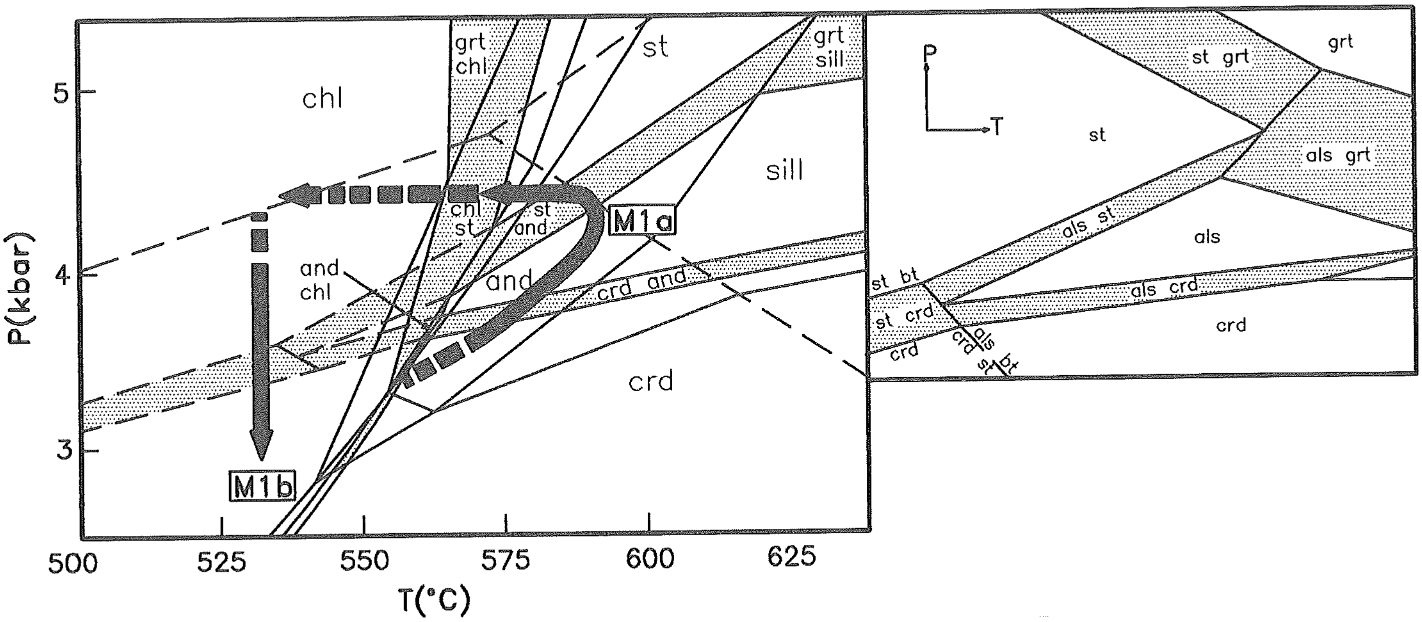


Figure 10: Calculated P-T pseudosection for X_{Fe} = 0.7 for the assemblage garnet + staurolite + aluminosilicate + cordierite + chlorite (excess biotite + muscovite + quartz + H_2O) (from Dymoke & Sandiford, 1992) showing P-T path inferred from reaction textures. Dashed lines indicate positions of staurolite, andalusite and cordierite equilibria at low T in H_2O-depleted system during cooling (see inset). The retrograde decompression path accompanied the development of the brittle fracture cleavage and pseudotachylite.

peak metamorphic conditions but while the crustal geothermal gradient was still sufficiently elevated (25-30 °Ckm^{-1}) that the rocks were still within the amphibolite facies.

SIGNIFICANCE OF THE P-T PATH

The metamorphism in the collar of the Vredefort Dome has traditionally been perceived as a localised effect developed only in the vicinity of the Dome. Several authors have gone so far as to propose that it was linked directly to the formation of the Dome. In this section the authors examine the implications that the new evidence presented above has for both the timing of the metamorphism and its origin. From this, an evaluation of the implications that this evidence has for the metamorphism observed in the Witwatersrand goldfields is undertaken.

Timing of the Metamorphic Event

As outlined earlier, geochronological studies have failed to establish unequivocally the timing relationship between the metamorphism observed in the Dome and the event responsible for the formation of the Dome, except to conclude that there is significant overlap in the results. Many authors have also concluded that field relationships are equivocal, with features such as those shown in Figure 4 suggesting that pseudotachylite in the Dome formed after the peak of metamorphism, while others, such as annealing of shock metamorphic features, suggest that significant metamorphic grades were achieved *after* the shock event (e.g. Schreyer, 1983; Stepto, 1990). The P-T path in Figure 10 indicates:

1. a clear relationship between the generation of pseudotachylite and exhumation of these rocks during doming;

2. that the peak of metamorphism (M1a) predated the doming event; and

3. that the doming event occurred while crustal geotherms were still elevated after the metamorphic peak.

The post-pseudotachylite M1b cordierite-biotite paragenesis developed in the Witwatersrand metapelites indicates that these rocks experienced temperatures of approximately 500-530 °C after doming. According to Martini (1992), the doming event occurred in crust experiencing a stable cratonic geotherm of 16 °C. However, the authors have found no evidence to suggest that the formation of the pseudotachylite and associated fracture cleavages was accomplished under zeolite facies conditions (Gibson et al., submitted). Thus, rather than M1b representing a separate, prograde metamorphic event, the M1b metamorphism is regarded as an artefact of high host-rock temperatures at the time of doming. This places tight constraints on the time that could have elapsed between the M1a metamorphic event and the formation of the Dome. It also implies a pre-doming geothermal gradient of > 30 °C/km. If this gradient is combined with Slawson's (1976) crust-on-edge model for the basement core of the Dome, this hypothesis can be tested by comparing the predicted temperature-vs-depth profile with the temperatures implied by the post-shock textures observed in the core of the Dome.

According to the crust-on-edge model, the rocks of the Steynskraal Metamorphic Zone must have been buried at depths of approximately 23-25 km immediately prior to doming. Extrapolation of the gradient obtained from the collar rocks predicts a pre-doming temperature of > 690 - 750 °C. This agrees with the estimates of Schreyer & Abraham (1978) who proposed that the post-pseudotachylite cordierite-orthopyroxene symplectites replacing garnet in samples from the SMZ formed at approximately 700 °C. Post-doming temperatures near the centre of the Dome (inferred pre-doming depths of > 25 km), as indicated by pyroxene exsolution textures (Schreyer et al., 1978) and fluid inclusion thermometry (Schreyer, 1983; Fricke et al., 1990), exceeded 800 °C. The differential annealing of shock textures documented by Fricke et al. (1990), Grieve et al. (1990) and Martini (1992) have provided qualitative confirmation of differential temperatures across the structure. It is thus concluded that, contrary to Martini (1992), a regional geothermal gradient exceeding 25-30 °C/km existed in the vicinity of the Witwatersrand Basin at the time of formation of the Dome. This gradient can explain the so-called Vredefort thermal event which resulted in the recrystallization of shock features and resetting of the remanent magnetism in the rocks of the Dome after they were upturned (Hart et al., 1995).

Modelling of thermal decay in crust subjected to elevated thermal gradients indicates that the high gradients are likely to be transient, with cooling to a stable geotherm typically occurring within 60 Ma of the metamorphic peak (McKenzie, 1978). Given that the age of the doming event has been reasonably well-fixed at approximately 2.025 Ga (Kamo et al., 1995; Spray et al., 1995), the information obtained from the P-T path in Figure 10 provided constraints on the maximum age limit for the M1a event. As shown in the following section, this is consistent with other evidence that indicates a likely syn-Bushveld Complex (2.065-2.054 Ga) timing for the metamorphism.

Origin of the M1a Metamorphism

The anticlockwise P-T path derived for the M1a metamorphism in the collar pelites indicates that anomalous heat flow conditions were achieved at mid-crustal levels during thickening of the upper crust. Similar P-T paths observed in low-P/high-T metamorphic belts have been explained in one of three ways:

1. tectonic thickening of previously thinned, hot crust (Vielzeuf & Kornprobst, 1984);

2. catastrophic thinning of the mantle during tectonic thickening of the crust (e.g., Sandiford & Powell, 1986; Loosveld, 1989); or

3. intrusion of voluminous high-level magmas during regional heating (Wells, 1980; Bohlen, 1987).

It is noteworthy that all three models propose regional-scale processes. A fourth possibility, that the metamorphism could have developed in a localised aureole beneath a sill immediately overlying these rocks, is not realistic given the lack of evidence of such a body in the exposures of these overlying levels in the surrounding collar of the Dome. A regional scenario is further supported by the evidence of (a) two syn-metamorphic fabrics in the rocks (Gibson, 1993) and (b) the post-doming metamorphic textures that indicate that a 30 °C/km geothermal gradient also existed over a large part of the lower crust now exposed in the core of the Dome.

There is no apparent evidence to support significant regional tectonic overthickening of the Kaapvaal Craton crust in the vicinity of the Vredefort Dome between 2.1 and 2.0 Ga. Lavas of the Roodekraal Complex (Fig. 2) that appear to have been extruded at approximately 2.2 Ga (Walraven, pers. comm., 1995) are still preserved in the vicinity of the Dome. The collar of the Dome also contains the full stratigraphic sequence known to have existed in the region, indicating that little, if any, erosion occurred between the deposition of the upper units of the Transvaal Supergroup and the formation of the Dome.

It is less clear whether thickening might have occurred after initial thinning and heating of the crust. While it could be argued that deposition of the Pretoria Group up until the Bushveld event might signify lower crustal thinning, and evidence exists of pre- to syn-Bushveld Complex compressional tectonics (e.g. Sharpe & Chadwick, 1982; Hartzer, 1995), no evidence has yet been found of major, ca. 2.1 Ga, extensional features in the region that could account for the extreme thinning required by this model.

The third model, involving magmatic thickening, meets not only the constraints of the P-T path, but also the geochronological constraints. Based on the latest data, the Bushveld Complex intruded the upper crust of the craton some 30-40 Ma before the formation of the Vredefort Dome. Although its thickness is variable, it is known to have extended at least as far as the Losberg Complex (Fig. 1) (Coetzee & Kruger, 1989) and it was accompanied by extensive sill intrusions throughout the region (Cawthorn et al., 1981). A combined thickness of 3-4 km of magmas emplaced in the upper crust (including sills emplaced below the Bushveld Complex, *sensu stricto*) could adequately account for the roughly 1 kbar of burial indicated by the P-T path in Figure 10a. Although it is impossible to establish directly the total thickness of Bushveld intrusions in the region of the Witwatersrand Basin, a combined thickness of 3-4 km for the mafic magmas, the precursor Rooiberg felsites and the Bushveld Granites appears to be reasonable given that the thickness of the Main Zone alone reaches 4 km along the present southern margin of the Complex some 150 km to the north of the Vredefort Dome (Walraven & Darracott, 1976).

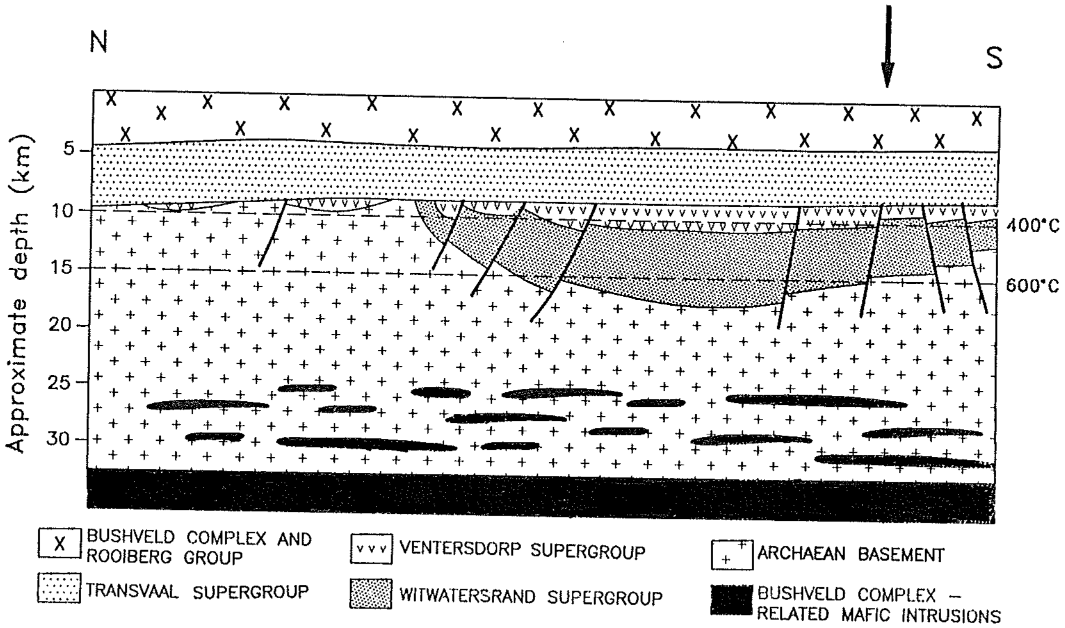
A MODEL FOR METAMORPHISM IN THE WITWATERSRAND BASIN

Figure 11a shows a schematic crustal cross-section through the Witwatersrand region immediately after intrusion of the Bushveld Complex. A thickness of 3-4 km for the magmas is deduced from the P-T path derived for the metapelites in the collar of the Vredefort Dome. A total thickness of 11-12 km for the Witwatersrand and Ventersdorp Supergroups and the Transvaal Supergroup in the vicinity of the Dome is in good agreement with observed stratigraphic thicknesses (P.L. Linton and P. Erikson, pers. comm., 1995). The predicted peak thermal array, shown schematically in Figure 11b, has been constructed assuming a relatively constant mid- to lower- crustal gradient of 40 °C/km. This high gradient was probably achieved largely through the intrusion of substantial volumes of mafic magma below and into the lower crust (Fig. 11a). Extrapolation of a 40 °C/km geothermal gradient into the lower crust implies that lower crustal levels could have been exposed to temperatures above their solidus. This would have resulted in significant melting of crustal material. The voluminous, hot, dry melts that formed the Rooiberg Felsites and Bushveld Granites are a likely consequence of these elevated temperatures.

Implications for the Witwatersrand Basin

A regional-scale, high geothermal gradient metamorphic event in the crust of the Kaapvaal Craton at the time of intrusion of the Bushveld Complex can explain the regional greenschist facies metamorphic assemblages documented by Phillips & Law (1994) throughout the Witwatersrand goldfields. It must be emphasized, however, that this model does not preclude earlier, pre-2.0 Ga metamorphic-hydrothermal events having affected the rocks of the Witwatersrand Basin. Events at 2.4-2.5 Ga (Frimmel, 1994) and/or approximately 2.3 Ga (Phillips et al., 1989), as suggested by geochronological studies, remain plausible, as does an event at ca. 2.2 Ga associated with the intrusion of the alkali

(a)



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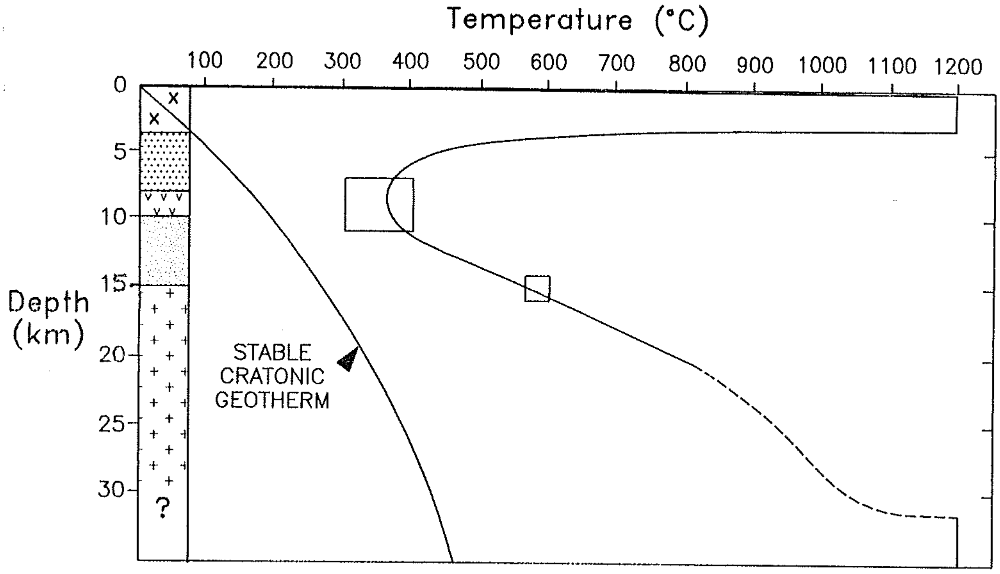

Figure 11: (a) Schematic crustal cross-section for the Kaapvaal craton in the vicinity of the Witwatersrand Basin after intrusion of the Bushveld Complex. The Bushveld Complex, including the Rooiberg Felsites, is assumed to have a simple, tabular shape. The 400 °C and 600 °C isotherms are shown, assuming a linear geothermal gradient of 40 °C/km. The future position of the Vredefort Dome is indicated by the arrow.
(b) Schematic peak metamorphic P-T array for crustal section in the vicinity of the arrow in Fig. 11(a), shortly after the intrusion of the Bushveld Complex. The array is constrained by the P-T estimates for the Witwatersrand goldfields (large box) and for the lower West Rand Group in the collar of the Vredefort Dome (small box) discussed in the text. An inverted array exists immediately below the Bushveld Complex, due to contact metamorphic effects.

granite plutons observed in the Basin. However, given the unusual circumstances necessary to achieve thermal gradients of 40 °C/km over large distances in the crust, the authors would disagree with Frimmel's (1994) statement that the Bushveld-related event represents only a retrogressive event following the main metamorphic event at 2.4-2.5 Ga. It appears more likely that any pre-Bushveld events represent more localised hydrothermal fluid events. This does not, however, diminish their efficacy as the agents for movement of gold within the Witwatersrand Basin.

CONCLUSIONS

1. The medium-grade metamorphism in the lower Witwatersrand Supergroup in the collar of the Vredefort Dome can be better-reconciled with a regional, crustal-scale event rather than a localised contact-metamorphic event, as has previously been proposed.

2. The P-T path followed by the rocks during metamorphism, and qualitative chronological relationships, establish that the metamorphism occurred during a regional magmatic event that also gave rise to the Bushveld Complex.

3. This event provides the most plausible explanation for the regional greenschist facies metamorphism observed in the Witwatersrand goldfields.

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