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The Transvaal Sequence: an overview

P. G. Eriksson, J. K. Schweitzer[1], P. J. A. Bosch[2],
U. M. Schereiber, J. L. Van Deventer[3] and C. J. Hatton[4]

Dept of Geology, University of Pretoria, Pretoria 0002, South Africa
[1]COMRO, P. O. Box 91230, Auckland Park 2006, South Africa
[2]Geological Survey of South Africa, Private Bag X112, Pretoria 0001, South Africa
[3]ISCOR, Private Bag X534, Thabazimbi 0380, South Africa
[4]AARL, P. O. Box 106, Crown Mines 2025, South Africa

Abstract - The 15 000 m of relatively unmetamorphosed clastic and chemical sedimentary and volcanic rocks of the 2550-2050 Ma Transvaal Sequence as preserved within the Transvaal and correlated Griqualand West basins of South Africa, and in the Kanye basin of Botswana are described. Immature clastic sedimentary and largely andesitic volcanic rocks of the Wolkberg, Godwan and Buffelsfontein Groups and the Bloempoort and Wachteenbeetje Formations probably represent rift-related sequences of Ventersdorp age. The thin sandstones of the Black Reef Formation, developed at the base of both the Kanye and Transvaal basin successions and correlated with the basal Vryburg siltstones of the Griqualand West Sequence, are considered here to be the basal unit of the Transvaal Sequence. The Black Reef fluvial deposits grade up into the epeiric marine carbonates of the Malmani Subgroup. These stromatolitic dolomites and interbedded cherts were laid down within a steepened carbonate ramp setting; transgressions from an initial Griqualand West compartment towards the northeast covered both the Kanye and Transvaal basins. Iron formations of the succeeding Penge Formation and Griqualand West correlates are envisaged as relatively shallow water shelf deposits within the carbonate platform model; siliceous breccias of the Kanye basin are interpreted as reflecting subaerial brecciation of exposed silica gels. The Duitschland Formation overlying the Penge iron formations is seen as a final, regressive clastic and chemical sedimentary deposit, as the Malmani-Penge sea retreated from the Transvaal basin.

The interbedded sandstones and mudstones of the unconformity-bounded Pretoria Group probably represent a combination of alluvial fan and fluviodeltaic complexes debouching into the largely lacustrine Transvaal and Kanye basins. A strong glacial influence in the lower Pretoria Group is reflected in the correlated Makganyene diamictites of the Griqualand West Sequence. Sedimentation across all three basins was interrupted by the extrusion of the Hekpoort-Ongeluk andesites. Upper Pretoria Group sediments of the Silverton and Magaliesberg Formations probably reflect a marine transgression. These rocks are not present in the Griqualand West basin, and were affected by Bushveld Complex-related thermal doming in the Transvaal basin; post-Magaliesberg sedimentation continued thereafter in separate eastern and western fluviodeltaic-lacustrine sub-basins.

The largely volcanic Rooiberg Group (sensu lato) began with catastrophic basin floor collapse and Leeuwpoort Formation fluvial sedimentation in the western sub-basin. The succeeding Smelterskop and Makeckaan Formations reflect a transition from fluvial deposition to volcanism, and are succeeded by the widespread and voluminous, predominantly felsitic lavas of the Dullstroom, Damwal and Selonsrivier Formations. The correlated Loskop, Glentig and Rust de Winter Formations which overlie the felsites conformably, represent the final sedimentary phase of the Transvaal basin.

INTRODUCTION

The Late Archaean-Early Proterozoic Transvaal Sequence provides one of the largest and best preserved examples of rocks from this geological time period in the world. In addition, researchers working on these rocks postulate that early plate tectonic processes may have played a role in their deposition, as did magmatic events leading up to the intrusion of the Bushveld Complex. The Transvaal Sequence also encompasses a time period when significant changes may have taken place in the composition of the terrestrial atmosphere.

The Transvaal Sequence is preserved within a main Transvaal or Bushveld basin within South Africa and southeastern Botswana and within the much smaller Kanye basin (Crockett, 1972) of Botswana (Fig. 1). Both basins show a degree of correlation with the Griqualand West Sequence of South Africa (Table 1), particularly in the case of

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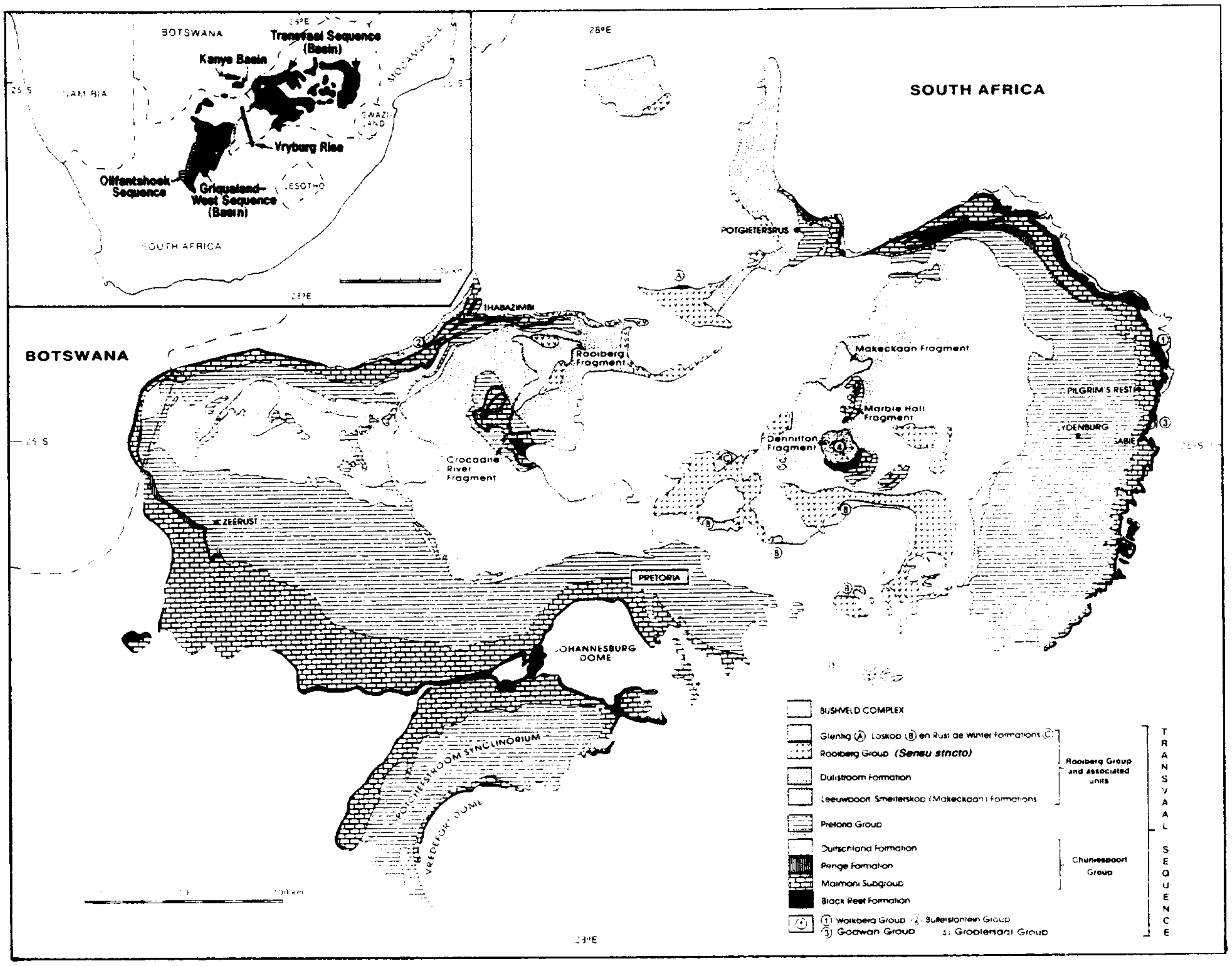


Fig. 1. Geological map of the Transvaal Sequence of the Transvaal-Bushveld basin, showing the distribution of the principal stratigraphic units. Inset map illustrates the location of the Transvaal-Bushveld, Kanye and Griqualand West basins. Also shown are the Bushveld Complex which intrudes the Transvaal rocks, the Vryburg rise between the Transvaal and Griqualand West basins, and isolated occurrences of Transvaal rocks surrounded by Bushveld intrusives at the so-called Rooiberg, Crocodile River, Dennilton, Marble Hall and Makeckaan

the chemical sedimentary rocks in the lower portions of the two sequences. This has led certain reseachers to use the term Transvaal Supergroup to include both sequences (for example, Beukes, 1986), an erroneous usage, as the lack of any contact between the Transvaal and Griqualand West basins precludes the use of an overall term according to normal stratigraphic terminology (SACS, 1980). The stratigraphic subdivision of the Transvaal Sequence adopted by the South African Committee for Stratigraphy (SACS) in 1980 is modified in this paper (Table 2), reflecting research carried out within the last decade. This review will concentrate on the succession within the Transvaal-Bushveld basin, and compare these rocks to those in the Griqualand West and Kanye basins.

The Transvaal Sequence in the Bushveld basin comprises up to 15 000 m (Button, 1986) of relatively undeformed and low grade metamorphosed mudrocks, sandstones, volcanics, dolomites and iron formations (Table 2). A basal clastic sedimentary and volcanic unit (Wolkberg Group and correlates) is found only within parts of the Bushveld basin, and is succeeded by a very widespread development of the Black Reef Formation clastic sedimentary rocks and the overlying chemical sedimentary unit, found in both Bushveld and Kanye basins as well as in the Griqualand West Sequence (Table 1). The overlying clastic sedimentary/volcanic unit (Pretoria Group of the Bushveld basin) is poorly represented in the Griqualand West Sequence and incompletely preserved within the Kanye basin (Table 1). The uppermost, largely volcanic unit is restricted to the Bushveld basin.

Modern research (Clendenin *et al.*, 1988b; Cheney *et al.*, 1990) has suggested that the pre-Black Reef Formation clastic sedimentary/volcanic rocks (Table 2) are correlates of the Late Archaean Ventersdorp Supergroup (Fig. 2), which has an age of approximately 2700 Ma (Armstrong *et al.*, 1991).

Table 1 Correlation of Griqualand West, Kanye & Transvaal/Bushveld Basins

Overall lithology	Griqualand West Sequence (Basin)		Kanye Basin (Transvaal Sequence) (Botswana)		TRANSVAAL SEQUENCE (Transvaal or Bushveld Basin): S.E. Botswana (Bushveld basin)	TRANSVAAL SEQUENCE (Transvaal or Bushveld Basin): South Africa (Transvaal basin)	
Volcanic & clastic sedimentary unit						Loskop clastics & volcanics Damwal/Selonsrivier Dullstroom lavas/Smelterskop & Makeckaan clastics (lavas) Leeuwpoort sandstones	Rooiberg Group (sensu lato)
Clastic sediments & volcanics					Woodlands volcanics/clastics	Rayton sandstones & shales & volcanics	Pretoria Group
	Postmasburg Group	Mooiwater dolomites & Manganiferous Hotazel ironstones			Sengoma sandstones	Magaliesberg sandstones	Pretoria Group
	Postmasburg Group				Sengoma shales	Silverton shales	Pretoria Group
	Postmasburg Group		Segwagwa Group	Mogapinyana & Gatsepane chert, sandstone & shale	Ditlhojana sandstones/conglomerates	Daspoort sandstones	Pretoria Group
	Postmasburg Group		Segwagwa Group			Strubenkop shales	Pretoria Group
	Postmasburg Group		Segwagwa Group			Dwaalheuwel/Droogedal sandstones	Pretoria Group
	Postmasburg Group	Ongeluk andesites	Segwagwa Group	Tsatsu andesites	Ditlhojana volcanics	Hekpoort andesite	Pretoria Group
	Postmasburg Group	Makganyene diamictites	Segwagwa Group	Ditlhojana & Tlaameng shales, sandstones & conglomerates	Ditlhojana shales [conglomerates / shales]	Boshoek conglomerates/sandstones	Pretoria Group
					Tsokwane sandstones	shales / sandstone / shales (Timeball Hill)	Pretoria Group
					Lephala [shales / conglomerates/sandstones]	Rooihoogte conglomerates/ sandstones	Pretoria Group
Regional unconformity							
Chemical sedimentary unit	Ghaap Group: Koegas Subgroup; Asbesheuwels Subgroup	Griquatown & Kuruman iron formations	Taupone Group	Masoke ironstone/ferruginous chert	Ramotswa shale	Duitschland carbonates, clastics	Chuniespoort Group
	Ghaap Group		Taupone Group			Penge iron formations	Chuniespoort Group
	Ghaap Group		Taupone Group	Kgwakgwe chert breccia			Chuniespoort Group
	Ghaap Group: Campbellrand Sub-grp	Campbellrand dolomites	Taupone Group	Ramonnedi dolomites	Magopane, Maholobota, Ramotswa – dolomites	Malmani dolomites	Chuniespoort Group
	Ghaap Group: Schmidtsdrif Subgrp	Lokamonna Formation (shale) Boomplaas Form. (dolomite)	Taupone Group				Chuniespoort Group
Clastic sediments	Ghaap Group: Schmidtsdrif Subgrp	Vryburg Formation		Black Reef Formation	Black Reef Formation	Black Reef Formation	
Clastic sediments & volcanics (Ventersdorp age)						Pre-Black Reef units - Wolkberg Group & correlates	

Table 2. Lithostratigraphy and geochronology of the Transvaal/Bushveld Basin

Sequence & Groups		Formation	Age (Ma)	Lithology	Stratigraphic correlation
TRANSVAAL SEQUENCE	Rooiberg (sensu lato)	Loskop/Glentig/Rust de Winter	2060 (U-Pb)	Mudrocks/sandstones/lavas	
		Selonsrivier Damwal	<2090	Felsites (minor sandstones & mudrocks)	
		Dullstroom / Smelterskop Leeuwpoort / (Makeckaan)	<2089 ± 15 (Rb-Sr)	Mafic & felsic lavas/ / Sandstones/mudrocks/ clastic sediments / lavas	Dullstroom thought to be basal part of Rooiberg Group
	Pretoria	Rayton/Woodlands: Houtenbek, Steenkampsberg, Nederhorst, Lakenvlei, Vermont		Mudrocks/sandstones/lavas/pyroclastic rocks/carbonate rocks	Pyroclastic Woodlands Formation of Botswana correlated with Rayton Formation of central Transvaal and five formations of eastern Transvaal
		Magaliesberg		Sandstones	
		Silverton		Mudrocks/volcanic rocks/carbonate rocks	
		Daspoort		Sandstones	
		Strubenkop		Mudrocks/sandstones	
		Dwaalheuwel/Droogedal		Sandstones/conclomerates	
		Hekpoort	2224 ± 21 (Rb-Sr)	Basaltic andesites	
		Boshoek		Conglomerates/sandstones	
		Timeball Hill	2263 (Rb-Sr)	Mudrocks/sandstones/minor tilloids	
		Rooihoogte		Breccias/conglomerates/sandstones/mudrocks	
	Chuniespoort	Duitschland		Mudrocks/carbonate rocks/minor volcanic rocks	Major unconformity and time gap of ± 150Ma
		Penge	2432 ± 31 (U-Pb SHRIMP)	Iron formations	
	Chuniespoort (Malmani Subgroup)	Frisco Eccles Lyttelton Monte Christo		Dolomites/chert/minor shales & sandstones	
		Oaktree	2557 ± 49 (U-Pb)		
		Black Reef	——	Sandstones/conglomerate	Probable base of Transvaal Sequence
Pre-Transvaal	Pre- Black Reef units	Buffelsfontein Group	Ages are unreliable due to weathered lavas & metamorphic resetting. Wolkberg probably about 2700 Ma.	Sandstones/mafic & felsic lavas/mudrocks	Provisionally considered to be equivalents of Ventersdorp Supergroup
		Godwan Group		Sandstones/lavas	
		Wachteenbeetje Fm.		Mudrocks/carbonate rocks/sandstones	
		Wolkberg Group		Mudrocks/sandstones/basalts	
		Upper Groblersal Group		Sandstones/lavas	
		Lower Groblersdal Group		Metamorphic rocks	Most likely pre-Wolkberg basement material

Age data from Burger and Walraven (1980), Beukes (1987), Jahn *et al.* (1990), Trendall *et al.* (1990) and Harmer and Von Gruenewaldt (1991).

This view is supported by the present authors and we thus consider the basal unit of the Transvaal Sequence to be the Black Reef Formation (Table 2). The Black Reef sandstones grade up into the basal carbonate rocks of the Oaktree Formation (Table 2); rocks at the equivalent stratigraphic level in the Griqualand West basin are dated at 2557 ±49 Ma (U-Pb) (Jahn *et al.*, 1990). Ages determined at various stratigraphic levels within the Transvaal Sequence (Table 2) culminate in an age of 2060 Ma (U-Pb) for lavas in the uppermost Rust de Winter Formation (Burger and Walraven, 1980). Walraven *et al.* (1990) have dated the intrusion of the mafic phase of the Bushveld Complex, which truncated the uppermost formations of the Pretoria Group and the Dullstroom lavas (Button, 1973, 1976), at 2061 ±27 Ma. An age constraint of approximately 2600/2500 - 2050 Ma may thus be placed on the Transvaal Sequence.

The intrusion of the Bushveld Complex detached the Rooiberg Group lavas and overlying Loskop Formation sedimentary rocks from the underlying Pretoria Group. Metamorphism of the Rooiberg felsites, which now form the roof of the complex, led to partial melting of the lavas and to the formation of leptites (French and Twist, 1983). Contact metamorphic effects on the Pretoria Group floor rocks included recrystallisation of sandstones to form quartzites, the formation of hornfelses or even partial melting and plastic flow of mudrocks (Button, 1986). The metamorphic aureole of the Bushveld Complex in the western outcrops of the Pretoria Group includes rocks belonging to the albite-epidote-hornfels, horn-

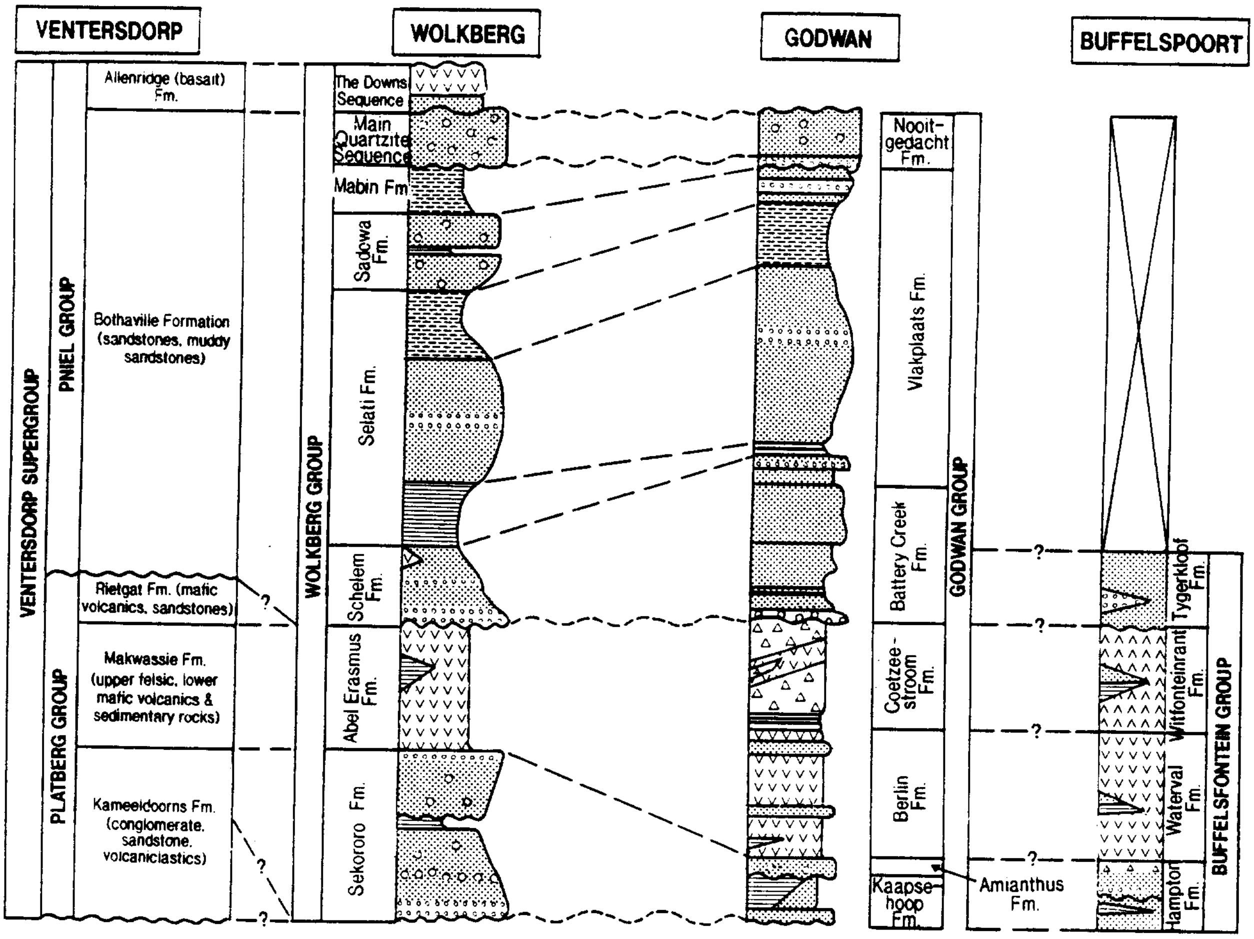


Fig. 2. Lithostratigraphy and correlation of the pre-Black Reef Formation units of the Transvaal Sequence. Note the proposed correlation of these units with the Ventersdorp Supergroup. Modified after Tyler (1979), Cheney *et al.* (1990), Myers (1990) and Bosch (1992).

blende-hornfels and pyroxene-hornfels facies (Engelbrecht, 1986). Cross-cutting Bushveld intrusives in the northeastern Transvaal basin formed metamorphic minerals in the Chuniespoort Group chemical sediments (Button, 1976, 1986). Bushveld related diabase sills commonly intrude the Pretoria Group strata, particularly the Silverton and succeeding formations (Table 2) (Button, 1976; Sharpe, 1981).

The Transvaal strata dip at angles up to 20° towards the centrally placed Bushveld intrusives (Fig. 1). Locally, diapiric structures are found within Transvaal sediments along the contact zone (Button, 1978, 1986) and complex deformation of the sequence occurred along the northern margin of the preserved basin, and in the south around the Vredefort Dome (Button, 1986). Isolated occurrences of Transvaal rocks surrounded by Bushveld intrusives, known as "fragments", are found at Rooiberg, Crocodile River, Dennilton, Marble Hall and Makeckaan (Fig. 1). These are ascribed either to pre-Bushveld updoming of the basin floor (Button, 1986; Hartzer, 1987), or may be related to a late Pretoria Group synsedimentary thermally induced palaeohigh (Eriksson *et al.*, 1990).

PRE-BLACK REEF FORMATION UNITS

Highly metamorphosed gneisses, schists and metavolcanics of the lower Dennilton Formation of the Groblersdal Group (Fig. 1) are inferred by the present authors to represent basement material to the pre-Black Reef units discussed here. The U-Pb age of 2460 Ma (Coertze *et al.*, 1978) obtained on rocks of this group is interpreted to be a reset metamorphic age. The overlying sedimentary and volcanic rocks of the Bloempoort formation are most likely correlates of the pre-Black Reef units (Hartzer, 1987). Both the 2600 m thick Wolkberg and 1700 m thick Buffelsfontein Groups are thickest over palaeovalleys in the earlier Archaean floor rocks and wedge out against basement highs (Button, 1973; Tyler, 1978; SACS, 1980). These successions and the correlated Godwan Group (Myers, 1990) most likely represent Ventersdorp-age, largely rift-controlled sedimentation and volcanism (Clendenin *et al.*, 1991) (Fig. 2), preserved around the present day northern and eastern margins of the succeeding Transvaal basin (Fig. 1). The 800 m of calcareous mudrocks, sandstones and dolomites of the Wachteenbeetje Formation underlying the Crocodile River fragment in the west-centre of the Transvaal basin (Fig. 1) are ascribed to deeper basinal conditions (Hartzer, 1987), possibly related to thermal subsidence.

The Wolkberg Group comprises basal immature sandstones and conglomerates of the Sekororo Formation (Fig. 2), laid down by braided stream and subordinate alluvial fan processes (Button, 1973; Bosch, 1992). These are overlain by andesitic lavas of the Abel Erasmus Formation; eruption was largely subaerial, with periodic pillow lavas and aqueous sedimentation indicating intermittent, localised lacustrine environments (Button, 1973; Bosch, 1992). The succeeding Schelem Formation (Fig. 2) is analogous to the Sekororo braided stream deposits; these lower three formations of the Wolkberg Group most likely represent a rifting environment, with the succeeding Selati, Mabin and Sadowa Formations probably being related to thermal subsidence. The latter three units comprise predominant mudrocks, immature and mature sandstones (Fig. 2), which are ascribed to deltaic and marginal marine sedimentation (Button, 1973); the basin may alternatively have been intracratonic (Bosch, 1992). The Wolkberg Group (*sensu stricto*, SACS, 1980) is thickest in the Selati trough of the northeastern Transvaal, where the Sadowa Formation is overlain by a "main quartzite sequence" and the succeeding "The Downs sequence" (Fig. 3), both informal stratigraphic units not yet recognised by SACS (Schwellnus *et al.*, 1962; Clendenin *et al.*, 1991). Fluvially deposited siliciclastics of the "main quartzite sequence" are unconformably overlain by pebbly fluvial sandstones and the Serala Basalt Member of "The Downs sequence" (Fig. 3) (Clendenin *et al.*, 1991). Both of these informal, unconformity-bounded units are considered here to be part of the Wolkberg Group (*sensu lato*) (Schwellnus *et al.*, 1962; Clendenin *et al.*, 1991) (Fig. 2) and formation status is recommended. Clendenin *et al.* (1991) related "The Downs sequence" to thermal subsidence following the rifting which initiated lower Wolkberg deposition.

The Buffelsfontein Group comprises basal immature fluvial sandstones, overlain by up to 1050 m of mafic to rhyolitic volcanics, and uppermost immature pebbly sandstones (Fig. 2) (Tyler, 1978). A similar rift environment to that inferred for the lower Wolkberg Group is thought to have controlled Buffelsfontein deposition. Palaeocurrents for the Wolkberg and Buffelsfontein Groups indicate sediment transport from the northeastern and northwestern Transvaal towards the central Transvaal, where the deeper water basinal deposits of the Wachteenbeetje Formation occur (Tankard *et al.*, 1982; Hartzer, 1987). The Godwan Group, a correlate of the Wolkberg Group in the eastern Transvaal (Fig. 2), consists of analogous immature arenites and medial pyroclastic rocks (Myers, 1990), with a maximum thickness of about 1500 m (SACS, 1980); a similar fluvial-subaerial volcanic palaeoenvironment is envisaged.

BLACK REEF FORMATION

The Black Reef Formation, here regarded as the basal unit of the Transvaal Sequence, comprises a thin veneer of arenaceous rocks, which unconformably overlie earlier lithologies, including the Wolkberg Group (Fig. 3) (Button, 1973; Clendenin *et al.*, 1991). The formation is preserved around the present day margins of the Transvaal basin (Fig. 1). Thicknesses are mostly between a few metres and 30 m, with a maximum of 60 m being recorded in eastern Botswana (SACS, 1980; Key,

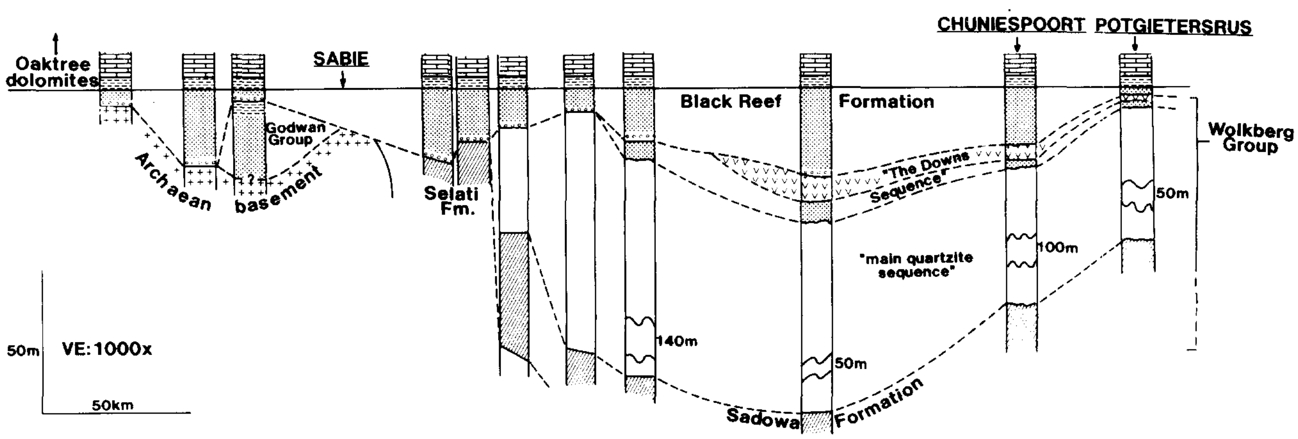


Fig. 3. Detailed lithostratigraphy of the upper part of the Wolkberg Group (sensu lato) illustrating the relationship of the Black Reef Formation with the underlying units. Modified after Clendenin *et al.* (1991).

1986; Henry *et al.*, 1990). In the eastern Transvaal the maximum thickness of 30 m is attained over the pre-existing Selati trough (Fig. 3) (Clendenin *et al.*, 1991). The Black Reef Formation is correlated with the basal Vryburg Formation of the Griqualand West Sequence (Table 1), composed of predominant siltstone, with subordinate mudstone, quartzose sandstone and lava (SACS, 1980).

Recent research on the Black Reef Formation has concentrated on the eastern Transvaal, where Henry *et al.* (1990) identified six lithologies: poorly sorted lenticular clast- and matrix-supported conglomerates, trough and planar cross-bedded sandstones, plane laminated arenites and laminated carbonaceous mudrocks. These facies are stacked in a basal, locally developed, upward-fining sequence, overlain by a sheet-like upward-coarsening facies sequence (Fig. 4). The uppermost mudrocks generally grade into the overlying dolomites of the Malamani Subgroup (Fig. 4) (Button, 1973, 1986). A similar basal conglomerate-predominant arenite-interbedded mudrock association is found in the west of the basin, with the arenites or mudrocks grading up into the overlying dolomites (Key, 1983, 1986).

Button (1973) interpreted the basal Black Reef conglomerates as fluvial deposits, with the succeeding mature quartz arenites being envisaged as a subtidal sheet sand, laid down as an epeiric protobasin developed over the Kaapvaal craton. A similar marginal marine-fluvial model is proposed by Key (1983, 1986). Henry *et al.* (1990) interpreted the lower upward-fining sequence (Fig. 4) as having been deposited in a sandy braided fluvial setting, locally channelised, analogous to Button's (1973) view. However, the upper upward-coarsening sequence (Fig. 4) is ascribed to a prograding braid-delta or braid-plain environment (Henry *et al.*,

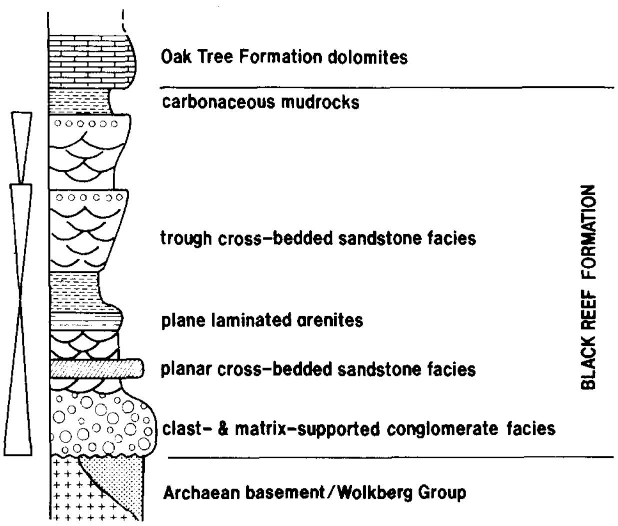


Fig. 4. Facies sequence of the Black Reef Formation in the eastern Transvaal basin. Modified after Henry *et al.* (1990).

1990). The transgressive mudrock (Clendenin, 1989) at the top of the formation (Fig. 4) reflects a tidal flat setting, marking the establishment of the epeiric marine carbonate platform of the Chuniespoort sea. Henry *et al.* (1991) ascribed the subtle unconformity at the base of the Black Reef Formation to northward tectonic tilting, related to thermal subsidence over the Kaapvaal craton following the Ventersdorp rifting; they saw these events as reflecting the closing stages of Ventersdorp age sedimentation, rather than being the protobasin deposits to the Chuniespoort sea, as envisaged by Button (1973) and Key (1983).

MALMANI SUBGROUP

The dolomites and interbedded cherts and mudrocks making up the Malmani Subgroup represent epeiric marine deposits which were deposited over a very large portion of the Kaapvaal craton, stretching well beyong the preserved outcrops within the Transvaal, Griqualand West and Kanye basins (Fig. 1). The widespread nature of the Chuniespoort Group epeiric rocks contrasts with the succeeding Pretoria, Postmasburg and Segwagwa Groups, which appear to have been largely restricted to these three basins. The Malmani Subgroup is correlated with the upper portion of the Schmidtsdrif Subgroup and the overlying Campbellrand Subgroup of the Griqualand West Sequence and with the Ramonnedi dolomites of the Kanye basin (Table 1). Button (1973) and SACS (1980) subdivided the Malmani Subgroup into five formations (Fig. 5), based largely on chert contents and the nature of the stromatolites in the rocks. Interbedded chert-in-shale breccias, mudrocks and sandstones are thin and commonly mark the location of marginal unconformities (Clendenin and Maske, 1986). On the basis of these unconformity-bounded sequences, Clendenin (1989) has subdivided the Malmani epeiric sea succession and overlying Penge and Duitschland Formations into five "packages"; the lowermost package is found only in the Griqualand West basin and the uppermost package includes the Penge Formation and correlated Griqualand West iron formations which gradationally overlie the dolomites (Fig. 5). The modified stratigraphic subdivision of Clendenin (1989) makes use of the stratigraphic terminology originally devised by Button (1973) and is supported by the present authors (Fig. 5). The packages are generally thickest in the Griqualand West basin and thin towards both the Transvaal and Kanye basins (Aldiss *et al.*, 1989; SACS, 1980; Clendenin and Maske, 1986). Stromatolites characterise the carbonate-chert sequence of the Malmani Subgroup, with forms varying from crinkly laminations, domical

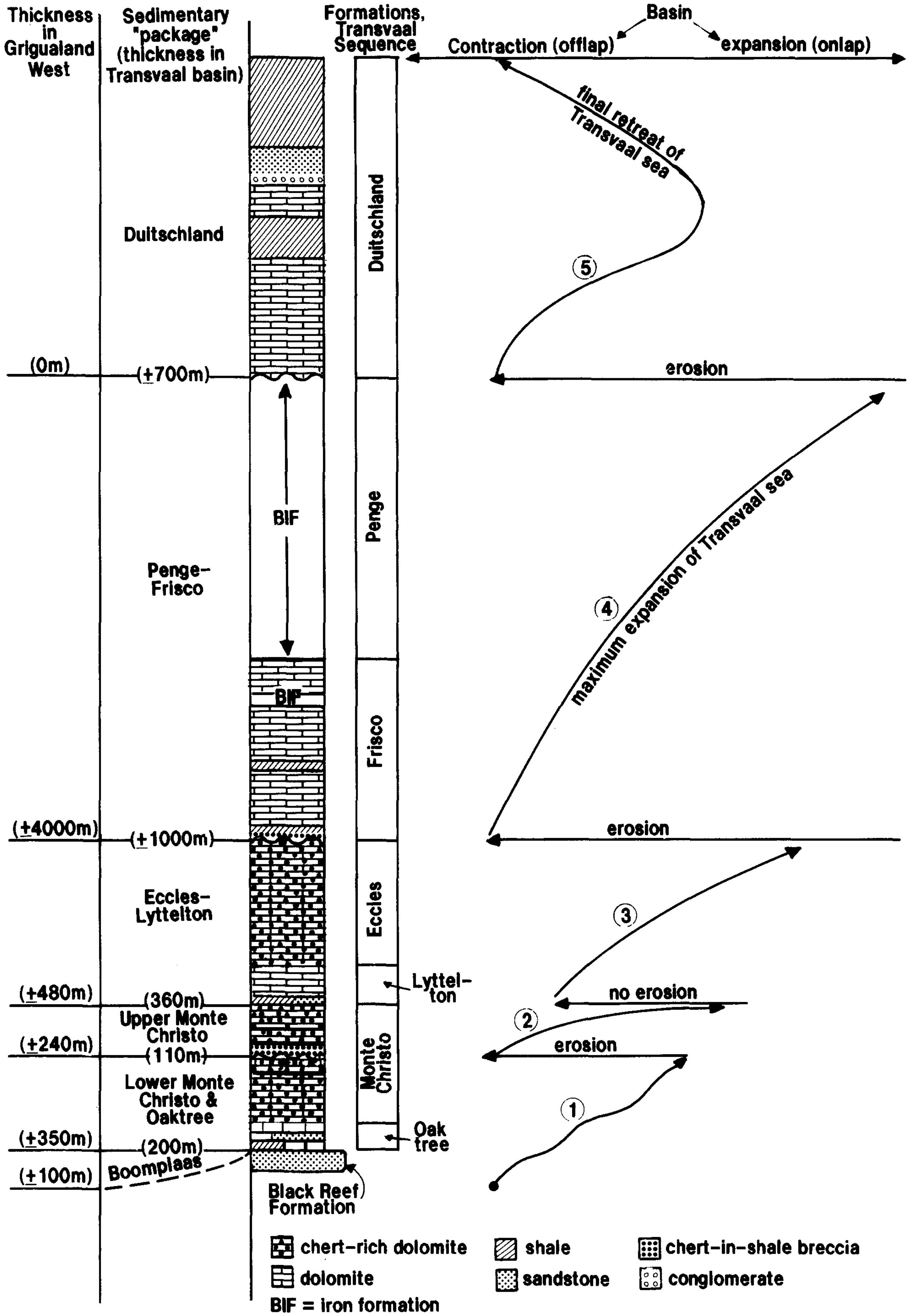


Fig. 5. Lithostratigraphy of the Malmani Subgroup, Penge and Duitschland Formations, Chuniespoort Group, Transvaal Sequence. Also shown are the "sedimentary packages" of Clendenin (1989), their thicknesses in the Transvaal and Griqualand West basins, and the transgressive expansion and regressive contraction of the proposed epeiric sea palaeoenvironment.

and columnar stromatolites to enormous, elongated algal mounds, up to 100 m long and 2 m high (Eriksson and Truswell, 1973). MacGregor *et al.* (1974), Nagy (1984) and Klein *et al.* (1987) successfully isolated algal microstructures in Malmani samples. Mechanically-formed sedimentary structures are common in the dolomites, including numerous ripple marks, oolites, intraclastic breccias and lesser cross-laminations (Button, 1973).

The basic, tidally-influenced palaeoenvironmental model supported by most researchers for the Malmani Subgroup was first suggested by Truswell and Eriksson (1973), Eriksson and Truswell (1974) and Eriksson *et al.* (1975), for a number of isolated localities. These workers proposed a modified version of the limestone shelf model, encompassing supratidal deposits, columnar stromatolites in an intertidal zone, a high energy zone above wave base with oolites and mechanically-formed sedimentary structures, and shallow and deep subtidal zones characterised by large stromatolitic domes and mounds (Fig. 6a). Limestones and dolomites characterise the intertidal, high energy and shallow subtidal zones, with dolomites being found in the deep subtidal facies (Fig. 6a) (Eriksson *et al.*, 1975).

Following upon the earlier work of Eriksson and co-authors, Beukes (1978, 1980, 1986, 1987) applied this basic tidally-influenced model to the carbonate rocks throughout the Transvaal and Griqualand West Sequences. Beukes (1987) differentiated between an initial carbonate ramp setting characterised by giant stromatolitic mounds, and which deepened towards a deep shelf environment in the west of the Kaapvaal craton, and a succeeding mature rimmed carbonate shelf model for the upper portion of the carbonate sequence (Fig. 6c). The latter encompassed eastern supratidal mudflats, passing westwards into broad zones of intertidal mudflats and shelf lagoonal settings, with a deeper euxinic basin in the far west (Fig. 6c) which exhibited turbiditic deposits as well (Beukes, 1987). Geophysical investigations of the western part of the Kaapvaal craton support Beukes' concept of a carbonate basin which was shallower in the east and deeper towards the west (Geerthsen *et al.*, 1991). However, the southwestern portion of the Griqualand West carbonate succession has been subjected to multiple folding and thrusting (Altermann and Hälbich, 1990), and the resultant thickening of sediments may be partly responsible for the observed geophysical trends. Altermann and Herbig (1991) and Hälbich *et al.* (1992) dispute the postulate that the carbonate basin became deeper towards the western margin of the Kaapvaal craton; they provide evidence of supratidal to intertidal flats in the southwestern region, analogous to the shallow water carbonate platform of Beukes (1987) to the east. The deepest portion of the basin is envisaged to have lain between these two shallow tidal platforms, co-incident with the Griquatown fault zone (Altermann and Herbig, 1991). Water depths within the carbonate platform/shelf model are estimated to have been up to 80 m in the euxinic deep basin (Klein *et al.*, 1987); REE chemistry of the carbonates also supports generally shallow water marine conditions, with the possibility of some freshwater mixing having taken place (Danielson, 1990).

Clendenin (1989) and co-workers (Clendenin and Maske, 1986; Clendenin *et al.*, 1988b) refined the early shelf model of Eriksson *et al.* (1975) to include Beukes' (1987) distal shallow basin (Fig. 6b); in addition, Clendenin (1989) related his carbonate ramp/steepened carbonate ramp model to syndepositional tectonics, which formed part of a successor basin sequence within the Kaapvaal craton during the late Archaean (Clendenin *et al.*, 1988a and b). Clendenin's broad facies belt model (Fig. 6b) inferred an arid and/or tropical palaeoclimate, supported by palaeomagnetic data (Windley, 1979), and tides whose height may have increased as open ocean tides encroached onto the wide, shallow carbonate platform (Klein, 1982; Pratt and James, 1986); facies distribution was determined largely by water depth and regressions and transgressions of the vast Transvaal-Griqualand West epeiric sea led to vertical stacking of deposits from the different facies within the carbonate ramp model shown in Figure 6b. Transgressions were predominantly towards the north-northeast, as the depository expanded from an initial Griqualand West compartment (Figs 5 and 6d). At the end of lower Monte Christo times, the sea retreated off the Kaapvaal craton, forming the erosional unconformity developed within this formation (Fig. 5). The north-northeastward transgressions were re-initiated three more times during Malmani deposition (Figs 5 and 6d), developing the sedimentary packages preserved in the rock record (Fig. 5). It should be noted that the final Malmani transgression had the greatest extent and continued on into the period of deposition of the succeeding Penge iron formations (Fig. 5); a final, fifth transgression-regression cycle was responsible for the deposition of the Duitschland Formation, the uppermost unit of the Chuniespoort Group (Fig. 5, Table 2). Dolomitization of primary limestones appears to have taken place shortly after deposition (Eriksson *et al.*, 1975); dolomitization was probably also related to meteoric waters which lowered the pH (Eriksson *et al.*, 1976).

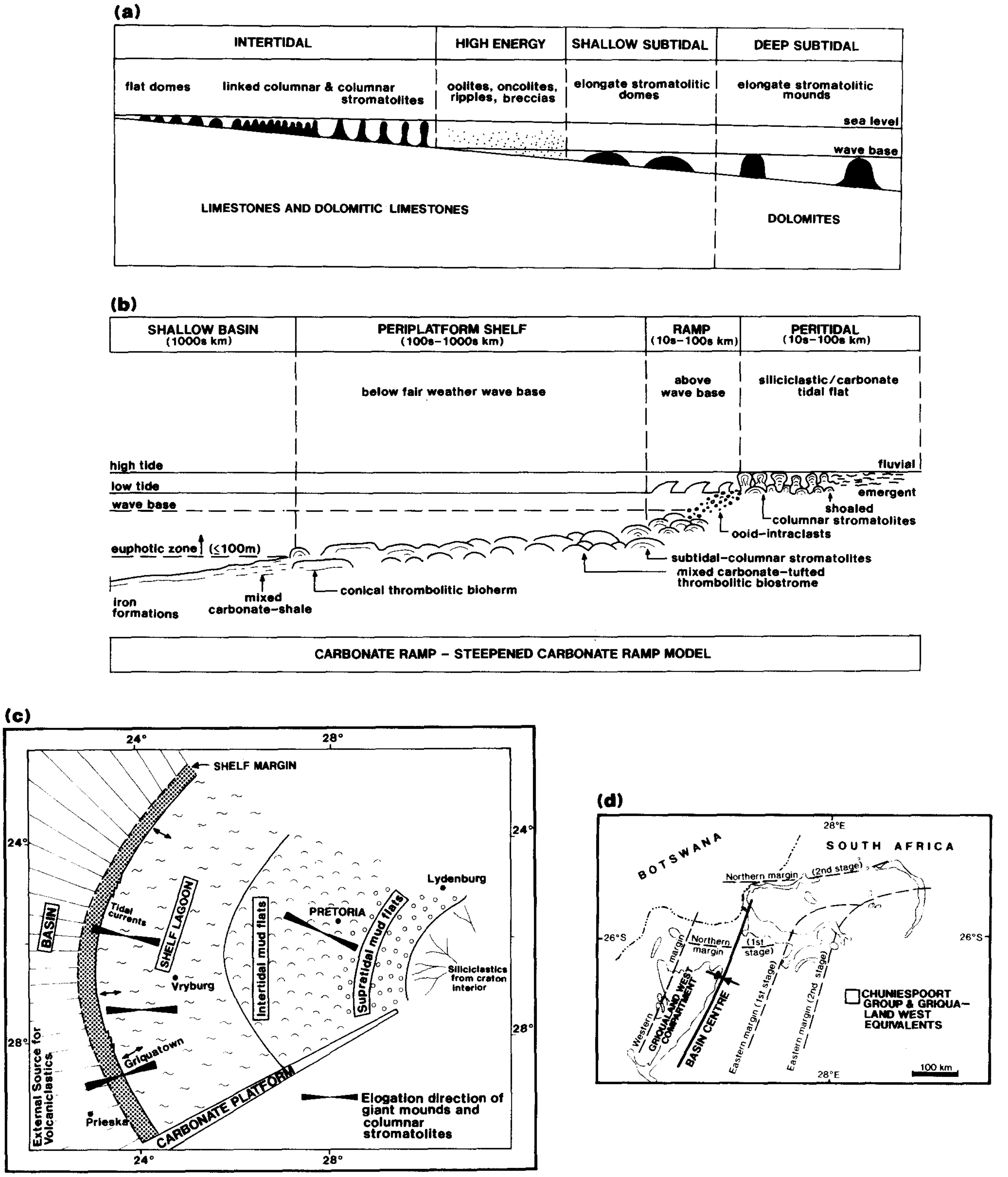


Fig. 6. Depositional models proposed for the Malmani Subgroup dolomites. (a) Limestone shelf model of Eriksson *et al.* (1975). (b) Carbonate ramp-steepened ramp model of Clendenin (1989). (c) Carbonate ramp - deep shelf model of Beukes (1987). d) Clendenin's (1989) proposed expansion of the Malmani epeiric sea from an initial Griqualand West compartment towards the north and northeast.

PENGE AND DUITSCHLAND FORMATIONS

The Penge and correlated iron formations of the Asbesheuwels Subgroup of Griqualand West (Table 1) overlie the Malmani dolomite sequence gradationally. As with the underlying dolomites, these iron formations formed part of a depositional system that covered a large portion of the Kaapvaal craton, including outcrops preserved in the Griqualand West, Transvaal and Kanye basins (Fig. 1).

The Penge Formation is composed predominantly of micro- to macro-banded iron formations with laterally persistent monomineralic or mixed mineral laminae of quartz, magnetite, hematite, stilpnomelane, riebeckite, minnesotaite, grunerite and ferruginous carbonate minerals (Beukes, 1973; Button, 1986; Van Deventer *et al.*, 1986). Stacked cycles of alternating ferruginous minerals are common (Beukes, 1978), with subordinate interbeds of carbonaceous mudrock and intraclastic iron formation breccias (Button, 1986). Shard structures, mostly in the basal Penge Formation (La Berge, 1966) suggest volcanic influences in the formation of these ferruginous lithologies. Bushveld Complex related contact metamorphism of the Penge iron formations has resulted in large scale recrystallisation and gruneritisation (Beukes, 1973, 1978). In the west of the Transvaal basin, there is a gradation into the Ramotswa Shale Formation of Botswana (Table 1), where ferruginous and siliceous mudrocks overlie the dolomites gradationally (Key, 1983). Their ferruginous nature appears to be due to diagenetic weathering of disseminated pyrite (Key, 1983). In the Kanye basin (Fig. 1), chert breccias are found at the stratigraphic level of the Penge Formation (Crockett, 1972) (Table 1); these rocks comprise equidimensional, angular chert and jaspilite fragments in a ferruginous matrix (Crockett, 1972; Aldiss *et al.*, 1989).

Beukes (1978, 1983) postulates that the Griqualand West-Transvaal iron formations were laid down within a clear water epeiric sea, covering much of the Kaapvaal craton, and bounded to the west and north by a deep basinal setting (Fig. 7). An external source of siliciclastic and volcanoclastic material is inferred, with authochthonous iron formations being deposited in the deep basinal areas and orthochemical and allochemical iron formations on the central shallow platform region (Beukes, 1978, 1983) (Fig. 7). This basin may have been fault-controlled (Button, 1973; Beukes, 1977, 1978, 1980). Mechanical reworking of iron formations was widespread on the platform, and the clastic sediments in the Koegas Subgroup (Table 1) may have been laid down within a fresh water lacustrine setting (Beukes, 1986).

The stratigraphic continuity on which Beukes' model relies, is challenged by Altermann and Hälbich (1990), who present clear evidence for early tectonic displacement and karstification of both iron formations and underlying carbonate rocks in the southwest of the craton. Hälbich *et al.* (1992) also demonstrate convincingly that shallow water and fresh water contamination conditions prevailed during genesis of the iron formations in this same southwestern region; in addition they dispute the importance of a volcanic influence in iron formation genesis. The shallow water setting proposed by Hälbich *et al.* (1992) is in agreement with REE patterns determined by Danielson (1990) and with Clendenin's (1989) carbonate ramp model in which iron formations are seen as a deeper water facies equivalent of the platform carbonates (Fig. 6b). Water depths of Clendenin's (1989) euphotic zone (Fig. 6b) were equivalent to those of a shelf setting rather than the deep basin envisaged by Beukes (1983). The gradational basal contact of the Penge Formation with the underlying Malmani dolomites supports a similar depositional setting for these chemical sediments (Eriksson and Clendenin, 1990). Shallow water conditions are also postulated for the ferruginous mudrocks and cherts found in the Bushveld and Kanye basins in Botswana (Table 1). The Ramotswa Formation mudrocks are ascribed to a regressive nearshore back-reef palaeoenvironmental setting and to mudflat deposition (Key, 1983). The Kgwakgwe breccias may reflect either tectonic formation along thrust soles (Rabie, 1958), karstification of chert-rich dolomite (Cullen, 1958), or dehydration of silica gels related to tectonic instability in the northwestern margin of the Penge basin (Crockett, 1972). Tectonic studies do not support the first hypothesis and the petrology of the breccias cannot be explained adequately by the second theory. The most plausible explanation is that tectonic instability of the basin margin led to uplift and subaerial exposure, thereby promoting brecciation of exposed silica gels and the non-deposition of iron formation (Crockett, 1972; Aldiss *et al.*, 1989).

The Penge Formation (and correlates) is seen by Clendenin (1989) as part of a Frisco-Penge "package", laid down as the Transvaal-Griqualand West sea underwent maximum expansion, thereby depositing deeper water iron formations over the Malmani carbonate platform (Fig. 5). This primitive early Proterozoic ocean is inferred to have been the source of both iron and silica, with possible subordinate contributions from external volcanic and terrestrial weathering sources (Beukes, 1983). The carbonaceous mudrocks, limestones and dolomites, with subordinate conglomerates, diamictites and lavas of the Duitschland Formation, which overlies the Penge Formation unconformably (Taussig and Maiden, 1986), are interpreted as a final, shallow, regressive facies of the Malmani-Penge epeiric sea (Fig. 5) (Clendenin, 1989).

THE PRETORIA GROUP

The Pretoria Group overlies the chemical sedimentary rocks of the Chuniespoort Group unconformably; the unconformity is commonly both

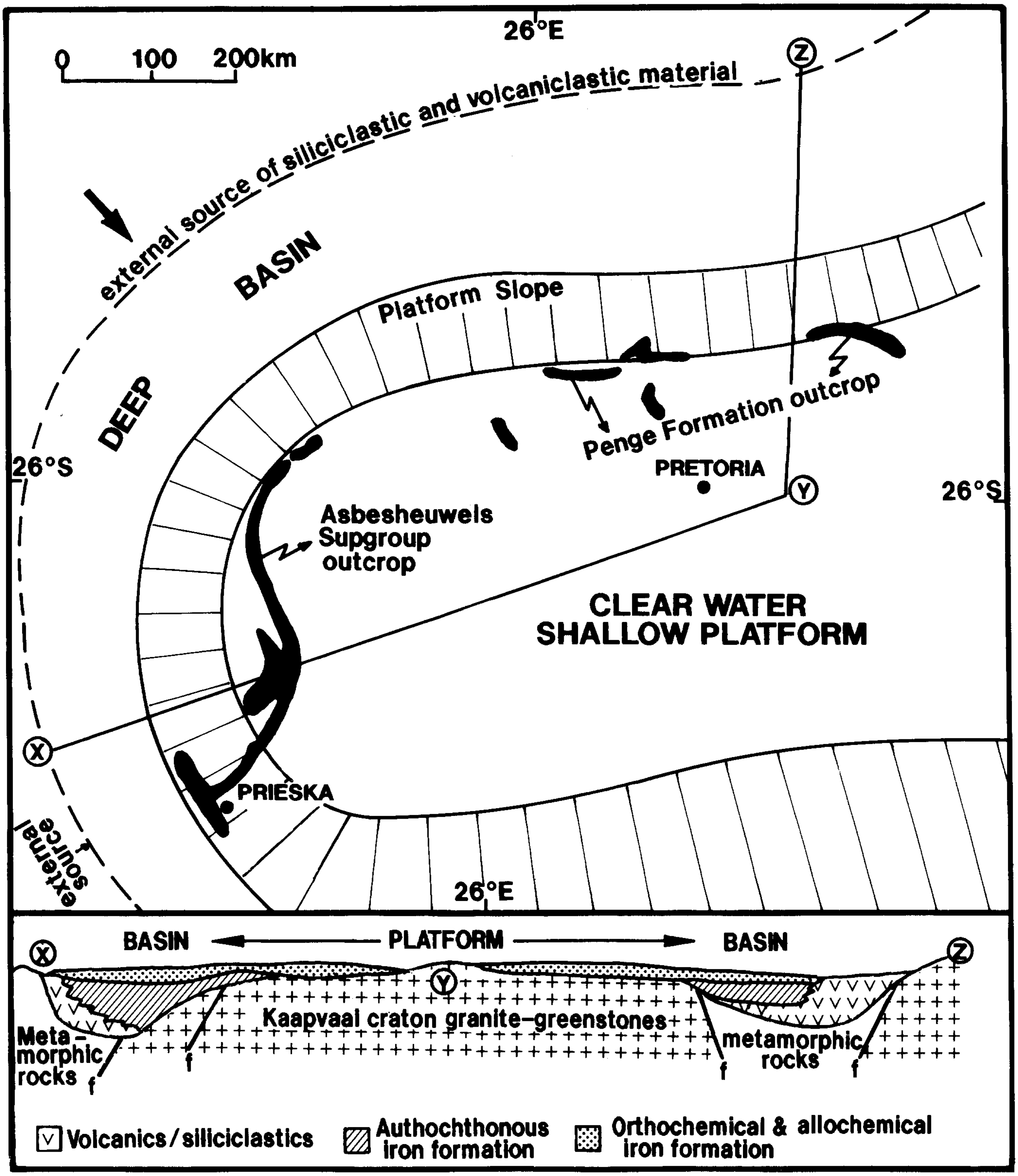
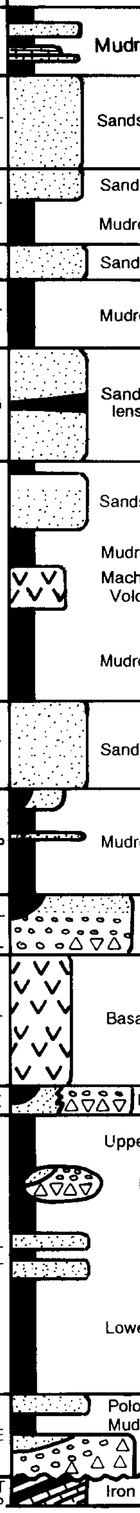


Fig. 7. Clear water epeiric platform - platform slope - deep basin model proposed for the Penge and correlated iron formations of the Transvaal and Griqualand West Sequences, after Beukes (1978, 1983).

angular and karstic (Button, 1973) and radiometric ages (Table 2) suggest a significant hiatus in Transvaal deposition. The lithostratigraphy of the Pretoria Group (Fig. 8) comprises a predominant alternation of mudrocks and sandstones, with less important volcanic horizons and subordinate diamictites and conglomerates. There are significant lithological differences across the basin, particularly in the upper part of the stratigraphy, and thicknesses are also variable (Fig. 8). Correlation of these rocks with the upper portion of the Griqualand West Sequence and with sedimentary rocks of the Kanye basin is poor (Table 1), with only the Hekpoort-Ditlhojana-Tsatsu-Ongeluk andesites



Formations		TRANSVAAL BASIN		
		WEST	CENTRE	EAST
HOUTENBEK	Mudrocks, sandstones, limestones			150-200m
STEENKAMPSBERG	Sandstones	Woodlands Formation in far west - interbedded mudrocks & sandstones, some conglomerates, significant andesitic pyroclastics & lavas. 800-1200m	Rayton Formation- interbedded mudrocks & sandstones, minor andesites & dolomites ±1200m	450-600m, erosive base in north
NEDERHORST	Sandstones Mudrocks			200-800m, sandstones are arkosic
LAKENVLEI	Sandstones			200-350m, Some arkosic sandstones
VERMONT	Mudrocks			500-700m
MAGALIESBERG	Sandstones (mudrock lenses & interbeds)	120-300m, significant mudrocks. Sandstones thicken to west & to east, wedge out into Silverton mudrocks	±300n Subordinate mudrocks thicken to west	±500m, subordinate mudrocks
SILVERTON	Sandstone lens Mudrocks Machadodorp Volcanic Member Mudrocks	13km x 150m sandstone lens present. ≤100m dolomites in west at top of fomation. Minor reworked tuffs. Total thickness : 400-800m & thins to west.	No significant sandstones Thin pyroclastic & dolomitic /chert member Total thickness ± 600m	Upper mudrocks ±1700m & thin to north Machadodorp Member ± 500m Lower mudrocks ±250m
DASPOORT	Sandstones	130-200m, pebbly sandstones & mudrocks in far west, thickens to west, locally Δ	±80m , pebbly sandstones common, locally ∇	0-±100m, pebbly sandstones, mudrocks & thicker in north. Ironstones & Fe-mudrocks in N.E.
STRUBENKOP	Mudrocks, lesser sandstones	100-135m, minor sandstones	±110m, significant sandstones, locally ∇	± 30-100m, thickens to south
DWAALHEUWEL or DROOGEDAL	Diamictite, conglomerate, sandstone	Droogedal - 10-50m	Absent or very thin	Dwaalheuwel - 50-100m, thins to south
HEKPOORT	Basaltic andesite	510-600ı	±400m. Air fall & reworked pyroclastics locally	±25-500m, pyroclastics common, thins to north
BOSHOEK	Diamictite, conglomerate, sandstone	0--10m	0-50m	≤100m Large channels
TIMEBALL HILL	Upper mudrocks Diamictite\conglomerate lens Klapperkop Quartzite Member Lower mudrocks	Upper mudrocks 200-500m, thicken westwards, no diamictite lens Quartzites 5-500m, thicken westwards Lower mudrocks 300-500m, thicken westwards	Upper mudrocks, ±100-200m, thick lens of diamictite/conglomerate Quartzites ±30-70m Lower mudrocks ±230-400m	Upper mudrocks ±400-600m, arkose wedge in north, thin diamicites, deformed mudrocks Quartzite 0-15-100m, thins to South Lower mudrocks 400-700m, thin southwards
ROOIHOOGTE	Polo Ground Quartzite Member Mudrock Bevets Conglomerates (breccia) Member	Polo Ground 0-1 Mudrocks 0-150m Bevets 0-150m	0-10m 0-18-250m 0-150m	Only Bevets Member, 0-30m, locally quartzite
CHUNIESPOORT GROUP	Iron Formation & dolomite	PALAEOKARST TOPOGRAPHY		

Fig. 8. Lithostratigraphy of the Pretoria Group, illustrating thicknesses and lithological variation across the basin.

providing a convincing marker unit. It is probable thus that the Pretoria Group of the Transvaal/Bushveld basin, the Segwagwa Group of the Kanye basin and the Postmasburg Group of the Griqualand West basin (Table 1) developed to a large extent in separate basins, a concept first developed by Crockett (1972) and supported more recently by Eriksson *et al.* (1988, 1991) and Eriksson and Clendenin (1990).

The basal Rooihoogte Formation of the Pretoria Group comprises a lowermost chert breccia-reworked conglomerate member, overlain by mudrocks and an uppermost arenaceous member (Visser, 1969; Button, 1973; Engelbrecht, 1986) (Fig. 8). The breccias are mostly *in situ* residual deposits overlying the palaeokarst landscape developed on the Malmani carbonates (Button, 1973), with some of them possibly representing fault scarp talus deposits (Eriksson, 1988). Both matrix-supported and clast-supported conglomerates are found, supporting the alluvial fan and fan-delta complexes proposed by Eriksson (1988). This postulate is in sharp contrast to the transgressive marine/basinal environment inferred by previous workers (Visser, 1969; Button, 1973, 1986; Beukes,1983), but is supported by recent detailed facies anlayses around the basin (Van der Neut, 1990; Schreiber, 1991); these workers were able to delineate a number of fan complexes and interpret the succeeding mudrock and sandstone members to be distal fan-delta and lacustrine sediments. Eriksson *et al.* (1991) suggest that a relatively steep northerly palaeoslope is indicated by the depth of karstic weathering in the underlying dolomites. They also consider that this may have bounded a half-graben system, whose low angle hanging wall lay to the north (Fig. 9a), and which provided the major source of detritus. This postulated half-graben appears to have controlled Pretoria Group sedimentation through to the deposition of the Daspoort Formation, with a northern basin boundary which was reactivated a number of times.

The Timeball Hill Formation consists of basal carbonaceous mudrocks, with micro-algal fossils (Nixon *et al.*, 1988), which grade upwards into rhythmically interbedded ferruginous mudrocks and fine-grained sandstones, in turn passing up into the Klapperkop Sandstone Member (Fig. 8); uppermost carbonaceous and ferruginous mudrocks complete the stratigraphy of this formation (Visser, 1969; Button, 1973; Eriksson 1973). The locally pyritic carbonaceous mudrocks indicate deep water anoxic suspension sedimentation (Eriksson, 1973), with the overlying mudrock-sandstone facies possessing sedimentary structures compatible with turbidity current re-sedimentation of distal delta deposits (Rust, 1961; Kuenen, 1963; Visser, 1969, 1972; Button, 1973). The upward-coarsening sandstones of the Klapperkop Member support tidal reworking of delta front sands (Visser, 1969; Button, 1973; Eriksson and Clendenin, 1990), with localised oolitic ironstones having developed in a shallow offshore setting. The Timeball Hill palaeoenvironment is thus thought to have comprised a relatively deep basin, filled by fluviodeltaic complexes advancing from the north, northwest and east; with the exception of the arkoses preserved in the northeast of the basin (Fig. 8), only the more distal deltaic and basinal sediments appear to have been preserved. This postulate is compatible with the suggestion of a half-graben deriving sediment largely from northerly sources, or characterised by sedimentation parallel to the bounding faults (Fig. 9a).

The diamictites within the upper Timeball Hill mudrocks and the conglomerates/diamictites of the succeeding Boshoek Formation in the western and central portions of the Transvaal basin (Fig. 8), are correlated by Visser (1971) with the Makganyene diamictites of Griqualand West; we suggest a broader correlation of the latter formation with the full thickness of both Timeball Hill and Boshoek Formations (Table 1). The diamictite-immature sandstone-conglomerate lithologies correlated by Visser are interpreted as being glacial, glaciofluvial and glaciomarine deposits (Visser, 1971). Schreiber *et al.* (1990) interpreted the Boshoek Formation, including the immature sandstones found in the northeast of the basin, as alluvial sediments. It is thus possible, particularly in view of palaeomagnetic data (Fig. 9b), that both the Timeball Hill and Boshoek Formations had an important glacial influence in their genesis. It is postulated here that the Timeball Hill lithologies compare closely with deep glacial lake deposits such as those in Lakes Constance and Geneva (Reineck and Singh, 1975), and that the coarse-grained diamictite-conglomerate-sandy assemblages discussed by Visser (1971) represent more proximal glacial ablation deposits. The centre of glaciation probably lay on the Vryburg rise, an intra-basinal high lying to the south of the Kanye basin and separating the Griqualand West and Transvaal basins (Visser, 1971) (Fig. 1).

The thick lavas of the Hekpoort Formation, with correlates extending across both the Kanye and Griqualand West regions (Table 1), have a basaltic to intracratonic andesitic chemistry (Sharpe *et al.*, 1983; Engelbrecht, 1986); interbedded pyroclastics and resedimented volcaniclastic rocks appear to be widespread, with subordinate thin beds of mudrock and chert (Button, 1973; Engelbrecht, 1986; Eriksson and Twist, 1986). The scarcity of pillow lavas and mudrock interbeds points to subaerial extrusion, with an uppermost basin-wide aluminous mudrock layer, mostly 1 - 2 m thick, being interpreted as either a palaeosol (Button 1973) or due to chemical leaching of the lavas (Engelbrecht, 1986). Martini (1990) identified a semi-arid cool climate playa lake deposit within this aluminous mudrock horizon. The lavas in the Transvaal basin thicken southwards, possibly reflecting a half-graben basin setting (Fig. 9a). The lavas are overlain with a sharp contact by the conglomerates, immature sandstones and minor mudrocks of the Droogedal/Dwaalheuwel Formations; the former represents a sheet of sediment entering the Transvaal basin from the northwest and the latter formation two lobes which entered from the north and east (Eriksson *et al.*, 1991). All three sheets thin and fine towards the Pretoria region, where these formations are absent and the Hekpoort lavas are succeeded by the mudrocks and subordinate sandstones of the Strubenkop Formation (Eriksson *et al.*, 1991). The shallow marine model proposed for the Dwaalheuwel Formation (Button, 1973) is disputed by Eriksson *et al.* (1989, 1991) who suggest alluvial fan and distal fan-delta deposition for both Droogedal and Dwaalheuwel Formations. The Strubenkop mudrocks overlie these sandstone sheets and are thickest where the Droogedal/Dwaalheuwel arenites wedge out, becoming thinner as the sandstones thicken towards the northwest, northeast and east of the basin (Eriksson et *al.*, 1991). This led Eriksson *et al.* (1991) to propose that the Strubenkop Formation represents a more distal lacustrine basin into which the Dwaalheuwel/Droogedal fan-deltas debouched. Oolitic ironstone lenses, mudcracks, ripple marks, graded bedding, flute casts, flaser bedding and minor channel-fills (Visser, 1969; Button, 1973; Engelbrecht, 1986) are compatible with the shallow marine (Visser, 1969), tidal flat (Button, 1973) and lacustrine (Eriksson and Clendenin, 1990) models proposed for the Strubenkop Formation. As suggested for the Rooihoogte and Timeball Hill Formations, the predominant northerly source of the Droogedal and Dwaalheuwel Formations may reflect a half-graben setting (Fig. 9a).

The recrystallised, cross-bedded and planar stratified sandstones of the Daspoort Formation overlie the Strubenkop mudrocks sharply, except in eastern Botswana, where the Ditlhojana Formation (as the Daspoort is known in the far west of the Transvaal basin) (Table 1) lies on thin mudrocks developed above the Hekpoort lavas (Key, 1983), indicating erosion of the preceding Droogedal arenites and most of the Strubenkop mudrocks (Eriksson *et al.* under review). Planar stratified mature sandstones and very thin ironstones/ferruginous mudrocks in the east of the basin led Button (1973) to propose a shallow marine beach-barrier palaeoenvironment. A locally erosive base and pebbly sandstones in the Pretoria region were taken by Visser (1969) to support a fluvial-beach model. The immature nature of the Pretoria region sandstones (Van der Neut, 1990) and the occurrence of significant pebbly sandstones and mudrocks in the north and far west of the basin (Key, 1983; G. Potgieter, 1992 pers. comm.) support the distal fan-fluvial braidplain model of Eriksson *et al.* (under review). The latter is based largely on lateral and vertical facies relationships, palaeocurrents and thickness trends observed across the entire preserved Transvaal-Bushveld basin, rather than relying on the localised data used by Visser (1969) and Button (1973). Source areas were largely to the north of the Transvaal basin, possibly supporting the half-graben model. The Daspoort Formation also indicates a period of tectonic reorganisation within the Transvaal depository, as isopach maps of the preceding and succeeding Pretoria Group formations are quite different (Eriksson *et al.*, 1991).

The thick, monotonous laminated mudrocks of the Silverton Formation encompass significant volcanic lithologies (Button, 1973). Carbonaceous mudrocks are relatively common, with subordinate chert, sandstone and dolomite lenses; carbonate rocks become important in the north of the basin (Button 1973; Engelbrecht, 1986). In the eastern Transvaal, a medial Machadodorp Member includes lower agglomerates and tuffs and upper pillowed basalts, with bomb sizes decreasing towards the north (Button, 1973); some agglomerates are found at the equivalent stratigraphic position in the Pretoria region (Visser, 1969) and reworked tuffs are found in the west of the basin (Eriksson *et al.*, 1990). An upward-coarsening, 13 km long and 150 m thick, sandstone lens in the west of the basin, with planar and cross-bedding and ripple marks (Engelbrecht, 1986; Eriksson *et al.*, 1991), supports the prodeltaic palaeoenvironment suggested by Button (1973). Eriksson and Clendenin (1990) and Eriksson *et al.* (1991) proposed lacustrine deltas and fan-deltas as an alternative model. In addition, the latter workers suggested that the Magaliesberg sandstones which overlie the Silverton Formation gradationally, may represent the shoreline of the Silverton basin; this postulate is supported by the essentially similar thickness distribution of both formations across the Transvaal basin (Eriksson *et al.*, 1991). The Magaliesberg sandstones also wedge laterally into Silverton mudrocks in the west of the basin (Crockett, 1972; Engelbrecht, 1986; Key, 1986; Eriksson and Clendenin, 1990). Herringbone cross-beds, carbonate rocks, flaser bedding and interference ripple marks point to a marine influence in the Silverton-Magaliesberg basin, which is also evident from the palaeosalinity data of Eriksson (1992). It is thus possible that the intracratonic half-graben setting envisaged for the lower Pretoria Group (Eriksson *et al.*, 1991) (Fig. 9a) may have given way to a transgressive epeiric sea in Silverton-Magaliesberg times. Palaeomagnetism (Fig. 9b) indicates movement away from polar latitudes in late Pretoria Group time, thereby supporting the possibility of a marine transgression. The thickness patterns of these two formations (Fig. 9c) point to doming in the centre of the depository, thought by Eriksson *et al.* (1988, 1990) to be a synsedimentary palaeohigh; the doming may have been thermal in nature, related to a rising plume of Bushveld magmas through the crust (Eriksson *et al.*, 1991). The sandstone isolith plot (Fig. 9c) indicates that sands may have been shed off the proposed dome, a postulate also supported by palaeocurrent data in the Magaliesberg Formation (Van der Neut, 1990; Schreiber, 1991). If such doming did occur, it may have led to a retreat of the Silverton epeiric sea, leaving the Magaliesberg shoreline sands subject to fluvial reworking. Such a postulate would be compatible with the beach-fluvial, delta-beach-shallow marine and tidal models proposed for the Magaliesberg Formation (Visser, 1969; Button, 1973; Eriksson *et al.*, 1987). Grain size patterns for the Magaliesberg Formation strongly support a fluvial influence in the deposition of these sandstones (Reczko *et al.*, 1992).

Continued doming in the centre of the Transvaal basin may have been responsible for the development of separate eastern and western intracratonic sub-basins in post-Magaliesberg times; a separate western lacustrine basin was first proposed by Crockett (1972) and more recently supported by Eriksson *et al.* (1988, 1991) and by Schreiber (1991). Lithostratigraphic differences between the far west and the east of the basin are also marked (Fig. 8). The five formations in the east of the

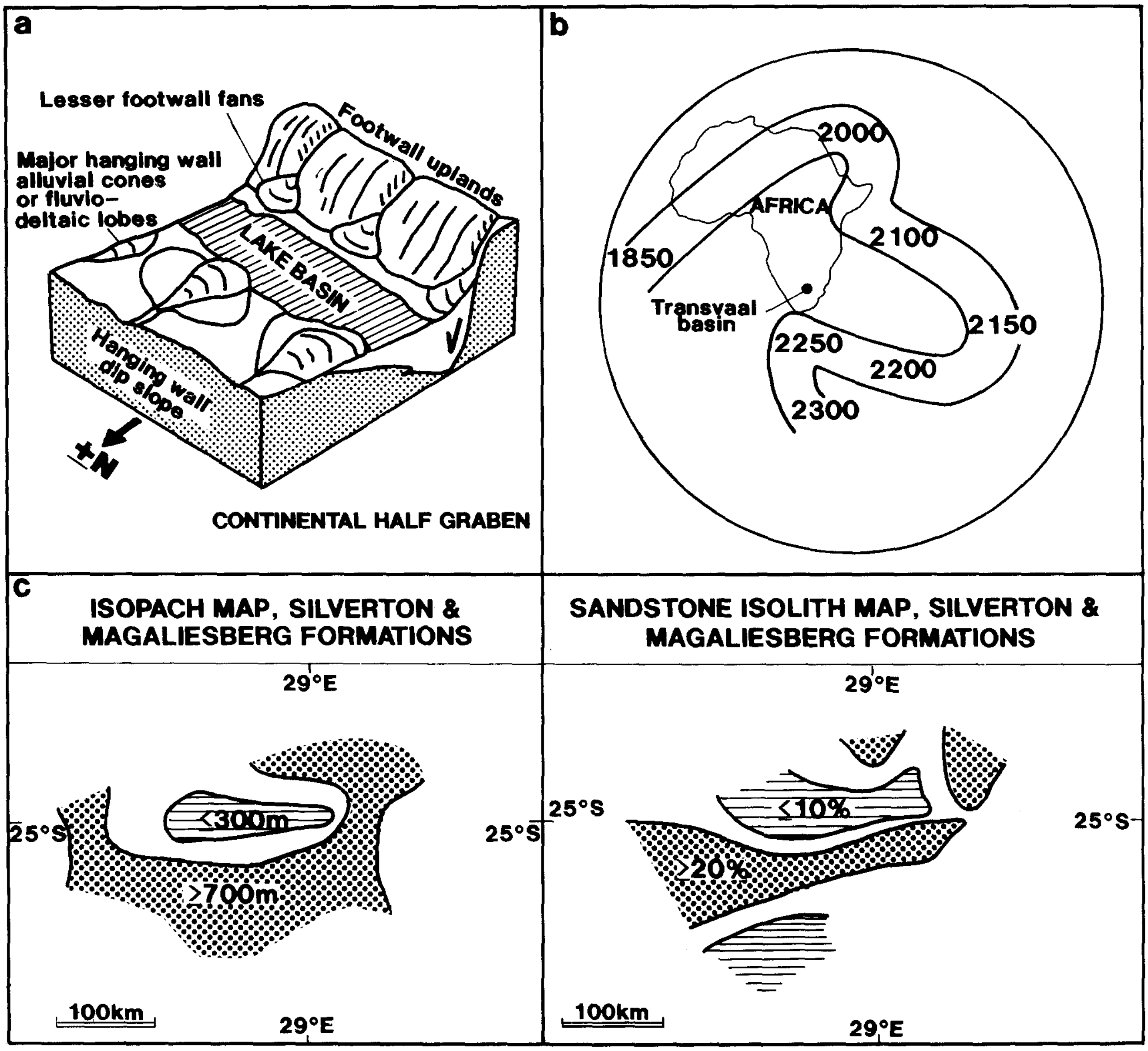


Fig. 9. Postulated controls on the Pretoria Group palaeoenvironment. (a) Half-graben tectono-sedimentary setting proposed by Eriksson et al. (1991). (b) Polar wandering curve for Africa for the Early Proterozoic; note location of Transvaal basin in cold latitudes at 2250 ma (Hekpoort Formation age) and movement away from polar latitudes in upper Pretoria Group times (Modified after Condie, 1989). (c) Thickness patterns of the Silverton and Magaliesberg Formations; note possible central basin palaeohigh indicated by thinner total sediment cover and surrounding zone of thicker sandstones.

eastern sub-basin are correlated with the Rayton Formation of the Pretoria region (SACS, 1980). The alternating mudrocks, arkosic and quartzitic sandstones, with subordinate carbonate and chert lithologies, which make up the five easternmost formations (Schreiber, 1991) include interbedded tuffaceous mudrock layers (Schreiber and Eriksson, in press). The beach-intertidal mudflat-shallow marine model of Button (1973) and Button and Vos (1977) does not explain adequately the common mudcracks, arkosic sandstones and overall upward-coarsening successions in these rocks. The latter are, perhaps, explained better by shallow lacustrine and wind-tidal flat deposition for the argillaceous units and by fan-delta and deltaic sedimentation for the interbedded arenites (Schreiber, 1991; Schreiber and Eriksson, in press). Source areas were located to the south, east and north of this steadily shrinking sub-basin (Schreiber, 1991). The analogous Rayton Formation succession of feldspathic arenites, quartzose sandstones and mudrocks, with subordinate dolomites and lavas, is ascribed to fluvial (Visser, 1969) or a combination of fan, fan-delta and fluvial sedimentation (Van der Neut, 1990). In the proposed western sub-basin, post-Magaliesberg rocks in the western Transvaal are obscured largely by Bushveld intrusives (Engelbrecht, 1986); further to the west in Botswana, the Woodlands Formation comprises interbedded sandstones and mudrocks, with significant andesitic pyroclastic rocks and lavas and some conglomerates (Key, 1983). Major tectonic disturbance of the Woodlands rocks has made palaeoenvironmental interpretation difficult; an intimate association between sedimentation, volcanism and tectonic instability is indicated (Crockett, 1969, 1972; Key, 1983). Crockett (1969, 1972) proposed catastrophic syn- to post-sedimentary collapse of the western sub-basin floor, resulting in gravity sliding of large blocks of sedimentary rocks towards the basin centre. Key (1983) relates this tectonism to the late-Transvaal intrusion of the Gaborone Granite to the west of the Bushveld basin; this postulate can perhaps be considered as related to the extrusion of the massive felsic lava province of the Rooiberg Group which succeeded Pretoria Group deposition in the east and centre of the Transvaal basin. No analogous late Pretoria sedimentary and volcanic rocks are preserved in either the Griqualand West or Kanye basins (Table 1).

There is a long-standing debate whether the sedimentary rocks of the Pretoria Group represent epeiric marine deposits or an intracratonic lacustrine basin (Du Toit, 1954; Visser, 1957; Willemse 1959; Visser, 1969; Crockett, 1972; Button, 1973; Button and Vos, 1977; Button, 1986; Eriksson and Clendenin, 1990). The epeiric postulate is supported by preserved tidal sedimentary structures, oolites, stromatolites and hummocky cross-bedding; however, these characteristics are equally compatible with the lacustrine alternative. The latter suggestion is in turn supported by common varves, the presence of palaeosols, predominantly arkosic to lithic sandstone petrography, inferred microtidal conditions typical of lakes (Schopf, 1980) and an absence of swash-formed ripple marks. The debate is discussed in some detail by Eriksson *et al.* (1991). Perhaps the best evidence in favour of an intracratonic basin is the Boron palaeosalinity data discussed by Eriksson (1992) (Fig. 10). Boron values determined by Böhmer (1977) from four widely spaced boreholes in the south of the Transvaal basin, when plotted against stratigraphic height, exhibit a very similar, in-phase variation, summed up in the palaeosalinity curve of Eriksson (1992) (Fig. 10). This in-phase variation from different localities supports a closed basin setting, and the rapid increases and decreases in Boron values further support inhomogeneous lacustrine basin chemistry rather than the far more uniform conditions typical of the marine environment (Schopf, 1980). The possible role of glacial and interglacial periods in the deposition of the Pretoria Group and the postulated marine incursion during Silverton times are supported by the palaeosalinity curve for the lower Pretoria Group (Fig. 10).

ROOIBERG GROUP

In this review we present a modified stratigraphy for the upper portion of the Transvaal Sequence (Tables 1 and 2) which is not in accord with SACS (1980), who consider the Rooiberg Group (sensu stricto) to comprise only the Damwal and Selonsrivier Formations; our stratigraphy reflects research carried out within the last five years on the Dullstroom Formation, Rooiberg Group (*sensu stricto*, SACS, 1980) and sedimentary units associated with these rocks, by Schweitzer and Hatton (in prep.). SACS (1980) considers the largely sedimentary Leeuwpoort, Smelterskop and Makeckaan Formations to belong to the upper Pretoria Group and various correlations with post-Magaliesberg units have been proposed, (e.g., Button, 1973; Schreiber, 1991). We consider the Leeuwpoort Formation, the basal unit outcropping within the Rooiberg "fragment" (Fig. 1), as the youngest post-Pretoria Group unit and assign it, tentatively, to the Rooiberg Group (*sensu lato*, as used here) (Table 2) as its basal unit; there do not appear to be any correlates elsewhere in the Transvaal basin, and no rocks equivalent to the Rooiberg Group (*sensu lato*) are found in either the Kanye or Griqualand West basins.

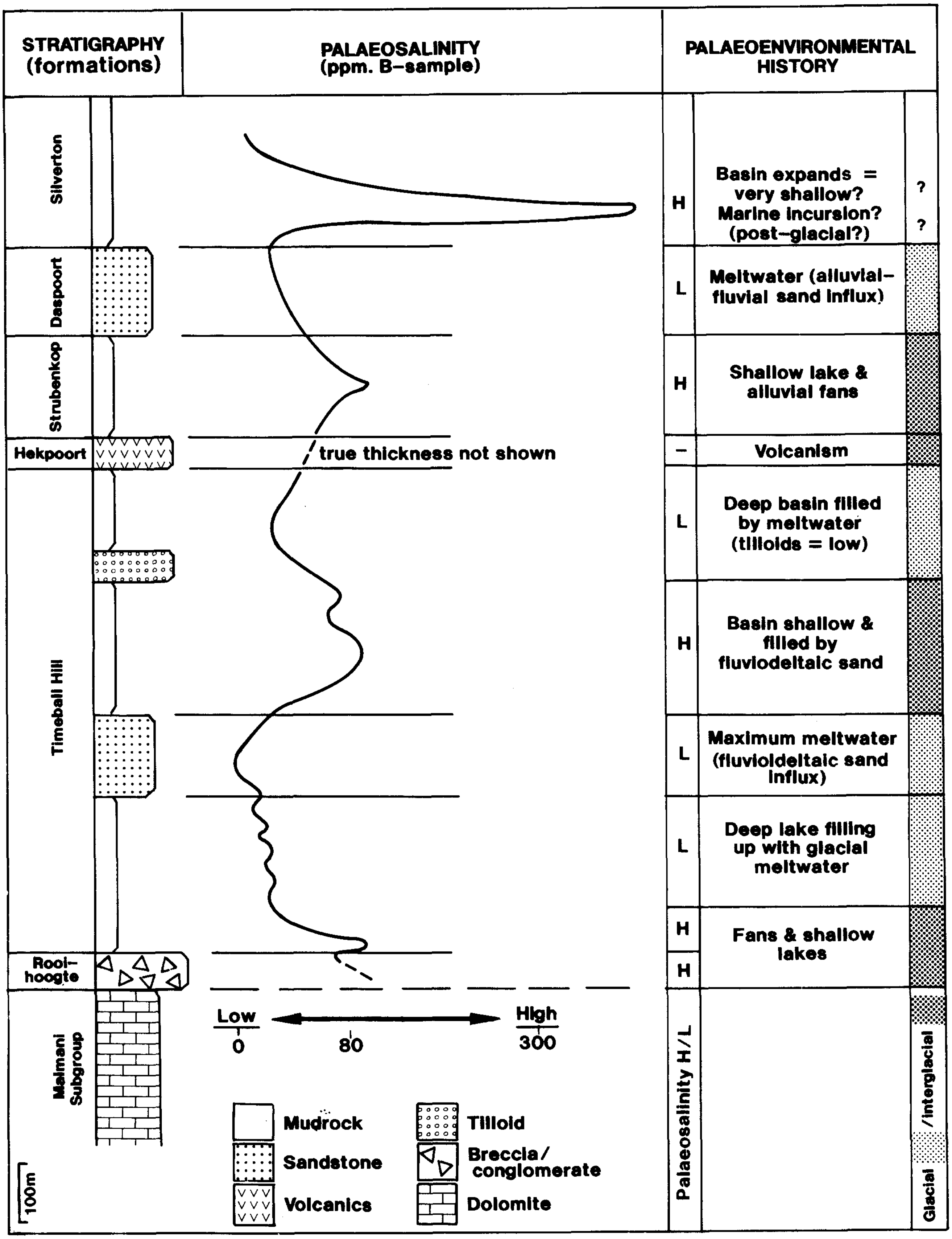


Fig. 10. Palaeosalinity curve for the lower Pretoria Group, based on Boron values determined in widely spaced boreholes by Böhmer (1977). Also shown is the possible role of glacial and interglacial periods on the Pretoria Group palaeoenvironment. Modified after Eriksson (1992).

The approximately 1500 m of conglomerates, pebbly and arkosic sandstones and uppermost mudrocks which make up the Leeuwpoort Formation (Fig. 11) (Stear, 1976; Richards, 1987), are generally interpreted to represent lower braided and upper meandering river deposits (Stear, 1977a and b; Phillips, 1982; Rozendaal *et al.*, 1986; Richards and Eriksson, 1988). The coarse immature fluvial sandstones which characterise the Leeuwpoort Formation (Fig. 11) do not compare well lithologically with the alternating sandstones and mudrocks of the post-Magaliesberg units in either the east or west of the Pretoria Group basin. We thus suggest that the Leeuwpoort Formation is part of the Rooiberg Group (*sensu lato*) and postulate that these coarse, immature fluvial deposits are probably related to the catastrophic collapse proposed by Crockett (1969, 1972) for the western Pretoria Group sub-basin floor, during and after deposition of the Woodlands Formation. The gravity sliding of large blocks of sedimentary rocks in Botswana (Crockett, 1972) may have been accompanied by more subdued immature fluvial deposition on the eastern margin of the collapsing sub-basin, today preserved at the base of the Rooiberg "fragment". Predominantly southwesterly palaeocurrent directions in the Leeuwpoort sandstones (Stear, 1977a; Richards, 1987) are compatible with this postulate.

The Smelterskop Formation conformably overlies the Leeuwpoort Formation (Richards and Eriksson, 1988) (Fig. 11) and comprises a thickness of about 280 m, with a basal quartzose sandstone member, four to five lenticular arkosic sandstone units and subordinate tuffaceous mudrocks, conglomerates and wackes (Stear,

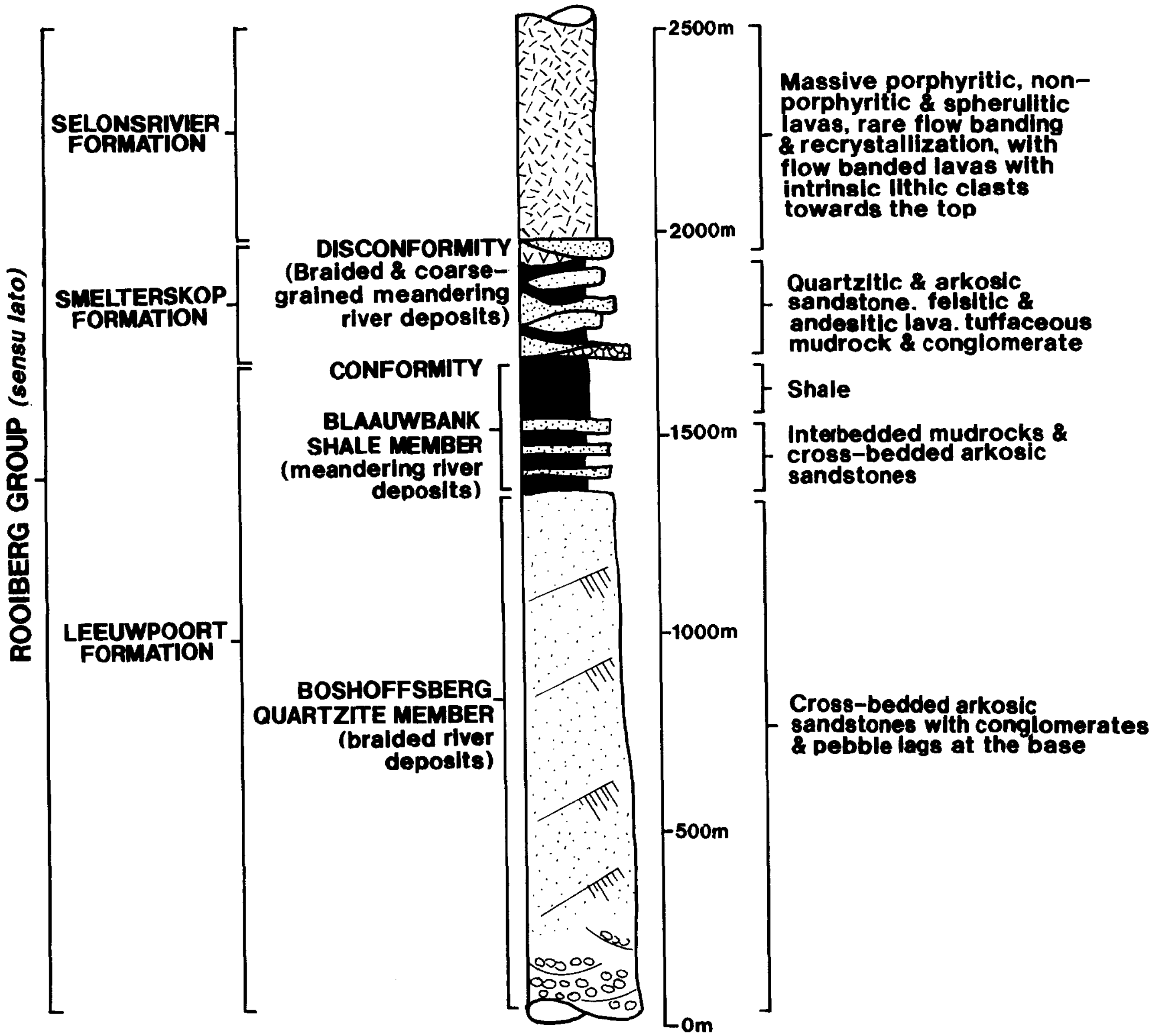


Fig. 11. Lithostratigraphy of the Leeuwpoort, Smelterskop and Selonsrivier Formations in the Rooiberg "fragment" (see Fig. 1 for location). Modified after Stear (1977a) and Richards (1987).

1977a, Richards, 1987). Braided fluvial and coarse-grained meandering river systems are thought to have laid these sediments down (Stear, 1977a; Richards and Eriksson, 1988). Impersistent felsitic and andesitic lava flows are intercalated with the sedimentary rocks of the Smelterskop Formation; the andesites are flow-banded and strongly vesicular, and the uppermost felsite is flow-banded and petrographically similar to the basal flows of the overlying Rooiberg (*sensu stricto*) lavas.

The Makeckaan Formation, preserved in the Makeckaan (or Stavoren) "fragment" comprises lower and upper feldspathic sandstone members with large scale cross-beds and ripple marks, separated by mature, recrystallised quartzitic sandstones and micaceous wackes (Fig. 12) (Rhodes, 1972). Andesites and felsites are intercalated in the upper arkosic member (Fig. 12), and the sedimentary succession of the Makeckaan Formation is succeeded concordantly by felsitic lavas (Mellor, 1905; Wagner, 1921, 1927; Schweitzer and Hatton in prep.), here assigned to the Dullstroom Formation. The felsites are spherulitic, vesicular and abundant matrix-supported lithic fragments suggest emplacement as ignimbrites. The Makeckaan sediments are interpreted as fluvial to neritic (Rhodes, 1972) or fluviodeltaic

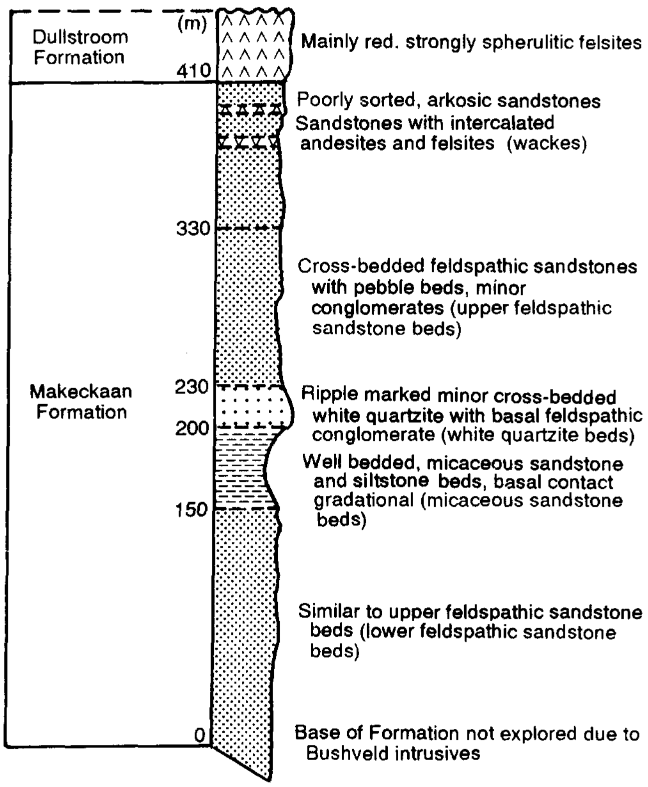


Fig. 12. Lithology of the Makeckaan Formation. Note the similarity to the rocks of the Smelterskop Formation (Fig. 11), and the overlying Dullstroom felsites. Modified after Rhodes (1972).

(Schreiber, 1991) deposits. The similarity between the apparently fluvially-deposited Smelterskop and Makeckaan feldspathic to quartzitic sandstones and their intercalated andesitic-felsitic lavas, leads us to suggest that these two formations may be correlated with each other, representing the transition from lower Rooiberg Group (*sensu lato*) fluvial sedimentation (Leeuwpoort Formation) to the vast outpourings of the main Rooiberg felsitic lavas. These two largely sedimentary formations are thus also, tentatively, correlated with the basal stage of the Dullstroom Formation (Figs 13 and 14) (Truter, 1949; Visser, 1969).

Lavas assigned to the Dullstroom Formation outcrop along a narrow strip in the eastern Transvaal basin and also overlie the Makeckaan Formation sediments in the Makeckaan "fragment" (Figs 13 and 14). The former is the type locality of the Dullstroom Formation, which unconformably (Cheney and Twist, 1991) succeeds the Houtenbek Formation of the Pretoria Group (Table 2). A maximum thickness of 1,4 km of this basalt-rhyolite association (Schweitzer, 1986) is preserved and the lavas are truncated in the north by Bushveld Complex intrusives (Fig. 13). Schweitzer (in prep.) and Schweitzer and Hatton (in prep.) distinguish a basal stage and an upper stage in the Dullstroom succession (Fig. 14). The former comprises about 300 m of fluidal and pyroclastic flows, with thin interbedded lenticular quartzitic and arkosic sandstones and mudrocks; the absence of pillow structures, and the presence of pipe amygdales and peperites suggests subaerial extrusion with localised shallow water conditions. Amygdaloidal low titanium andesites predominate, with flow-banded, amygdaloidal and spherulitic rhyolites referred to as the basal rhyolites (Fig. 14), occurring locally at the base. Volcaniclastic sediments and debris flow deposits are also found locally. The upper Dullstroom stage comprises more uniform lavas with almost no intercalated sedimentary rocks. Three major flow types are distinguished: porphyritic high titanium basalts with amygdaloidal and brecciated flow tops; high magnesium amygdaloidal and sparsely porphyritic flow-banded felsites; and low titanium andesites similar to those in the basal stage (Fig. 14). These different flow types are closely associated and interlayered, with a general upward increase in high magnesium felsites at the expense of the high titanium and low titanium flows.

A large xenolith within Bushveld mafic intrusives to the north of the Dullstroom outcrops (locality 2, Fig. 13) appears to comprise altered equivalents of the uppermost low magnesium felsites of the Dullstroom Formation and succeeding high Fe-Ti-P lavas at the base of the Damwal Formation

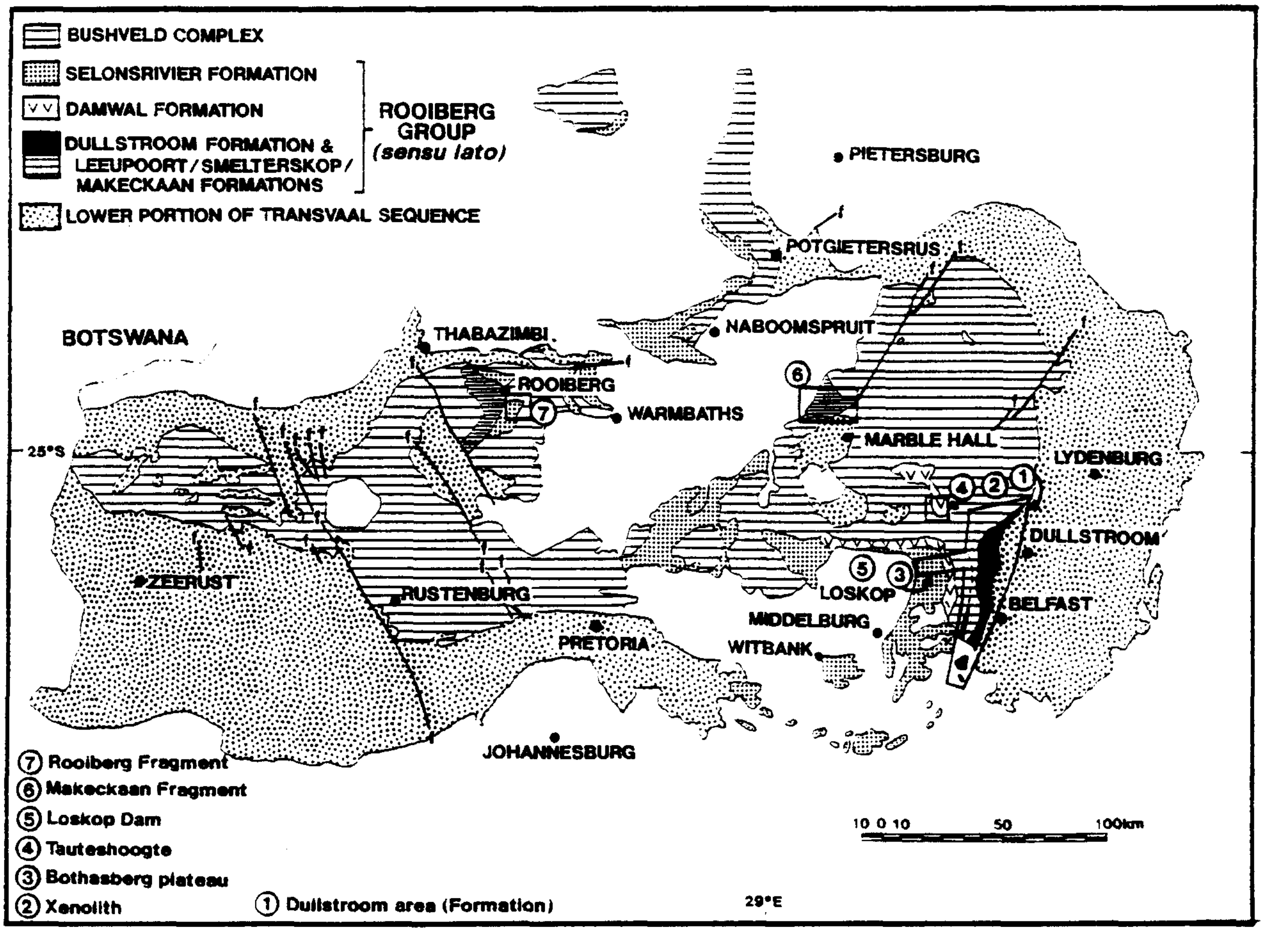


Fig. 13. Map of the Transvaal basin, South Africa, illustrating the distribution of rocks belonging to the Rooiberg Group (sensu lato, as used here). Note localities 1 to 7, as used also in Fig. 14.

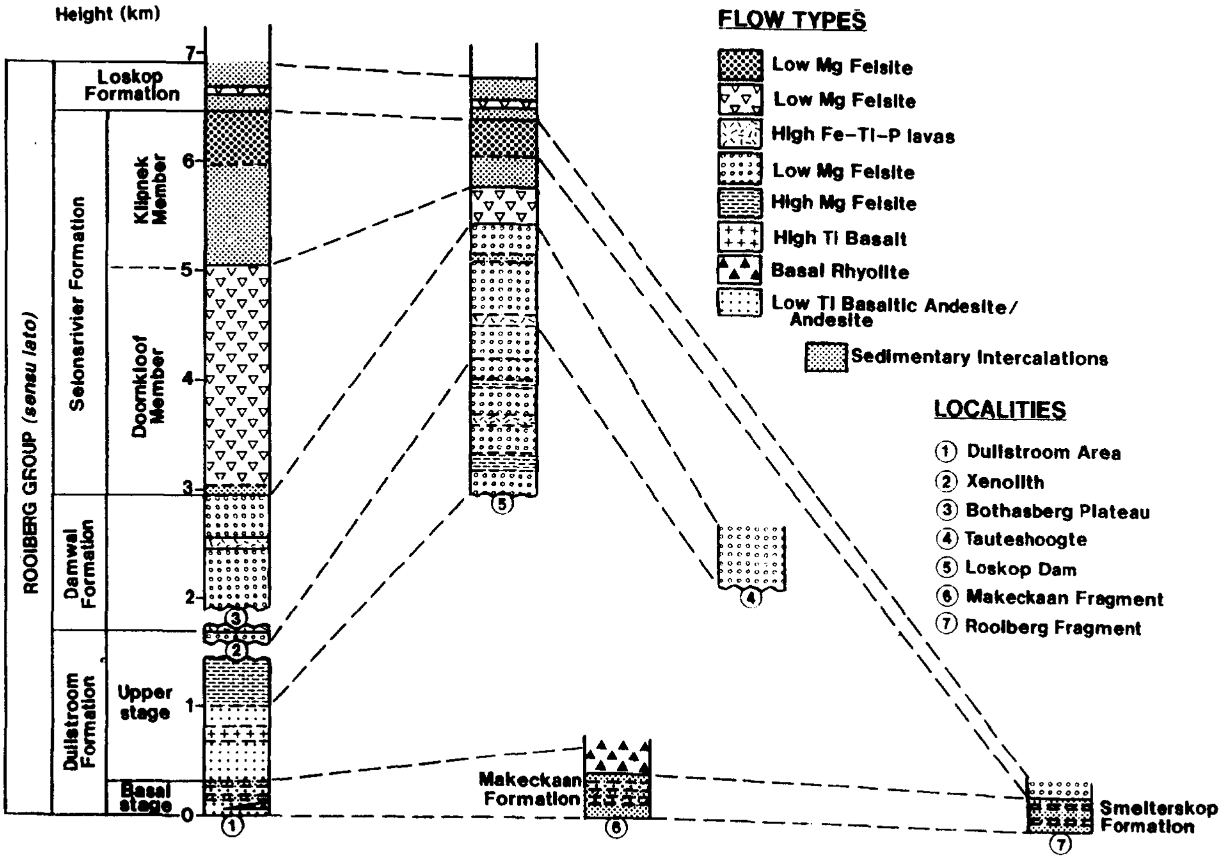


Fig. 14. Lithostratigraphy and correlation of the volcanic formations of the Rooiberg Group (*sensu lato*). See Fig. 13. for locations 1 to 7. Also note the different flow types identified within the volcanic formations.

(Fig. 14). The Damwal Formation is found within the Bothasberg, Tauteshoogte and Loskop Dam areas (respectively, localities 3, 4 and 5 in Fig. 13) (Fig. 14). The succeeding Selonsrivier Formation, which is subdivided into a lower Doornkloof and upper Klipnek Member, is found at Bothasberg, Loskop Dam and overlying the Smelterskop Formation in the Rooiberg "fragment" (Figs 13 and 14). These two felsitic formations were first defined by Clubley-Armstrong (1977, 1980), and together comprise a succession of very extensive siliceous eruptives within the central and eastern parts of the Transvaal basin (Fig. 13); the original volume of these eruptives is estimated to have reached up to 300 000 km^3 (Twist and French, 1983), although this is probably an overestimate. The Damwal Formation comprises largely dark flows of low magnesium felsite, with subordinate high Fe-Ti-P lavas, whereas the Selonsrivier Formation is characterised by red-coloured, flow-banded flows of low magnesium felsite in the Doornkloof Member and intercalated sedimentary rocks and low-Mg felsites in the Klipnek Member (Fig. 14). Twist (1985) distinguished nine major units separated by sedimentary intercalations within the felsite succession at Loskop Dam, placing the Damwal-Selonsrivier contact at the boundary of units 5 and 6; the same stratigraphic level was inferred to mark the transition to an oxygen-rich atmosphere in the upper Transvaal Sequence (Twist and Cheney, 1986). However, Eriksson and Cheney (1992) showed that the colouration of the felsites in the Selonsrivier Formation was formed diagenetically, and proposed that the transition to an atmosphere rich in free oxygen lay rather at the Loskop Formation-Waterberg Group unconformity. The lavas of the Damwal and Selonsrivier Formations are predominantly rhyolitic to dacitic in composition, with both high- and low-Mg types occurring (Twist and French, 1983; Twist and Harmer, 1987). Extrusion was subaerial and feldspathic sandstone interbeds with some preserved sedimentary structures such as cross-bedding, channel-fills, ripple marks and mudcracks (Clubley-Armstrong, 1977), point to intermittent periods of deposition between volcanic eruptions, possibly by a combination of fluvial channels, small lakes and gravity flows.

Many authors (Button, 1976; Coertze *et al.*, 1977; Sharpe *et al.*, 1983; Twist and French, 1983) support the view expressed here, that the Dullstroom and Damwal-Selonsrivier volcanic events originally formed one continuous eruptive succession, implying that the latter two formations are also older than the intrusive mafic suite of the Bushveld Complex (Von Gruenewaldt, 1971, 1972); this view is confirmed by geochemical data (Schweitzer, 1986, Schweitzer and Hatton in prep.) (Fig. 15). The sedimentary rocks of the correlated Loskop, Glentig and Rust de Winter Formations (Fig. 1) overlie the Selonsrivier felsites conformably (SACS, 1980; Cheney and Twist, 1991); they represent the final sedimentary phase of the Transvaal basin. Minor interbedded volcanic flow deposits range in composition from basic to siliceous and probably represent continued eruptions of magmas related to the Damwal-Selonsrivier lavas (Clubley-Armstrong, 1977; Coertze *et al.*, 1977). This leads us to suggest here, tentatively, that these uppermost sedimentary formations also be considered part of the Rooiberg Group (*sensu lato*) (Tables 1 and 2).

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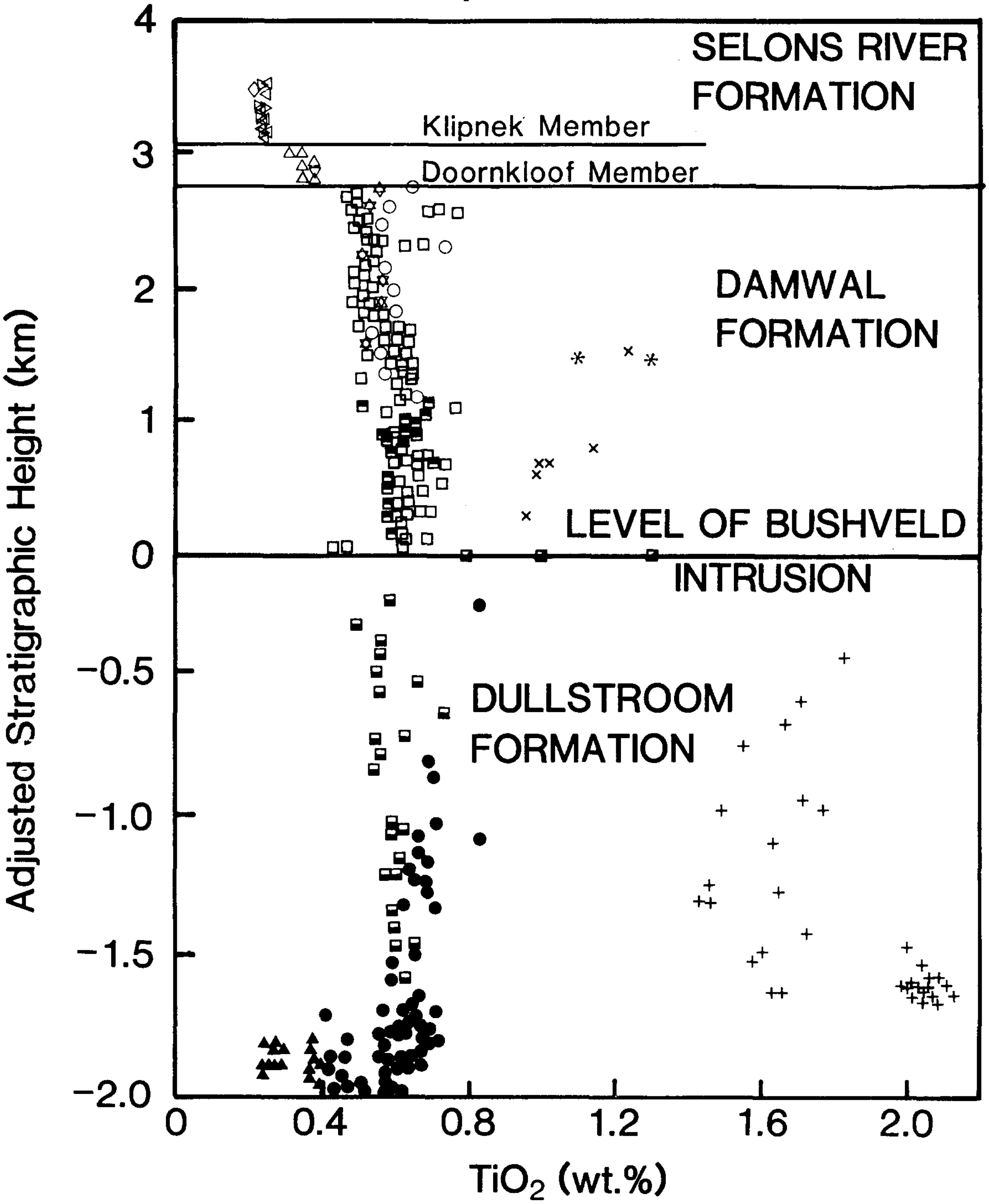

◁ Low Mg Felsite (LMF$_{Klip}$) - Loskop Dam

◇ Low Mg Felsite (LMF$_{Klip}$) - Bothasberg Plateau

▷ Low Mg Felsite (LMF$_{Klip}$) - Rooiberg Fragment

△ Low MG Felsite (LMF$_{Doorn}$) - Loskop Dam

▽ Low Mg Felsite (LMF$_{Doorn}$) - Bothasberg Plateau

× High Fe-Ti-P lavas - Loskop Dam

✲ High Fe-Ti-P lavas - Bothasberg Plateau

□ Low Mg Felsite (LMF$_{Dam}$) - Loskop Dam

○ Low Mg Felsite (LMF$_{Dam}$) - Bothasberg Plateau

✡ Low Mg Felsite (LMF$_{Dam}$) - Tauteshoogte

▤ High Mg Felsite (HMF) - Loskop Dam

◪ Xenolith

▣ High Mg Felsite (HMF) - Dullstroom Area

\+ High Ti Basalt (HTI) - Dullstroom Area

▲ Basal Rhyolite - Dullstroom Area

● Low Ti Basaltic Andesite/Andesite (LTI$_{ba/a}$)

Fig. 15. Plot of TiO_2 content of Rooiberg lavas against stratigraphic height to illustrate continuity of the Dullstroom-Damwal-Selonsrivier volcanic event. Modified from Schweitzer (in prep.).

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