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REGIONAL METAMORPHISM DUE TO ANOROGENIC INTRACRATONIC MAGMATISM

R.L. GIBSON and G. STEVENS

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by

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ABSTRACT

The Vredefort Dome, the central uplifted portion of the ca. 300 km diameter, 2.02 Ga Vredefort Impact Structure, exposes an ~20km deep profile through the crust of the Kaapvaal Craton of southern Africa which displays evidence of greenschist- to granulite-facies low-pressure metamorphism. Reaction textures in the medium- to high- grade metamorphic rocks indicate an anticlockwise IBC P-T evolution, and peak metamorphic temperatures consistent with a mid- to upper- crustal geotherm of 40-50° C/km. The metamorphism is attributed to craton-wide intraplating of mantle-derived magmas into the deep crust during the 2.05-2.06 Ga Bushveld Event. Felsic volcanic rocks and high-level intrusions in the Bushveld Complex represent anatectic derivatives from the deeper, high-grade parts of the metamorphic terrane. The intracratonic setting, the absence of evidence of concomitant orogenic deformation, and the preservation of the pre-existing diamondiferous lithospheric root beneath the craton, suggest that the thermal event was triggered by the transient positioning of the Kaapvaal Craton over a mantle plume.

Mantle-derived magmas are widely regarded as playing a major role in advecting heat into the crust in order to effect regional high-temperature low-pressure metamorphism (e.g., Wells 1980; Harley 1989; De Yoreo et al. 1991). In most cases of low-P metamorphism documented in the literature, the root is inferred to be a disturbance in the structure of the lithosphere, and immediately underlying asthenosphere, brought about by plate tectonic processes such as subduction (Bohlen 1987), tectonic thinning of the lithosphere (Sandiford & Powell 1986) or removal of the mantle lithosphere from the crust due to tectonic overthickening (Loosveld 1989; Sandiford & Powell 1991). In this paper we document an early Proterozoic regional, greenschist- to granulite-facies, low-P metamorphic terrane in the Kaapvaal Craton (South Africa) which occurs in an intracratonic setting, spatially and temporally unrelated to orogenic activity. We attribute this event to intraplating of mafic-ultramafic magmas generated by mantle plume activity. The preservation of some of the magmatic rocks and voluminous felsic extrusive and intrusive rocks in proximity to the metamorphic rocks provides a rare example where the link between mantle magmatism, regional low-P metamorphism and crustal anatexis can be studied.

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REGIONAL SETTING

The Kaapvaal Craton of southern Africa (Fig. 1) is one of the oldest known fragments of Archaean continental crust, comprising a nucleus of subsidiary arc-like oceanic terranes (greenstones) and tonalite-trondjhemite-granodiorite basement ranging in age from 3.7 to 3.1 Ga (De Wit et al. 1992). Following consolidation of the cratonic nucleus at 3.1 Ga, four major unconformity-bounded supracrustal sequences (the Dominion Group and the Witwatersrand, Ventersdorp and Transvaal Supergroups) were deposited on the craton in Late Archaean to Early Proterozoic times (Table 1). Between 2.7-2.65 Ga, the craton was docked with the Zimbabwe Craton to the north along the Limpopo Belt (Fig. 1).

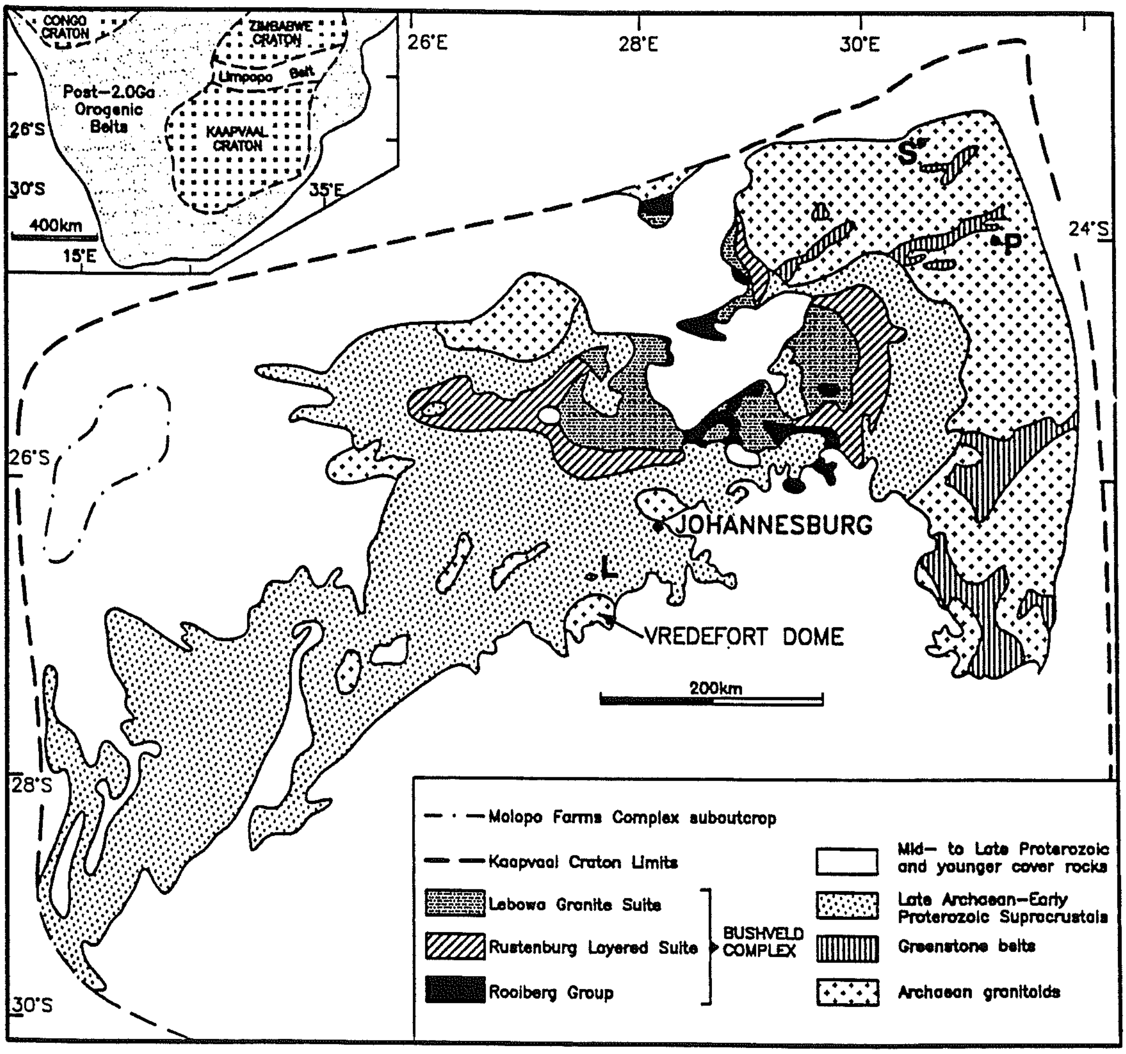

Figure 1: Regional geology of the Kaapvaal Craton showing the locations of the Bushveld Complex and related intrusions (S: Schiel Alkaline Complex; P: Phalaborwa Alkaline Complex; L: Losberg Complex), and the Vredefort Dome. The sub-outcrop limits of the Molopo Farms Complex, which is not exposed at surface, are shown in the west of the craton.

Following the formation of the Archaean to Proterozoic basins, the Kaapvaal Craton experienced a major magmatic episode, the Bushveld Event, at ca. 2.05-2.06 Ga (Table 1). The most voluminous component of this magmatism comprises ultramafic to mafic intrusive rocks in the Bushveld (Rustenburg Layered Suite) and Molopo Farms Complexes (Fig. 1) which are located within the upper levels of the Transvaal Supergroup, as well as abundant sills in the immediately underlying supracrustal succession. Geochronological data (Table 2) indicate an age of 2061 $\pm$ 27 Ma for the Rustenburg Layered Suite (Walraven et al. 1990). Coetzee & Kruger (1989) obtained a similar age (2041 $\pm$ 41 Ma) for a small outlier of mafic

Table 1: Generalized Archaean to Early Proterozoic evolutionary scheme for the Kaapvaal Craton (after Stanistreet & McCarthy 1991; De Wit et al. 1992; Robb & Meyer 1995; Cheney 1996)

Age (Ga)	Event	Interpretation
2.02	Vredefort	large meteorite impact
2.05-2.06	Bushveld	voluminous mafic-ultramafic intrusions, felsic volcanics and intrusions
2.2-1.8	Ubendian	SE-directed subduction beneath Kaapvaal Craton along NW margin, culminating in orogenesis at 1.8 Ga
2.43-2.22	Upper Transvaal	~3-4 km arenaceous-argillaceous foreland basin sedimentation
2.58-2.43	Lower Transvaal	~1-2 km dolomite epicontinental shelf sedimentation
2.70-2.65	Limpopo	Alpine convergence and amalgamation of Kaapvaal and Zimbabwe Cratons
2.71	Ventersdorp	~4 km tholeiitic flood basalts and subsidiary rift sedimentation
2.84-2.71	Upper Witwatersrand	~3 km arenaceous-rudaceous fluvio-deltaic foreland basin sedimentation
2.98-2.91	Lower Witwatersrand	~4 km argillaceous-arenaceous subtidal passive margin sedimentation
3.09-3.07	Dominion	~2.5 km bimodal volcanics and subsidiary rift sedimentation
3.7-3.1	Archaean craton	mafic-ultramafic volcanic arcs and sediments + tonalite-trondjhemite-granodiorite-granite intrusions

and ultramafic rocks located 100 km south of the main outcrops of the Bushveld Complex, the Losberg Complex (Fig. 1).

The Rustenburg Layered Suite is intruded, and overlain by rocks of the Lebowa Granite Suite (Fig. 1), dated at 2054 $\pm$ 2 Ma (Walraven & Hattingh 1993). Both Suites intrude a sequence of heterogeneous, predominantly felsic, volcanics and minor intercalated sedimentary rocks (Rooiberg Group) and granophyre sills (Rashoop Granophyre Suite) (Fig. 1) which were originally regarded as being over 100 Ma older than the Bushveld intrusions (Coertze et al. 1978). However, recent field and geochemical studies (Hatton & Schweitzer 1995; Schweitzer et al. 1995) suggest that the eruption of the Rooiberg Group was penecontemporaneous with the intrusion of the Rustenburg Layered Suite. This is supported by a U-Pb zircon age of 2061 $\pm$ 2 Ma for a Rooiberg Group lava (Table 2; F. Walraven, unpublished data).

In addition to the mafic-ultramafic and felsic magmatism in the central and western portions of the craton, two alkaline complexes of similar age (Table 2), Phalaborwa (2060.6 $\pm$ 0.5 Ma; Reischmann 1995) and Schiel (2059 $\pm$ 35 Ma; Walraven et al. 1992), intrude the Archaean basement in the northeast of the craton (Fig. 1).

Table 2: Detailed geochronology of the Bushveld Complex

Unit	Age (Ma)	Technique	Source
Lebowa Granite Suite	2054 ± 2	single-zircon Pb-evaporation	Walraven & Hattingh (1993)
Rustenburg Layered Suite	2061 ± 27	Rb-Sr whole-rock	Walraven et al. (1990)
Rashoop Granophyre Suite	2060 ± 2	single-zircon Pb-evaporation	F. Walraven (unpubl.)
Rooiberg Group	2061 ± 2	single-zircon Pb-evaporation	F. Walraven (unpubl.)

The Bushveld Event was closely followed by the formation of the Vredefort Dome in the south-central parts of the craton (Fig. 1) at ca. 2.02 Ga (Table 1). Apart from limited tectonic reactivation of cratonic lineaments, and localised magmatic activity and intracratonic basin sedimentation in some parts of the craton, most subsequent tectonomagmatic activity in the region was confined to the craton boundaries where lateral accretion of material onto the craton occurred towards the end of the early Proterozoic Ubendian Event (ca. 1.8 Ga; Master 1991) and during the 1.2-1.0 Ga Namaqua, and ca. 500 Ma Pan-African Events (Thomas et al. 1994). As a result, much of the Archaean and Early Proterozoic structure of the craton has been preserved largely intact.

GENERAL GEOLOGY OF THE VREDEFORT DOME

The Vredefort Dome is an 80-km-wide structure located 120 km southwest of Johannesburg, partially covered by Phanerozoic sediments and dolerites of the Karoo Supergroup (Figs. 1 & 2). Its core is comprised of heterogeneous pre-3.1 Ga granitoid gneisses and subsidiary metamorphosed supracrustal rocks and mafic intrusions. Stepto (1990) divided this core into an inner domain, dominated by leucogneisses (Inlandsee Leucogranofels Terrane), and an outer domain dominated by heterogeneous granodioritic, trondjhemitic and granitic gneisses and granites (Outer Granite Gneiss Terrane) (Fig. 2). The core is surrounded by a collar of generally subvertical to overturned supracrustal rocks of the post-3.1 Ga late Archaean to early Proterozoic sedimentary-volcanic sequences of the Dominion Group and the Witwatersrand, Ventersdorp and Transvaal Supergroups (Fig. 2). The steep dip of the supracrustal strata in the collar shallows outward into a rim synclinorium that surrounds the dome. The stratigraphically highest unit exposed in the core of this syncline is the mafic to ultramafic Losberg Complex (Fig. 1). Other intrusive rocks in the supracrustal succession include Bushveld-related mafic sills in the Transvaal Supergroup beneath the Losberg Complex, Ventersdorp-age mafic sills in the Witwatersrand Supergroup (Pybus et al. 1995), and several small peralkaline and dioritic complexes (Fig. 2). Walraven & Elsenbroek (1991) estimated a ca. 2.2 Ga age for these complexes, but recently Moser (1997) obtained a U/Pb single-zircon age of 2078 $\pm$ 12 Ma.

A variety of unusual features in the rocks of the Vredefort Dome (shatter cones, planar microdeformation lamellae in quartz, coesite and stishovite) indicates that they have been exposed to shock pressures in excess of 20 kbar, consistent with an origin by meteoritic impact (see review in Reimold & Gibson 1996). The Vredefort Dome is thus interpreted as the rebound-induced central uplift of a large, ca. 250-300 km diameter impact structure that

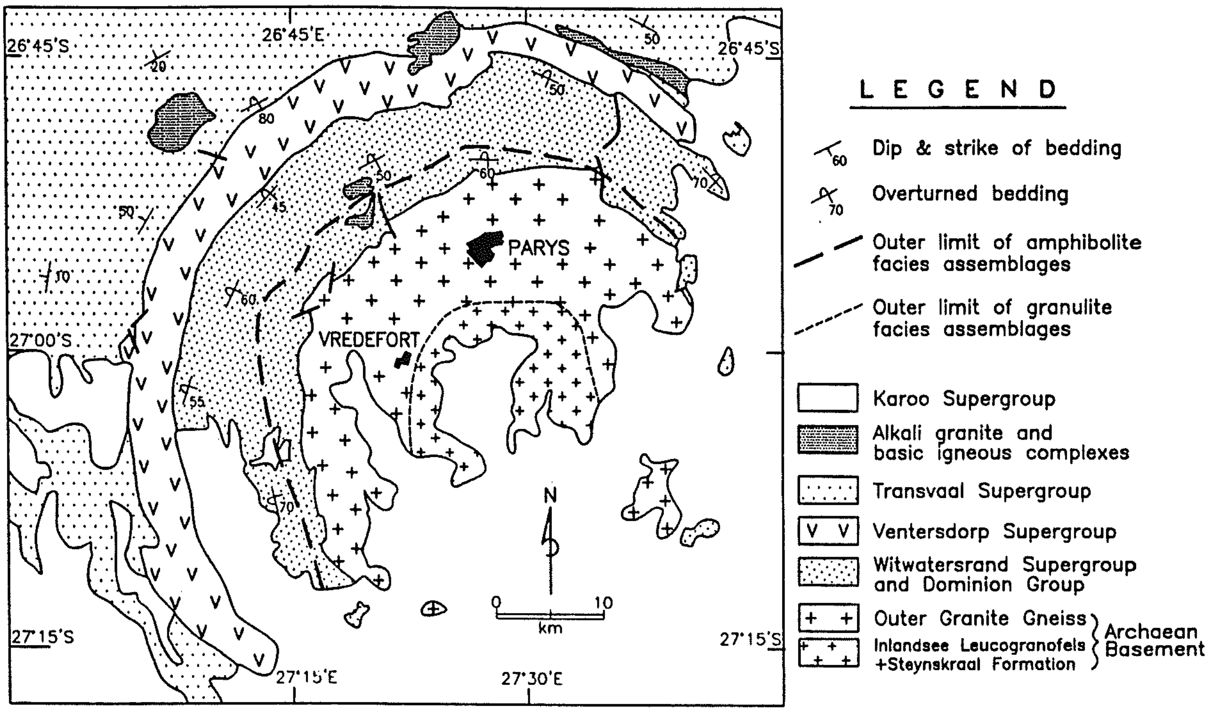

Figure 2: Simplified geological map of the Vredefort Dome showing the distribution of the amphibolite- and granulite-facies metamorphic zones.

encompasses most of the Witwatersrand Basin (Henkel & Reimold 1995; Reimold & Gibson 1996, and references therein). One of the effects of the impact-related deformation was the generation of abundant, extremely voluminous pseudotachylitic breccias (Reimold & Colliston 1994; Gibson et al. 1997b) similar to those observed in the Sudbury impact structure in Canada (Thompson & Spray 1996). U-Pb dating of shock-metamorphosed zircons from the basement rocks, and zircons from these pseudotachylitic breccias, indicates an age of 2023 ± 4 Ma for the impact event (Kamo et al. 1996), corroborating earlier Ar-Ar results of ca. 2.02 Ga from other pseudotachylitic and impact breccias (Walraven et al. 1990; Allsopp et al. 1991; Trieloff et al. 1994; Spray et al. 1995).

The geometry of the Dome indicates that it exposes a cross-section through a large part of the crust of the Kaapvaal Craton. Based on geochemical studies in the core of the Dome, Slawson (1976) proposed that the exposures represent an almost complete crustal section, a view shared by Hart et al. (1990) who suggested that ultramafic rocks situated close to the surface in the centre of the Dome represented upper mantle. Recent geophysical modelling of the impact structure (Henkel & Reimold 1995) and quantitative thermobarometry (see below) suggest, however, that only upper- and mid-crustal levels are exposed. In the remainder of this paper we describe the metamorphic features observed in these rocks and relate these to a model for their formation.

GREENSCHIST- TO GRANULITE-FACIES ROCKS EXPOSED IN THE VREDEFORT DOME

In the goldfields at the margins of the Witwatersrand Basin, which surrounds the Vredefort Dome, the regional grade of metamorphism has been established as lower greenschist facies (350 ± 50 °C, 2-3 kbar; Phillips & Law 1994). In the Dome itself, metamorphic facies are distributed broadly concentrically, with grade increasing radially inwards (Fig. 2). The peak metamorphic assemblages are affected by the shock deformation phenomena and pseudotachylitic breccias (Gibson & Wallmach 1995; Gibson et al. 1997a; Stevens et al. 1997a), indicating that this metamorphism predated the shock event (Fig.3a,b).

Greenschist-facies Zone

The greenschist-facies zone in the outer collar of the Dome is located in dolomites (lower Transvaal Supergroup), mafic lavas and sills (Ventersdorp Supergroup) and orthoquartzites with minor pelites (upper Witwatersrand Supergroup). The pelites contain chloritoid and aluminosilicate minerals (Nel 1927; Bisschoff 1982) and Bt + Chl + white mica (mineral abbreviations after Kretz 1983), indicative of lower- to mid-greenschist-facies conditions. This is confirmed by assemblages in the mafic lavas of the Ventersdorp Supergroup and the Ventersdorp-age mafic sills within the upper Witwatersrand Supergroup (Chl + Act + Pl (relic) + Ep; Bisschoff 1982).

Amphibolite-facies Zone

The amphibolite facies zone straddles the boundaıy between the supracrustal rocks in the collar of the Dome and the Archaean basement in the core (Fig. 2). In the collar, this zone includes pelites, orthoquartzites and banded iron formations of the lower Witwatersrand Supergroup, the mafic lavas of the Dominion Group and intercalated Ventersdorp mafic sills. The boundary between the greenschist- and amphibolite-facies zones is defined by the appearance of fibrous hornblende, with clinozoisite partially replacing calcic plagioclase in the mafic sills (hornblende-zoisite zone; Bisschoff 1982). In the lowermost parts of the Witwatersrand Supergroup, the sills contain a Hbl + Pl (andesine) + Qtz + Spn ± Zo paragenesis (hornblende-andesine zone; Bisschoff 1982), indicative of mid-amphibolite-facies metamorphic grades. A similar assemblage occurs in the mafic Dominion Group lavas (Jackson 1994). The most informative assemblages are those in the metapelites, which involve biotite, chlorite, muscovite, cordierite, andalusite, garnet and/or staurolite, and which occur in various combinations depending on bulk rock composition (Bisschoff 1982; Gibson & Wallmach 1995). The rocks contain two poorly-defined, syn-metamorphic foliations orientated oblique to bedding (Gibson 1993), but no evidence has been found of related outcrop- or larger- scale folding and the preservation of delicate sedimentary structures indicates only low ductile strain in these rocks (Fig. 3a). Kyanite and fibrolitic sillimanite occur as rare phases, always in association with andalusite. Minerals show no evidence of compositional zoning (Gibson & Wallmach 1995).

In the Archaean basement in the outer parts of the core of the Dome (Outer Granite Gneiss, Stepto 1990; Fig. 2), amphibolitic xenoliths contain Hbl + Bt + Pl + Qtz assemblages, consistent with upper amphibolite facies conditions. There are no other rock types containing diagnostic assemblages.

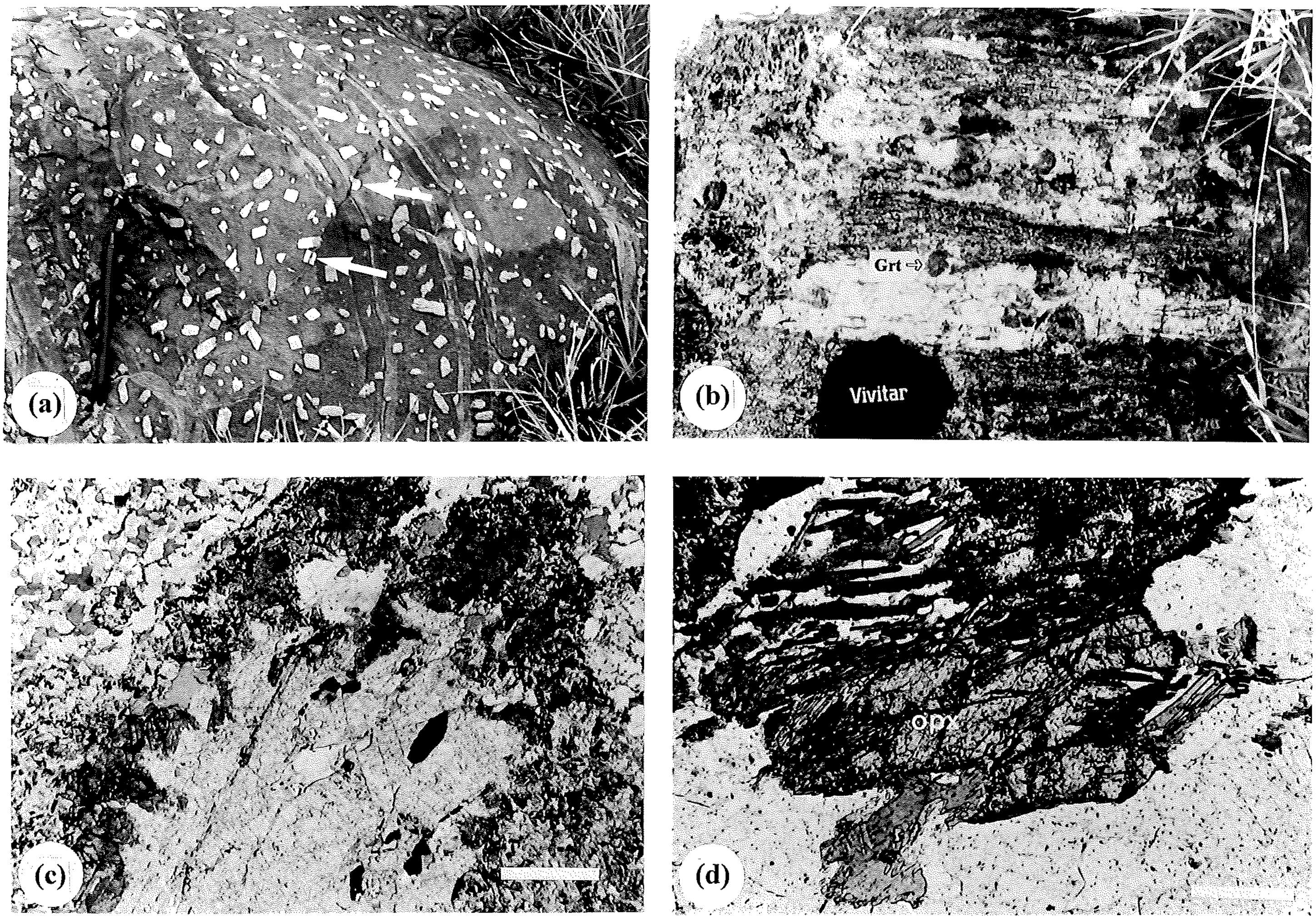

Figure 3: Metamorphic features of metapelites from the Vredefort Dome.
(a) Andalusite metapelite from the lower Witwatersrand Supergroup (amphibolite-facies zone). Note the well-preserved sedimentary bedding defined by the light grey quartz-rich layers and the thin pseudotachylite veinlet cutting andalusite porphyroblasts (arrowed).
(b) Pelitic stromatic migmatite from the granulite facies zone showing large garnet crystals produced by fluid-absent biotite melting.
(c) Staurolite aggregates replacing andalusite porphyroblast (centre); lower Witwatersrand Supergroup, amphibolite facies zone. Scale bar 500 mm.
(d) Retrogressive replacement of peak metamorphic orthopyroxene (opx) by vermicular biotite-quartz intergrowths related to leucosome recrystallization in granulite facies migmatite. Scale bar 150 mm.

Granulite-facies Zone

The amphibolite- to granulite-facies transition coincides with the gradational transition between the Outer Granite Gneiss and the Inlandsee Leucogranofels terranes (Fig. 2). Outcrop within the granulite-facies zone is poor and field relationships are complicated by extensive structural disruption caused by the impact event. Within the leucogneiss are found xenoliths of a dismembered supracrustal succession comprising pelitic, iron formation and mafic granulites. Peak metamorphic assemblages are coarse-grained and texturally well equilibrated and, in the pelites, are associated with anatectic features (Fig. 3b). The peak assemblage in the iron formations is Grt + Opx + Qtz + Mag ± Cpx whereas the mafic granulite assemblages comprise Cpx + Hbl + Pl + Mag ± Opx (Stepto 1979). In the metapelites, coarse-grained garnet, cordierite and orthopyroxene crystals occur in association with stromatic leucosomes (Fig. 3b) or in anhydrous, alumonous restitic assemblages. The garnet commonly contains inclusions of relic sillimanite and biotite and, more rarely, hercynitic spinel. The evidence points to the attainment of conditions consistent with incongruent biotite melting such as: Bt + Sil + Qtz + Pl = Grt + Crd ± Opx ± Kfs + M. The restitic assemblages e.g., Grt + Qtz; Grt + Opx; Crd + Kfs + Rt, indicate that temperatures in some parts of the terrane exceeded the upper limit of biotite stability.

Peak P-T conditions, P-T Paths and Geothermal Gradients

The calculation of peak P-T conditions in the amphibolite-facies metapelites in the collar of the Dome is hampered by a pervasive retrograde event that postdates the impact shock event. This led to overgrowth of the peak metamorphic assemblages by a secondary paragenesis, and to chemical re-equilibration of garnet and biotite (Gibson & Wallmach 1995; Gibson et al. 1997b). However, two critical prograde to peak reaction textures are preserved, and allow constraints to be placed on the P-T conditions and P-T path:

1. in magnesian bulk rock compositions (XMg ~ 0.5-0.6), andalusite occurs as a rim around cordierite porphyroblasts, suggesting the reaction: Crd + Ms = And + Bt + Qtz; and

2. in intermediate XMg bulk compositions (XMg ~ 0.3-0.45), staurolite replaces andalusite by the reaction: And + Bt = St + Ms + Qtz (Fig. 3c). Staurolite also pseudomorphs cordierite porphyroblasts in these rocks.

Both of the above reactions have shallow positive dP/dT slopes, with the products occurring on the high-P side (Holland & Powell 1990) (Fig. 4). The textures thus indicate an anticlockwise P-T path for the rocks and peak P-T conditions of 570-600 °C at 4.0-4.5 kbar (Fig. 4). Similar P-T conditions were estimated by Jackson (1994), from assemblages in the mafic Dominion Group lavas. A broadly isobaric cooling path (IBC) is inferred for these rocks as estimates of the depth of overburden at the time of the meteorite impact (estimated to be 14 km from reconstructions using shatter cone orientations; Manton 1965) are similar to the maximum burial depths of the rocks during the metamorphic peak (Fig. 4).

In the metapelitic granulites, the restitic, biotite-free nature of some of the assemblages indicates that they experienced peak temperatures in excess of the upper limit

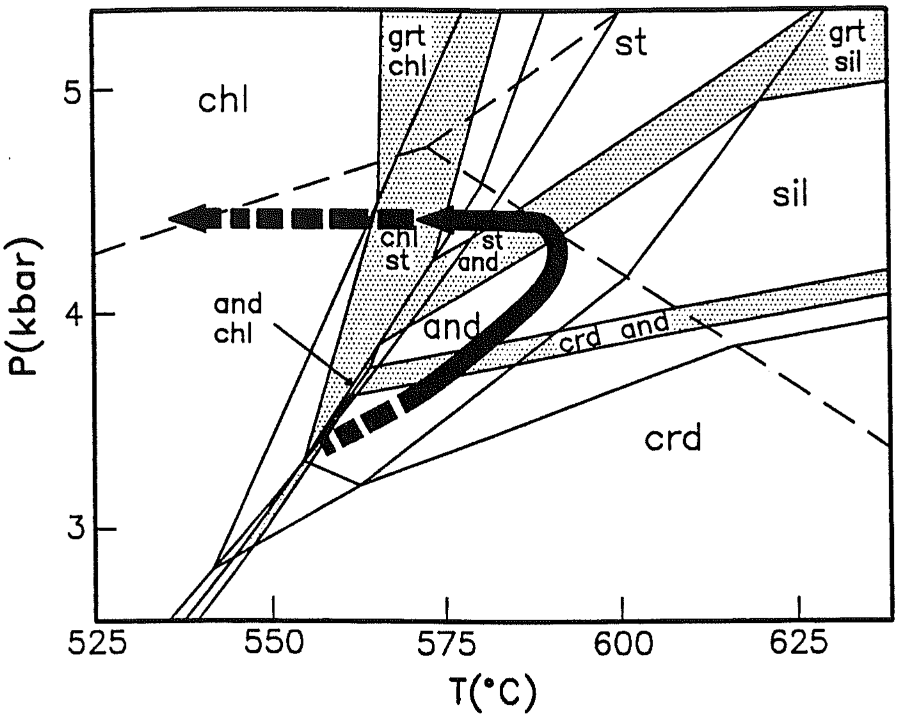

Figure 4: P-T path for amphibolite facies metapelites from the lower Witwatersrand Supergroup in the Vredefort Dome, based on reaction textures. P-T pseudosection for X_{Mg} = 0.3 for the assemblage Grt + St + Als + Crd + Chl + Bt (excess Ms + Qtz + H_2O) from Dymoke and Sandiford (1992).

of biotite stability. Experimental studies suggest that this limit is reached at ~920 °C in pelites containing biotite with normal Ti and F contents (Vielzeuf & Hollaway 1988; Stevens et al. 1997b). The presence of peak metamorphic biotite in some of the migmatites indicates that not all of the granulites experienced these temperatures. However, minimum temperatures of ~850 °C are necessary to produce significant melt proportions in pelitic protoliths by biotite breakdown (Stevens et al. 1997b). Schreyer (1983) estimated similar peak temperatures of > 875 °C for the mafic granulite assemblages and Schreyer et al. (1978) estimated peak temperatures of > 800 °C in the iron formation granulites, based on clinoeulite-ferroaugite exsolution textures after ferropigeonite.

Mineral compositional data for the pressure estimates (Table 3) were obtained from the JEOL JSM 6400 SEM in the Geology Department at the University of Manchester (for operating conditions, see Stevens et al. 1997a). Using the program THERMOCALC (Powell & Holland 1988) with the updated dataset of Holland & Powell (1990), the assemblage Grt-Crd-Opx-Qtz in samples VT596 and VGS3 yields peak metamorphic pressures of 5.1 ± 1.6 kbar and 4.8 ± 1.3 kbar respectively at 900 °C. The Grt + Sil = Spl + Qtz geobarometer of Nichols et al. (1992) yielded similar pressures of 4.9 ± 0.9 kbar at 800 °C and 6.3 ± 0.8 kbar at 900 °C respectively, for sample VT600.

The near-peak prograde P-T path for the granulites can be constrained from the textural relations between spinel and garnet. Spinel occurs only as inclusions within garnet, indicating that the reaction Grt + Sil = Spl + Qtz was crossed from the spinel to the garnet side. As this reaction has a positive dP/dT slope, and as the spinel occurs as inclusions in

Table 3: Mineral compositional data for geobarometry

	Garnet			Cordierite		Orthopyroxene		Spinel	
	VT596	VGS3	VT600	VT596	VGS3	VT596	VGS3	VT600	VT600
SiO_2	37.82	37.57	38.88	49.81	50.41	50.77	51.12	-	-
Al_2O_3	20.68	20.39	20.64	34.12	33.97	3.12	3.11	56.99	57.63
Cr_2O_3	0.41	0.32	0.41	-	-	-	-	2.31	2.37
Fe_2O_3	1.90	1.51	2.07	-	-	-	-	1.96	1.55
FeO	32.77	31.98	28.41	4.48	4.33	28.75	28.53	24.62	25.33
V_2O_5	-	-	-	-	-	-	-	1.42	0.82
MnO	0.93	1.04	0.84	-	-	-	-	0.11	0.11
MgO	5.72	5.66	8.62	10.84	10.94	16.92	17.24	7.99	7.75
ZnO	-	-	-	-	-	-	-	4.61	4.24
CaO	1.05	1.53	1.33	-	-	0.16	-	-	-
Na_2O	-	-	-	0.31	0.36	-	-	-	-
Total	101.28	100.00	101.19	99.56	99.38	99.72	100.00	100.01	99.80
Si	5.94	5.97	6.00	4.98	3.98	1.96	1.95	-	-
Al	3.83	3.82	3.74	4.01	3.98	0.15	0.14	15.01	15.17
Cr	0.05	0.04	0.05	-	-	-	-	0.41	0.42
Fe^{3+}	0.22	0.18	0.24	-	-	-	-	4.60	4.73
Fe^{2+}	4.31	4.25	3.66	0.37	0.36	0.93	0.91	0.33	0.26
V	-	-	-	-	-	-	-	0.21	0.12
Mn	0.12	0.14	0.11	-	-	-	-	0.02	0.02
Mg	1.35	1.34	1.98	1.62	1.62	0.97	0.98	2.66	2.58
Zn	-	-	-	-	-	-	-	0.76	0.70
Ca	0.18	0.26	0.22	-	-	0.01	-	-	-
Na	-	-	-	0.06	0.07	-	-	-	-
Total	16.00	16.00	16.00	11.04	11.04	4.02	3.98	24.00	24.00
O	24.00	24.00	24.00	18.00	18.00	6.00	6.00	32.00	32.00
Mg#	23.85	23.97	35.11	81.41	81.81	51.05	51.85	36.64	35.29

the peak-metamorphic garnet (indicating that the reaction occurred prior to or during the attainment of peak metamorphic temperatures), the rocks must have evolved along an anticlockwise P-T path (Fig. 5).

Post-peak, retrograde cooling in the migmatitic metapelites involved partial rehydration of cordierite from the peak assemblage to produce Bt + Sil + Qtz, and orthopyroxene to produce Bt + Qtz (Fig. 3d). The absence of similar retrograde effects in the melt-depleted restitic granulites suggests that this reaction was related to crystallization of the leucosomes. No evidence has been found of concurrent garnet breakdown through pressure-sensitive equilibria, suggesting that cooling was approximately isobaric.

Both the peak and the retrograde granulite assemblages are overprinted by a post-impact, lower-P, lower-granulite-facies event which produced fine-grained symplectic partial replacement textures involving Opx + Crd ± Spl (Stevens et al. 1997a).

The peak P-T estimates obtained from the amphibolite- and granulite-facies metapelitic assemblages indicate that the synmetamorphic geothermal gradient in the upper and middle crust must have reached values of 40-50 °C/km, at least locally. This is similar to the estimate for the greenschist-facies metamorphism observed in the Witwatersrand goldfields (Phillips & Law 1994) and implies extremely high heat flux through the craton at this time.

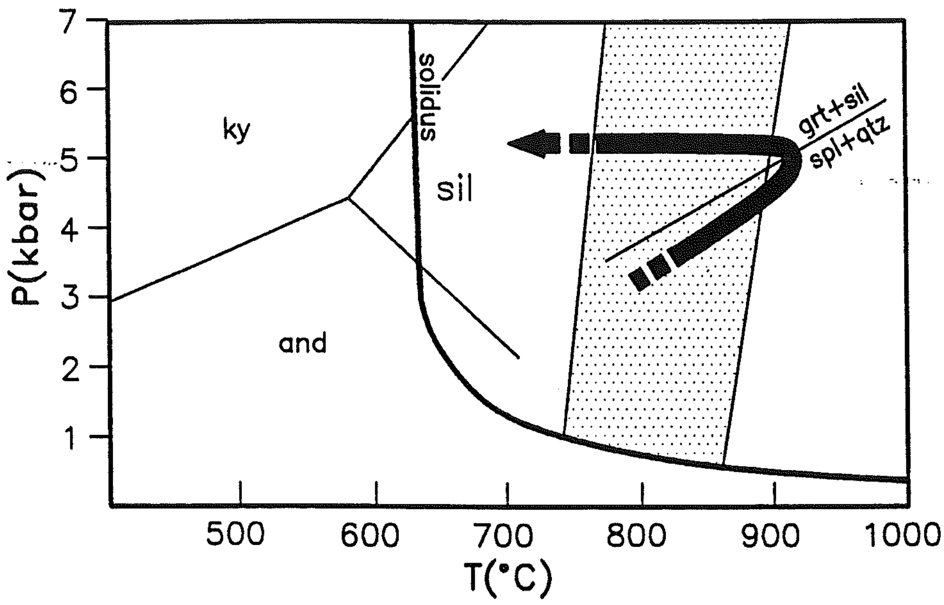

Figure 5: P-T path for metapelitic granulites in the Vredefort Dome. Stippled field represents the experimentally-determined field of biotite-melt co-existence (after Stevens et al. 1997b). Slope of Grt + Qtz + Spl + Sil reaction from Nichols et al. (1992).

DISCUSSION

Timing of the Metamorphism

The textural evidence from the metapelites in the Vredefort Dome (Fig. 3a) indicates that the metamorphic peak occurred prior to the impact event. The available geochronological data from the amphibolite- and granulite-facies rocks, however, suggests a temporal overlap between the metamorphism and the impact event:

1. Ar-Ar mineral (Allsopp et al. 1991; Gibson, unpublished data) and Rb-Sr whole-rock data (Hart et al. 1981; Walraven et al. 1990; Allsopp et al. 1991) display a spread between ~2070 Ma and ~1950 Ma, with the main cluster between ~2060 Ma and ~2010 Ma (Reimold et al. 1995); and

2. a SHRIMP U-Pb zircon age of 2017 ± 5 Ma obtained from a small leucogranite body in the granulite facies gneisses (Gibson et al. 1997b) is indistinguishable from the age obtained for the impact event (2023 ± 4 Ma; Kamo et al. 1996).

Additional, circumstantial evidence supporting a close temporal relationship between the metamorphism and the impact event is provided by textural studies of the impact-related shock deformation features (e.g. Grieve et al. 1990; Martini 1992) and quantitative and semi-quantitative geothermometry from metamorphic assemblages that postdate the impact event (Schreyer 1983; Fricke et al. 1990; Gibson & Wallmach 1995; Gibson et al. 1997a, b; Stevens et al. 1997a). These indicate unusually high post-impact temperatures in the rocks of the Dome, compared with the rocks exposed in the comparably-sized Sudbury impact structure in Canada (e.g. Dressler 1984). Such elevated post-impact temperatures are best explained if the crustal geotherm at the time of impact exceeded that expected in stable cratonic crust. Gibson et al. (1997a) estimated a minimum value for the pre-impact geotherm in the Vredefort region of ~25 °C/km, in contrast to a value of 15-16 °C/km predicted from heat-flow studies (Grieve et al. 1990; Jones (1988) in Martini 1992). Such an elevated pre-impact geotherm suggests the occurrence of a transient thermal event on the craton shortly before the 2.02 Ga impact.

Further evidence is provided by the persistence, until after the impact event, of anatectic melts derived during the metamorphic peak in the granulites (Gibson et al. 1997b; Stevens et al. 1997a). This indicates that parts of the granulite-facies terrane were still at temperatures above the granite solidus at the time of impact.

From this, we conclude that the metamorphic and impact events must have occurred within a period of no more than a few tens of millions of years. As the Bushveld Complex predates the impact by only some 30 Ma (2.05-2.06 Ga; Table 1), the most plausible explanation is that the metamorphism was linked to the Bushveld magmatic event.

Bushveld Magmatothermal Event

Given the temporal link between the greenschist- to granulite-facies metamorphism and the Bushveld magmatism indicated by the geochronological studies, the anticlockwise P-T paths documented for the medium- and high- grade rocks are inferred to reflect magmatic thickening of the upper crust due to shallow-level intrusions and/or volcanism (the lateral equivalents of the Bushveld Complex and the underlying mafic sills). The thermobarometric data for the rocks in the Vredefort Dome indicate, however, that the heat source for the mid-crustal metamorphism was located at depth. Thus, the Bushveld Event must have also included voluminous lower- to sub- crustal mantle-derived intrusions.

Mid- to lower-crustal harzburgitic intrusions which display a chemical affinity with the Rustenburg Layered Suite have been described from borehole core from the central granulite terrane in the Vredefort Dome (Merkle & Wallmach 1998). In addition, deep-crustal reflection seismic data from the vicinity of the Vredefort Dome indicate several major, laterally-extensive, subhorizontal lower crustal reflectors that could represent mafic to ultramafic sills (Durrheim et al. 1991). The main evidence in support of voluminous lower crustal intraplating of mantle magmas during the Bushveld Event comes, however, from bulk and mineral chemical studies of the Rustenburg Layered Suite. Cawthorn & Davies (1983) and Kruger (1994) noted that the parental magmas for the Rustenburg Layered Suite are enriched in Si, K, Rb and $^{87}Sr/^{86}Sr$, indicating contamination by a siliceous source. The mineral chemistry and crystallization sequence observed in the Rustenburg Layered Suite also suggests that the parental magmas underwent varying degrees of crystal fractionation in deeper-level magma chambers prior to their final emplacement (Cawthorn & Walraven 1997). The remarkably uniform chemistry of individual magma pulses across the Bushveld Complex indicates that the magmas spent sufficient time at these deeper levels to achieve thorough mixing with the crustal contaminants.

Cawthorn & Walraven (1997) conservatively estimated the volume of mafic to ultramafic rock preserved within the present outcrop limits of the Bushveld Complex to be

384 000 km^3. This figure does not include eroded material, the Molopo Farms Complex (Fig. 1) which has an estimated volume of 200 000 km3 (Reichardt 1994), or the sills in the underlying Transvaal Supergroup which attain a cumulative thickness of 2.5 km in places (Sharpe 1984). Modelling of the crystallization history of the Rustenburg Layered Suite also indicates that a volume of magma at least equivalent to that comprising the presently-preserved Rustenburg Layered Suite must have escaped from the Bushveld Complex magma chamber before it could crystallize (Cawthorn & Walraven 1997). In addition to this, the crystallized residuum left in the deeper-level magma chambers must be included. Thus, the evidence suggests a cumulative volume of mantle-derived magma in excess of 1 to 1.5 x 10^6 km^3. This volume is comparable to the volumes of mafic volcanic rocks found in major flood basalt provinces such as the Deccan and North Atlantic Tertiary Province, although the estimates for the latter do not include the intrusive rocks inferred to underlie these provinces (White & McKenzie 1995). The most recent geochronological data available for the Bushveld Complex (Table 2), although not as precise as those obtained for some flood basalt provinces, suggest that rates of magma production during the Bushveld Event were of the same order of magnitude (i.e. km^3/yr) (White & McKenzie 1995; Cawthorn & Walraven 1997).

Experimental studies by Cawthorn & Biggar (1993) suggest that the parental magmas of the Rustenburg Layered Suite intruded at temperatures in excess of 1300 °C. Given that the magmas underwent fractional crystallization at lower crustal levels prior to their final emplacement, this implies that their initial temperatures were even higher. Experimental and theoretical work by Huppert & Sparks (1988) has shown that, in the case of intraplating at such temperatures and in such volumes, the lower crust in contact with the intraplated magmas would have undergone widespread partial melting on a time-scale of the order of 10^2 to 10^3 years. This is in marked contrast to terranes dominated by conductive heat transfer where prograde heating through the fluid-absent melting interval is either extremely unlikely, or in the case of crust with a prior elevated heat flux, will take tens of millions of years (e.g. England & Thompson 1984). Geochemical studies of the felsic volcanic rocks and granophyres (Schweitzer et al. 1995) and the Lebowa Granite Suite (Kleeman & Twist 1989) in the Bushveld Complex have established both that they were derived from crustal sources, and that they were emplaced at the same time as the Rustenburg Layered Suite. This, together with their exceptional volumes (> 110 000 km^3 to 300 000 km^3 for the lavas of the Rooiberg Group; Twist & French 1983; Schweitzer et al. 1997), and a minimum of approximately 90 000 km^3 for the Lebowa Granite Suite (Kleeman & Twist 1989) supports the existence of a regionally extensive, intraplating-driven, heating event in the lower crust of the Kaapvaal Craton. The emplacement temperature of the granites (in excess of 900 °C; Kleeman & Twist 1989) suggests that they were derived from levels similar to those exposed in the central granulite-facies terrane in the core of the Vredefort Dome.

To summarize, the greenschist- to granulite- facies metamorphic terrane exposed in the Vredefort Dome represents part of a much more extensive, craton-wide, low-P metamorphic terrane that developed in response to the intrusion of voluminous mantle-derived magmas at ca. 2.05-2.06 Ga. In the next section we discuss the constraints on the origin of the mantle magmatism which gave rise to this event.

Constraints on the Origin of the Bushveld Magmatic Event

Voluminous mafic magmatism is generally attributed to decompression melting of the upper mantle asthenosphere (e.g. McKenzie & Bickle 1988) associated with thinning of the overlying lithosphere or convection within the asthenosphere itself. In recent years, a variety of mechanisms for lithospheric thinning have been proposed to explain how such decompression might also account for regional low-P metamorphism in which magmatic intraplating or underplating was involved (e.g. Bird 1979; Houseman et al. 1981; Sandiford & Powell 1986, 1991; Loosveld 1989). Most of these studies invoke lithospheric plate tectonic processes (subduction, tectonic overthickening or tectonic thinning) as the cause of the decompression.

In the case of the Bushveld Event, it is apparent from regional studies in the Kaapvaal Craton that the magmatism and metamorphism could not have been associated with orogenic activity, as the rocks of the Transvaal Supergroup (1) show no evidence of significant pre-Bushveld deformation, and (2) their preservation over large parts of the Craton indicates a lack of significant erosional exhumation which is a normal consequence of crustal thickening. At 2 Ga, the Craton formed part of a larger continental mass incorporating, inter alia, the Zimbabwe Craton and several other smaller cratons to the north, the Grunehogna Block of Antarctica to the east and, possibly, the Pilbara Craton to the south (Master 1991; Cheney 1996). According to Master (1991), between 2.2 Ga and 1.8 Ga this continental mass formed the overriding plate above a SE-dipping subduction zone whose trench was located some 1200 km to the NW of the central parts of the Kaapvaal Craton. Subduction was finally terminated at ca 1.8 Ga (Ubendian Orogeny; Master 1991). During the Bushveld Event the Kaapvaal Craton experienced extensional and strike-slip reactivation of Archaean structures, consistent with NE-SW directed extension which A. Friese (pers. comm., 1997) attributes to plate boundary forces related to this subduction event. The preservation of the volcanic rocks and shallow-level intrusive rocks of the Bushveld Complex indicates, furthermore, that the significant magmatic thickening related to the Bushveld Event must have been compensated by concomitant crustal thinning.

The volume of magma estimated to have been involved in the Bushveld Event is similar to that of a small- to medium-sized flood basalt event (White & McKenzie 1995). White & McKenzie (1995) suggest that the optimum conditions for such an event involve the interaction of a mantle plume with lithosphere that has been thinned to between 110 and 50 km. However, some authors (e.g. Richards et al. 1989) maintain that such volumes of melting might be achieved within abnormally hot plumes without the need for lithospheric thinning. It is unfortunate that fractionation of the magmas that produced the Rustenburg Layered Suite, and their contamination by crustal material (Cawthorn & Davies 1983; Kruger 1994; Cawthorn & Walraven 1997), preclude direct estimation of the depth of formation of the parent magmas, which might resolve this issue. Homogeneous thinning of the Kaapvaal lithosphere to the values proposed by White & McKenzie (1995) can, however, be ruled out as the Craton contains abundant Late Proterozoic to Cretaceous diamondiferrous kimberlites which contain a predominantly ca. 3.1 Ga diamond population (Richardson et al. 1984). This indicates that a lithospheric root in excess of 140 km must have existed beneath the Craton in the Archaean and that it must have survived the Bushveld Event. One way to reconcile this with the White & McKenzie (1995) model is if lithospheric thinning was heterogeneous and the arrival of a plume beneath the lithosphere led to localised decompression melting in 'thinspots' (Thompson & Gibson 1991). Limited published geophysical data indicate that the crustal thicknesses in the central parts of the Craton vary by up to 30%, from 32 km to 45 km (Green et al. 1995). If similar amounts of thinning also affected the underlying mantle lithosphere, it could indicate regions where the Kaapvaal lithosphere was only 100 km thick. However, this value is at the lower limit of the range required for major melting proposed by White & McKenzie (1995). To compensate for this, the plume head would have to have been exceptionally hot and voluminous, conditions which are most likely to be achieved in juvenile plumes (Richards et al. 1989; Griffiths & Campbell 1990; White & McKenzie 1995). Either the transient nature of such young plumes, which rapidly degenerate into cooler, smaller, steady-state plumes (Richards et al. 1989; Griffiths & Campbell 1990; White & McKenzie 1995), or migration of the craton away from the plume source, might explain why the Bushveld Event never proceeded to the development of oceanic crust.

Figure 6 shows the envisaged model for the Bushveld magmatothermal event, in which a hot juvenile plume reached the base of the Kaapvaal lithosphere at ca 2.06 Ga, leading to partial melting within the plume head and in asthenosphere beneath thinned regions of the lithosphere (Fig. 6a). These 'thinspots' may already have existed in the Kaapvaal lithosphere prior to the Bushveld Event. However, the tectonic reconstructions (A. Friese, pers. comm., 1997) allow for syn-Bushveld extensional thinning. Once these mantle melts had formed, they rose to levels immediately below, or within, the crust where partial fractional crystallization occurred (Fig. 6b). The heat released from these magmas resulted in an elevated crustal geotherm, to values approaching 40-50 °C/km, and regional metamorphism of the adjacent crust. Crustal anatectic magmas generated by this event rose to form the felsic volcanic rocks of the Rooiberg Group and the shallow-level intrusions of the Rashoop Granophye and Lebowa Granite Suites. At the same time, the partially fractionated and contaminated mafic magmas were remobilized and rose to intrude shallow crustal levels as the Rustenburg Layered Suite and its extensions in the Molopo Farms and Losberg Complexes (Fig. 6c). These rocks were preserved as a result of concomitant crustal thinning to compensate for the magmatic overthickening. This thinning prevented exhumation of the deeper levels of the metamorphic terrane until the Vredefort impact event at 2.02 Ga in the south-central parts of the Craton, explaining the IBC paths observed in the metamorphic rocks. The absence of evidence for the formation of oceanic crust suggests that the magmatic event was transient, either due to degeneration of the mantle plume from its initial state to a steady-state, or due to migration of the craton away from the plume source.

Implications for Regional Low-P Metamorphism

It is clear from studies of recent and present-day plume activity (e.g. Hawaii, Kerguelen, Réunion) that mantle plumes can exist independently both of the movement of lithospheric plates and of lithospheric plate boundaries. Starting plumes, in particular, appear capable of inducing voluminous, transient, mantle-derived intraplating events in the crust which, under the right conditions, could generate regional low-P metamorphism and extensive crustal anatexis in intraplate settings unrelated to orogenic or other tectonic activity. This does not preclude such magmatothermal events being associated with large-scale crustal deformation; magmatic intraplating may trigger deformation in the thermally weakened crust, due to plate boundary forces. Given that starting plumes may produce thermal anomalies over

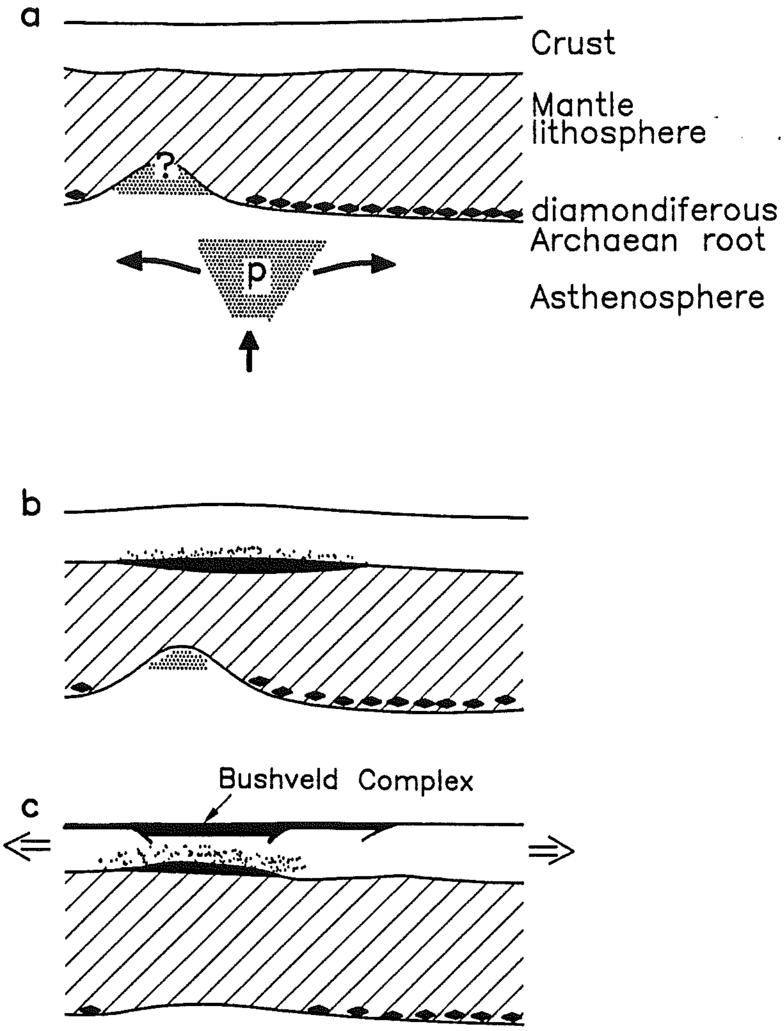

Figure 6: Proposed model for the Bushveld magmatic and metamorphic event. (a) Rising plume beneath the thick Kaapvaal lithosphere results in decompression melting (dark stipple) in the plume head (p) and, possibly, in thin spots. (b) Mantle magmas (black) underplate and intraplate the deep Kaapvaal crust and undergo partial fractional crystallization, leading to medium- to high-grade metamorphism, with widespread anatexis at the highest grades (stipple). (c) Felsic anatectic melts and remaining mafic-ultramafic melts from the deep crustal magma chambers rise and are emplaced as the Bushveld Complex. Thermal weakening of the crust leads to thinning, assisted by rifting due to extensional boundary forces, to accommodate the magmatic overthickening.

regions 1500-2000 km or more across (White & McKenzie 1995), such deformation could be of regional extent.

CONCLUSIONS

1. A greenschist- to granulite-facies low-P metamorphic terrane exposed in the central parts of the Kaapvaal Craton, in the Vredefort Dome, is linked to intracratonic,

anorogenic mantle-derived magmatism at 2.05-2.06 Ga. The highest-grade rocks in the Dome correspond to the upper levels of the anatectic lower crust that generated the felsic magmas, preserved as volcanic rocks and shallow-level granites in the Bushveld Complex.

2. The anticlockwise P-T paths deduced for the medium- and high-grade zones in the metamorphic terrane are consistent with magmatic loading by mantle-derived magmas and felsic melts derived from the lower crust.

3. The volumes of mafic and felsic magma are consistent with the metamorphic event being regionally extensive.

4. The lack of erosion of the Bushveld volcanics and high-level intrusions, and the IBC paths for the metamorphic rocks, indicate crustal extension and thinning concomitant with magmatic thickening.

5. Evidence that diamondiferous, lower lithospheric levels (beneath large parts of the Craton) survived the Bushveld Event indicates that extensive mantle lithospheric thinning may not be a pre-requisite for low-P metamorphism.

6. The intracratonic setting of the metamorphic rocks, and the lack of evidence of temporally related orogenic activity and lithospheric thinning, suggests a mantle plume as the ultimate cause of the Bushveld magmatothermal event. The coincidence of felsic and mafic magmatism indicates that prograde heating of the lower and middle crust is likely to have been extremely fast (<10 Ma).

7. Exhumation of the amphibolite- to granulite-facies mid-crustal levels of the terrane was unrelated to the magmatothermal event itself and required the highly unusual mechanism of meteorite impact.

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