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10 THE TRANSVAAL SUPERGROUP AND ITS PRECURSORS

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INTRODUCTION

The late Archaean to early Proterozoic Transvaal Supergroup is preserved within three structural basins on the Kaapvaal Craton of southern Africa: the Transvaal and Griqualand West Basins in South Africa and the Kanye Basin in Botswana (Figs. 1, 2). The Griqualand West Basin can further be subdivided into the Ghaap Plateau and Prieska sub-basins. The succession preserved in the three structural basins is generally similar, with correlation being best for the lower chemical sedimentary units, and poorest for the upper clastic sedimentary and volcanic lithologies (Table 1). With the exception of volcanic precursor rocks of the Kanye Formation, the so-called protobasinal Transvaal rocks and parts of the Griqualand West Basin, the geometry of

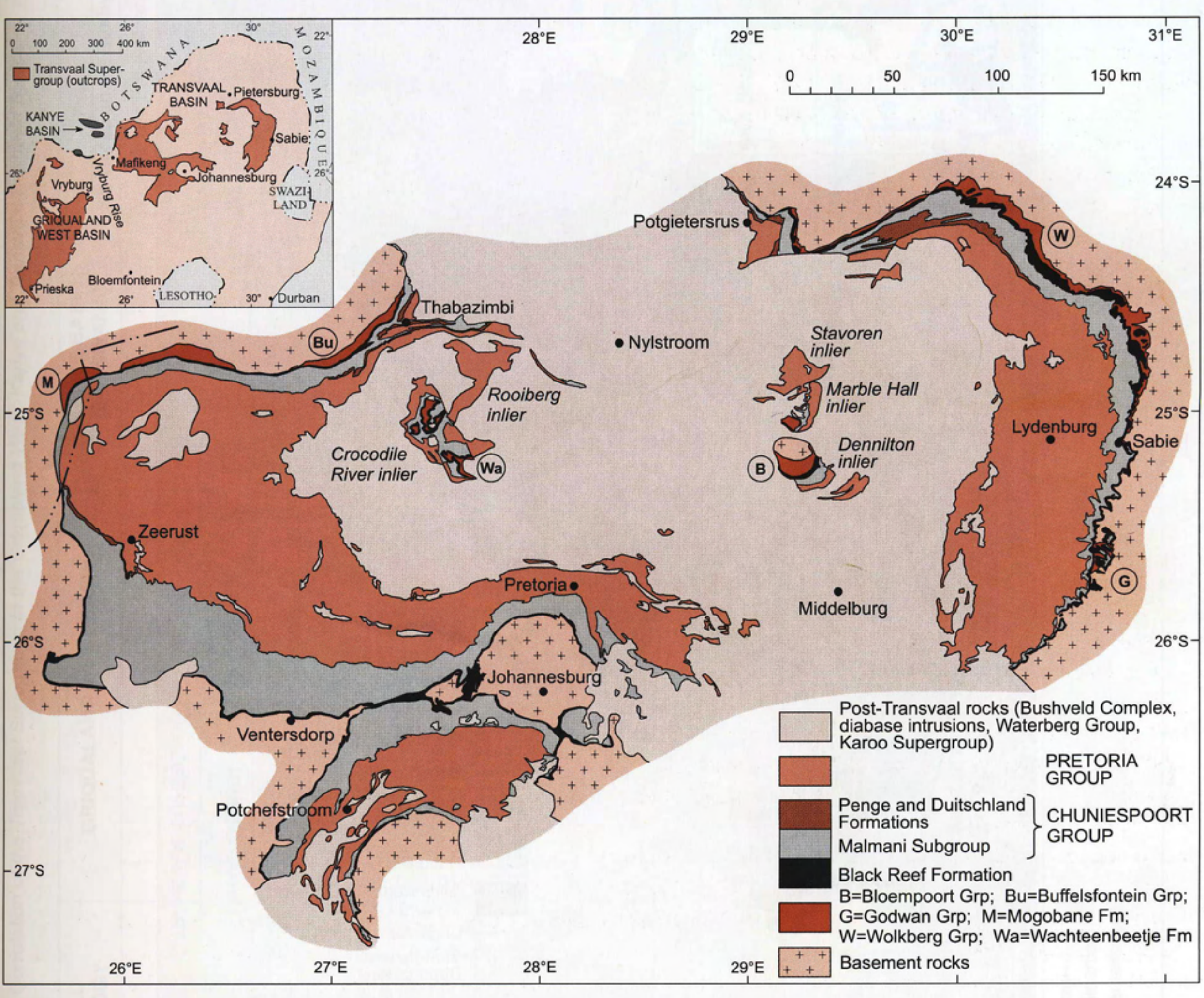

Fig. 1 Geological map of the Transvaal Basin of the Transvaal Supergroup. Inset map shows the entire distribution of the supergroup, including that of the Kanye Basin in Botswana.

the supergroup within the three basins is essentially stratiform. The southwestern portion of the Griqualand West Basin has been extensively deformed during several events. This deformation encompasses up to seven phases between >2500 Ma and 1100 Ma. In the area between Prieska, Westerberg and Griquatown (Fig. 2) stacking of thrust packages affected the stratigraphy and thickness of various formations (Altermann and Hälbich, 1990, 1991). Major thrust faulting in the area of the centrally placed Maremane Dome (Fig. 2) thrust the Gamagara Formation of the succeeding Olifantshoek Supergroup over carbonate rocks of the Ghaap Group, with, in turn, thrusting of Postmasburg Group formations over these younger manganiferous shales; this thrust faulting also extends southwards to the Rooinekke region (Beukes, 1986; Grobbelaar *et al.*, 1995). Metamorphic grade in the southwestern Griqualand West Basin, reflected by illite crystallinity, decreases from low to very low grade from the west to the east (Altermann *et al.*, 1992).

Transvaal Supergroup rocks in the Transvaal Basin were intruded by the approximately 2060 Ma Bushveld Complex (Walraven and Martini, 1995), with the Magaliesberg Formation of the Pretoria Group (Table 1) forming the floor rocks in most areas. Lavas and subordinate sedimentary rocks of the Rooiberg Group (Table 1) locally form the floor to the Bushveld Complex in the southeastern part of the basin, and elsewhere comprise the roof rocks to this major intrusion (Eriksson *et al.*, 1995). Five isolated but extensive occurrences of Transvaal rocks, surrounded by intrusive Bushveld rocks, constitute the Rooiberg, Crocodile River,

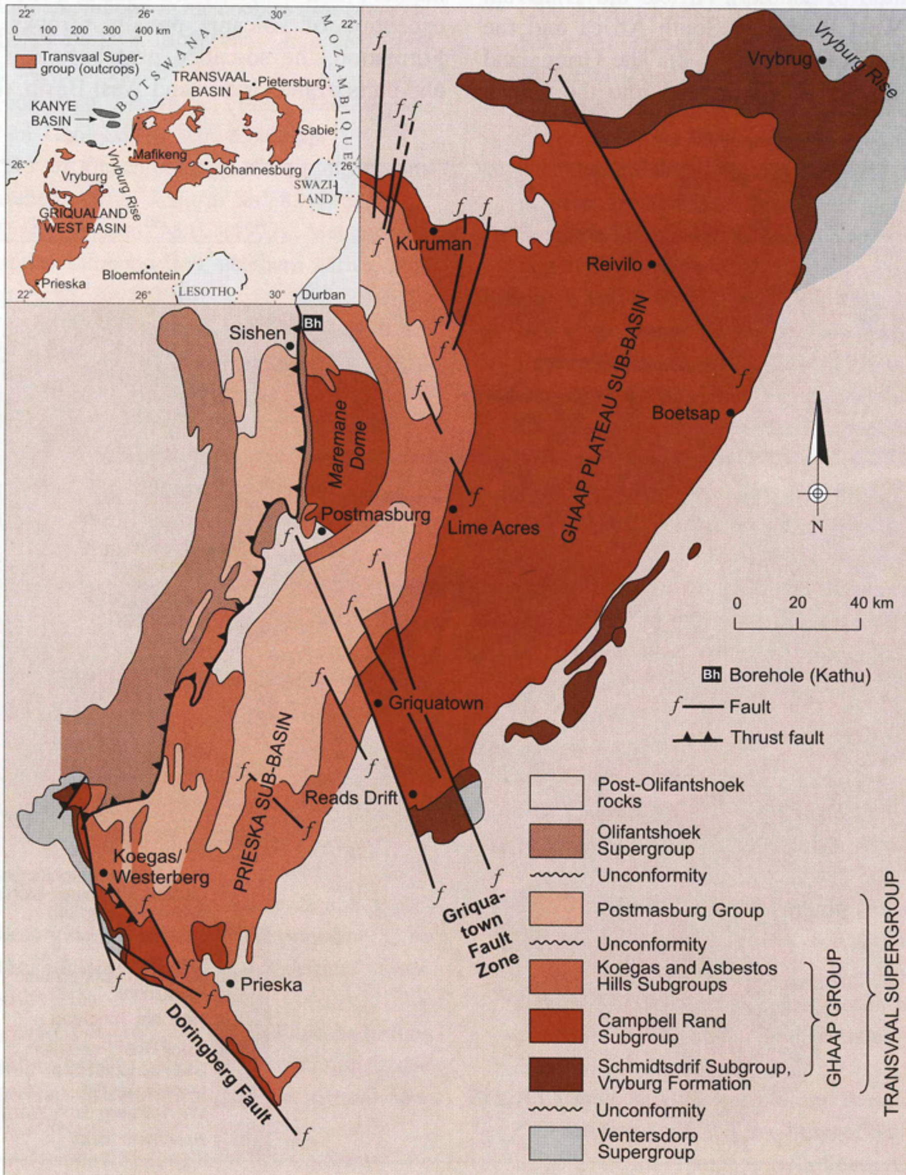

Fig. 2 *Geological map of the Griqualand West Basin of the Transvaal Supergroup.*

Table 1 *Correlation of the Transvaal Supergroup in the Griqualand West, Kanye and Transvaal/Bushveld Basins.*

OVERALL LITHOLOGY	**GRIQUALAND WEST BASIN**		**KANYE BASIN (BOTSWANA)**		**TRANSVAAL BASIN**		
					SE Botswana ("Bushveld basin")	**South Africa**	
Clastic sediments and volcanic rocks	**Postmasburg Group**				Woodlands Fm (volcanic and clastic rocks)	Rayton Fm, etc. (sandstone, shale, volcanic rocks)	**Pretoria Group**
		Mooidraai Fm (dolomites), Hotazel Fm (manganiferous ironstone)			Sengoma Quartzite Fm (sandstone)	Magaliesberg Fm (sandstone)	
					Sengoma Argillite Fm (shale)	Silverton Fm (shale, lava)	
			Segwagwa Group	Mogapinyana and Gatsepane Fms (chert, sandstone, shale)	Ditlhojana Quartzite Fm (sandstone, conglomerate)	Daspoort Fm (sandstone)	
						Strubenkop Fm (shale)	
						Dwaalheuwel Fm (sandstone)	
		Ongeluk Fm (andesite)		Tsatsu Fm (andesite)	Ditlhojane Volcanic Fm (volcanic rocks)	Hekpoort Fm (andesite)	
				Ditlhojana and Tlaameng Fms (shale, sandstone, conglomerate)	Ditlhojane Shale Fm: Conglomerate (minor)	Boshoek Fm (conglomerate, sandstone)	
		Makganyene Fm (diamictite)			Ditlhojane Shale Fm: Shale (predominant)	Timeball Hill Fm: Shale	
					Tsokwane Quartzite Fm (sandstone)	Timeball Hill Fm: Sandstone	
					Lephale Fm: Shale	Timeball Hill Fm: Shale/lava	
					Lephale Fm: Conglomerate, sandstone	Rooihoogte Fm (conglomerate, sandstone)	
Regional unconformity							
Chemical sediments	**Ghaap Group**	*Koegas Subgroup* / *Asbestos Hills Subgroup*	**Taupone Group**	Masoke Fm (ironstone, ferruginous chert)	Ramotswa Shale Fm (shale)	Duitschland Fm (carbonate and clastic rocks)	**Chuniespoort Group**
						Penge Fm (iron-formation)	
		Campbell Rand Subgroup: Various formations (dolomite)		Kgwakgwe Fm (chert breccia)	Magobane Fm, Maholobota Fm, Ramotswa Dolomite Fm (dolomite)	Malmani Subgroup (dolomite)	
		Schmidtsdrif Subgroup: Clearwater Fm (shale), Boomplaas Fm (dolomite)		Ramonnedi Fm (dolomite)			
Clastic sediments	Vryburg Fm (mixed)			Black Reef Fm (quartzite)	Black Reef Quartzite Fm (quartzite)	Black Reef Fm (quartzite)	
Clastic sediments, volcanic rocks (Ventersdorp age?)						Pre-Black Reef units – Wolkberg Group and correlates	

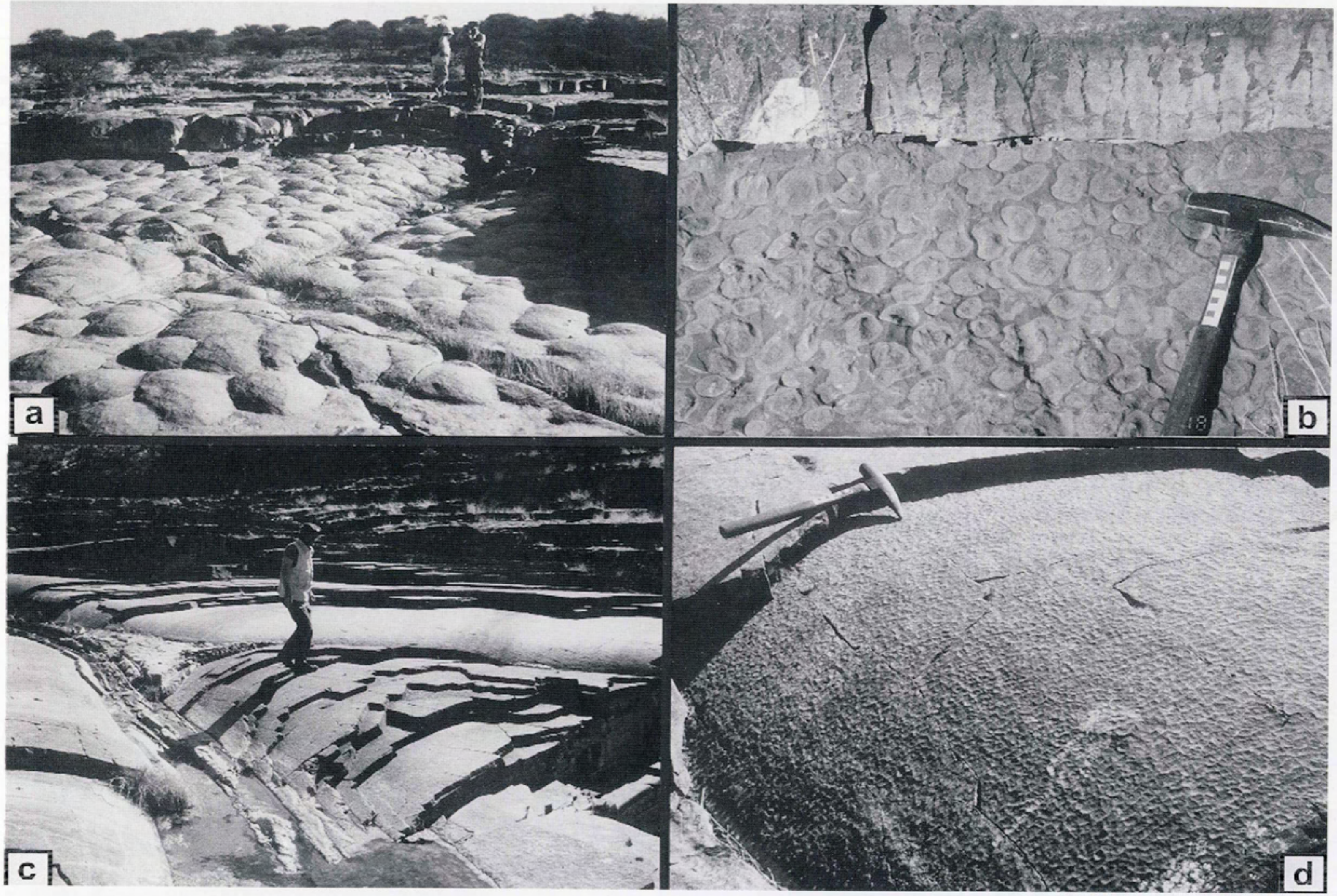

Fig. 3 *Stromatolites in the Campbell Rand Subgroup, Ghaap Group, at Boetsap in the Northern Cape. (a) Closely spaced domical stromatolites with fine internal laminations: shallow, low-energy, subtidal palaeoenvironment. In this outcrop, these stromatolites are overlain by columnar stromatolites (b), with coarse sediment trapped between the columns, reflecting an intertidal palaeoenvironment of higher energy than (a). (c) Elongated large stromatolite domes with wavy laminated internal structure, reflecting a subtidal, current-influenced environment. The elongation is parallel to palaeocurrent flow. (d) Detail of the surface of elongated domes (as shown in (c)), illustrating "rippled" tufted structure. The tufts are accentuated by trapped fine sediment, forming ripple-like crests perpendicular to stromatolite elongation and to palaeocurrent flow direction.*

Stavoren (Makeckaan), Marble Hall and Dennilton inliers (Fig. 1); they are interpreted as floor-attached domes and roof pendants (Hartzer, 1995). In most areas, Transvaal Supergroup strata dip towards the centrally located Bushveld lithologies. Several other smaller inliers occur, scattered in both the eastern and western Bushveld Complex.

The Transvaal Supergroup overlies the Archaean basement, the Witwatersrand and Ventersdorp Supergroups, and, in the far western Transvaal and Kanye Basins, rocks belonging to the Kanye Formation and the Gaborone Granite Suite. Kanye Formation rhyolitic lavas and Gaborone Suite granites occur in eastern Botswana, and on the South African side of the border in the Mafikeng and Derdepoort areas; correlatable lavas and granites may be present beneath Transvaal rocks in the Dennilton inlier (Walraven *et al.*, 1994; Hartzer, 1995). These acidic precursors to the Transvaal Supergroup are dated at 2830–2770 Ma (Walraven *et al.*, 1994).

The Transvaal Supergroup encompasses one of the world's earliest carbonate platform successions (Beukes, 1987; Altermann and Wotherspoon, 1995), with very well preserved and extensive stromatolites (Fig. 3) and an excellent record of cyanobacterial and bacterial evolution, recording the early history of life on earth (Klein *et al.*, 1987; Altermann and Schopf, 1995). These carbonates are overlain by economically important, laterally extensive banded iron-formation (BIF) containing some of the world's largest iron and asbestos deposits. Succeeding sedimentary and volcanic units include the Hekpoort-Ongeluk flood basalt, which, in the Griqualand West Basin, is overlain by the world's largest manganese deposit. Mississippi Valley-type base metal (Martini *et al.*, 1995; Altermann, 1997) and structurally controlled gold deposits (Tyler and Tyler, 1996) also occur within the supergroup. There is also potential for sedimentary-exhalative ore deposits (Reczko *et al.*, 1995a).

PROTOBASINAL ROCKS

The protobasinal rocks (a purely descriptive term) comprise a number of separate stratigraphic units: Mogobane Formation (in Botswana), rocks of the Tshwene-Tshwene belt, Buffelsfontein Group, Wachteenbeetje Formation, Bloempoort Group, Wolkberg Group and Godwan Group (Fig. 4). Each protobasinal unit has its own stratigraphy and none of them are in contact with each other. Although these units are generally laterally restricted, and are mostly relatively thick for their preserved strike length (Fig. 4), Hartzer (1994) has shown that the upper parts of the Wolkberg and Bloempoort

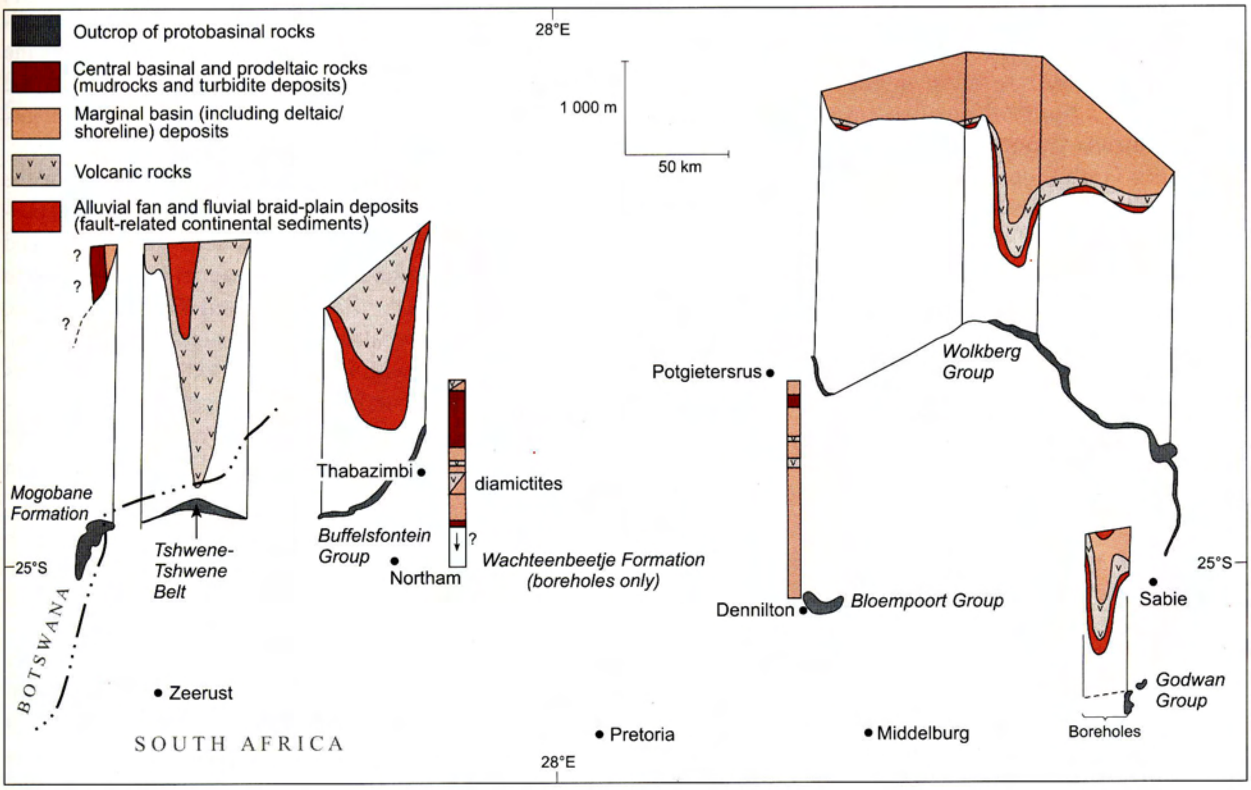

Fig. 4 Outcrop areas, geometry and inferred depositional systems for the protobasinal rocks of the Transvaal Supergroup.

Groups can be directly compared. The Bloempoort Group and the Wachteenbeetje and Mogobane Formations crop out very poorly, and are known mainly from boreholes (Hartzer, 1994).

Wolkberg Group

The largest protobasinal succession, the Wolkberg Group, comprises eight units (Button, 1973a; Bosch *et al.*, 1993). Lenticular, coarse-grained, immature sandstone and conglomerate characterise the basal **Sekororo Formation**, which is overlain by andesitic basalts of the **Abel Erasmus Formation**. The succeeding **Schelem Formation** has a similar lithology to the Sekororo Formation, and is overlain by the **Selati Formation** (mudrock and sandstone), the **Mabin Formation** (sandstone and subordinate mudrock) and the **Sadowa Formation** (mudrock and sandstone). The **Langkrans Formation** (quartzitic sandstone) and the **Serala Formation** (basalt) complete the Wolkberg Group.

Button (1973a) proposed a combination of fluvial sedimentation, volcanism, deltaic deposition, and littoral to shelf settings for the group. More recently, Bosch *et al.* (1993) suggested that the Wolkberg Basin may have been closed (i.e., non-marine); Clendenin *et al.* (1991) postulated braided fluvial deposition for at least the uppermost part of the Langkrans Formation. A strong measure of tectonic control is envisaged, with initial mechanical rifting having been followed by thermal subsidence (Eriksson and Reczko, 1995).

Godwan Group

The Godwan Group, located to the south of the Wolkberg Group, comprises an analogous succession of clastic sedimentary and lesser volcanic rocks (Fig. 4). Mass-flow diamictites and pyroclastic rocks are more prominent here than in the Wolkberg Group (Myers, 1990). Inferred depositional conditions and the tectonic framework were similar to those controlling Wolkberg sedimentation (Eriksson and Reczko, 1995).

Buffelsfontein Group

The Buffelsfontein Group in the northwest comprises a thick basal succession (the **Hampton Formation**) of immature sandstones, conglomerates and shales (Tyler, 1979a). This is overlain by predominantly acid lavas and pyroclastic rocks of the **Waterval** and **Witfonteinrant Formations**. These lavas are depleted in Na and enriched in K, probably reflecting marked alteration (Tyler, 1979a). Immature, coarse, clastic sedimentary rocks of the **Kransberg Formation** complete the Buffelsfontein succession. Tyler (1979a) interpreted the sedimentary rocks to be fluvial and shallow-marine deposits; this has been modified to an overall fluvial model by Eriksson and Reczko (1995, *cf.* Fig. 4). SHRIMP ages of 2657 and 2659 Ma have been reported for the upper lavas of the Buffelsfontein Group (minutes of 32nd meeting of SACS, 1993).

Tshwene-Tshwene Belt, Bloempoort Group and Wachteenbeetje Formation

Analogous rocks in the Tshwene-Tshwene belt to the west of the Buffelsfontein Group comprise basal, predominantly mafic lavas, succeeded by very immature, coarse, clastic sedimentary rocks (Fig. 4) which include volcaniclastic lithologies. The mafic lavas are chemically comparable to the Ventersdorp Supergroup lavas, with fluvial sedimentation and mass-flow deposition being inferred for the sedimentary rocks (Tyler, 1979b; Eriksson and Reczko, 1995). Further to the west, along both sides of the Botswana border, poorly exposed outcrops of argillaceous and lesser arenitic sedimentary rocks of the **Mogobane Formation** occur; depositional conditions are thought to have varied from shallow basin margin settings to deep basinal environments with turbidites (Hartzer, 1994). Hartzer assigned an essentially analogous model to the sedimentary successions of the **Wachteenbeetje Formation** and the **Bloempoort Group** (Fig. 4). Both comprise predominant upward-coarsening cycles of carbonaceous claystone-siltstone and sandstone, with subordinate coarse-grained clastic rocks, carbonates and volcanic rocks (Hartzer, 1994). Eriksson and Reczko (1995) suggested that the Mogobane and Wachteenbeetje Formations and the Bloempoort Group may have been deposited in a single, narrow, rift-valley-type closed basin.

In a general overview of the tectonic setting of the Transvaal Supergroup, Eriksson and Reczko (1995) postulated that the protobasinal rocks may have been laid down in small strike-slip or extensional basins related to collision between the Zimbabwe and Kaapvaal Cratons, approximately coeval with the Ventersdorp volcanism. They also agreed with Hartzer (1994) that the lithologies and depositional conditions of the protobasinal successions are broadly analogous and can be related to a fairly simple set of palaeoenvironmental conditions, given in Figure 4.

BLACK REEF AND VRYBURG FORMATIONS

The **Black Reef Formation**, consisting predominantly of relatively mature quartz arenites with lesser conglomerates and subordinate mudrocks, forms a thin veneer of arenaceous rocks unconformably overlying older successions. An impersistent basal conglomerate is succeeded by thicker sandstones and thin mudrocks, the whole forming an upward-fining sequence (Henry *et al.*,

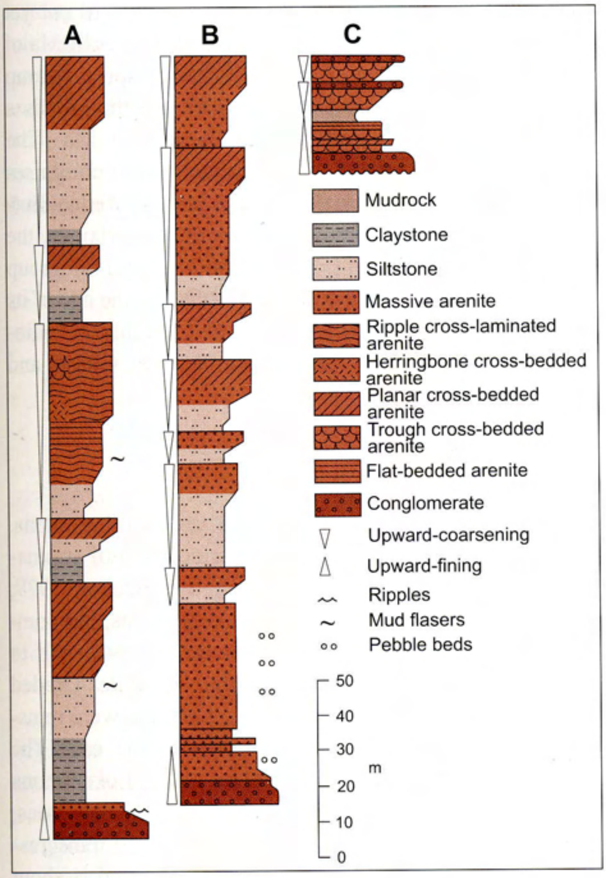

Fig. 5 Typical vertical sections through the Black Reef Formation, illustrating lithology and inferred palaeoenvironmental conditions. A: Crocodile River inlier (Hartzer, 1994); B: Dennilton inlier (Hartzer, 1994); C: South of Sabie (Henry et al., *1990).*

1990). This lower sequence tends to fill irregularities in the floor rocks (Button, 1973a; Els *et al.*, 1995), and is followed by upward-coarsening sequences comprising mudrocks and sandstones (Key, 1983; Henry *et al.*, 1990) (Fig. 5). The formation forms a very widespread thin sheet sandstone, generally varying between a few metres and 30 m in thickness, with up to 60 m occurring locally in the west of the basin (Fig. 6). Anomalously large thicknesses (up to 210 m) were obtained in the Marble Hall, Dennilton and Crocodile River inliers (Fig. 6), with the formation exhibiting repeated upward-coarsening mudrock-sandstone cycles (Hartzer, 1994; Fig. 5).

Depositional models for the Black Reef sandstones suggest a combination of initial fluvial sedimentation followed by shallow-marine conditions as an epeiric sea (later to form the succeeding carbonate-BIF platform succession) advanced onto the Kaapvaal Craton (Button, 1973a; Key, 1983; Els *et al.*, 1995). An alternative, purely fluvial model (Clendenin, 1989; Henry *et al.*, 1990; Eriksson and Reczko, 1995) envisages the basal black shales of the succeeding carbonate rock sequence as representing the onset of this marine transgression. This alternative view is, perhaps, supported by an inferred palaeodrainage divide with divergent palaeocurrent directions, identified by Els *et al.* (1995) in the southwest of the basin (Fig. 6). Hartzer (1994) proposed fluvial and succeeding beach sedimentation for the thick Black Reef deposits in the inliers; an alternative continental braid-delta model was presented by Eriksson and Reczko (1995). The erosional surface underlying the Black Reef Formation is seen as representing widespread removal of pre-existing lithologies and deposi-

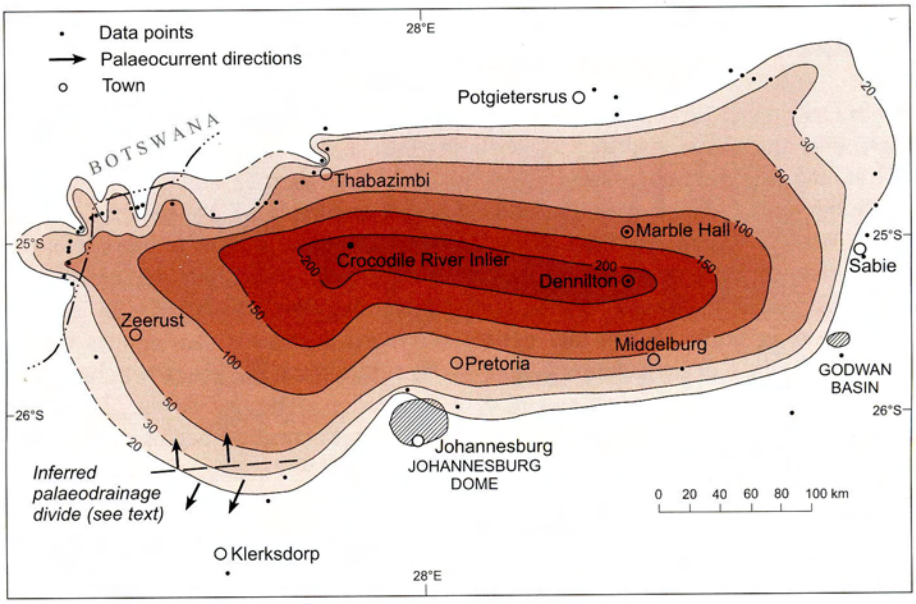

Fig. 6 Isopach map of Black Reef Formation thicknesses (in metres).

tion of a sheet sandstone as a new base level was established under gentle, northward-directed tectonic compression (Clendenin, 1989). Further support for such compression is derived from pre-Black Reef compressional shearing in the Johannesburg Dome (Hilliard, 1994), deformation of the Godwan Group succession (Fig. 4) prior to Black Reef deposition, and from Els *et al.*'s (1995) palaeodrainage divide (Fig. 6).

The Black Reef Formation has traditionally been regarded as a time equivalent of the Vryburg Formation in Griqualand West (SACS, 1980). However, Walraven and Martini (1995) suggested, on the basis of radiometric age dating of the overlying Oaktree Formation, that disconformities might be hidden above and within the Black Reef Formation, and that such a correlation with the Vryburg Formation might thus be incorrect.

The **Vryburg Formation** directly overlies the Ventersdorp Supergroup in Griqualand West. It locally exhibits a basal, transgressive conglomerate, with quartzites, shales and subordinate stromatolitic carbonates passing up in places into basaltic to andesitic amygdaloidal lavas dated at 2642 ± 3 Ma (Walraven and Martini, 1995). It has been interpreted as a fluvial to marginal marine deposit, shed from western and northwestern source areas (Beukes, 1979). The formation is generally ~100 m thick in outcrop, but may be up to 340 m thick in places or even 900 m if the underlying Geelbeksdam (lava) and Kalkput (quartzite) Members are regarded as part of the formation (Smit *et al.*, 1991). In a stratigraphic borehole at Kathu (Fig. 2) the formation is at least 230 m thick and comprises shales, quartzites and laminated carbonates deposited under shelf conditions, and thus reflects a much deeper environment than known from surface outcrops (Altermann and Siegfried, 1997).

CHUNIESPOORT AND GHAAP GROUPS

The Ghaap Group in Griqualand West is subdivided, in stratigraphic order, into the Schmidtsdrif, Campbell Rand, Asbestos Hills and Koegas Subgroups (Fig. 7). The Schmidtsdrif Subgroup is similar over the entire preserved Griqualand West Basin, whereas the Campbell Rand Subgroup comprises differing sedimentary facies in the Prieska and Ghaap Plateau sub-basins. The Asbestos Hills Subgroup is uniform over the entire Griqualand West Basin, but the Koegas Subgroup is preserved only in the southern part of this basin and along the southwestern rim of the Maremane Dome (Fig. 2). The Ghaap Group is younger than 2642 ± 3 Ma (zircon age for the Vryburg lavas below the Schmidtsdrif Subgroup – Walraven and Martini, 1995). The youngest radiometric age of the Ghaap Group is 2432 ± 31 Ma for the base of the Daniëlskuil Formation of the Asbestos Hills Subgroup (Trendall *et al.*, 1990; also discussed in Barton *et al.*, 1994). The Koegas Subgroup has not yet been dated, but must be older than the 2222 ± 13 Ma of the Ongeluk Formation of the Postmasburg Group (Cornell *et al.*, 1996). The Ghaap Group thus spans a time period of probably more than 200 Ma. The Chuniespoort Group of the Transvaal Basin comprises the Malmani Subgroup, followed by the Penge Formation, which is, in turn, unconformably overlain by the Duitschland Formation (Fig. 7). The Malmani Subgroup is dated between ~2600 and ~2500 Ma, but the age of its base remains uncertain. Table 2 lists the results of radiometric dating carried out to date on the Ghaap and Chuniespoort Groups.

Schmidtsdrif Subgroup

The Schmidtsdrif Subgroup comprises two formations. The lower, **Boomplaas Formation** consists of stromatolitic and oolitic platform carbonates (Beukes, 1979, 1983), up to 100 m thick in surface outcrops, but comprising 185 m of shale and carbonate rocks of shelf to deep lagoonal origin in the Kathu borehole. Graded oolite beds encountered in this environment were transported from the shallower platform in the east. The **Clearwater Formation** (equivalent to the Lokammona Formation of Beukes, 1987), consisting of shales, tuffites and BIF-like cherts, is interpreted as a transgressive deposit over the Boomplaas platform and is about 50 m thick in outcrops and in the Kathu borehole.

Malmani and Campbell Rand Subgroups

The Malmani Subgroup in the Transvaal Basin is up to 2000 m thick and is subdivided into five formations, based on chert content, stromatolite morphology, intercalated shales and erosion surfaces (Button, 1973b; Eriksson and Truswell, 1974). It starts above the Black Reef Formation with the **Oaktree Formation**. This unit is transitional from siliciclastic sedimentation to platform carbonates and consists of 10–200 m of carbonaceous shales, stromatolitic dolomites and locally developed quartzites. A tuff layer in the upper Oaktree has been dated at 2585 Ma (Martin *et al.*, 1998). Based on this date, these authors suggested that a hiatus may exist between this formation and the Black Reef Formation (assuming that the 2642 Ma date obtained for the Vryburg Formation also applies to the Black Reef) or that the latter is much younger than the Vryburg Formation. The following, 300–500-m-thick **Monte Christo Formation** begins with an erosive breccia and continues with stromatolitic and oolitic platformal dolomites. The **Lyttelton Formation** follows the Monte Christo, with 100–200 m of shales, quartzites and stromatolitic dolomites, and is, in turn, overlain by the cherty dolomites of the **Eccles Formation**, up to 600 m thick, and which include a series of erosion breccias. The erosion breccias within the Eccles are locally auriferous, mineralisation being attributed to the hydrothermal remobilisation of fluids by the Bushveld Complex (Tyler and Tyler, 1996). An erosion breccia separates the Eccles from the overlying **Frisco Formation**, comprising mainly stromatolitic dolomites. The latter becomes more shale-rich towards the top, reflecting deepening depositional conditions, and is up to 400 m thick.

Eriksson and Truswell (1974) and Eriksson *et al.* (1975) developed a tidal palaeoenvironmental model for the Malmani Subgroup, ranging from supratidal flat stromatolitic mats to intertidal columnar stromatolites, with a subtidal zone characterised by giant stromatolitic domes. Clendenin (1989) recognised unconformity-

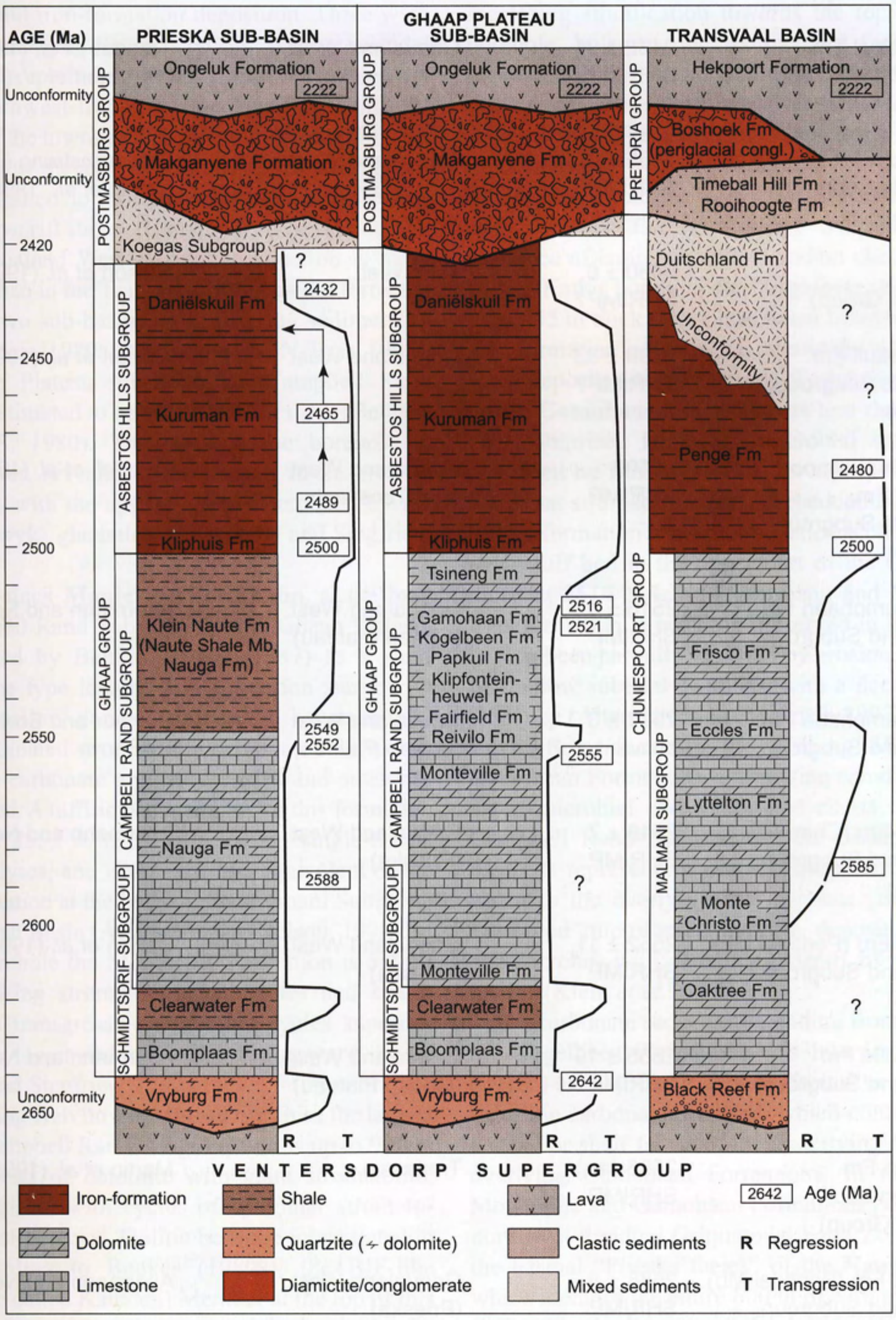

Fig. 7 *Transgression-regression interbasinal correlation for the Chuniespoort and Ghaap Groups (Altermann and Nelson, 1998). For lithological descriptions, see text. Note that the vertical scale of the columns in this figure refers to time and not to the thickness of the strata. See Table 2 for further information on the dates.*

Table 2 *Summary of radiometric ages from the Ghaap, Chuniespoort, Postmasburg and Pretoria Groups.*

Stratigraphic unit	Age (Ma ± 2s)	Basin, sub-basin	Author(s)
Hekpoort Fm (Pretoria Group)	2224 ± 21 Rb-Sr	Transvaal	Burger and Coertze (1973–74)
Ongeluk Fm (Postmasburg Group)	2222 ± 13 Pb-Pb	Griqualand West	Cornell *et al.* (1996)
Lower Daniëlskuil Fm (Asbestos Hills Subgroup, Ghaap Group)	2432 ± 31 SHRIMP	Griqualand West	Trendall *et al.* (1990); *cf.* Barton *et al.* (1994)
Upper Kuruman Fm (Asbestos Hills Subgroup, Ghaap Group)	2465 ± 7 SHRIMP	Griqualand West	R.A. Armstrong (*in* Martin *et al.*, 1998)
Penge Fm (Chuniespoort Group)	2480 ± 6 SHRIMP?	Transvaal	Nelson *et al.* (1999)
Lower Daniëlskuil Fm (Asbestos Hills Subgroup, Ghaap Group)	2489 ± 33 SHRIMP?	Griqualand West	Nelson *et al.* (1999)
Penge Fm (Chuniespoort Group) and Kuruman Fm (Asbestos Hills Subgroup, Ghaap Group)	~2500 SHRIMP	Griqualand West and Transvaal	Trendall *et al.* (1995)
Uppermost Gamohaan Fm (Campbell Rand Subgroup, Ghaap Group)	2516 ± 4 SHRIMP	Griqualand West (Ghaap Plateau)	Altermann and Nelson (1998)
Uppermost Gamohaan Fm (Campbell Rand Subgroup, Ghaap Group)	2521 ± 3 convent.	Griqualand West (Ghaap Plateau)	Sumner and Bowring (1996)
Upper Nauga Fm (Chert Mb) (Campbell Rand Subgroup, Ghaap Group)	2549 ± 7 SHRIMP	Griqualand West (Prieska)	Altermann and Nelson (1998)
Upper Nauga Fm (Peritidal Mb) (Campbell Rand Subgroup, Ghaap Group)	2552 ± 11 SHRIMP	Griqualand West (Prieska)	Barton *et al.* (1994)
Upper Monteville Fm (Campbell Rand Subgroup, Ghaap Group)	2555 ± 19 SHRIMP	Griqualand West (Ghaap Plateau)	Altermann and Nelson (1998)
Upper Oaktree Fm (Malmani Subgroup, Chuniespoort Group)	2585 (av.) SHRIMP	Transvaal	Martin *et al.* (1998)
Middle Nauga Fm (Peritidal Mb) (Campbell Rand Subgroup, Ghaap Group)	2588 ± 6 SHRIMP	Griqualand West (Prieska)	Altermann and Nelson (1998)
Vryburg Formation	2642 ± 2 Evap.	Griqualand West (Ghaap Plateau)	Walraven and Martini (1995)

bounded sequences in the Chuniespoort Group and subdivided the Malmani epeiric sea sediments into four time-stratigraphic units called "packages". He identified a fifth "package" of basinal sediments in Griqualand West and proposed a carbonate ramp depositional model for the Malmani and Campbell Rand Subgroups. In this model, the southwesterly inclined carbonate ramp extended into a shallow basin, with mixed carbonate/shale and iron-formation deposition. Three wide facies belts, ranging from shallow basin, subtidal periplatform to peritidal facies, were recognised. A northeast-southwest-trending line, passing approximately east of the towns of Vryburg and Prieska (Fig. 2), was identified as the axis of this basin. The sedimentation was attributed to extensional tectonics reflecting three successive rift stages (Clendenin *et al.*, 1988).

In the Griqualand West Basin, the situation is more complicated than in the Transvaal Basin, as the former is divided into two sub-basins with differing sedimentary histories (Beukes, 1980a; Altermann and Nelson, 1998). In the Ghaap Plateau sub-basin the Campbell Rand Subgroup is estimated to be about 1600 m thick (Beukes, 1980a; SACS, 1980). The stratigraphic borehole at Kathu, however, revealed the subgroup to be at least 2500 m thick, with the uppermost part removed by the Palaeozoic Dwyka glaciation (Altermann and Siegfried, 1997).

The 200-m-thick **Monteville Formation**, at the base of the Campbell Rand Subgroup on the Ghaap Plateau, was interpreted by Beukes (1980a, 1987) as a shelf deposit. At the type locality this formation starts with giant stromatolitic domes, succeeded by microbial laminites (laminated stromatolitic carbonate rocks) with fenestrae and carbonate argillites; shales and siltstones are intercalated. A tuffitic bed at the top of this formation was dated at 2555 ± 19 Ma, based on single-zircon SHRIMP analyses, and is thus the age equivalent of the Oaktree Formation at the base of the Malmani Subgroup in the Transvaal Basin (Altermann and Nelson, 1998). In the Kathu borehole the Monteville Formation is 540 m thick, comprising stromatolitic carbonates and shales arranged into transgressive-regressive cycles superimposed on a lower-order shallowing-upward cycle (Altermann and Siegfried, 1997).

The following **Reivilo Formation** occupies the largest part of the Campbell Rand Subgroup and is up to 900 m thick. It consists of dolomite with giant stromatolitic domes intercalated with cycles of columnar stromatolites and fenestral facies. Oolitic beds are intercalated in places. According to Beukes (1980a), the BIF-like Kanguru (also called Kamden) Member at the top of this formation marks a renewed transgressive phase over the Campbell Rand platform.

With the **Fairfield Formation**, composed of clastic laminated carbonate beds passing upward into columnar stromatolites and fenestrated laminites, shallow platform conditions returned to the Ghaap Plateau sub-basin. The overlying **Klipfonteinheuwel** and **Papkuil Formations** also comprise platform carbonates with columnar stromatolites, oolite beds, and laminated and sapropelic facies. The following **Klippan Formation** typically starts with small domal stromatolites, a few centimetres high, which pass upwards into coarse-crystalline microbial laminites with drusy quartz concentrations and increasing silicification towards the top. In the Kathu borehole, however, the sedimentary facies is of deep lagoonal affinity, interpreted as a transgression and evidenced by marls and dark, carbonaceous, oncolitic dolomicrites. The overlying **Kogelbeen Formation** is represented by varying dolomite, limestone and chert lithologies. Cycles of domal to columnar stromatolites and laminites follow, and are overlain by a thick sequence of laminites with abundant chert replacement. In the Kathu borehole the Kogelbeen Formation is at least 445 m thick. The Lime Acres Member at the top of this formation contains economically important limestone deposits (Altermann and Wotherspoon, 1995).

The **Gamohaan Formation** is less than 100 m thick and comprises laminated microbial mat carbonates, described by Beukes (1980a) as sapropelic. Domal and columnar stromatolites are interbedded in the lower part of this formation and are associated with some dolarenite. A tuff bed in the upper part of the Gamohaan was dated at 2516 ± 4 Ma (Altermann and Nelson, 1998). This formation is not fully preserved in the Kathu core, having been partially removed by erosion. The facies are of shallow subtidal character with a deepening-upward tendency (Altermann and Siegfried, 1997). The **Tsineng Formation** (formerly the Tsineng Member of the Gamohaan Formation), comprising some 30 m of laminated microbial carbonates and cherts, terminates the Campbell Rand Subgroup in the Ghaap Plateau sub-basin. It represents a deepening-upward facies, transitional to the overlying BIF deposits (Beukes, 1980a). Preserved microfossils indicate deposition within the photic zone, with maximum depth of a few tens of metres (Klein *et al.*, 1987).

The carbonate sequence extending from the Reivilo to the Kogelbeen Formations was interpreted by Beukes (1980a) as the "Ghaap Plateau facies", a preserved stromatolitic carbonate platform, which contrasted with the basinal or shelf facies of the underlying Monteville and overlying Gamohaan Formations. In this model, the Monteville and Gamohaan Formations pass south of the northwest-trending Griquatown Fault Zone (Fig. 2) into the basinal "Prieska facies" of the **Nauga Formation**, which includes the entire 600 m of carbonates that accumulated in the Prieska sub-basin. The Nauga Formation carbonates are overlain by a thick sequence of shales (**Klein Naute Formation**, i.e., Buisvlei Bed of SACS, 1980), with some thick chert beds of great lateral continuity. The Nauga and Klein Naute Formations were thought to reflect shelf or deeper basinal conditions, with turbiditic carbonate deposition from the Ghaap Plateau platform.

An alternative model, in which the Ghaap Plateau sub-basin experienced the highest subsidence rates, was proposed by Altermann and Herbig (1991). The subsidence was matched by stromatolitic growth and shallow-marine conditions thus prevailed throughout. South of the Griquatown Fault (Prieska sub-basin), peritidal flats, which were often exposed to erosion, prevented the accumulation of a thick pile of carbonate strata. The decline in carbonate sedimentation was accompanied by siliciclastic influx, evident in the Klein Naute shales. The synsedimentary hinge character of the Griquatown Fault, which separates the Ghaap Plateau and Prieska sub-basins (Beukes 1980a, 1987), was questioned by Altermann (1997) who argued that this fault zone is mainly of strike-slip affinity, with some vertical displacement.

In the Prieska sub-basin, rapid lateral and vertical facies changes within the lower Nauga Formation have been described by Altermann and Nelson (1998). Vertically, the Nauga Formation can be subdivided into four informal members:

(1) a basal member, with carbonate and some siliciclastic sediments of peritidal origin;
(2) a second peritidal member, consisting of widespread stratiform mats and tidal channel carbonate sand bodies. A tuff layer in this member gave an age of 2588 ± 6 Ma (Table 2);
(3) a chert member, in which the tidal flat features give way to lagoonal platy dolomicrites, dolarenites and microbial laminites and laterally persistent chert marker horizons. A tuff bed at the top of this member was dated at 2549 ± 7 Ma (Altermann and Nelson, 1998); and
(4) an overlying proto-BIF member, consisting mainly of carbonates with thin microbial mats and three laterally persistent BIF-like cherts (proto-BIF of Button, 1976a), deposited below wave base.

The approximately 150-m-thick Klein Naute Formation, comprising finely laminated shales intercalated with rare thin tuffites and prominent chert beds, overlies the above succession and is considered to represent shelf deposition, probably below the storm wave base.

The comparison of ages for the various formations of the Campbell Rand and Malmani Subgroups necessitated a new stratigraphic correlation across the Transvaal, Prieska and Ghaap Plateau depositories (Fig. 7). Carbonate platforms of the Ghaap Plateau and the Transvaal basins lasted for at least 30 Ma longer than in the Prieska area, where the Klein Naute shales were deposited between roughly 2550 Ma and 2500 Ma. The correlation of the Klein Naute Formation of the Prieska sub-basin with the Reivilo to Gamohaan Formations of the Ghaap Plateau sub-basin, and of the upper Nauga Formation carbonates with the Monteville Formation of the Ghaap Plateau and Oaktree Formation of the Transvaal Basin, led to a new model of basin development (Altermann and Nelson, 1998), discussed briefly below. The Boomplaas Formation above the Vryburg Formation originated as the first carbonate platform between 2642 and 2588 Ma in Griqualand West. The subsequent transgression resulted in deposition of the Clearwater Formation shales, which were followed, on the Ghaap Plateau, by the Monteville Formation, when platform conditions returned before 2555 Ma. In the Transvaal Basin the Oaktree Formation sedimentation commenced somewhat prior to 2550 Ma. This suggests a transgression, progressing from the west or southwest towards the east or northeast. The major transgressive step, at around 2550 Ma, drowned the tidal flats in the Prieska sub-basin, where the Klein Naute shales were deposited after 2549 Ma, and shifted the main site of carbonate sedimentation to the east, until the next major transgressive step, after 2516 Ma (Gamohaan Formation), also drowned the Ghaap Plateau and Transvaal Basin carbonate platforms (Fig. 7).

Asbestos Hills Subgroup and Penge Formation

Whereas in the Transvaal Basin banded iron-formation (BIF) is confined to the Penge Formation, there are three successive BIF units in the Asbestos Hills Subgroup of the Griqualand West Basin. Zircons from stilpnomelane layers (former tuffites) at the base of the BIF in both basins (Kuruman and Penge Formations) have been dated at ~2500 Ma (single-zircon U-Pb SHRIMP technique), and thus the beginning of BIF sedimentation is contemporaneous (Trendall *et al.*, 1995).

In Griqualand West the **Kliphuis Formation** is transitional from the Campbell Rand carbonates, cherts and shales to the Kuruman BIF proper. It comprises an intercalation of sideritic shales and sideritic and haematitic cherts with a fairly uniform thickness of 8–13 m over the entire basin. The Kliphuis was formerly regarded as the basal member of the Kuruman Formation (Beukes, 1980b).

The **Kuruman Formation** displays full BIF macrocycles (Fig. 8), each beginning with a bed of stilpnomelane lutite, followed by a whitish chert and then by sideritic and haematitic chert of the so-called siderite-silicate facies rhythmite. Further upward this facies becomes more magnetite-rich, changing into a colour siderite-magnetite (Beukes, 1980b, c) facies rhythmite, until the oxide facies rhythmite, in which magnetite is the main Fe-mineral in chert, is reached. Another whitish, Fe-poor chert completes the macrocycle and is covered by a stilpnomelane bed, starting the next macrocycle (Beukes, 1980b, c). The thickness of individual macrocycles can vary from one metre to several tens of metres.

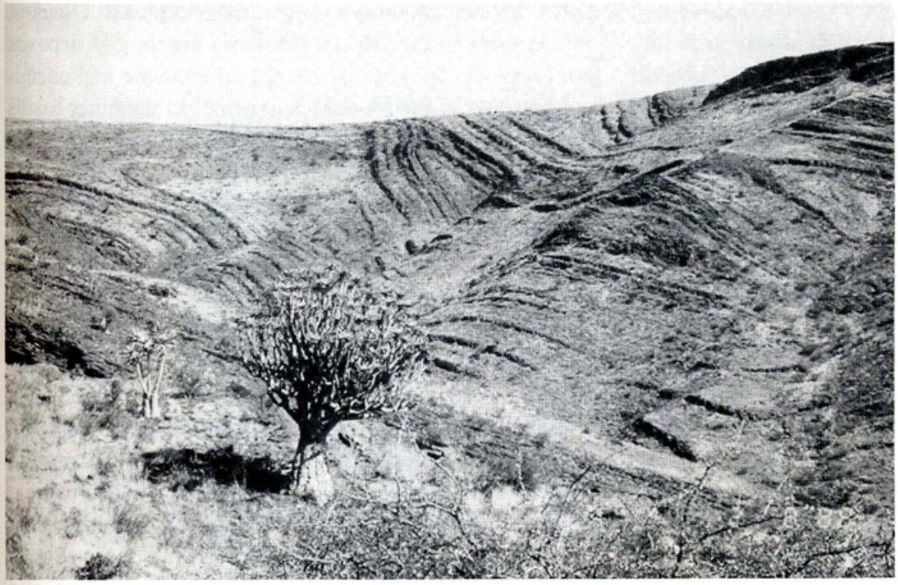

Fig. 8 *Outcrops of folded BIF at Koegas (Prieska sub-basin), with preserved macrocycles delineated by resistant chert layers. East-verging to recumbent folds are here thrust over less deformed, open folds.*

The Kuruman Formation has been drilled at the Westerberg/Koegas asbestos deposits (Prieska sub-basin) where it reaches a thickness of 750 m. Several members were recognised in the drill core, but are difficult to follow up in outcrops, where such a thick pile of BIF can nowhere be reconstructed. North of the Griquatown Fault (Ghaap Plateau sub-basin), the Kuruman BIF is maximally 250 m thick. The thickness difference between the two sub-basins was explained by Beukes (1980b) in a depositional model in which the Prieska sub-basin accumulated thicker BIF sediments, via stronger subsidence and for a longer period of time. This model thus implies that BIFs and carbonates represent laterally equivalent facies in the pelagic Prieska sub-basin and the carbonate platform of the Ghaap Plateau sub-basin respectively (compare also Klein and Beukes, 1989). Altermann and Hälbich (1990, 1991), on the other hand, proposed that the thickness increase of the Kuruman BIF in the Prieska sub-basin is rather due to the much stronger folding and thrust faulting in this region, and that it is thus not of primary depositional but of structural origin.

The **Daniëlskuil Formation**, overlying the Kuruman Formation, is regarded as a reworked (by currents or waves) Kuruman-type BIF. It is rather granular and not as finely laminated as the underlying Kuruman BIF, is rich in BIF intraclasts, and displays abundant sedimentary structures. Chemically, however, it is very similar to the Kuruman BIF (Beukes and Klein, 1990; Horstman and Hälbich, 1995). It is some 200 m thick and, together with the Kuruman Formation, is interpreted as a shallowing-upward sequence, consisting, in turn, of lower-order cycles described in detail by Beukes (1983) (compare also Klein and Beukes, 1989; Beukes and Klein, 1990).

The **Penge Formation** in the Transvaal Basin correlates lithologically and time-stratigraphically with the Kuruman Formation. The Penge Formation is in part thermally metamorphosed by the Bushveld Complex and exhibits concomitant mineralogical variations (Miyano *et al.*, 1987). It is capped by an early Proterozoic erosion surface and is, in places, unconformably covered by the siliciclastic Duitschland Formation. Whether the Duitschland Formation correlates with the Daniëlskuil BIF or with sediments of the Koegas Subgroup is not known; it is, however, incorporated in the Chuniespoort Group (SACS, 1980).

Genesis of BIF

Banded iron-formation lacks actualistic equivalents and its inferred genesis is thus controversial (Trendall and Morris, 1983; Appel and La Berge, 1987). Rare earth and isotope geochemistry point towards a hydrothermal, oceanic source of the iron in the Asbestos Hills BIF (Klein and Beukes, 1989; Beukes and Klein, 1990; Beukes *et al.*, 1990). The sedimentary environment is, however, somewhat enigmatic, because of the problematic lack of sedimentary structures, enormous lateral continuity of the macrocycles and laminations in the BIF, the huge amount of silica of uncertain origin and questionable mode of precipitation, the peculiar presence of diagenetic massive riebeckite lutite, and the conspicuous absence of siliciclastic sediments in the BIF basins. None of the proposed genetic models can satisfy all of these observations.

Button (1976a) explained BIF sedimentation as an end member of a chemical sedimentary cycle starting with carbonates, followed by sideritic ironstones and cherts, and ending with silicic ironstones. Iron precipitation was seen as occurring in a barred basin or lagoon in the most marginal zones of the basin by Eriksson *et al.* (1976).

Beukes' (1980b, 1983) sedimentary model is governed by the interpretation of the northwest-southeast-trending Griquatown Fault Zone as a synsedimentary growth fault, separating the basinal facies in the southwest from platform environments northeast of this fault. This model was further developed by Klein and Beukes (1989) and Beukes *et al.* (1990) into a facies transition model of carbonate deposition in shallow-marine environments, advancing through shales, siderite-, magnetite- and finally, haematite-BIF sedimentation with progressive deepening of the basin. The precipitation of iron-formation was explained as having been triggered along a chemocline separating oxygenated near-surface ocean water from upwelling deep water with a hydrothermal Fe^{2+} source.

Hälbich *et al.* (1992, 1993) proposed that BIF deposition occurred in an intracratonic, gradually shallowing basin with mixed sea-water/fresh-water conditions. Weathering of Ventersdorp Supergroup volcanic rocks and fluvial transport were assumed to be the source of iron and silica in the basin. A biogenetic model for BIF deposition was proposed by Klemm (1979). Klemm (2000) subsequently proposed a non-biogenetic, evaporative model for BIF deposition in an anaerobic, shallow-marine environment. In this model, silica is precipitated, supported by an increase of fO_2 (oxygen fugacity) and oxidation of Fe^{2+}, while Eh-pH changes control haematite and magnetite generation.

Because most workers agree today that BIF deposition reflects a transgression that drowned the Campbell Rand and Malmani carbonates (Beukes, 1980b; Altermann and Nelson, 1998), the models of Button (1976a) and Hälbich *et al.* (1992, 1993) are no longer tenable. The interpretation of BIF deposition is further complicated by the Maremane Dome structure (Fig. 2) in northern Griqualand West, where the Kuruman BIF has a setting that is distinct from the rest of the basin. Because of updoming and karst development in the underlying carbonates around this dome during BIF deposition, the ferruginous **Wolhaarkop Formation** (chert breccia), which grades upwards into a distorted, laminated and haematitic **Manganore Formation**, was deposited in localised depressions and sinkholes (Plehwe-Leisen and Klemm, 1995). This unique setting led to the development of the rich Sishen-type iron ores. Massive riebeckite lutite beds in the Kuruman and Daniëlskuil Formations are recrystallised to crocidolite (blue asbestos) along shear zones and in open fold hinges and were intensively mined, mainly in the Koegas/Westerberg region (Fig. 2). Silicification of the asbestos fibres during Phanerozoic weathering periods led to the formation of mineable tiger's eye deposits.

Koegas Subgroup and Duitschland Formation

The post-BIF sedimentation in Griqualand West indicates further shallowing-upward conditions (Beukes, 1983). Only in the Prieska sub-basin are the BIF deposits conformably covered by mixed siliciclastic and chemical deposits of the Koegas Subgroup. In the other basins a regional unconformity cuts into the BIFs. The Koegas Subgroup embraces five formations, ranging from 6 to 360 m in thickness in their type area and consisting of shales, siltstones and quartzites with some stromatolitic carbonates and thin jasperoidal iron-formation. The **Pannetjie Formation** (50 m) comprises shoreline to deltaic sediments and is overlain by the transgressive iron-formation of the **Doradale Formation** (6 m), followed by prograding deltaic to tidal flat siliciclastics (**Kameelfontein** and **Naragas Formations**, 80 m and 360 m thick respectively). The **Heynskop Formation** (~100 m) consists of jaspilite, banded iron-formation, mudrock, quartzite and stromatolitic dolomite. According to Beukes (1983), iron-formation units in the Koegas Subgroup (Doradale Formation and Rooinekke and Nelani Members of the Heynskop Formation) each form the base of an upward-coarsening BIF (distal) to siliciclastic (proximal) progradational sedimentary cycle. Stromatolitic carbonates within the Koegas Subgroup were suggested to be of lacustrine origin (summarised in Eriksson *et al.*, 1995).

The **Duitschland Formation** (?700 m thick) of the Transvaal Basin may correlate with the Koegas Subgroup. It consists of carbonaceous mudrocks, limestones and dolomites, with subordinate conglomerates, diamictites and lavas, which were interpreted by Clendenin (1989) as a final, shallow, regressive facies of the Malmani-Penge epeiric sea. However, in contrast to the Koegas Subgroup, the Duitschland unconformably overlies the BIF of the Penge Formation (Taussig and Maiden, 1986).

PRETORIA AND POSTMASBURG GROUPS

The Pretoria Group is approximately 6–7 km thick and comprises predominant mudrocks alternating with quartzitic sandstones, significant interbedded basaltic-andesitic lavas, and subordinate conglomerates, diamictites and carbonate rocks (Fig. 9), all of which have been subjected to low-grade metamorphism. A general sheet-like geometry is evident for most of the nine lower formations, with certain sandstone and lava units exhibiting more wedge-like three-dimensional forms (Fig. 10). Erosion has removed many formations in the south of the basin, and the Magaliesberg Formation has also been affected by the intrusion of the Bushveld Complex in many areas. In the eastern part of the basin, five mudrock and sandstone formations overlie the Magaliesberg arenites, and are correlated with single formations in the central (Rayton Formation) and far western (Woodlands Formation – Key, 1983) regions

FORMATIONS	Lithology	WESTERN AREA	CENTRAL AREA	EASTERN AREA	SOUTHERN AREA	Inferred palaeoenvironments
Houtenbek	Mudrock (tuffaceous in places), sandstone, limestone	Rayton/Woodlands Formation in far west (mudrock, sandstone) ≤200 m	Rayton Formation (mudrock, sandstone, minor andesite and dolomite) ~1200 m	150–200 m	Absent	Fan, fan-delta, delta, shallow lacustrine
Steenkampsberg	Sandstone			450–600 m		
Nederhorst	Sandstone (arkosic in places) Mudrock (tuffaceous in places)			200–800 m		
Lakenvalei	Sandstone			200–350 m		
Vermont	Mudrock (tuffaceous in places)			500–700 m		
Magaliesberg	Sandstone with mudrock lenses and interbeds	150–430 m, significant mudrock, sandstones thicken westwards and eastwards	260–340 m, subordinate mudrocks thicken westwards	~225–550 m, subordinate mudrock	≤340 m, mostly eroded	Regressive sandy shoreline, braid-delta, high-energy tidal flat
Silverton	Carbonate rocks Lydenburg Shale Member (commonly tuffaceous) Machadodorp Volcanic Member (pyroclastic rocks, basalt) Boven Shale Member	~500–1328 m, reworked tuffs common, thins westwards, uppermost carbonates ~117–167 m thick, Machadodorp Member absent, basal shale generally thin	~450–850 m, Machadodorp Member 1–2 m thick, upper shales thin	~1040–2230 m, lower shales generally thin, Machadodorp Member ~57–517 m thick	≤1365 m, Machadodorp Member thin (≤6 m), mostly eroded	Relatively deep-water, transgressive epeiric sea; volcanic activity, mainly in the east
Daspoort	Sandstone, mudrock	~65–120 m, sandstone pebbly in far west	~40–110 m, pebbly sandstone common	~10–120 m, sandstone pebbly, thicker in north, ironstones in northeast	~45–80 m, sandstone pebbly	Distal fan, fluvial braid-plain, braid-delta; transgressive epeiric sea in the east
Strubenkop	Mudrock, subordinate sandstone	~50–360 m, minor sandstone	~100–150 m, significant sandstone, minor tuff	~30–145 m, thickens to north and south	~80–185 m, thickens southwards	Transgressive lacustrine
Dwaalheuwel	Sandstone, conglomerate, subordinate mudrock	~15–70 m, basal conglomerate in north	≤3–4 m, lenticular, absent in places	~40–110 m, minor conglomerates in north	Absent	Alluvial fan, fan-delta
Hekpoort	Basaltic andesite, pyroclastic rocks	~190–890 m, thins northwards	~340–630 m, air-fall and reworked pyroclastics relatively common	~90–500 m, thins northwards, pyroclastics common	~430–1140, significant tuffs (200–300 m thick), thickens southwards	Volcanic
Boshoek	Sandstone, conglomerate, diamictite	~35–70 m, significant conglomerates	≤2 m, mostly absent	~20–80 m, large channels	~30–60 m, localised diamictite	Alluvial fan, slump deposits
Timeball Hill	Upper mudrock unit Diamictite/conglomerate/arkose lens	Mudrock 200–430 m, thickens westwards	Mudrock 130–350 m, thick lens of diamictite/conglomerate	Mudrock ~225–750 m, thickens northwards, thick arkose/diamictite lenses in north and northeast	Mudrock ~130–300 m	Relatively deep lacustrine (with suspension sedimentation and turbidity currents), distal fluvio-deltaic, basal volcanism in south and southwest
	Klapperkop Quartzite Member	Quartzite ~90–620 m, thickens westwards	Quartzite ~40	Quartzite ~70–230 m, thins southwards	Quartzite ~40–100 m, thins southwards	
	Lower mudrock unit	Mudrock 160–460 m, thickens westwards	Mudrock ~220–350 m	Mudrock ~300–580 m, thins to south, thin tuff bed	Mudrock ~80–540 m, thickens southwards	
	Bushy Bend Lava Member	Minor basal lavas			Bushy Bend Member ≤90 m	
Rooihoogte	Polo Ground Quartzite Member Mudrock, subordinate carbonate rocks Bevets Conglomerate/Breccia Member	~17–232 m, basal conglomerate thick in north, shale thick in south	~10–50 m, breccia and conglomerate lenticular, Polo Ground Member thin	≤~2–140 m, thickest in Dennilton and Marble Hall fragments	~14–150 m, thick breccia	Karst-fill, alluvial fan, lacustrine
Chuniespoort Group	Iron-formation, dolomite	Palaeokarst topography				

Fig. 9 *Schematic summary profile for the Pretoria Group, illustrating variation in lithology and thickness across the Transvaal Basin, as well as inferred depositional conditions.*

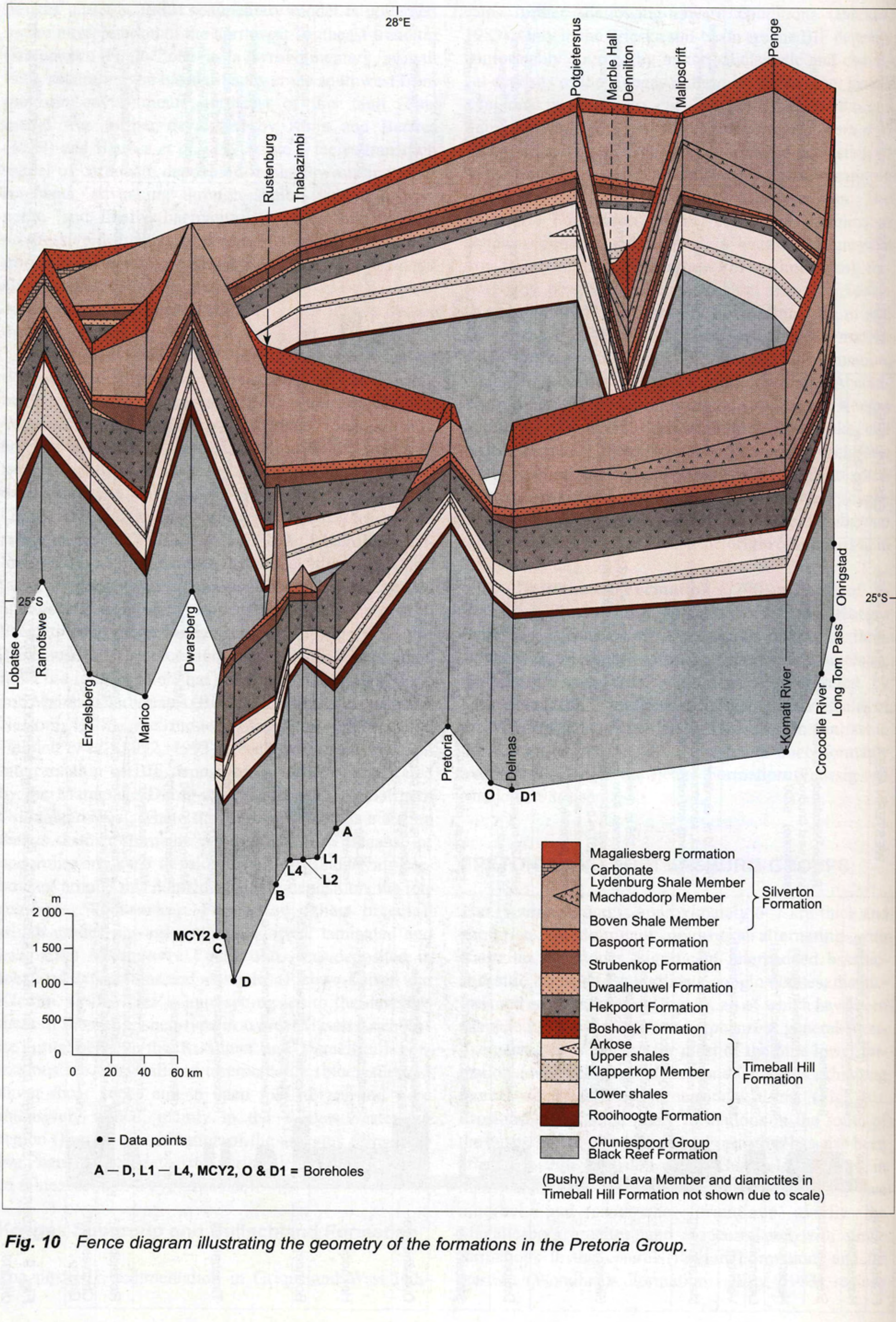

Fig. 10 Fence diagram illustrating the geometry of the formations in the Pretoria Group.

(Fig. 9 and Table 1). In the northeast of the preserved basin, intrusive Bushveld rocks have cut down from the uppermost Houtenbek Formation to the Silverton Formation (Button, 1976b). Isolated outcrops of correlatable Pretoria Group formations overlying Chuniespoort carbonates and BIF are to be found in the Rooiberg, Crocodile River, Stavoren, Marble Hall and Dennilton inliers (Hartzer, 1994). Only one radiometric age is available for the Pretoria Group, with lavas of the Hekpoort Formation being dated at 2224 ± 21 Ma (Rb-Sr whole rock, Burger and Coertze, 1973–74).

The Postmasburg Group in the Griqualand West Basin, up to 1.5 km thick, may be correlated with the lower part of the Pretoria Group (Table 1) (Eriksson *et al.*, 1993a). The basal diamictites of the **Makganyene Formation** cut deeply down into the underlying Ghaap Group (Beukes, 1986; Altermann and Hälbich, 1991) and are thought to be of glacial origin (Fig. 11) (Visser, 1971; De Villiers and Visser, 1977). Analogous, possibly glacial diamictites and associated wackes occur in the Timeball Hill Formation of the Pretoria Group (Visser, 1969). Extrusion of tholeiitic basaltic-andesitic lavas of the **Ongeluk Formation** disconformably over the Makganyene Formation (Altermann and Hälbich, 1991) at 2222 ± 13 Ma (Cornell *et al.*, 1996), formed part of a major Hekpoort-Ongeluk flood-basalt volcanic event (Reczko *et al.*, 1995b), which also occurred in the Kanye Basin of Botswana (Fig. 1). Pillow lavas (Grobler and Botha, 1976), as well as hyaloclastites and massive flows, support subaqueous extrusion for the middle and upper part of the Ongeluk lavas (Cornell and Schütte, 1995). On the other hand, the basal flows of the Ongeluk Formation exhibit abundant pipe amygdales and flow structures, indicating subaerial extrusion over the Makganyene Formation. The correlated Hekpoort Formation lavas appear to have been largely subaerial (Oberholzer, 1995). The Ongeluk lavas are overlain by the jaspillites and inferred volcanic-exhalative manganese deposits of the **Hotazel Formation**. Clastic-textured and stromatolitic dolomites of the **Mooidraai Formation** (Beukes, 1986) complete the restricted succession of rocks in the Postmasburg Group (compared with the Pretoria Group). Preliminary whole-rock Pb-Pb dating of dolomite in the adjacent Mooidraai Formation (2394 ± 26 Ma) by Bau *et al.* (1999) suggest the possibility that the Ongeluk-Hekpoort lavas may be significantly older than previously thought.

The basal **Rooihoogte Formation** of the Pretoria Group overlies a deeply weathered karstic palaeotopography developed on the Chuniespoort Group carbonates. Manganese-rich debris, known locally as "wad", fills palaeosinkholes in many areas. A basal chert breccia, often reworked to form a chert conglomerate, is known as the Bevets Member. This member is best developed in the northwest of the basin, where up to 250 m of conglomerate occurs. Chert breccias are best developed in the Pretoria region, with chert-rich sandstones being more characteristic in the east of the basin (Button, 1973a; Eriksson, 1988). Thick mudrocks, succeeded by thin immature sandstones of the Polo Ground Member, form the upper part of the Rooihoogte succession in the west of the basin (Eriksson, 1988).

In many areas the different facies in this unit exhibit highly variable vertical and lateral relationships; in the Dennilton and Marble Hall inliers, dolomites are also found. The conglomerates and sandstones are thought to reflect a combination of alluvial fan and fluvial braid-plain sedimentation (Schreiber, 1990; Eriksson *et al.*, 1991, 1993a). The mudrocks and dolomites probably formed in shallow lacustrine basins (Eriksson *et al.*, 1995). An alternative marine-transgression model (Button, 1973a) is not supported by the general immaturity of these rocks, nor by their variable vertical and lateral facies arrangements.

The thick shales and subordinate sandstones of the succeeding **Timeball Hill Formation** are generally ascribed to a fluvio-deltaic basin-fill sedimentation system (Eriksson, 1973). A basal black shale facies, associated basin-wide in the subsurface with lavas and pyroclastic rocks of the Bushy Bend Lava Member (Eriksson *et al.*, 1994), reflects suspension sedimentation and fumarolic eruptions. Succeeding rhythmically interbedded mudstones/siltstones and fine-grained sandstones are interpreted as turbidite deposits (Kuenen, 1963; Button, 1973a). These grade up into the medial Klapperkop Quartzite Member (Fig. 9), interpreted as fluvio-deltaic sandstones which fed the more distal turbidites (Eriksson *et al.*, 1993a); tidal reworking of these sands is also inferred (Visser, 1969; Button, 1973a). An upper shale member repeats the black shale and rhythmite facies, and also includes a more proximal, inferred fluvial arkose wedge in the Mokopane (Potgietersrus) area, and diamictites and associated wackes near Pretoria (Visser, 1969; Button, 1973a). The latter probably reflect slumping of periglacial detritus (Visser, 1969), which was possibly coeval with the Makganyene diamictites in Griqualand West. Upper Timeball Hill shales in the east of the basin have suffered extensive soft-sediment deformation (Button, 1973a) due to the onset of tectonic instability, which subsequently led to the succeeding immature fan deposits of the Boshoek Formation and the flood basalts of the Hekpoort Formation.

The **Boshoek Formation** comprises two lobes of immature conglomerates, sandstones and subordinate mudrocks which entered the basin from the northeast and the northwest and were laid down on distal fans and sandy fluvial braid-plains; a shallow lacustrine central basin is also inferred (Schreiber *et al.*, 1990). An irregular palaeotopography and disconformities are locally developed below the Boshoek Formation, which displays a lenticular geometry throughout the Transvaal Basin.

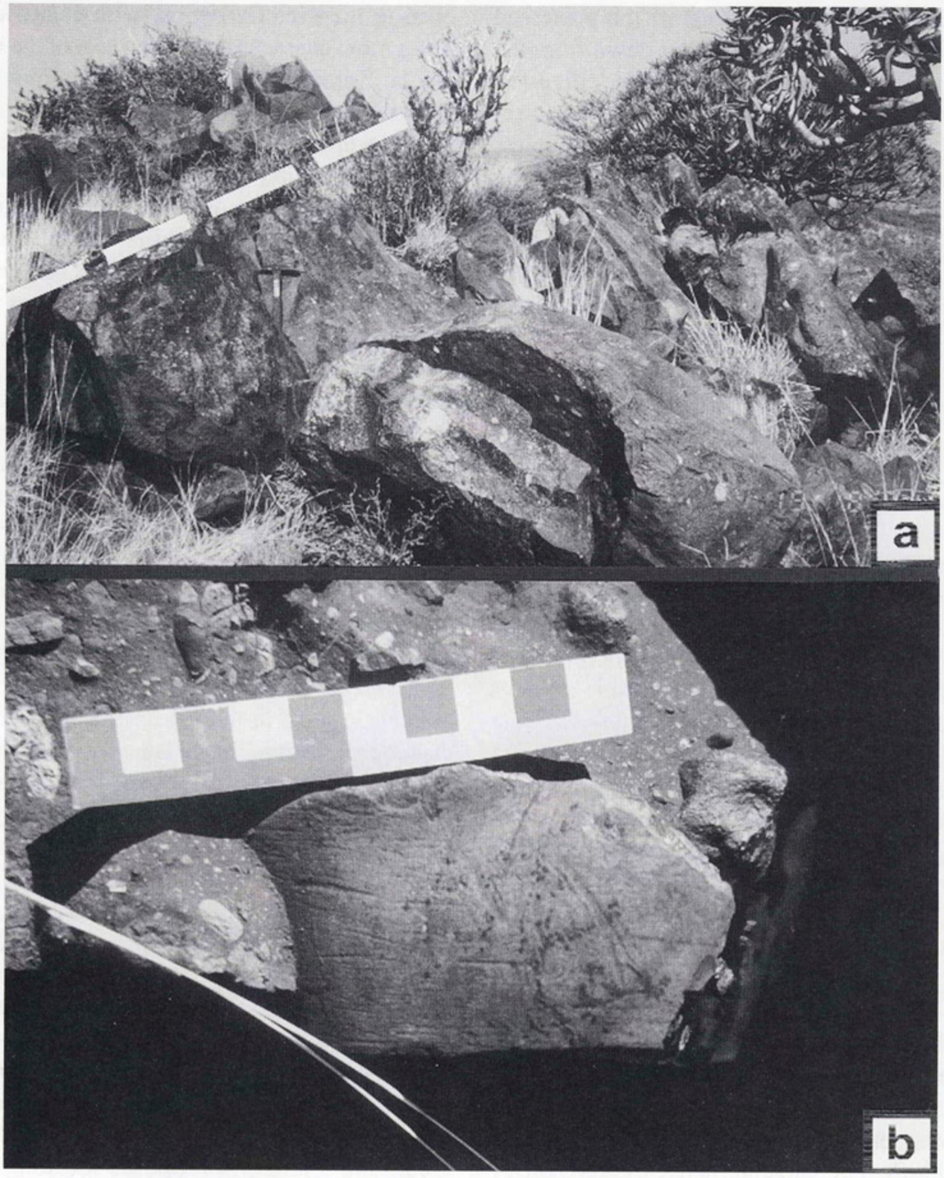

***Fig. 11** (a) The Makganyene Formation tillite, just below the Ongeluk Formation lavas (above the dashed white line) on the farm Libertas, at Koegas. (b) Striated clast in the Makganyene Formation diamictites, north of Koegas. Such striations are abundant in these sediments and, together with the various lithofacies, clearly indicate glaciation. The scale is 10 cm long.*

The subaerial volcanism of the **Hekpoort Formation** began in the south, with the first lavas extruding during deposition of the Boshoek sandstones (Reczko *et al.*, 1995b). In excess of 1100 m of basaltic-andesitic lava is preserved in the south, with over 800 m in the west of the basin, but these rocks thin to the northeast where less than 50 m is found. Subaerial fissure eruptions appear to have dominated, with locally important pyroclastic systems (Oberholzer, 1995). Intermittent hiatuses in volcanism were marked by small lacustrine shale deposits. An uppermost, widespread palaeosol postulated by Button (1973a) has been disputed by more recent workers (Engelbrecht, 1986; Oberholzer, 1995).

Immature, predominantly sandy distal fan and fluvial braid-plain deposits of the **Dwaalheuwel Formation** overlie the Hekpoort Formation (Schreiber and Eriksson, 1992). Two lobate sedimentation systems entered the basin, one from the northeast and the other from the northwest. Impersistent, thin sandstones are present in the central portions of the basin and the unit is absent in the south (Fig. 9) (Eriksson *et al.*, 1995).

The succeeding **Strubenkop Formation** comprises alternating mudstones and siltstones, with subordinate interbedded, immature, fine-grained sandstones. The succession is generally upward-coarsening and pyrite/limonite speckling is common (Button, 1973a; Schreiber, 1990; Van der Neut, 1990). The Strubenkop is interpreted as the lateral facies equivalent of the Dwaalheuwel sandstones, probably reflecting lacustrine conditions which prograded over the Dwaalheuwel braid-plain deposits (Eriksson *et al.*, 1991, 1993a), although other researchers have seen it as being a shallow-marine deposit (Button, 1973a).

The **Daspoort Formation** overlies the Strubenkop shales unconformably (Eriksson *et al.*, 1993b). Although characterised by mature quartz arenites and subordinate mudrocks and ironstones in the east of the basin (Button, 1973a), elsewhere it predominantly comprises more immature sandstones, pebbly arenites, conglomerates and mudrocks (Eriksson *et al.*, 1993b). The latter authors discussed the merits of alternative shallow-marine and fluvial models for these rocks. The Daspoort Formation probably reflects both settings, and is inferred to mark the beginning of a major marine transgression, responsible for the deposition of the succeeding Silverton Formation shales and Magaliesberg Formation sandstones (Eriksson *et al.*, 1995). This scenario is supported by the local occurrence of thin stromatolitic carbonates and cherts at the top of the Daspoort, usually passing into a condensed, transgressive dolomite or chert, before being covered by shales of the Silverton Formation.

The lower Boven Shale Member of the **Silverton Formation** is best developed in the eastern part of the basin, and is characterised by high-alumina shales. The upper Lydenburg Shale Member is thick across much of the basin, being characterised by tuffaceous, high CaO-MgO-MnO shales (Reczko, 1994). A medial Machadodorp Volcanic Member in the east of the basin comprises tholeiitic basalts (Reczko *et al.*, 1995b), which appear to have been shallow subaqueous eruptives, associated with subordinate carbonate rocks (Button, 1973a). The Silverton Formation may be seen as a highstand facies tract, reflecting the advance of an epeiric sea onto the Kaapvaal Craton from the east; the underlying Daspoort Formation may, in this context, represent a lowstand facies tract or a transgressive systems tract.

The **Magaliesberg Formation**, which gradationally overlies the Silverton shales, has generally been interpreted as a shallow-marine sandstone, due to its supposedly mature petrography and common tidal and other shallow-water structures (Visser, 1969; Button, 1973a). More recently, Eriksson *et al.* (1995) have demonstrated a more immature character of these sandstones, and have argued that they represent a regressive shoreline with a combination of braid-deltas and high-energy tidal channels. The Magaliesberg sandy coastline would thus represent a regressive systems tract, reflecting the retreat of the Silverton epeiric sea off the Kaapvaal Craton.

The post-Magaliesberg formations (Fig. 9) comprise an alternation of quartzitic sandstones and shales, with subordinate carbonate rocks, tuffs and lavas (Schreiber, 1990; Van der Neut, 1990). In the eastern part of the basin the succession comprises (from the base up) the **Vermont Formation** (mudrock), the **Lakenvalei Formation** (sandstone), the **Nederhorst Formation** (mudrock, overlain by sandstone in the south), the **Steenkampsberg Formation** (sandstone) and the **Houtenbek Formation** (mudrock, sandstone and limestone). The equivalent strata in the central and western parts of the basin have not been subdivided and are known as the **Rayton** and **Woodlands Formations** respectively. These units have been variously interpreted: Button (1973a) and Button and Vos (1977) suggested a shallow-marine setting for these rocks, while Schreiber (1990) and Schreiber and Eriksson (1992) discussed the alternative of a closed basin, with tidal flat, lacustrine and fan-delta/delta depositional systems. The generally immature feldspathic nature of many of the post-Magaliesberg sandstones is, perhaps, more supportive of a closed basin and tectonically unstable conditions (Eriksson and Reczko, 1995).

Post-Magaliesberg sedimentary rocks in the Stavoren (Makeckaan) inlier were assigned by Hartzer (1994) to the **Makeckaan Subgroup**, comprising four formations which he correlated with the Nederhorst, Steenkampsberg and Houtenbek Formations in the eastern Transvaal Basin as well as the subdivisions of the Rayton Formation. Immature conglomerates and strongly feldspathic sandstones of the **Leeuwpoort Formation** in the Rooiberg inlier are overlain by quartzite, shale and andesitic lava of the Smelterskop Formation (Stear, 1977); the latter unit has been correlated with the Dullstroom Formation and included in the Rooiberg Group by Hartzer (1994). Overall fluvial sedimentation is generally inferred for most of these rocks (Stear, 1977; Richards and Eriksson, 1988). Hartzer (1994) correlated the Leeuwpoort succession with the Nederhorst-Houtenbek succession further east.

There has been a lively debate for many years whether the Pretoria Group was a shallow-marine basin or a closed continental (non-marine) depository (see Eriksson *et al.*, 1991, 1993a, for a summary of views). In the absence of invertebrate fossils, these two alternatives are very difficult to assess, but qualified use of boron contents indicates the distinct possibility that the Rooihoogte to Strubenkop Formations were laid down in a closed basin, succeeded by transgressive/regressive marine sediments of the Daspoort, Silverton and Magaliesberg Formations (Eriksson and Reczko, 1995). The tectonic setting of the Pretoria Group basin is inferred to lie in the rift-to-intracratonic-sag-type continuum (Reczko *et al.*, 1995b).

ROOIBERG GROUP

The volcanic Rooiberg Group overlies the Pretoria Group unconformably (Table 1) and is the uppermost unit of the Transvaal Supergroup in the Transvaal Basin according to SACS (1980). However, since it is closely associated with the igneous activity represented by the Bushveld Complex, many believe that it should form part of this unit rather than the Transvaal Supergroup. In view of its uncertain stratigraphic status the Rooiberg Group is described in a separate chapter (P.C. Buchanan, this volume).

MINERAL DEPOSITS

Gold

Sporadic palaeoplacer gold was mined in the past from the Black Reef Formation (e.g., Söhnge, 1986). The Sabi-Pilgrim's Rest goldfield in Mpumalanga had produced almost 185 Mt of ore at an average grade of *c.* 8 g/t by 1988 (Meyer, 1988), and is still producing. These quartz-gold lodes are associated with pyritic carbonaceous mudrocks and cross-cutting veins within the Malmani Subgroup dolomites; hydrothermal remobilisation of gold from the host mudrocks is envisaged (Meyer, 1988).

Dolomite and limestone

Metallurgical dolomite is mined in the Pretoria and Potchefstroom areas from the Malmani Subgroup. Limestone is also mined locally, such as at Lime Acres in the Northern Cape (Altermann and Wotherspoon, 1995).

Base metals

Mississippi Valley-type Pb-Zn-F deposits are fairly common within the dolomitic succession of the Campbell Rand and Malmani Subgroups, such as those at Pering, Bushy Park, Genadendal, Leeuwbosch and within the Zeerust fluorspar field (see review by Martini *et al.*, 1995). They comprise a mixture of stratiform, pipe-like, ring-shaped and irregular ore bodies, that most likely formed during the karstic weathering period separating Chuniespoort-Ghaap chemical sediment deposition and the clastic sedimentation of the succeeding Pretoria-Postmasburg Groups (Martini *et al.*, 1995). These authors concluded that heated brines circulating through pre-Transvaal rocks leached metals under conditions of tectonic tension, and that ore minerals were deposited within karstic solution cavities once mixing with surficial waters occurred. Within the Pretoria Group there is potential for stratiform sulphide deposits of the sedimentary exhalative type, with the lower Timeball Hill Formation, in particular, exhibiting chemical enrichment in base metals, Ba and, locally, even tuff-hosted PGE contents (Reczko *et al.*, 1995a).

Iron

Large deposits of iron ore are mined in the Thabazimbi area in Limpopo Province and in the Sishen ore field in the Northern Cape. From 1933 to 1958 Thabazimbi was South Africa's main supplier, and until 1993 had produced about 113 Mt of high-grade ore (Hälbich *et al.*, 1993); production continues there, with 2.4 Mt being produced in 1994 (Vermaak, 1997). Supergene enrichment of these ores within the Penge Formation has enhanced ore values above primary chemical sedimentary concentration (Van Deventer *et al.*, 1986). Since 1958 the giant Sishen Mine has been the country's main producer, with an annual production of 20.8 Mt in 1994 (Vermaak, 1997), the ore being obtained from the Kuruman and overlying Daniëlskuil Formations in the Asbestos Hills Subgroup. Here, the iron ores generally consist of hard, high-grade material, and conglomerate as well as massive and laminated types are found (Hälbich *et al.*, 1993).

Manganese

The Kalahari (north) and Postmasburg (south) Manganese Fields in the Northern Cape are located in the Hotazel and Gamagara Formations, where well-known mines include the Mamatwan, Black Rock and Wessels Mines. At about 9 billion tons of Mn metal with >68% Mn, it is the world's largest manganese resource (Kleyenstüber, 1984; Grobbelaar and Beukes, 1986; Grobbelaar *et al.*, 1995), with an estimated 81.4% of global reserves (Vermaak, 1997). Models inferred for this enormous deposit include a sedimentary origin involving transgression-regression across a shallow, epeiric, marine shelf related to an oceanic redox boundary (e.g., Beukes, 1983), as well as a hydrothermal volcanic exhalative model related to the underlying Ongeluk Formation rocks (e.g., Cornell and Schütte, 1995).

Asbestos and andalusite

Metamorphic ores from the Transvaal Supergroup include large reserves of asbestos minerals in banded iron-formation of the Chuniespoort and Ghaap Groups, with the largest deposits being in Mpumalanga where about 2.3 Mt occurs (Beukes and Dreyer, 1986), and *c.* 35 Mt of high-grade andalusite, largely in feldspathic mudrocks of the Timeball Hill Formation in the aureole of the Bushveld Complex (Hammerbeck, 1986).

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