Nature of and controls on volcanism in the c. 2.32 – 2.06 Ga Pretoria Group, Transvaal Supergroup, Kaapvaal Craton, South Africa

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Abstract

The time of deposition of the Pretoria Group between 2.32 and 2.06 Ga on South Africa's Kaapvaal Craton was characterized by the first major increase in atmospheric oxygen. It was accompanied by the extrusion of significant thicknesses of volcanic deposits, namely the Bushy Bend lavas of the Timeball Hill Formation, the Hekpoort Formation and the Machadodorp Volcanic Member of the Silverton Formation, marking the three major volcanic events within the Transvaal Supergroup which are thought to be precursors to the succeeding Bushveld Complex magmatism. The Bushy Bend Lava Member of the Timeball Hill Formation is characterized by subaqueous basaltic-andesitic fissure eruptions with fumarolic activity and probably minor subaerial explosive eruptions in the hinterland of the Transvaal Supergroup basin. The subaerial volcanism of the Hekpoort Formation appears to have been dominated by fissure eruptions with a preponderance of lava flows and locally important pyroclastic material. Intermittent hiatuses in volcanism were marked by local lacustrine shale deposition. Finally, the Machadodorp Lava Member can be interpreted as the deposits of several seamounts aligned along a fissure, probably within an extensional environment.

The geodynamic control inferred for the evolution of the Pretoria Group basin encompasses two cycles of prerift uplift, subsequent mechanical rifting and long lived thermal subsidence. The limited extent of the Bushy Bend lavas in the south of the Pretoria Group depository attest to the likelihood that volcanism accompanying the first rifting event was short-lived. In contrast, the second rifting cycle postulated for the Pretoria Group basin has a strong association with widespread and large scale volcanism of the Hekpoort-Ongeluk flood basalts, which may have been plume-related. The scale of the second cycle volcanism and its importance in influencing the upper part of the Pretoria Group basin-fill is underlined by the Machadodorp volcanism, probably related to hot spot volcanism as the Ongeluk-Hekpoort plume waned.

Keywords: Palaeoproterozoic within-craton volcanism; Pretoria Group; Kaapvaal Craton; South Africa

Introduction

The deposition of the Pretoria Group (Transvaal Supergroup) on South Africa's Kaapvaal Craton marks a radical change in environment during the Palaeoproterozoic (Eriksson et al., 2006). What was initially a shallowly submerged epeiric continental platform on which cyanobacteria flourished and formed the thick stromatolitic carbonate beds of the underlying Chuniespoort Group, became covered by large volumes of Pretoria Group clastic sediments. After a period of uplift (80 My according to Eriksson et al. (2001), possibly up to c. 200 My according to Mapeo et al., 2006) following the end of chemical (carbonate and BIF) sedimentation, a clastic epicontinental sea formed on the craton, into which sediments from various sources, mainly of Palaeoproterozoic age with minor inputs from Archaean provenances (Mapeo et al., 2006) were deposited. Deposition of clastic sediment was accompanied by the extrusion of significant thicknesses of volcanic deposits, thereby largely hindering the further growth of stromatolites. Studies on pyrites in marine shales of the Rooihoogte and Timeball Hill Formations at the base of the Pretoria Group suggest that the first major increase in global atmospheric oxygen may have occurred during the deposition of these formations (Bekker et al., 2004).

Alternating mudstone and quartzose sandstone formations characterize the volcano-sedimentary succession of the Pretoria Group (Schreiber, 1991), in which three main volcanic units are identified. These volcanic units comprise the Bushy Bend Lava Member of the Timeball Hill Formation, the Hekpoort Formation, and the Machadodorp Member of the Silverton Formation, in ascending stratigraphic order. In addition, thin tuff and tuffaceous shale beds are scattered throughout the Silverton and Vermont Formations, to a lesser extent in the Nederhorst and Houtenbek Formations, and are locally found at the base of the Magaliesberg Formation (Reczko, 1994). A generalized lithostratigraphic profile for the Pretoria Group is presented in Fig. 1. The Pretoria Group comprises 14 formations, starting with the basal Rooihoogte Formation and ending with the Houtenbek Formation at the top where it is unconformably capped by the Rooiberg Group lavas of the Bushveld Large Igneous Province (Lenhardt and Eriksson, this volume). The upper five formations are only preserved in the east of the Transvaal Basin, with a single western equivalent lacking any continuous outcrops in between (Eriksson et al., 1998; Eriksson et al., 2001). The depositional age of the onset of Pretoria Group sedimentation is partially constrained by a Re/Os age of 2322±16 Ma derived from mudstones from the lower part of the Timeball Hill Formation (Hannah et al., 2004) and a SHRIMP U–Pb detrital zircon age of 2324±17 Ma from the Timeball Hill Formation (Dorland et al., 2004), near the base of the group. The timing of the cessation of Pretoria Group sedimentation is unknown as yet. A minimum age is given by the extrusion of the extensive Rooiberg Group lavas (2061±2 Ma; Walraven, 1997) at the top of the Transvaal succession, that represent the oldest part of the Bushveld Large Igneous Province (Ernst and Bell, 2010).

The depositional palaeoenvironment of the Pretoria Group is thought to have been either an epeiric marine setting (Willemse, 1959; Visser, 1969; Button, 1973, 1986; Button and Vos, 1977; Eriksson et al., 2001, 2006) or an intracratonic basin with short-lived marine incursions (e.g., Crockett, 1972). Thin, lenticular diamictites found within the Timeball Hill Formation (Eriksson and Altermann, in press) reflect fluvioglacial to predominantly glaciomarine periglacial deposits (Visser, 1971; De Villiers and Visser, 1977; Eriksson et al., 1993; Evans et al., 1997). The Palaeoproterozoic glaciation, however, is a complex issue with highly divergent views – such as those supporting the "Snowball Earth hypothesis" (e.g., Kirschvink, 1992; Hoffman et al., 1998) and those diametrically opposed to this (e.g., Williams, 2004; Young, 2004). It is beyond the purposes of this paper to become involved in a debate as yet essentially unresolved.

Eriksson et al. (1991) proposed a continental rift tectonic setting with half-graben development for the Pretoria Group. This model encompasses a polycyclic rift stage, followed by thermal subsidence, with expansion of the basin during Silverton Formation times (Eriksson and Reczko, 1995). The rift hypothesis has been applied by von Gruenewald and Harmer (1993) for the Pretoria Group volcanic rocks, which are thought to be precursors to the succeeding Bushveld Complex magmatism. More recently, a geodynamic model encompassing two second-order rifting-thermal subsidence cycles has been espoused (Catuneanu and Eriksson, 1999; Eriksson et al., 2001, 2006).

This article briefly reviews the existing literature on the Pretoria Group volcanism in conjunction with new or as yet unpublished field and minor new geochemical data (a major basin-wide geochemical study of the entire Pretoria Group, including its volcanic units was performed by Reczko, 1994), and examines the emplacement of the three main volcanic units within the Pretoria Group; by studying the various eruptional, transport and depositional processes inferred, we will attempt an explanation for the tectonic control of the volcanism. This will be related to geodynamic models for Pretoria Group basin evolution.

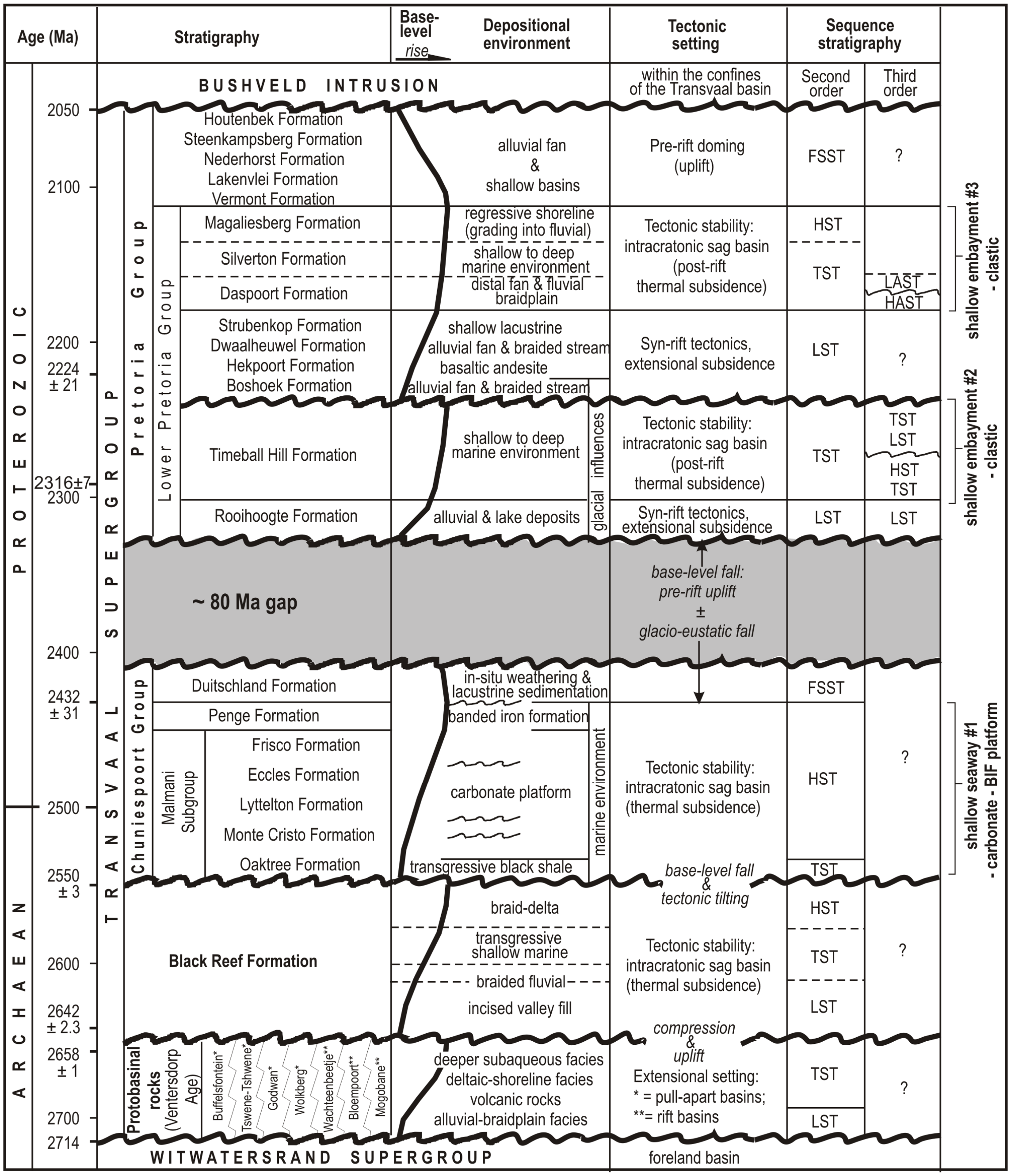

Figure 1. Summary of the Transvaal Supergroup stratigraphy (Eriksson et al., 2001).

Geological setting of the Pretoria Group and its three main volcanic units

A hiatus between 80 My (based on minimum age of Chuniespoort Group of 2432±31 Ma [Trendall et al., 1990], and maximum age of Pretoria Group of 2316±7 Ma [Hannah et al., 2004]; Eriksson and Reczko, 1995; Eriksson et al., 1995, 2006) and 200 My (based on detrital zircon data of major sandstones in the Pretoria Group (Timeball Hill, Daspoort and Magaliesberg formations); Mapeo et al., 2006) marks the transition from the underlying Chuniespoort Group to the deposition of the mudstone–sandstone, lavas and pyroclastics with intercalated subordinate conglomerate, diamictite and carbonate beds of the Pretoria Group of the Transvaal Basin (Dankert and Hein, 2010) (Table 1). Erosion has widely reduced the area that was formerly covered by the Pretoria Group and it is now mainly confined to an area extending from eastern Botswana (where it is correlated with the Segwagwa Group; Mapeo et al., 2006) around the Transvaal Basin, including the smaller Potchefstroom sub-basin along its southern margin (Reczko et al., 1995a), to the eastern escarpment of the Mpumalanga Province. Correlated successions occur within the small Kanye basin of Botswana and in the Griqualand West basin in the Northern Cape Province of South Africa where it is known as the Postmasburg Group (Fig. 2a). In the centre of the Transvaal basin, the Pretoria Group is covered by rocks of the Bushveld Complex, and is only exposed along the margins thereof in a number of floor- and roof-attached (Hartzer, 1995) "fragments", lying to the northeast and northwest of Pretoria, and in a basin-marginal main outcrop zone (Fig. 2b). The Pretoria Group reaches a maximum thickness of up to 7500 m in the east of the Transvaal basin (Eriksson et al., 1993). All rocks have been subjected to low-grade metamorphism. A general sheet-like geometry is evident for most of the lower formations, with certain sandstone and lava units exhibiting more wedge-like three-dimensional forms (Eriksson et al., 2006). Palaeocurrent studies indicate variable source regions for clastic sedimentary rocks (Schreiber et al., 1991; Eriksson and Reczko, 1995; Eriksson et al., 1995, 2001). The Timeball Hill Formation has inferred northwesterly and northeasterly sources whereas both Hekpoort and Magaliesberg Formations have more prominent westerly sources as suggested by thickness patterns (Eriksson et al., 2001; Mapeo et al., 2006). Large populations of detrital zircons from the Pretoria Group yield an average age of 2781±7 Ma which is identical to the age of the Gaborone Igneous Complex in Botswana (Grobler and Walraven, 1993; Walraven and Grobler, 1996). These authors therefore suggest that this igneous complex was the main source of the clastic sedimentary rocks, indicating a proximal source for the basal succession of the Pretoria Group. They propose that a further major source of the Pretoria Group sedimentary rocks was either the Zimbabwe Craton or a Palaeoproterozoic terrain, which was located north or northwest of the Kaapvaal Craton and west of the Zimbabwe Craton (c.f., Mapeo et al., 2006). Detrital zircons from Pretoria-equivalents in the Kanye basin date at 2250±14/15 Ma for the Timeball Hill sandstones near the base, 2236±13 Ma in the medially-situated Daspoort Formation, and 2193±20 Ma in the upper Magaliesberg Formation (Mapeo et al., 2006; comparable data in Dorland et al., 2004). So far there are no known data concerning source areas for the volcanic rocks of the Pretoria Group.

The up to 1100 m thick Timeball Hill Formation, overlying the Rooihoogte Formation and unconformably covered by the Boshoek Formation, is mainly made up of shales, quartzose sandstones (locally recrystallised to quartzites), lesser conglomerate lenses and diamictites (Visser, 1969; Button, 1973, 1986; Eriksson, 1973; Eriksson and Clendenin, 1990). The formation is the product of dominantly shallow marine sedimentation (Catuneanu and Eriksson, 2002) fed by fluvio–deltaic systems advancing into an epeiric embayment basin open to the ocean on the southeast (Visser, 1971; Eriksson, 1973; Coetzee, 2002; Eriksson and Reczko, 1998). No regional unconformity bounded units are known and the facies architecture is essentially comprised of sheet-like units (Eriksson et al., 1994a). The shallow water basin which developed in early Timeball Hill times was characterized by organic-rich (presumably cyanobacterial remains) prodeltaic muds settling out of suspension associated with widespread volcanic activity, which included localised deposition of lava, pyroclastic eruptions and fumarolic emissions, resulting in a basal black shale facies (2316±7 Ma; Re-Os; Hannah et al., 2004) throughout most of the basin (Eriksson et al., 1994a). The volcanic rocks associated with the lower part of the Timeball Hill Formation (the Bushy Bend lava Member) reach their maximum thicknesses in the Potchefstroom region in the southern part of the basin from where they thin towards the north (Coetzee, 2002). In the Potchefstroom area, boreholes have penetrated up to 39 m (with a mean thickness of 30 m) of highly altered lava over an area of approximately 30 km by 15 km (Eriksson et al., 1994a, b; Reczko et al., 1995b). These lavas lie immediately above the Rooihoogte Formation conglomerates and are in turn succeeded by black pyritic shales of the Lower Timeball Hill Formation (Eriksson et al., 1994a, b, 1995). In the eastern part of the Transvaal Basin, the Bushy Bend Member consists of a highly altered tuff, only a few meters thick (Reczko et al., 1995b).

The 300-830 m thick Hekpoort Formation in the Transvaal basin crops out over an area of c. 100,000 km^2. Within the Griqualand West basin and as part of the Postmasburg Group, which is the western equivalent of the Pretoria Group (Button, 1973; Tankard et al., 1982; Beukes, 1983), is the Ongeluk Formation, correlated with the Hekpoort through similar ages: 2224±21 Ma (whole-rock Rb-Sr isochron; Walraven and Martini, 1995) for the Hekpoort Formation and 2222±12 Ma (whole rock Pb-Pb isochron; Cornell et al., 1996) for the Ongeluk Formation, respectively. Together with the comagmatic Ongeluk Formation of the Griqualand West basin (Eriksson and Reczko, 1995) the tholeiitic andesites of the Hekpoort Formation are thought to have originally covered c. 500,000 km^2 of the Kaapvaal Craton (Cornell et al., 1996). In most parts of the Transvaal basin, the Hekpoort Formation sharply overlies the subaerially emplaced Boshoek conglomerates and sandstones with a sharp contact (Cheney, 1996), and is unconformably succeeded by the Dwaalheuwel continental sandstones (Eriksson et al., 1993). In the central-southern parts of the basin the Hekpoort Formation sharply overlies mudrocks of the Timeball Hill Formation, and is similarly overlain by Strubenkop Formation mudrocks (Fig. 1 and Table 1). In the Griqualand West basin, the basaltic lava flows of the Ongeluk Formation overlie glacigenic rocks of the Makganyene Formation and are characterized by evidence for deposition in a subaqueous setting along the submerged western margin of the Kaapvaal Craton (Grobler and Botha, 1976; Gutzmer et al., 2003). The Hekpoort Formation dips gently to the N-NW at c. 15° and is characterized by a complex interplay of subaerial lava flows, pyroclastic deposits, and their reworked counterparts (Oberholzer, 1995) representing deposition in a subaerial palaeoenvironment (Button, 1973; Res, 1993; Eriksson and Reczko, 1995; Oberholzer, 1995). Although there has been no regional study so far to establish the proportions of lava flows and volcaniclastic detritus in the Hekpoort Formation, Sharpe et al. (1983) suggested that lava flows predominate. In the east of the preserved Transvaal basin, Button (1973) found that volcaniclastic rocks only made up approx. 10% of the formation, and that they were concentrated towards its base. In the central-southern part, however, Oberholzer and Eriksson (2000) found that volcaniclastic rocks present a significant portion of the lithology of the formation.

The predominantly argillaceous Silverton Formation (Fig. 1) has a basin-wide extent and varies in total thickness from c. 2000 m in the east of the Transvaal Basin to several hundred meters in the west (Button, 1973). Erosional removal characterizes most of the southern and central preserved occurrences (Eriksson et al., 2002). The Silverton Formation is characterized by various mudrocks (approximately 80%), locally significant volcanic rocks (the Machadodorp Member), and minor carbonates, cherts and sandstones (Schreiber, 1991). According to Catuneanu and Eriksson (1999) the Silverton shales reflect the transgressive systems tract of an epeiric sea. A lowermost, westerly arenaceous facies association is ascribed to a braid-deltaic and turbidity current deposition, whereas the predominant argillaceous deposits are interpreted as sub-storm wave base pelagic sediments (Eriksson et al., 2002, 2008).

The lower contact of the formation is gradational, locally with mudstones interlayered with cm- to dm-thick beds of immature sandstone resting on the arenaceous Daspoort Formation (Fig. 3). The upper contact with the overlying arenaceous Magaliesberg Formation is gradational and upward-coarsening (Button, 1973; Schreiber, 1991; van der Neut, 1990). Furthermore, gradational contacts between the lower Boven Shale Member, the medial Machadodorp Volcanic Member and the upper Lydenburg Shale Member are observed within the Silverton Formation. The three units average 250 m, 300 m, and 1250 m in thickness, respectively, the greatest thicknesses being encountered in the east of the Transvaal basin with the lavas being restricted to that portion of the depository (Schreiber, 1991; Eriksson et al., 2002).

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)	Ditlhojana Volcanic Fm (volcanic rocks)		Hekpoort Fm (andesite)		
					Ditlhojana and Tlaameng Fms (shale, sandstone, conglomerate)	Ditlhojana Shale Fm	Conglomerate (minor)	Boshoek Fm (conglomerate, sandstone)		
		Makganyene Fm (diamictite)					Shale (predominant)	Shale	Timeball Hill Fm	
						Tsokwane Quartzite Fm (sandstone)		Sandstone		
						Lephale Fm	Shale	Shale/lava		
							Conglomerate, sand-stone	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		

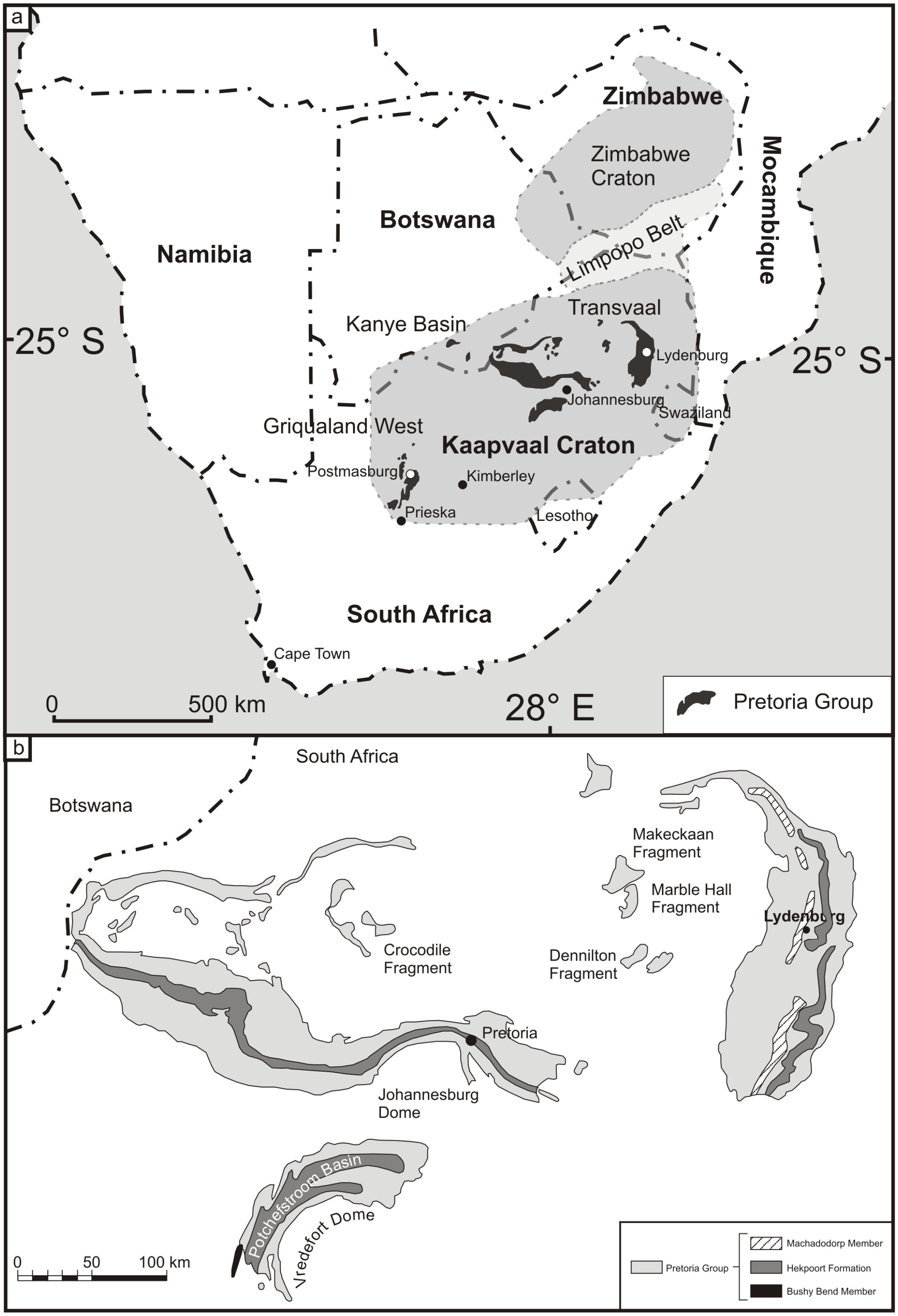

Figure 2. Map showing (a) the locations of the Kaapvaal Craton and the Transvaal Basin in South Africa, and (b) the locations of the three main volcanic units (Bushy Bend Lava Member, Hekpoort Formation, Machadodorp Member) within the Transvaal Basin.

The Bushy Bend Lava Member (Timeball Hill Formation)

The up to 90 m thick Bushy Bend Lava Member is characterized by a sharp contact with the conglomerates of the underlying Rooihoogte Formation and a gradational upper contact with the Lower Shale Member of the Timeball Hill Formation. The lithology of the Bushy Bend Lava Member is dominated by lava with subordinate tuffs and clastic sedimentary rocks. The lavas vary from fine crystalline to amygdaloidal. Single lava flows with thicknesses varying between 0.3 and 12 m can be distinguished by their chilled and undulating flow bottoms (Eriksson et al., 1994a, b); rounded and deformed amygdales below and above the flow contacts enhance the discrimination of individual flows. Amygdales are filled with zeolite, chlorite, quartz and/or calcite (Eriksson et al., 1994b). Locally, flows are separated by beds of several metres-thick laminated tuffs with interbedded siltstone (see Fig. 3, borehole C). Phenocrysts consist of plagioclase and amphibole pseudomorphously developed after clinopyroxene (Eriksson et al., 1994b), and the phenocrysts are set in a fine-grained matrix of similar material. The lavas are epidotized and sericitized, and in addition show extensive veining and/or brecciation with veinlets filled by calcite and, sporadically, quartz.

The only geochemical data for the Bushy Bend lavas have been provided by Eriksson et al. (1994b), Reczko et al. (1995b) and Coetzee (2002). The first two research groups report high K contents that are interpreted as either high primary K contents or, alternatively, related to a syn- or post-extrusive phase of K-metasomatism. Coetzee (2002) classifies the Bushy Bend lavas by means of trace elements as subalkaline to alkaline basalts and favours post-exstrusive K-metasomatism as the cause for the high K content.

In addition to the main occurrence of the predominant lavas near Potchefstroom (where they are only known from boreholes) several authors have described analogous volcanic rocks in other parts of the basin that are likely to be correlates of the Bushy Bend Lava Member. Key (1983) has found a 1 m-thick pyritic bed in the lower Timeball Hill Formation in eastern Botswana, where the sulphides are associated with feldspathic tuffs. Klop (1978) found a chert interbed in the Gopane area of the southwestern Transvaal basin in the basal pyritic black shales of the Timeball Hill Formation, which contains possible altered glass shards. In the Pretoria region, van der Neut (1990) described a mudclast conglomerate immediately underlying the Timeball Hill black shales, which he interpreted as reworked and weathered products of basaltic rocks by means of XRF and XRD analyses. In the east of the Timeball Hill basin, a several dm-thick weathered tuff occurs within the basal black shales, and extends for approx. 60 km along strike, north of Carolina (Eriksson et al., 1994b; Reczko et al., 1995b).

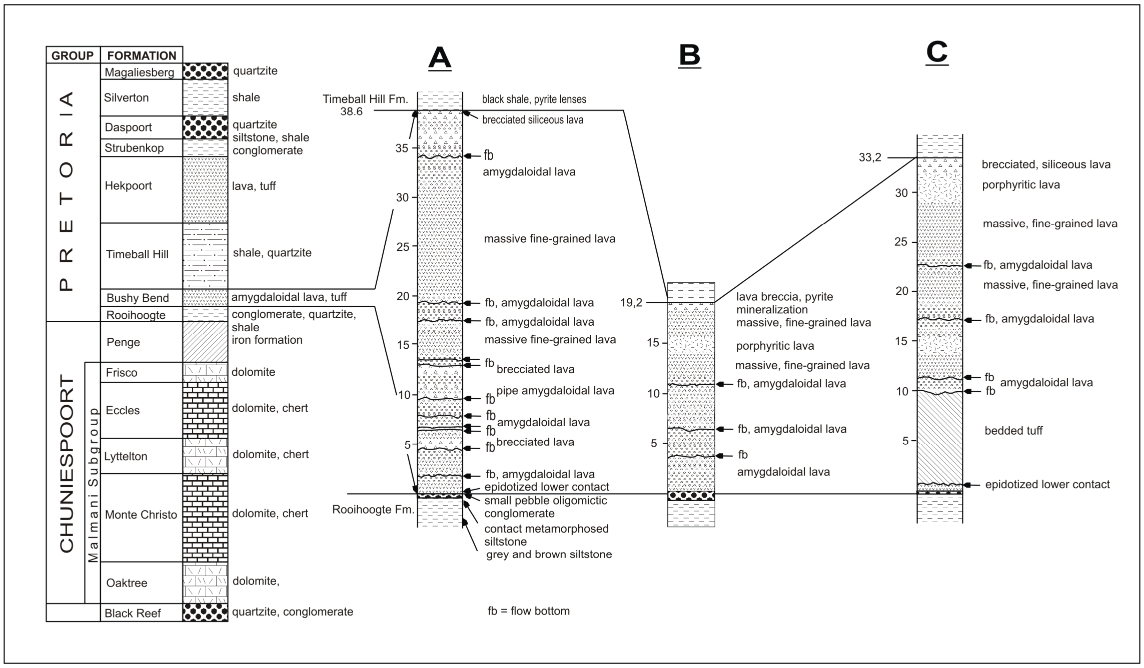

Figure 3. Lithostratigraphy of the Bushy Bend Lava Member within three boreholes in the Stilfontein area, Potchefstroom region (after Eriksson et al., 1994b).

The Hekpoort/ Ongeluk Formations

The Hekpoort Formation

The Hekpoort/ Ongeluk volcanism marks a major volcanic event within the Transvaal Supergroup. Although limited regional data (Button, 1973; Sharpe et al., 1983; Engelbrecht, 1986) indicate that volcaniclastic rocks within the Hekpoort Formation are subordinate to basaltic-andesitic lava flows, in the preserved Transvaal Basin (TB) the work of Oberholzer (1995) and Oberholzer and Eriksson (2000) shows that the formation in the southwestern part of the basin has essentially equal proportions of these lithologies. In excess of 1100 m of basaltic-andesitic lava is preserved in the south of the TB, with over 800 m in the west of the basin, but these rocks thin to the northeast where less than 50 m is found. Concomitant with the northerly thinning is a reduction in the number of individual lava flows (Button, 1973).

Single lava flows reach thicknesses up to 60 m, appear massive to amydaloidal, and can locally develop several meters-thick flow top breccias. Amygdales are typically developed in the upper part of the flows and are often coalesced and typically filled by quartz or lined by chlorite and zeolites. The relatively fine-grained lavas contain microphenocrysts of feldspar and chloritized pyroxene set in a groundmass of chlorite and feldspar laths (Oberholzer, 1995). Secondary phases are characterized by chlorite, K-feldspar, quartz, actinolite, epidote and pyrite; the replacement of magmatic by secondary phases is more pervasive in amydaloidal lavas than it is in massive lavas (Coetzee, 2002).

The few metres-thick flow-top breccias are grey-green in colour and are composed of angular to sub-angular lava clasts, 0.1 – 1m in diameter, set in a fine-grained matrix. The clasts are composed of the same mineral assemblage as the lavas but are described as being finer grained in texture. Coetzee (2002) describes the matrix between the clasts as composed of partially devitrified volcanic glass shards and lava particles. Furthermore, he observed flow lamination in the matrix between clasts in several locations.

Primary and secondary volcaniclastic rocks, ranging from lapilli tuffs and pyroclastic breccia to coarse-grained diamictites occur as lens-shaped bodies intercalated with the Hekpoort lavas. The lapilli tuffs are characterized by well-developed, normally graded bedding, with fragment sizes in the range of 2-64 mm. The clasts are well rounded. Compositions of the fragments are basaltic-andesitic, scoracious material and lithic fragments of mudstone and sandstone. The latter rocks were probably derived from the underlying Timeball Hill and Rooihoogte Formations, and possibly from minor such lithologies in the preceding Chuniespoort Group (Oberholzer and Eriksson, 2000). The massive lapilli-tuff is interpreted as an ash-flow deposit and such rocks are described by many authors as the most common ignimbrite lithofacies (e.g. Ross and Smith, 1961; Sparks, 1976; Wilson and Walker, 1982; Branney and Kokelaar, 2002; Lenhardt et al., 2011).

According to Oberholzer (1995) the pyroclastic breccias are rich in block- and bomb-sized fragments, some exceeding 1 m in diameter. The clasts are characterized by a high degree of variation in both rounding and sphericity. Fragments range from dense basaltic-andesitic through vesicular scoracious material to lithic fragments of mudstone, sandstone and chert. These rocks exhibit no internal stratification and instead have a massive appearance. The matrix of both pyroclastic lithofacies types exhibits a grey-green colour, and consists of very fine- to coarse-grained tuffaceous material. The pyroclastic breccia is interpreted to be derived from block-and-ash flows (Oberholzer and Eriksson, 2000). Block-and-ash flow deposits generally result from small-volume pyroclastic flows, generated by explosive disruption or the sudden gravitational collapse of lava flows or domes (Wright et al., 1980; Francis, 1994; Freundt et al., 2000). Block-and-ash flow deposits are mostly restricted to magmas of dacitic to andesitic composition (e.g., Saucedo et al., 2002; Cole et al., 2002; Busby et al., 2004). However, basaltic block-and-ash flows are also possible (e.g., Kobayashi et al., 2003; Miyabuchi et al., 2006).

The coarse-grained diamictites are characterized by a high degree of rounding of the block-sized fragments, as well as by a sub-horizontal planar fabric with a shaly appearance (Oberholzer and Eriksson, 2000). Fragment sizes range from 1-80 cm, with an average of 15 cm. Another important feature of these rocks is the high degree of alteration they have undergone. This alteration is apparent from the mineralogy, where primary minerals such as plagioclase and clinopyroxene are replaced by epidote group minerals like clinozoisite as well as amphibole, zeolites and clay minerals. These rocks are interpreted as coarse volcanic debris flows (Eriksson and Twist, 1986; Oberholzer and Eriksson, 2000), possibly deposited by lahar processes that acted throughout deposition of the Hekpoort Formation (e.g. Walton and Palmer, 1988). The term lahar as used here describes all rapidly flowing mixtures of rock debris and water (other than normal streamflow) from a volcano (cf., Smith and Lowe, 1991).

Non-volcaniclastic sedimentary rocks are rarely encountered within the Hekpoort Formation. Button (1973) describe a thin shale interbedded with the lavas southeast of Machadodorp (Fig. 2). The shale is described as being green-grey with a “flinty texture and […] a sub-conchoidal fracture”, which is interpreted as being probably due to the heat of the overlying lava flow. Button (1973) further described laminated shales from the Penge area (NE of the TB), which had been altered to andalusite- and biotite-bearing hornfels by the intrusion of the Bushveld Complex. Thin lenses of clastic sedimentary material have been found in a number of localities. These occurrences, mainly shales but in one case also argillaceous quartzite, are not laterally persistent, and are generally to be found in the lower third of the formation. Their lenticular shape suggests that they were deposited in local lakes developed on the surfaces of lava flows during periods of volcanic quiescence (Button, 1973).

To date, the Hekpoort Formation is best studied in the southwestern part of the Transvaal basin between Skeerpoort and Hartebeeshoek where it reaches thicknesses of up to 600 m. Here, Oberholzer and Eriksson (2000) describe the basaltic-andesitic formation as follows: stratigraphically, a lower unit of approx. 100 m and made up of lava flows is succeeded by lenticular volcaniclastic rocks, with a maximum thickness of 50 – 100 m and extending along strike for c. 5000 m (Fig. 4). A 400 – 450 m thick volcanic succession follows, composed of lava flows in the southwest, and lenticular volcaniclastic rocks in the northeast. Estimated maximum thickness of the NE lens is 450 m, laterally extending for at least 8 km. The volcanic succession is overlain by a thin, 9 m thick tuffaceous mudrock, followed by 50 – 100 m of lava, with a laterally restricted, 100 – 150 m thick lens of volcaniclastic rocks (Fig. 4).

Oberholzer (1995) describes all Hekpoort lavas as basalts to basaltic andesites which plot according to Coetzee (2002) into the within-plate tholeiitic basalt field. Furthermore, the lavas are characterized by high MgO (> 8 wt%), low TiO_2 (< 0.5 wt%) and intermediate SiO_2 (average of 53 wt%) values. Although the pyroclastic rocks are relatively highly altered, geochemical data show no significant differences in immobile elements to that of the lava flows (Oberholzer and Eriksson, 2000). The variable silica contents in the mineralogically relatively uniform Hekpoort lavas may point to secondary processes of silica enrichment (Reczko, 1994). The silica contents of individual samples may have been increased compared to their primary contents due to a changing degree of silicification related to alteration processes (Reczko et al., 1995b).

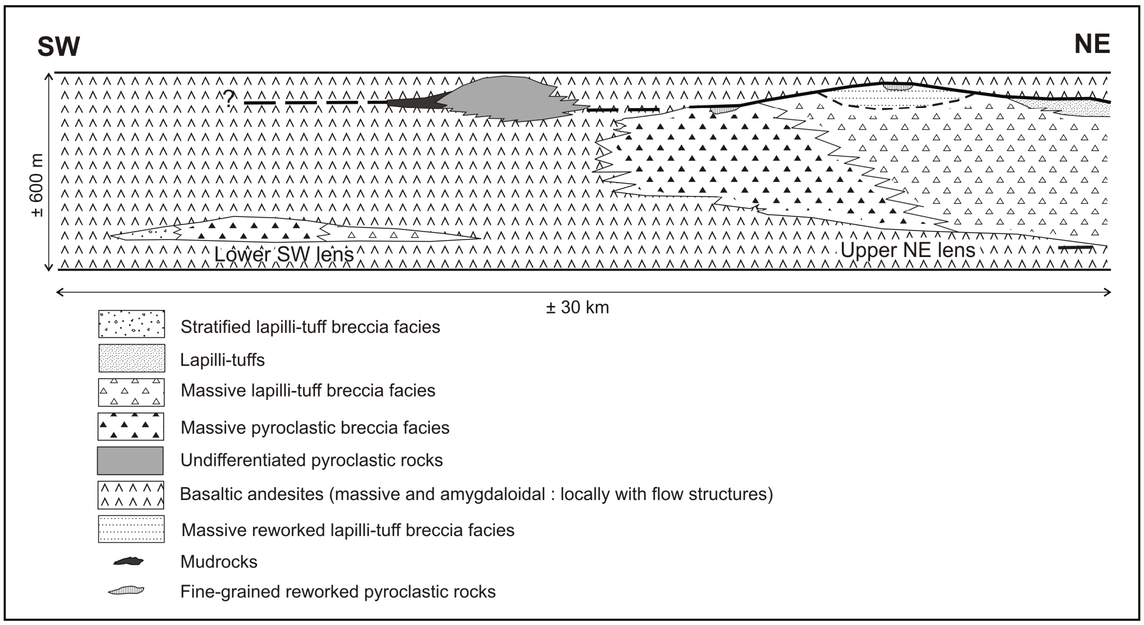

Figure 4. Schematic cross-section through the Hekpoort Formation between Skeerpoort and Hartebeeshoek, parallel to the NE-SW strike of the unit (Oberholzer and Eriksson, 2000).

The Ongeluk Formation

In the Griqualand West basin (GWB), the Hekpoort Formation correlate is the volcanic Ongeluk Formation of the Postmasburg Group. The Ongeluk Formation is characterized by an up to 900-m-thick succession of massive basaltic andesites, pillow lavas and hyaloclastites indicating subaqueous extrusion of the volcanic rocks along the submerged western margin of the Kaapvaal Craton (Grobler and Botha, 1976; Cornell et al., 1996). According to Cornell et al. (1996) the majority of the Ongeluk volcanic rocks are basaltic andesites and exhibit tholeiitic rather than calc-alkaline character, with low alumina levels of 14.5%. The Ongeluk Formation disconformably overlies the Makganyene Formation glacigenic lithologies (Altermann and Hälbich, 1991). The volcanic rocks of the Ongeluk Formation are conformably overlain by Superior-type banded iron-formations, manganese-formations and shallow marine dolostones that constitute the Voelwater Subgroup of the Transvaal Supergroup in Griqualand West (Beukes, 1983).

Dark-green, massive basaltic andesitic lava flows are the most prominent rock type in the Ongeluk Formation. The basal portion of the flows tends to be free of amygdales which become more abundant towards the top. Flow breccias are often found near the tops of the flows (Cornell et al., 1996). According to Gutzmer et al. (2003) the basaltic andesites of the Ongeluk Formation are typically microcrystalline and dark grey to greenish-grey in colour. Igneous textures are mostly well preserved and range from predominantly subophitic and intergranular to intersertal and glomeroporphyritic. Subhedral prismatic augite crystals (up to 0.75 mm long) are typically intergrown with euhedral plagioclase laths of similar size. Lath-shaped labradorite phenocrysts (1.5 mm long) and glomerophyritic clusters of stubby Ti-augite phenocrysts (0.5 mm in size) occur as minor constituents (Schutte, 1992). Interstices between labradorite and augite crystals are filled by a submicroscopic groundmass containing very thin needles of plagioclase. Trace amounts of zircon, titanomagnetite, ilmenite and hematite occur. The primary igneous mineral assemblage has suffered variable degrees of alteration. Ti-augite is partly altered to a fine-grained mass of pumpellyite, chlorite and quartz, whereas labradorite is replaced by albite and fine-grained pumpellyite. The groundmass is converted to a dense mass of pumpellyite, chlorite, albite and quartz. Primary pore spaces are lined by chalcedony and filled by coarse crystalline Mg–Fe chlorite, pumpellyite and rare calcite. Minute grains of chalcopyrite, pyrite and galena occur in the altered groundmass (Gutzmer et al., 2003).

Pillow lavas are common throughout the Ongeluk Formation and are characterized by very dark-green rims, cooling cracks, and rounded three-dimensional forms, with either vesicular or massive cores (Cornell et al., 1996). The hyaloclastites within the Ongeluk Formation are described as consisting of fragments of massive lava in a matrix of angular glass shards (Cornell et al., 1996). No graded bedding was observed. Alteration of hyaloclastite is evidenced by the development of brown palagonite and epidote or piedmontite along shard boundaries, whereas the matrix consists largely of zeolites and chlorite (Cornell et al., 1996). The term hyaloclastite is used for clastic aggregates formed by non-explosive fracturing and disintegration of quenched lavas and intrusions (McPhie et al., 1993; Batiza and White, 2000). The origin of the fragments is interpreted as having been in response to thermal stress, built up during rapid cooling, and stress imposed on chilled outer parts of lava flows and intrusions by continued movement of the ductile interior (Kokelaar, 1986). The quench fragmentation usually affects subaerially erupted magma that flows into water (e.g., Waters, 1960; Moore et al., 1973) and lava that erupted subaqeously (e.g., Dimroth et al., 1978; Bergh and Sigvaldason, 1991; Kano et al., 1991). The clasts are formed *in situ* and normally show a typical jigsaw-texture with clasts that fit more or less neatly together (McPhie et al., 1993). Resedimented hyaloclastites usually show evidence for transport such as sedimentary structures and an absence of the jigsaw-texture. Neither bedding nor jigsaw texture was observed by Cornell et al. (1996) for which reason only a short distance of transport can be inferred.

The Machadodorp Volcanic Member (Silverton Formation)

The Machadodorp Member lavas and volcaniclastic rocks of the Silverton Formation reach a maximum thickness of 500 m (300 m on average), and a strike length of approx. 220 km (Button, 1973; Schreiber, 1991) in the eastern part of the basin; sporadic lavas and tuffs occur at the same stratigraphic level in the centre and west of the Transvaal basin (Eriksson et al., 1990). The basal part of the Machadodorp Member is formed by a coarsening-upward sequence of volcaniclastic rocks, ranging from relatively fine-grained tuffs to tuff-breccias and fluidal-clast breccias. The volcaniclastic rocks are overlain by sheet-lavas, with local occurrences of pillows (Lenhardt and Eriksson, 2011). Bedding within the underlying Boven Shale and the fine-grained tuffs at the base of the Machadodorp Member generally dip towards the NW, at angles up to 18° - 20°.

The Machadodorp Member lavas are characterized by their flat, sheet-like morphology. They exhibit a dense core with vesicles increasing in number and size (up to 3 cm) within the carapace. In drill cores, the member exhibits thicknesses up to 250 m. Button (1973) described several localities with prominent pillow structures, especially NW of Carolina and S of Lydenburg. However, the majority of the lavas can be described as sheet-lavas. The lavas have a dark grey colour and a fine-grained, aphyric texture. Therefore, no macroscopic distinction is apparent in the field. In thin section, however, they exhibit a composition of clinopyroxene, plagioclase and olivine microphenocrysts embedded in a microlithic matrix of tremolite-actinolite and epidote (Lenhardt and Eriksson, in prep.). Where present, olivines have been variably serpentinized and chloritized. Sericitization of plagioclases and chloritization and calcitization of pyroxenes are common. The massive lavas represent non-channelized sheet flows and are characteristic of higher effusion rates and temperatures compared to pillowed flows (Yamagishi, 1991; Gregg and Fink, 1995, 2000; Gregg and Smith, 2003). According to Crow and Condie (1990) the Machadodorp volcanics are exclusively tholeiitic basalts and exhibit flat to LREE depleted patterns similar to MORB (cf., Button, 1974). The basalts are slightly enriched in Ba, Th and Rb (Navaya et al., 2011), suggesting either crustal contamination or derivation from a MORB-like source to which a very minor subduction zone component had been added. However, any subduction zone geochemical component must have been acquired during earlier subduction regimes when the craton and lithosphere grew by successive arc collisions largely before ca. 2.8 - 3 Ga (cf., Crow and Condie, 1990).

The volcaniclastic rocks of the Machadodorp Member consist of fine-grained tuffs, accretionary lapilli tuffs, agglomerates or fluidal-clast breccias, and planar-bedded lapilli tuffs. The fine-grained, dark grey to black tuffs can be found in the lower part of the Machadodorp Member, on top of the Boven Shale Member and represent the onset of volcanic activity within the Silverton Formation. Macroscopically, some of the tuffs can easily be confused with the underlying shales. The well sorted, laminated tuffs are distributed over a wide area and are interpreted as the fine products of subaqueous eruptions that settled through the water column.

The accretionary lapilli tuff consists of grey to brownish-beige accretionary lapilli in a lithified fine-grained matrix of recrystallized fine ash of the same colour. The rocks are poorly sorted and defined by alternations and gradations from lapilli-rich to lapilli-poor mudstone layers. The pellets are mostly spheroidal and 0.3 mm to 1.0 cm in size. Many of them are in mutual contact. The lapilli have been subjected to diagenetic and metamorphic changes which make the determination of the original morphology of the fine ash difficult. Lithic clasts are absent. The accretionary lapilli are interpreted to represent phreatomagmatic fall units. Accretionary textures typically form where water vapour facilitates adhesion of small ash grains, and do not form in the presence of excessive water (Schumacher and Schmincke, 1995), such as where tephra grains are entrained in water (White, 2000). They are generally considered to be restricted to the proximal facies (Schumacher and Schmincke, 1991). Although accretionary lapilli commonly have been used to identify subaerially erupted tephra, recent studies (e.g. White and Houghton, 2000) show that their formation is also possible in subaqueous settings within the steam envelope of an eruption column (c.f., Martin et al., 2004).

The fine- to medium-grained, planar-bedded lapilli tuffs reach thicknesses up to several 10s of metres and show diffuse alignment of the clasts. The poor sorting and the limited extent of the tuffs are consistent with deposition or redeposition from a nearby subaqueous eruption. They are interpreted as reflecting a combination of subaqueous sediment gravity flow processes and eruption driven ballistically emplaced ejecta (c.f., Mueller and White, 1992; Mueller et al., 2000).

The fluidal-clast breccia or agglomerate is composed of basaltic clasts that are fluidally shaped to blocky and splintery forms, and which are moderately to highly vesicular. These rocks are poorly sorted, internally massive and clast-supported, and occurrences reach thicknesses of several 10s of metres. Variations in the ratio of vesicular to non-vesicular clasts are gradational over thicknesses of several metres. The fluidal clasts are interpreted as volcanic bombs and fluidal lapilli that formed by the tearing-apart of relatively low-viscosity lava ribbons, jetted upward from vents during subaqeous Hawaiian-style fire-fountain eruptions (cf., Macdonald, 1972; Allen et al., 1997; Simpson and McPhie, 2001). Similar fluidal-clast breccias have been identified on the modern seafloor (Smith and Batiza, 1989), and in several other submarine volcanic successions, in Japan (Yamagishi, 1987; Cas et al., 1996), Sweden (Allen et al., 1996), Canada (Mueller and White, 1992), Germany (Schmincke and Sunkel, 1987) and Iceland (Kokelaar and Durant, 1983). The fire-fountain breccias rarely extend beyond tens of metres from the source vent (Kokelaar and Durant, 1983; Kokelaar, 1986) and are thus good indicators of proximity to a volcanic centre.

The best outcrops of the Machadodorp Member can be found in the eastern part of the Transvaal basin between Carolina in the South and Burgersfort in the North. Here, along a c. 220 km long ridge (Fig. 2b) an intercalation of tuffs and fine grained Boven Shale deposits is overlain by fine- to medium-grained, planar-bedded tuffs. Laterally and more proximal to the ridge, these planar-bedded tuffs grade into a fluidal-clast breccia with dominantly irregular shaped, stretched and deformed clasts, that exhibit an increase in blocky clasts up-section. Locally, layers of accretionary lapilli can be found intercalated with the planar-bedded tuffs. The volcaniclastic deposits are covered by sheet lavas and local occurrences of pillows.

Discussion and Conclusions

Nature and controls of the Pretoria Group volcanism

The 2324±17 Ma (Dorland et al., 2004) Timeball Hill Formation, incorporating in places up to 1500 m of sedimentary rocks, reflects a variety of different depositional settings (Button, 1973), which are at least partly relevant to the genesis of the Bushy Bend lavas at the base of this formation. The lowermost part of the Timeball Hill formation consists of black shales that were entirely deposited subaqueously, probably by gravity settling from suspension in a shallow epeiric sea; succeeded by predominant laminated mudrocks and siltstone-fine sandstone sheets that are interpreted as low density turbidites reworked from distal delta complexes with localised contourites (Eriksson et al., 2008) (Fig. 5). Other more sandy sedimentary rocks in a medial stratigraphic position, and showing cross-bedding, upward-coarsening, and flute or groove casts, are interpreted as emanating from fluvial deltas debouching into the epeiric Silverton sea (Visser, 1969, 1971; Button, 1973, 1986; Eriksson 1973; Eriksson and Clendenin, 1990; Schreiber, 1991; van der Neut 1990; Eriksson et al., 1993). The upper part of the formation repeats the predominant argillaceous – fine sandy facies found below the medial deltaic sandstones. Therefore, the Timeball Hill Formation is regarded as an ancient deltaic complex (Button, 1973), analogous to the marine basin-centre model (Eriksson et al., 1994a).

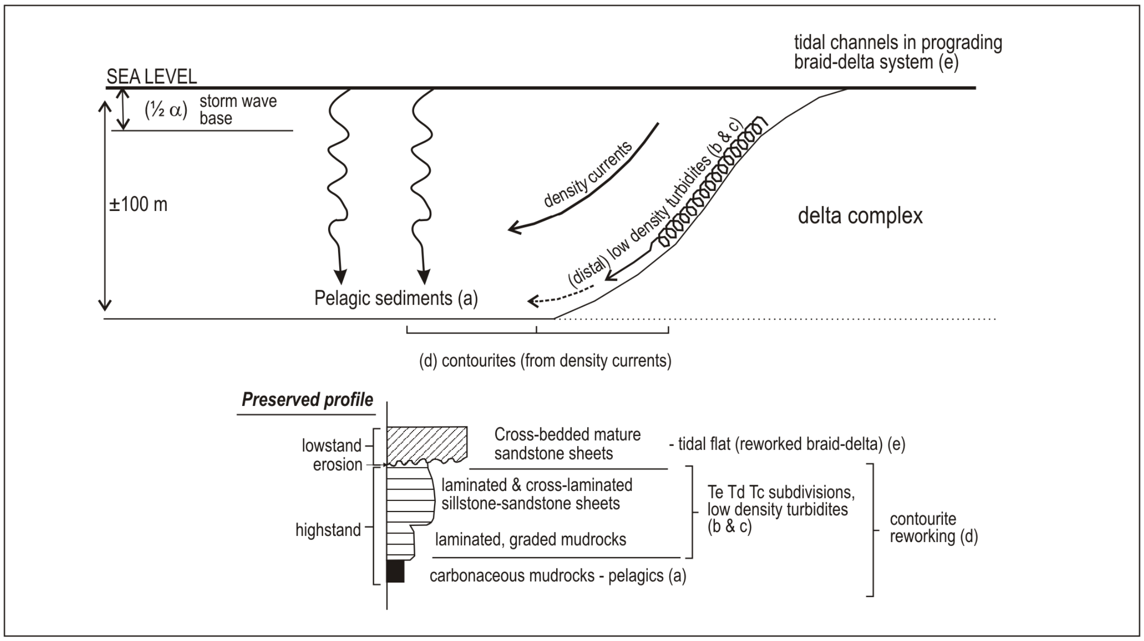

Figure 5. Clastic epeiric sea model inferred for the Palaeoproterozoic Timeball Hill Formation, Transvaal Supergroup (Eriksson et al., 2008).

The tabular, commonly amygdaloidal (sometimes exhibiting pipe amygdales), flow-banded, and locally porphyritic, andesitic-basaltic lava flows of the Bushy Bend Member, at the base of the Timeball Hill Formation, are generally interpreted as having been extruded under subaerial conditions (Eriksson et al., 1994b). According to Eriksson et al. (1994a) this suggestion is compatible with the chilled bases of individual lava flows and the lack of pillow lavas. On the other hand, the uppermost brecciated lavas at the contact with the overlying pyritic black shales may reflect more rapid subaqueous cooling of eruptive magma (Eriksson et al., 1994a). The occurrence of pipe amygdales and flow banding or the lack of pillow lavas cannot be seen as unambiguous evidence for a possible subaerial extrusion. Amygdales reflect the late degassing history of lavas (Orth and McPhie, 2003), representing vesicles that were formed by exsolution of volatiles from the lava and have been partially or completely infilled with secondary minerals (McPhie et al., 1993). Pipe vesicles, slender cylindrical cavities that when filled by secondary minerals become pipe amygdales, are commonly attributed to the exsolution of gas into bubbles that are attached to the zone of solidification of a lava (Philpotts and Lewis, 1987; Godinot, 1988). As this zone advances into the cooling lava, bubbles continue to grow, forming pipes perpendicular to the solidification front (McPhie et al., 1993). These pipe vesicles are commonly found near the bases of subaerial pahoehoe lava flows (Wilmoth and Walker, 1993) but also occur subaqeously in basaltic sheet flows (Dimroth et al., 1978) and pillow lavas (Kawachi and Pringle, 1988; Yamagishi et al., 1989; Walker, 1992) depending on confining pressure (Moore and Schilling, 1973; Cousineau and Dimroth, 1982). Flow banding can be envisaged as the result of mingling of two compositionally different lavas from a single eruptive fissure (Robins et al., 2010) and/or the stretching of a magma that contained pre-existing zones (vesicles or proto-vesicles) of contrasting water concentration, as the magma flowed in the conduit and on the surface (Seaman et al., 2009). It is frequently developed in evolved viscous lavas such as rhyolite and dacite but has also been observed in submarine basalt sheet flows (Expedition 309 Scientists, 2005), including some very thin glassy ones (Maicher et al., 2000) and basaltic pillow lavas (Robins et al., 2010). Finally, it can be said that submarine lava flow morphologies are usually dependant on effusion rates, angle of slope and tensile strength of the lava (Gregg and Smith, 2003) and that an increase in effusion rate normally favours the formation of sheet flows over pillows or lobes (e.g. Fink and Griffiths, 1990; Griffiths and Fink, 1992; Gregg and Fink, 1995). The fact that in borehole B, (Fig. 3), Bushy Bend lava fragments are found in the lowermost shales, and in borehole C shale fragments occur within the uppermost crystalline Bushy Bend lavas implies that the lowermost pyritic black shales of the Timeball Hill Formation and the uppermost lavas are approximate time equivalents (Eriksson et al., 1994a) and therefore also favours a subaqueous over a subaerial deposition for the Bushy Bend Lava Member. The combination of a greater number of lava flows and their greater cumulative thickness in the southern part of the Transvaal basin near Potchefstroom suggests that the flows probably issued from volcanic centres, most likely fissure vents, situated in this area (Button, 1973). The limited extent of the Bushy Bend lavas and the thin, generally tuffaceous volcanic rocks found at the same stratigraphic level in other parts of the basin suggest that short-lived, probably subaqueous and subaerial, eruptions occurred at many localities around the depository. It is thus suggested that the transition from the subaerial alluvial fans and shallow lacustrine conditions of the Rooihoogte Formation (Eriksson, 1988), to the black shales of the Lower Timeball Hill Formation, was accompanied by basaltic-andesitic fissure eruptions (Fig. 6) with fumarolic activity and probably minor subaerial explosive eruptions in the hinterland of the basin. Such volcanism could have provided a source of both iron and sulphur to form the features observed in the ferruginous black pigmentation of the Timeball Hill shales, and would most likely have been associated with sulphate reducing bacterial organisms which led to the formation of localized carbonaceous black shales (Eriksson et al., 1994a).

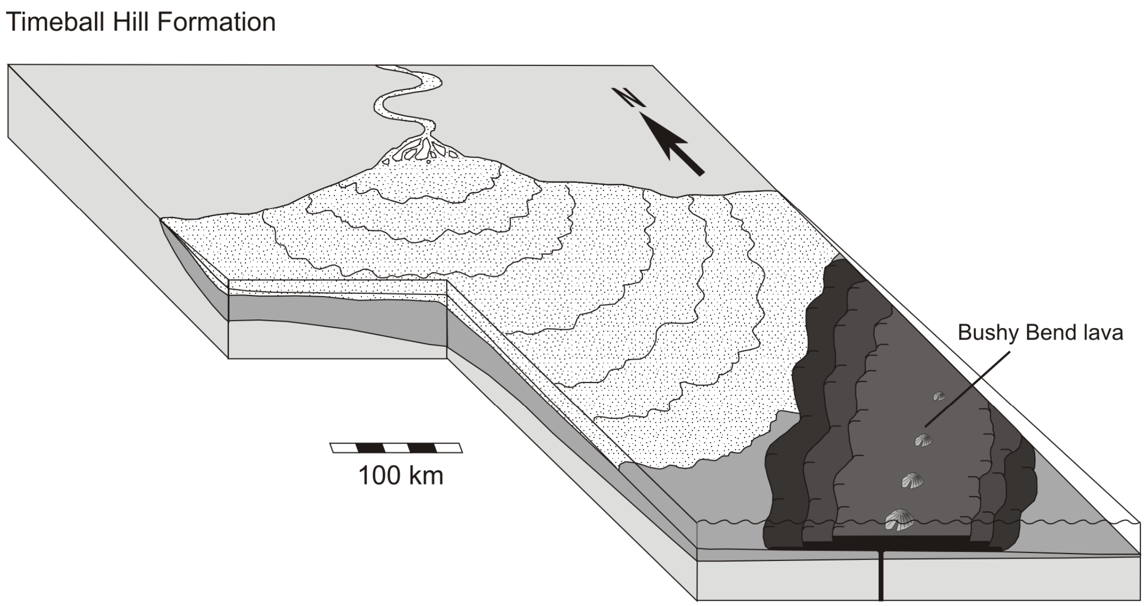

Figure 6. Reconstruction of the palaeoenvironment of the Timeball Hill Formation including the Bushy Bend lavas (modified after Oberholzer, 1995).

The uppermost black shales of the Timeball Hill Formation in the Potchefstroom and Pretoria areas are overlain directly by the Hekpoort lavas, which include tuffaceous shales in their lower portions (Visser, 1969; Engelbrecht et al., 1986; Eriksson et al., 1993). A similar volcanic source of black shale colouration may thus be postulated for these rocks (Eriksson et al., 1994a). The 2224±21 Ma (Walraven and Martini, 1995) Hekpoort Formation is described a being largely deposited in a subaerial setting (Visser, 1969; Button, 1973; Sharpe et al., 1983; Engelbrecht, 1986; Eriksson and Twist, 1986; Schreiber, 1991; Res, 1993; Oberholzer, 1995; Reczko et al., 1995b) and is seen, together with the comagmatic Ongeluk Formation, as a major continental flood basalt event on the Kaapvaal craton (Cornell and Schütte, 1995; Reczko et al., 1995b). According to Button (1973) no structures are known from the upper portions of the Hekpoort Formation that would suggest a subaqueous deposition. However, the occurrence of epiclastic sedimentary rocks such as sandstones and mudstones showing cross-bedding and cross-lamination, respectively, suggest the presence of open water bodies such as localized ponds and lakes where these sediments could have been deposited (Button, 1973). 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). Although a basin-wide study of the proportion of lava flows to volcaniclastic rocks in the Hekpoort Formation is lacking, previous studies support a preponderance of lava flows (Button, 1973; Sharpe et al., 1983; Engelbrecht, 1986; Schreiber, 1991) with the exception of the southwestern part of the Transvaal basin showing approximately equal proportions of volcaniclastic rocks and lava flows (Oberholzer, 1995). Therefore, subaerial fissure eruptions appear to have dominated, with locally important pyroclastic systems (Oberholzer, 1995) (Fig. 7). Intermittent hiatuses in volcanism were marked by small lacustrine shale deposits. Presumably, the partly basaltic character of the Hekpoort lavas and the concomitant low volatile content was responsible for the apparent lack of pyroclastic rocks on a regional scale (Oberholzer and Eriksson, 2000). The more calc-alkaline basaltic andesitic Hekpoort magmas reported by Oberholzer (1995) in the southwestern part of the Transvaal basin possibly formed by differentiation from basaltic magmas, accompanied by modification through fluids and volatiles (e.g. Fisher and Schmincke, 1984), thus promoting a more explosive style of volcanism and allowing the formation of varying proportions of lava flows and pyroclastic rocks.

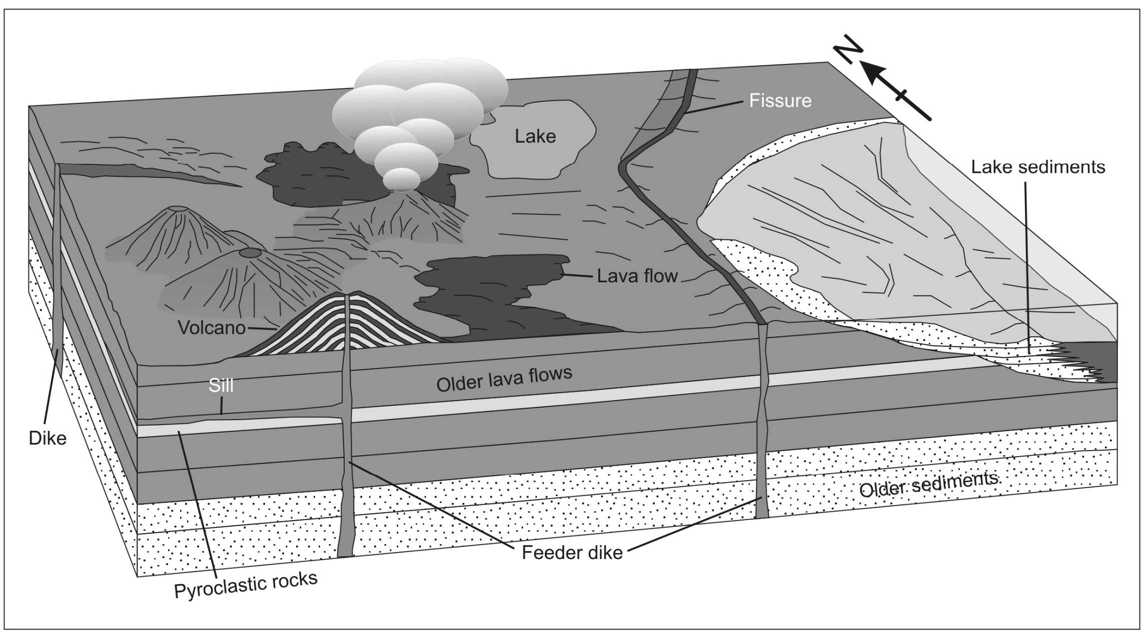

Figure 7. Reconstruction of the palaeoenvironment of the Hekpoort Formation.

Pillow lavas (Grobler and Botha, 1976), as well as hyaloclastites and massive lava flows, support subaqueous extrusion for the Ongeluk Formation volcanic rocks (Cornell and Schütte, 1995). Interpretation of the Hekpoort–Ongeluk volcanics as mantle plume-related continental flood basalts is supported by their geochemistry and preserved geometry, and processes within a replenished, fractionated, tapped, assimilated (RFTA; Arndt et al., 1993) magma chamber have been discussed by Reczko et al. (1995b).

Although the Hekpoort and Ongeluk Formations are frequently correlated with each other in literature, recently doubt has been expressed whether they are indeed comparable. According to Moore et al. (2001) correlation of the Pretoria and Postmasburg Groups has essentially hinged around similarities between thick basaltic-andesite volcanic units in the two sequences (eastern Hekpoort Formation, western Ongeluk Formation), in both composition and model age (Sharpe et al., 1983; Cornell et al., 1996). The Ongeluk volcanics contain pillow lavas and hyaloclastites, indicative of subaqueous conditions, and show extensive alteration that has been ascribed to sea-floor processes (Cornell et al., 1996), whereas the Hekpoort volcanics comprise subaerial coarse and fine graded volcaniclastics and associated amygdaloidal lavas, also showing intense alteration/metamorphism, probably related to the intrusion of the Bushveld Complex (Oberholzer and Eriksson, 2000). Based on the Ongeluk–Hekpoort correlation, lithological differences between the Pretoria and Postmasburg Groups have been ascribed to facies changes from a proximal north-eastern coarse-clastic environment (Pretoria Group) to a distal chemical setting in the south-west (Postmasburg Group) (Button, 1986). Extensions of the Transvaal basin into southern Botswana, however, indicate no change in the subaerial volcaniclastic nature of the Hekpoort volcanics or in the coarse clastic nature of sedimentary rocks immediately overlying the volcanic unit (Key, 1983). Possible extensions of the Postmasburg Group into southern Botswana are obscured by younger Kalahari cover, whereas undoubted Pretoria Group sedimentary rocks occur in the Kanye and Molopo areas (Crockett, 1972), preserved as an erosional remnant surrounding the syn-Bushveld Molopo Farms igneous complex (Gould et al., 1987). This continuity of Pretoria Group clastic sedimentary rocks into the Kanye basin, places it in close juxtaposition with, if not superposition on, the Postmasburg Group, contradicting the proposed broad facies model (see discussion in Moore et al., 2001). Regarding the geochemical similarities of the Hekpoort and Ongeluk Formations, Myers et al. (1987) have demonstrated that a fundamentally uniform trace element pattern exists for 13 different basaltic volcanic sequences that have erupted onto the Kaapvaal Craton between 3.0 and 2.1 Ga, relating this to repeated melting of uniformly metasomatised subcontinental lithosphere. This, together with significant degrees of subsequent alteration, makes discrimination of individual basaltic andesite units in the Transvaal Supergroup difficult on geochemical grounds alone (Moore et al., 2001). Besides these objections against a comparability of the Hekpoort and Ongeluk Formations, Bau et al. (1999) regard the ages for the Ongeluk Lava, and by extension for the Hekpoort andesite, as unreliable. Little credibility can therefore be attributed to correlations of the Ongeluk and Hekpoort Formations on the basis of available whole-rock model ages (Bau et al., 1999). It has thus been argued that there appear to be no compelling geological, geochemical or geochronological reasons to correlate these two volcanic formations (Moore et al., 2001); further detailed studies would seem to be necessary to resolve these divergent viewpoints.

The abundant carbon within the black shales, the presence of stromatolitic carbonates and of inferred, localised high-density turbidity current deposits in the Silverton Formation (Schreiber, 1991; Eriksson et al., 1993) support deposition of the formation in shallow to intermediate water-depth not far from the shoreline. Therefore, deposition in the lower shoreface, below normal wave base or proximal offshore setting with increased suspension sedimentation is proposed (c.f., Reinson, 1984). Eriksson et al. (2008) postulate an analogous model for the Silverton deposits, as having been laid down within an offshore mud belt and transition zone to the more proximal coastal sand belt, below a storm wave base of ca. 8-10 m (Fig. 8). These inferred settings are relevant to the formation of the Machadodorp lavas which are enclosed by Silverton sedimentary lithologies.

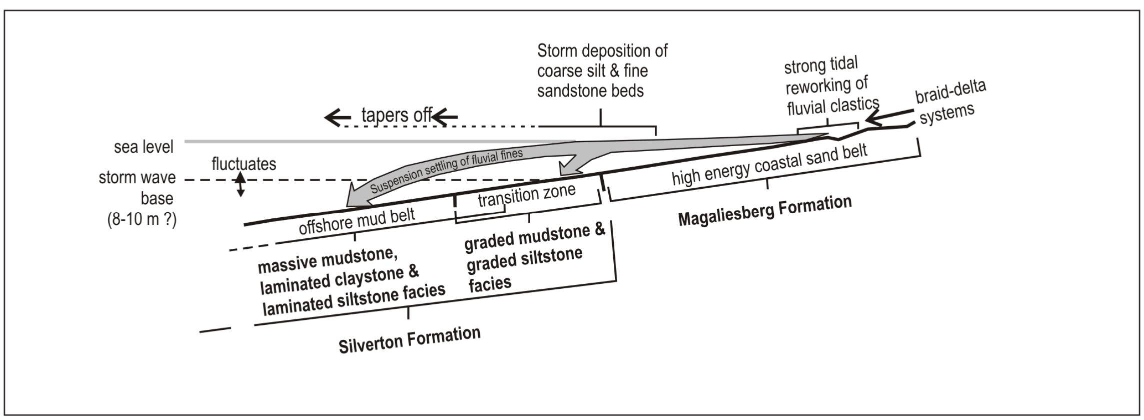

Figure 8. Clastic epeiric sea model interpreted for the Palaeoproterozoic Silverton Formation, Transvaal Supergroup (Eriksson et al., 2008).

The volcanic activity of the Machadodorp Member was heralded by a few isolated, thin tuff beds, intercalated with the mudrocks of the underlying Boven Shale Member. These pyroclastics, which had been interpreted as subaqeous fall deposits are described as being the major deposits developed in the Potgietersrus and Marble Hall areas (De Waal, 1963; Button, 1973; Schreiber, 1991). The features of the overlying fluidal-clast breccia dominating the ridge between Carolina and Burgersfort indicate that its clasts were still hot and plastic at the time of deposition. In a subaqeous environment, fluidal clasts can only be formed and preserved if hot clasts are isolated from the surrounding seawater, so that hydroclastic fragmentation is avoided (Kokelaar, 1986). Steam that evolved during subaqeous eruptions can provide such essentially water-free sub-environments, within steam envelopes, allowing fluidal pyroclasts to form and be preserved (Mueller and White, 1992). If the accretionary lapilli interpretation in proximal vent areas is correct, then a wet steam-rich cupola can be inferred (c.f., White, 1996; Mueller et al., 2000). Subaqeous hawaiian-style lava fountaining, an intermediate eruptive style between calm effusive and violent plinian eruptions, commonly produces fluidal clasts that remain unbroken and hot until deposited, due to insulation from the ambient medium, water, by steam jackets around individual clasts or clusters, and by transport and deposition under a steam cupola (c.f., Smith and Batiza, 1989; Mueller and White, 1992; Ransom et al., 1999; Simpson and McPhie, 2001; Busby, 2005). The mixture of fluidal clasts with blocky clasts in the upper part of the Machadodorp deposits studied suggests that water gained access to the margins of the erupting fountain and during transport, and implies a diminished insulation of any steam envelope (Mueller and White, 1992). The increasing content of blocky clasts further up-section is thought to have resulted from fragmentation of magma or lava in a brittle or semi-brittle state, either by magmatic (e.g., Heiken, 1978) or hydrovolcanic (e.g., Wohletz, 1983; Heiken and Wohletz, 1991) processes. The blocky clasts were formed as water gained access to the lava fountain, and caused additional fragmentation before or very shortly after initial separation in the fountain, while the vesiculating lava was still slightly plastic (Mueller and White, 1992).

The extensive distribution of the fluidal-clast breccia, at least 220 km along strike, and indicating a close proximity to a volcanic source (e.g., Kokelaar and Durant, 1983; Kokelaar, 1986), suggests the presence of several vents between Carolina in the South and Burgersfort in the North. In subaerial settings, fountaining commonly occurs at vents along fissures (Macdonald, 1972; Wilson and Head, 1981). A similar scenario may be expected in subaqueous settings (Simpson and McPhie, 2001; Scott et al., 2002). The eruption must have been sufficiently strong to create a steam envelope in which ballistic material had time to be deposited, but not voluminous enough to develop thick pyroclastic flow deposits with heat retention structures and welding (Mueller et al., 2000).

The stratified lapilli tuff exhibits a change in the fragmentation process from magmatic to hydrovolcanic, attributable to significant ingress of water into the eruption column. They probably formed due to the collapse of a fountaining subaqeous eruption and its steam envelope, either complete or only at the margins, producing hot, high-concentration density currents (c.f., Lowe, 1982; White, 2000).

The overlying sheet lavas most likely resulted from effusion that accompanied fountaining and may indicate fluctuations in the magma discharge rate (Simpson and McPhie, 2001). In particular, relative to fountaining episodes, the lavas were probably generated during periods of reduced discharge (e.g. Griffith and Fink, 1992). An alternative explanation is given by Head and Wilson (2003) who propose that the volatile build-up at the top of the reservoir would leave a complementary volatile-depleted magma below. Thus, after the volatile-rich layer was discharged during the hawaiian-style eruption event, it could be followed by a very volatile-depleted effusive phase, and the vesicle-poor lavas overlying pyroclastic-rich cones might therefore be a distinctive signature of this eruption setting. The Machadodorp lavas are unlikely to be fountain-fed lavas, given the non-welded character of the associated fluidal-clast breccia (cf., Simpson and McPhie, 2001). The dominance of sheet lavas over pillow lavas affirms the prediction that the relatively high effusion rates associated with hawaiian-type eruptions would lead to lava flows that would be characterized by lobate sheets, rather than pillows (e.g., Head et al., 1996; Gregg and Fink, 1995; Head and Wilson, 2003).

In general, the Machadodorp Volcanic Member can thus be interpreted as the deposits of several seamounts (e.g., Keating et al., 1987; Wessell and Lyons, 1997; Schmidt and Schmincke, 2000) aligned along a fissure, probably within an extensional environment (Button, 1973 ; Reczko et al., 1995b) (Fig. 9). The volcanic facies constituting seamounts often overlie deep water sediments and/or are interstratified with sedimentary material deposited as suspension fallout during volcanism (Fisher, 1984). According to Head and Wilson (2003) landforms anticipated from these eruptions might include cones surrounding the vent with rim crests within a few meters of the vent, possible lava ponds within the cone, and an apron of pyroclastic deposits surrounding the vent. The cone and the flanking deposits should consist of interlayers of pyroclastic flows and lava flows with sheet flow morphology, rather than pillow lava morphology dominating. At greater radial distances from the vent, one would predict that flows and layers of agglutinated pyroclasts would dominate proximally in the cone, giving way to bedded pyroclastics and interlayered lava flows distally. Seamounts, although primarily associated with divergent plate boundaries (e.g. Buck et al., 1998; Macdonald, 1998; Perfit and Chadwick, 1998; Head et al., 1996), have also been related to back-arc, arc, and hot spot volcanism (Corcoran, 2000). These subaqueous features can vary from 0.05-10 km in thickness and attain diameters as large as 100 km (Corcoran, 2000). Their depths are generally recognized to be approx. 200-1000 m and less, depending on magma composition and volatile content (e.g., Kokelaar, 1986; Bonatti and Harrison, 1988; Gill et al., 1990; Oshima et al., 1991; Heikinian et al., 1991; Binard et al., 1992; Wright, 1996, 1999; White, 1996; Kano, 1998; Fiske et al., 1998, 2001; Hunns and McPhie, 1999).

The sporadic tuffs in the west of the basin contrast with the eastern Machadodorp Lava Member (Eriksson et al., 1998) and most certainly did not arise from the same eruption mechanisms. However, to date they have not been studied in detail.

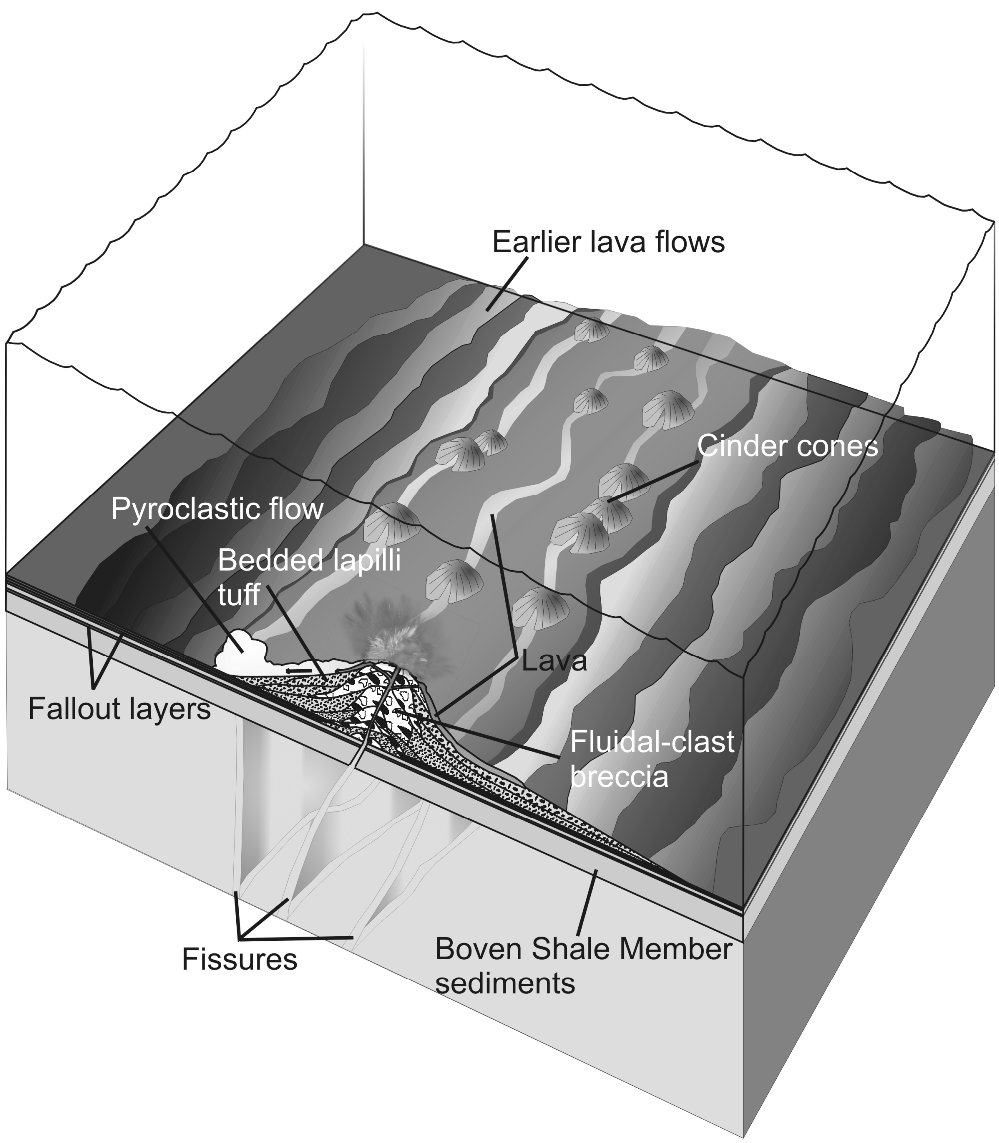

Figure 9. Reconstruction of the palaeoenvironment of the Machadodorp Member.

Pretoria Group basin evolution

The lithostratigraphy, interpreted depositional environments and sequence stratigraphic framework proposed for the Pretoria Group have been investigated in detail in published work (e.g., Eriksson et al., 1991, 2001, 2006; Eriksson and Reczko, 1995; Catuneanu and Eriksson, 1999; Moore et al., 2001) (Fig. 1). Geodynamic control inferred for the evolution of the Pretoria Group basin encompasses two cycles of prerift uplift - subsequent mechanical rifting – long lived thermal subsidence; in this model, two thick predominantly argillaceous successions ascribed to epeiric marine deposition (respectively, the Timeball Hill and younger Silverton Formations) are separated by thinner fluvial (alluvial) deposits and volcanic units related to the uplift and rifting phases (Catuneanu and Eriksson, 1999; Eriksson et al., 2002, 2008) (Fig. 1). Glaciation during Pretoria Group sedimentation is inferred to have been related to global palaeo-atmospheric and geodynamic events and shows no obvious relationship with the evolution of the basin itself (Eriksson et al., 2001).

Eriksson et al. (2001) suggested that a generally weak spatial relationship of the lower Pretoria rifting cycle as evidence for volcanism supported a role for cratonic-scale plate tectonic processes in the evolution of this first basinal geodynamic cycle. In this paper, this postulate can be expanded upon: the limited extent of the Bushy Bend lavas in the south of the Pretoria depository and the relatively poorly developed and discontinuous, mostly tuffaceous lithologies at the equivalent stratigraphic level around the rest of the basin, attest to the likelihood that volcanism accompanying the first rifting event was short-lived, with both subaerial and subaqueous eruptions scattered around the preserved basin margin. Basaltic-andesitic fissure eruptions accompanied by fumarolic activity are thus inferred within the basin at the Rooihoogte-Timeball Hill transition, with limited subaerial explosive eruptions probably having occurred in the basin’s hinterland. The limited scope of volcanism accompanying the rifting within the first cycle of Pretoria basin evolution is thus confirmed in this paper. The Kaapvaal craton at this time in its evolution has no known plate tectonic associations with larger supercontinental plates nor are plate collisions known which might have enabled rifting related to plate tectonic processes in general (e.g., Eriksson et al., in press). The cause of this first rifting cycle might thus indeed lie in sub-cratonic magmatic-thermal processes, even though the surficial volcanic expression thereof was limited.

In contrast, the second rifting cycle postulated for the Pretoria Group basin has a strong association with widespread and large scale volcanism of the Hekpoort-Ongeluk flood basalts, which might well have been plume-related (e.g., Cornell and Schütte, 1995; Reczko et al., 1995b; Eriksson et al., 2001, 2006). Reczko et al. (1995b) have argued for a model encompassing genetic processes within a replenished, fractionated, tapped, assimilated (RFTA; Arndt et al., 1993) magma chamber, due to the geochemistry and preserved geometry of the inferred comagmatic Hekpoort–Ongeluk volcanics. Ongoing debate on the chronology of the two flood basalts (e.g., Moore et al., 2001; current paper under review by these authors), one (Hekpoort) subaerial and the other (Ongeluk) demonstrably essentially subaqueous might place the latter in a stratigraphic position preceding Hekpoort eruption. However, any such debate aside, a major and widespread plume-related volcanic episode would still remain related to this second cycle of rifting in Pretoria Group evolution. The scale of this second cycle volcanism and its importance in influencing the upper part of the Pretoria Group basin-fill (Daspoort-Silverton-Magaliesberg epeiric sea succession; cf. Eriksson et al., 2002, 2008) is underlined by the Machadodorp volcanism, which is related in this paper to extension, the development of an extensive fissure and a set of inferred seamounts, probably related to hot spot volcanism as the Ongeluk-Hekpoort plume possibly waned.

Globally, the period c. 2.5-2.0 Ga is seen by some workers to have been characterized by the occurrence of ultramafic intrusions, dyke swarms, and layered mafic complexes, indicating a continental break-up (Heaman, 1997). This viewpoint is one coloured by thinking related to the inferred extended breakup of the Kenorland supercontinent from c. 2.45 – 2.2 Ga (e.g. Aspler and Chiarenzelli, 1998); however, supercontinentality should not be seen as a universally applicable state of the Palaeoproterozoic Earth (e.g., Eriksson et al., in press). In addition, Condie et al. (2009) argue in favour of a global-scale magmatic shutdown in the c. 2.45-2.2 Ga period, which provides an alternative model for the apparently extended breakup of Kenorland. That the deposition of the Pretoria Group sediments was accompanied by the extrusion of significant thicknesses of volcanic rocks in this same time period, also suggesting that rifting was active at this time (Eriksson and Reczko, 1995), stresses its global importance against this background of variable views of Earth's geodynamic evolution in the Palaeoproterozoic. Although the syn-Transvaal Supergroup deformation on the Kaapvaal Craton is generally regarded by structural geologists as being extensional, producing small, deep pull-apart basins which acted as depositories for the Transvaal Supergroup (e.g. Eriksson et al., 1996; Bumby et al., 1998), the sheet-like nature of the entire volcano-sedimentary pile is more indicative of thermal subsidence, rather than deposition in a fault-bounded trough. Therefore, it seems more likely that the actual process of syn-Pretoria Group subsidence was transitional between thermal subsidence and incipient rifting (Eriksson and Reczko, 1995; Reczko et al., 1995b).

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References

Allen, R.I., Hundstrom, I., Ripa, M., Simeonov, A., Christofferson, H., 1996. Facies analysis of a 1.9 Ga, continental margin, back-arc, felsic caldera province with diverse Zn-Pb-Ag-(Cu-Au) sulfide and Fe oxide deposits, Bergslagen Region, Sweden. Econ. Geol. 91, 979-1008.

Allen, R.L., Weihed, P., Svenson, S., 1997. Setting of Zn-Cu-Au-Ag massive sulfide deposits in the evolution and facies architecture of a 1.9 Ga marine volcanic arc, Skellefte District, Sweden. Econ. Geol. 91, 1022-1053.

Altermann, W., Hälbich, I.W., 1991. Structural history of the southwestern corner of the Kaapvaal Craton and the adjacent Namaqua realm: new observations and a reappraisal. Precambr. Res. 52, 133-166.

Arndt, N.T., Czamanske, G.K., Wooden, J.L., Fedorenko, V.A., 1993. Mantle and crustal contributions to continental flood volcanism. Tectonophysics 223, 39-52.

Aspler, L.B., Chiarenzelli, J.R., 1998. Protracted breakup of Kenorland, a Neoarchean supercontinent? Geochronologic, tectonostratigraphic and sedimentologic evidence from the Paleoproterozoic. Sed. Geol. 120, 75-104.

Batiza, R., White, J.D.L., 2000. Submarine lavas and hyloclastite, in: Sigurdsson, H., Houghton, B.F., McNutt, S.R., Rymer, H., Stix, J. (Eds.), Encyclopedia of Volcanoes. Academic Press, San Diego, pp. 361-382.

Bau, M., Romer, R.L., Lüders, V., Beukes, N.J., 1999. Pb, O, and C isotopes in silicified Mooidraai dolomite (Transvaal Supergroup, South Africa): implications for the composition of Paleoproterozoic seawater and ‘dating’ the increase of oxygen in the Precambrian atmosphere. Earth Planet. Sci. Lett. 174, 43-57.

Bekker, A., Holland, H.D., Wang, P.-L., Rumble III, D., Stein, H.J., Hannah, J.L., Coetzee, L.L., Beukes, N.J., 2004. Dating the rise of atmospheric oxygen. Nature 427, 117-120.

Bergh, S.G., Sigvaldason, G.E., 1991. Pleistocene mass-flow deposits of basaltic hyaloclastite on a shallow submarine shelf, South Iceland. Bull. Volcanol. 53, 597-611.

Beukes, N.J., 1983. Palaeoenvironmental setting of iron formations in the depositional basin of the Transvaal Supergroup, South Africa, in: Trendall, A.F., Morris, R.C. (Eds.), Iron Formations: Facts and Problems. Elsevier, Amsterdam, pp. 131-209.

Binard, N., Hekinian, R., Cheminee, J.L., Stoffers, P., 1992. Styles of eruptive activity on interplate volcanoes in the Society and Austral hot spot regions: Bathymetry, petrology, and submersible observations. J. Geophys. Res. 97, 13999-14015.

Bonatti, E., Harrison, C.G.A., 1988. Eruption styles of basalt in oceanic spreading ridges and seamounts: Effect of magma temperature and viscosity. J. Geophys. Res. 93, 2967-2980.

Branney, M.J., Kokelaar, P., 2002. Pyroclastic density currents and the sedimentation of ignimbrites. Mem. Geol. Soc. Lond. 27, 143 pp.

Buck, W.R., Delaney, P.T., Karson, J.A., Lagabrielle, Y. (Eds.), 1998. Faulting and Magmatism at Mid-Ocean Ridges. Geophysical Monograph 106. American Geophysical Union, Washington, DC.

Bumby, A.J., Eriksson, P.G., van der Merwe, R., 1998. Compressive deformation in the floor rocks to the Bushveld Complex (South Africa): evidence from the Rustenburg Fault Zone. J. Afr. Earth Sci. 27, 307-330.

Busby, C., 2005. Possible distinguishing characteristics of very deepwater explosive and effusive silicic volcanism. Geology 33, 845-848.

Busby, C.J., Garrison, N., Gans, P., Wagner, D., 2004. Mid-Miocene tholeiitic basalt in the ancestral Cascades arc: Geology, volcanology and geochronology of the Lovejoy Basalt. Geol. Soc. Am., Abstracts with Programs 36(4), p. 97.

Button, A., 1973. A regional study of the stratigraphy and development of the Transvaal Basin in the eastern and northeastern Transvaal. PhD thesis, University of Witwatersrand, Johannesburg.

Button, A., 1974. Low-potash pillow basalts in the Pretoria Group, Transvaal Supergroup. Trans. Geol. Soc. S Afr. 77, 99-104.

Button, A., 1986. The Transvaal sub-basin of the Transvaal Sequence, in: Anhaeusser, C.R., Maske, S. (Eds.), Mineral Deposits of Southern Africa. Geol. Soc. S. Afr., Johannesburg, pp. 811-817.

Button, A., Vos, R.G., 1977. Subtidal and intertidal clastic and carbonate sedimentation in a macrotidal environment: an example from the lower Proterozoic of South Africa. Sed. Geol. 18, 175-200.

Cas, R., Moore, C.L., Scutter, C., Yamagishi, H., 1996. Fragmentation processes and water depth significance of Miocene submarine fire fountain deposits, Ryugazaki, Hokkaido, Japan. AGU 1996 Western Pacific Geophysics Meeting. Abtract W 126.

Catuneanu, O., Eriksson, P.G., 1999. The sequence stratigraphic concept and the Precambrian rock record: An example from the 2.7-2.1 Ga Transvaal Supergroup, Kaapvaal craton. Precambr. Res. 97, 215-251.

Catuneanu, O., Eriksson, P.G., 2002. Sequence stratigraphy of the Precambrian Rooihoogte–Timeball Hill rift succession, Transvaal Basin, South Africa. Sed. Geol. 147, 71-88.

Cheney, E.S., 1996. Sequence stratigraphy and plate tectonic significance of the Transvaal succession of southern Africa and its equivalent in Westerrn Australia. Precambr. Res. 79, 3-24.

Coetzee, L.L., 2002. Genetic stratigraphy of the Paleoproterozoic Pretoria Group in the Western Transvaal. MSc thesis, Rand Afrikaans University.

Cole, P.D., Calder, E.S., Sparks, R.S.J., Clarke, A.B., Druitt, T.H., Young, S.R., Herd, R.A., Harford, C.L., Norton, G.E., 2002. Deposits from dome-collapse and fountain-collapse pyroclastic flows at Soufrière Hills Volcano, Montserrat, in: Druitt, T.H., Kokelaar, B.P. (Eds.), The Eruption of Soufrière Hills Volcano, Montserrat, from 1995 to 1999. Geol. Soc. Lond., Mem. 21, pp. 231-262.

Condie, C.K., O'Neill, C., Aster, R., 2009. Evidence and implications for a widespread magmatic shutdown for 250 My on Earth. Earth Planet. Sci. Lett. 282, 294-298.

Corcoran, P.L., 2000. Recognizing distinct portions of seamounts using volcanic facies analysis: examples from the Archean Slave Province, NWT, Canada. Precambr. Res. 101, 237-261.

Cornell, D.H., Schütte, S.S., 1995. A volcanic-exhalative origin for the world's largest (Kalahari) maganese field. Mineral. Depos. 30, 146-151.

Cornell, D.H., Schütte, S.S., Eglington, B.L., 1996. The Ongeluk basaltic andesite formation in Griqualand West, South Africa: submarine alteration in a 2222 Ma Proterozoic sea. Precambr. Res. 79, 101-123.

Cousineau, P., Dimroth, E., 1982. Interpretation of the relations between massive, pillowed and brecciated facies in an Archean submarine andesitic volcano – Amulet Andesite, Rouyn-Noranda, Canada. J. Volcanol. Geotherm. Res. 13, 83-102.

Crockett, R.N. 1972. The Transvaal System in Botswana: its geotectonic and depositional environment and special problems. Trans. Geol. Soc. S. Afr. 75, 275-292.

Crow, C., Condie, K.C., 1990. Geochemistry and origin of early Proterozoic volcanic rocks from the Transvaal and Soutpansberg successions, South Africa. Precambr. Res. 47, 17-26.

Dankert, B.T., Hein, K.A.A., 2010. Evaluating the structural character and tectonic history of the Witwatersrand Basin. Precambr. Res. 177, 1-22.

De Villiers, P.R., Visser, J.N.J., 1977. The glacial beds of the Griqualand West Supergroup as revealed by four deep boreholes between Postmasburg and Sishen. Trans. Geol. Soc. S. Afr. 80, 1-8.

De Waal, S.A., 1963. Die plooikomplex van die Sisteem Transvaal noord van Marble Hall en die meegaande metamorfe en intrusiegesteentes. MSc thesis, University of Pretoria.

Dimroth, E., Cousineau, P., Leduc, M., Sanschagrin, Y., 1978. Structure and organisation of Archean basalt flows, Rouyn-Noranda area, Quebec, Canada. Can. J. Earth Sci. 15, 902-918.

Dorland, H.C., Beukes, N.J., Gutzmer, J., Evans, D.A.D., Armstrong, R.A., 2004. Trends in detrital zircon provenance from Neoarchaean–Palaeoproterozoic sedimentary successions on the Kaapvaal craton. Abstract book, Geoscience Africa 2004 Congress, Johannesburg, South Africa, pp. 176-177.

Engelbrecht, J.P., 1986. Die Bosveld Kompleks en sy vloergesteentes in die omgewing van Nietverdiend, Wes- Transvaal. PhD thesis, University of Pretoria.

Engelbrecht, C.J., Baumbach, G.W.S., Matthysen, J.L., Fletcher, P., 1986. The West Wits Line, in: Anhaeusser, C.R., Maske,S. (Eds.), Mineral Deposits of Southern Africa, Vol. I, Geol. Soc. S Afr., Johannesburg, pp. 599-648.

Eriksson, K.A. 1973. The Timeball Hill Formation – a fossil delta. J. Sed. Petrol. 43, 1046-1053.

Eriksson, P.G., 1988. The sedimentology of the Rooihoogte Formation, Transvaal Sequence. S Afr. J. Geol. 91, 477-489.

Eriksson, P.G., Altermann, W., in press. Palaeoproterozoic-aged glacigenic deposits: Transvaal Supergroup (Kaapvaal craton) (part of Chapter 7.2 Huronian age glaciation), in: Melezhik, V.A., Fallick, A.E., Kump, L.R., Lepland, A., Prave, A.R., Strauss, H.W, (Eds.), Reading the Archive of Earth's Oxygenation. Springer, Berlin, 2011.

Eriksson, P. G., Clendenin, C.W., 1990. A review of the Transvaal Sequence, South Africa. J. Afr. Earth Sci. 10, 101-116.

Eriksson, P.G., Reczko, B.F.F., 1995. The sedimentary and tectonic setting of the Transvaal Supergroup floor rocks to the Bushveld Complex: tectonic setting derived from basin-fill geometry, inferred volcanic processes and sedimentation systems. J. Afr. Earth Sci. 21, 487-504.

Eriksson, P.G., Reczko, B.F.F., 1998. Contourites associated with pelagic mudrocks and distal delta-fed turbidites in the Lower Proterozoic Timeball Hill Formation epeiric basin (Transvaal Supergroup), South Africa. Sed. Geol. 120, 319-335.

Eriksson, P.G., Twist, D., 1986. A note on a lahar deposit in the Hekpoort Formation, Transvaal Sequence, near Pretoria. Trans. Geol. Soc. S Afr., 89, 415-418.

Eriksson, P.G., Twist, D., Snyman, C.P., Burger, L., 1990. The geochemistry of the Silverton Shale Formation, Transvaal Sequence. S. Afr. J. Geol. 93. 454-462.

Eriksson, P.G., Schreiber, U.M., van der Nuit, M., 1991. A review of the sedimentology of the early Proterozoic Pretoria Group, Transvaal sequence, South Africa: implications for tectonic setting. J. Afr. Earth Sci. 13, 107-119.

Eriksson, P.G., Schweitzer, J.K., Bosch, P.J.A., Schreiber, U.M., van Deventer, J.L., Hatton, C.J., 1993. The Transvaal Sequence: An overview. J. Afr. Earth Sci. 16, 25-51.

Eriksson, P.G., Merkle, R.R.W., Schreiber, U.M., Engelbrecht, J.P., Res, M., Snyman, C.P., 1994a. Early Proterozoic black shales of the Timeball Hill Formation, South Africa: volcanogenic and palaeoenvironmental influences. J. Afr. Earth Sci. 97, 325-337.

Eriksson, P.G., Engelbrecht, J.P., Res, M., Harmer, R.E., 1994b. The Bushy Bend lavas, a new volcanic member of the Pretoria Group, Transvaal Sequence. S. Afr. J. Geol. 97, 1-7.

Eriksson, P.G., Engelbrecht, J.P., Res, M., Harmer, R.E., 1995. An overview of the geology of the Transvaal sequence and the Bushveld Complex, South Africa. Mineral. Depos. 30, 98-111.

Eriksson, P.G., Reczko, B.F.F., Corner, B., Jenkins, S.L., 1996. The Kanye axis, Kaapvaal craton, southern Africa: a postulated Archaean crustal architectural element inferred from three-dimensional basin modelling of the lower Transvaal Supergroup. J. Afr. Earth Sci. 22, 223-233.

Eriksson, P.G., van der Merwe, R., Bumby, A.J., 1998. The Palaeoproterozoic Woodlands Formation of eastern Botswana-northwestern South Africa: lithostratigraphy and relationship with Transvaal Basin inversion structures. J. Afr. Earth Sci. 27, 349-356.

Eriksson, P.G., Altermann, W., Catuneanu, O., Van der Merwe, R., Bumby, A.J., 2001. Major influences on the evolution of the 2.67–2.1 Ga Transvaal Basin, Kaapvaal Craton. Sed. Geol. 141-142, 205-231.

Eriksson, P.G., Altermann, W., Eberhardt, L. Ahrend-Heidbrinck, S. Bumby, A.J., 2002. Palaeoproterozoic epeiric sea palaeoenvironments: the Silverton Formation (Pretoria Group Transvaal Supergroup), South Africa, in: Altermann, W., Corcoran, P.L. (Eds.), Precambrian Sedimentary Environments: a modern approach to ancient depositional systems. Blackwell, Oxford, pp. 351-368.

Eriksson, P.G., Altermann, W., Hartzer, F.J., 2006. The Transvaal Supergroup and its precursors, in: Johnson, M.R., Anhaeusser, C.R., Thomas, R.J. (Eds.), The Geology of South Africa. Geol. Soc. S Afr., Johannesburg/Council for Geoscience, Pretoria, pp. 237-260.

Eriksson, P.G., Bose, P.K., Catuneanu, O., Sarkar, S., Banerjee, S., 2008. Precambrian clastic epeiric embayments: examples from South Africa and India, in: Pratt, B.R., Holmden, C. (Eds.), Dynamics of Epeiric Seas. Geological Association of Canada Spec. Pap. 48, St. Johns, Newfoundland, pp. 119-136.

Eriksson, P.G., Rigby, M.J., Bandopadhyay, P.C., Steenkamp, N.C., 2011. The Kaapvaal craton (South Africa): no evidence for a supercontinental affinity prior to 2.0 Ga? Int. Geol. Rev., in press. doi: 10.1080/00206814.2010.527638.

Ernst, R.E., Bell, K., 2010. Large Igneous provinces (LIPs) and carbonatites. Miner. Petrol. 98, 55-76.

Evans, D.A., Beukes, N.J., Kirschvink, J.L., 1997. Low-latitude glaciation in the Palaeoproterozoic era. Nature 386, 262-266.

Expedition 309 Scientists, 2005. Superfast spreading rate crust 2: a complete in situ section of upper oceanic crust formed at a superfast spreading rate. IODP Prelim. Rep. 309. doi:10:2204/iodp.pr.309.2005.

Fink, J.H., Griffiths, R.W., 1990. Radial spreading of viscous-gravity currents with solidifying crust. J. Fluid Mech. 221, 485-509.

Fisher, R.V., 1984. Submarine volcaniclastic rocks, in: Kokelaar, B.P., Howells, M.F. (Eds.), Marginal Basin Geology: Volcanic and Associated Sedimentary and Tectonic Processes in Modern and Ancient Marginal Basins. Geol. Soc. Lond., Spec. Publ. 16, 5-27.

Fisher, R.V., Schmincke, H.-U., 1984. Pyroclastic Rocks. Springer, Berlin.

Fiske, R.S., Cashman, K.V., Shibatas, A., Watanabe, K., 1998. Tephra dispersal from Myojinsho, Japan, during its shallow submarine eruption of 1952-1953. Bull. Volcanol. 59, 262-275.

Fiske, R.S., Naka, J., Iizasa, K., Yuasa, M., Klaus, A., 2001. Submarine silicic caldera at the front of the Izu-Bonin arc, Japan: Voluminous seafloor eruptions of rhyolite pumice. Geol. Soc. Am. Bull. 113, 813-824.

Francis, P.W., 1994. Volcanoes: A Planetary Perspective. Oxford University Press, Oxford.

Freundt, A., Wilson, C.J.N., Carey, S.N., 2000. Ignimbrites and block-and ash flow deposits, in: Sigurdsson, H., Houghton, B., McNutt, S.R., Rymer, H., Stix, J. (Eds.), Encyclopedia of volcanoes. Academic Press, San Diego, pp. 581-599.

Gill, J., Torssander, P., Lapierre, H., Taylor, R., Kaiho, K., Koyama, M., Kusakabe, M., Aitchison, J., Cisowski, S., Dadey, K., Fujioka, K., Klaus, A., Lovell, M., Marsaglia, K., Pezard, P., Taylor, B., Tazaki, K., 1990. Explosive deep water basalt in the Sumisu backarc rift. Science 248, 1214-1217.

Godinot, A., 1988. Comment on ‘Pipe vesicles in Hawaiian basaltic lavas: their origin and potential as paleoslope indicators’. Geology 16, 90.

Gould, D., Rathbone, P.A., Kimbell, G.S., 1987. The geology of the Molopo Farms Complex, southern Botswana. Bulletin of the Geological Survey Botswana 23, 178.

Gregg, T.K.P., Fink, B.J.V., 1995. Quantification of submarine lava-flow morphology through analog experiments. Geology 23, 73-76.

Gregg, T.K.P., Fink, J.H., 2000. A laboratory investigation into the e!ects of slope on lava flow morphology. J. Volcanol. Geotherm. Res. 96, 145-159.

Gregg, T.K.P., Smith, D.K., 2003. Volcanic investigations of the Puna Ridge, Hawai’i: relations of lava flow morphologies and underlying slopes. J. Volcanol. Geotherm. Res. 126, 63-77.

Griffiths, R.W., Fink, J.H., 1992. Solidification and morphology of submarine lavas: A dependence on extrusion rate. J. Geophys. Res. 97, 19,729-19,737.

Grobler, N.J., Botha, B.J.V., 1976. Pillow lavas and hyaloclastites in the Ongeluk andesite formation in a road-cut west of Griquatown, South Africa. Trans. Geol. Soc. S. Afr. 79, 53-57.

Grobler, D.F., Walraven, F., 1993. Geochronology of the Gaborone Granite Complex extension in the area north of Mafikeng, South Africa. Chem. Geol. 105, 319-337.

Gutzmer, J., Banks, D.A., Lüders, V., Hoefs, J., Beukes, N.J., von Bezing, K.L., 2003. Ancient sub-seafloor alteration of basaltic andesites of the Ongeluk Formation, South Africa: implications for the chemistry of Paleoproterozoic seawater. Chem. Geol. 201, 37-53.

Hannah, J.L., Bekker, A., Stein, H.J., Markey, R.J., Holland, H.D., 2004. Primitive Os and 2316 Ma age for marine shale: implications for Paleoproterozoic glacial events and the rise of atmospheric oxygen. Earth Planet. Sci. Lett. 225, 43-52.

Hartzer, F.J., 1995. Transvaal Supergroup inliers: Geology, tectonic development and relationship with the Bushveld Complex, South Africa. J. Afr. Earth Sci. 21, 521-547.

Head, J.W., Wilson, L., 2003. Deep submarine pyroclastic eruptions: theory and predicted landforms and deposits. J. Volcanol. Geotherm. Res. 121, 155-193.

Head, J.W., Wilson, L., Smith, D.K., 1996. Mid-ocean ridge eruptive vent morphology and structure: Evidence for dike widths, eruption rates, and evolution of eruptions and axial volcanic ridges. J. Geophys. Res. 101, 28265-28280.

Heaman, L.M., 1997. Global mafic magmatism at 2.45 Ga: remnants of an ancient large igneous province. Geology 25, 299-302.

Heiken, G.H., 1978. Plinian-type eruptions in the Medicine Lake Highland, California and the nature of the underlying magma. J. Volcanol. Geotherm. Res. 4, 375-402.

Heiken, G., Wohletz, K., 1991. Fragmentation processes and explosive volcanic eruptions, in: Fisher, R.V., Smith, G.A. (Eds.), Sedimentation in Volcanic Settings. Soc. Econ. Paleontol. Mineral., Spec. Publ. 45, 19-26.

Heikinian, R., Bideau, D., Stoffers, P., Cheminee, J., Muhe, R., Puteanus, D., Binard, N., 1991. Submarine intraplate volcanism in the South Pacific: Geologic setting and petrology of the Society and Austral region. J. Geophys. Res. 96, 2109-2138.

Hoffman, P.F., Kaufman, A.J., Halverson, G.P., Schrag, D.P., 1998. A Neoproterozoic Snowball Earth. Science 281, 1342-1346.

Hunns, S.R., McPhie, J., 1999. Pumiceous peperite in a submarine volcanic succession at Mount Chalmers, Queensland, Australia. J. Volcanol. Geotherm. Res. 88, 239-254.

Kano, K., 1998. A shallow-marine alkali-basalt tuff cone in the Middle Miocene Jinzai Formation, Izumo, SW Japan. J. Volcanol. Geotherm. Res. 87, 173-191.

Kano, K., Takeuchi, K., Yamamoto, T., Hoshizumi, H., 1991. Subaqueous rhyolite block lavas in the Miocene Ushikiri Formation, Shimane Peninsula, SW Japan. J. Volcanol. Geotherm. Res. 46, 241-253.

Kawachi, Y., Pringle, I.J., 1988. Multiple-rind structure in pillow lava as an indicator of shallow water. Bull. Volcanol. 50, 161-168.

Keating, B., Fryer, P., Batiza, R., Boehlert, W. (Eds.), 1987. Seamounts, Islands and Atolls. Geophysical Monograph 43, Amer. Geophys. Union, Washington, DC.

Key, R.M., 1983. The geology of the area around Gaborone and Lobatse, Kweneng, Kgatleng, Southern and South East Districts. District Memoir Geological Survey Botswana 5, 230p.

Kirschvink, J.L., 1992. Late proterozoic low-latitude global glaciation: The snowball Earth, in: Schopf, J.W., Klein, CA. (Eds.), The Proterozoic Biosphere: a multidisciplinary study. Cambridge Univ. Press, Cambridge, 51-52.

Klop, A.C.C., 1978. The metamorphosed sediments of the Pretoria Group and the associated rocks northwest of Zeerust, western Transvaal. MSc thesis, University of Pretoria.

Kobayashi, T., Ishihara, K., Suzuki-Kamata, K., Fukushima, D., 2003. Sakurajima volcano and pyroclastic deposits in southern Kyushu. IUGG 2003 Field Trip Guidebook. Volcanol. Soc. Jpn., pp. 105-131.

Kokelaar, P., 1986. Magma-water interactions in subaqeous and emergent basaltic volcanism. Bull. Volcanol. 48, 275-291.

Kokelaar, B.P., Durant, G.P., 1983. The submarine eruption and erosion of Surtla (Surtsey), Iceland. J. Volcanol. Geotherm. Res. 19, 239-246.

Lenhardt, N., Eriksson, P.G., 2011. Lithofacies of the volcanic Machadodorp Member (Silverton Formation), South Africa. 23rd Colloquium of African Geology, Johannesburg, South Africa, 08.-14.01.11., Abstracts volume, p. 239.

Lenhardt, N., Eriksson, P.G., this volume. Volcanism of the Palaeoproterozoic Bushveld Large Igneous Province: the Rooiberg Group, Kaapvaal Craton, South Africa. Precambr. Res.

Lenhardt, N., Hornung, J., Hinderer, M., Böhnel, H., Torres-Alvarado, I.S., Trauth, N., 2011. Build-up and depositional dynamics of an arc front volcaniclastic complex: The Miocene Tepoztlán Formation (Transmexican Volcanic Belt, Central Mexico). Sedimentology 58, 785-823.

Lowe, D.R., 1982. Sediment gravity flows: II Depositional models with special reference to the deposits of high-density turbidity currents. J. Sediment. Petrol. 52, 279-297.

Macdonald, G.A., 1972. Volcanoes. Prentice Hall, New Jersey.

Macdonald, K.C., 1998. Linkages between faulting, volcanism, hydrothermal activity and segmentation on fast spreading centers, in: Buck, W.R., Delaney, P.T., Karson, J.A., Lagabrielle, Y. (Eds.), Faulting and Magmatism at Mid-Ocean Ridges. Geophysical Monograph 106, American Geophysical Union, Washington, DC, pp. 27-58.

Maicher, D., White, J.D.L., Batiza, R., 2000. Sheet hyaloclastite: density current deposits of quench and bubble-burst fragments from thin, glassy sheet lava flows, Seamount Six, Eastern Pacific ocean. Mar. Geol. 171, 75-94.

Mapeo, R.B.M., Armstrong, R.A., Kampunzu, A.B., Modisi, M.P., Ramokate, L.V., Modie, B.N.J., 2006. A ca. 200 Ma hiatus between the Lower and Upper Transvaal Groups of southern Africa: SHRIMP U–Pb detrital zircon evidence from the Segwagwa Group, Botswana: Implications for Palaeoproterozoic glaciations. Earth Planet. Sci. Lett. 244, 113-132.

Martin, U., Breitkreuz, C., Egenhoff, S., Enos, P., Jansa, L., 2004. Shallow-marine phreatomagmatic eruptions through a semi-solidified carbonate platform (ODP Leg 144, Site 878, Early Cretaceous, MIT Guyot, West Pacific). Mar. Geol. 204, 251-272.

McPhie, J., Doyle, M., Allen, R., 1993. Volcanic textures – A guide to the interpretation of textures in volcanic rocks. Centre for Ore Deposit and Exploration Studies, University of Tasmania, Hobart, pp. 197.

Miyabuchi, Y., Watanabe, K., Egawa, Y., 2006. Bomb-rich basaltic pyroclastic flow deposit from Nakadake, Aso Volcano, southwestern Japan. J. Volcanol. Geotherm. Res. 155, 90-103.

Moore, J.G., Schilling, J.G., 1973. Vesicles, water and sulphur in Reykjanes Ridge basalts. Contrib. Mineral. Petrol. 41, 105-118.

Moore, J.G., Phillips, R.L., Grigg, R.W., Peterson, D.W., Swanson, D.A., 1973. Flow of lava into the sea, 1969-1971, Kilauea Volcano, Hawaii. Geol. Soc. Am. Bull. 84, 537-546.

Moore, J.M., Tsikos, H., Polteau, S., 2001. Deconstructing the Transvaal Supergroup, South Africa: implications for Palaeoproterozoic palaeoclimate models. J. Afr. Earth Sci. 33, 437-444.

Mueller, W., White, J.D.L., 1992. Felsic fire-fountaining beneath Archean seas: pyroclastic deposits of the 2730 Ma Hunter Mine Group, Quebec, Canada. J. Volcanol. Geotherm. Res. 54, 117-134

Mueller, W.U., Garde, A.A., Stendal, H., 2000. Shallow-water, eruption-fed, mafic pyroclastic deposits along a Paleoproterozoic coastline: Kangerluk volcano-sedimentary sequence, southeast Greenland. Precambr. Res. 101, 163-192.

Myers, R.E., Cawthorn, R.G., McCarthy, T.S., Anhaeusser, C.R., 1987. Fundamental uniformity in the trace element patterns of the volcanics of the Kaapvaal Craton from 3000 to 2100 Ma: evidence for the lithospheric origin of these continental tholeiites. Geol. Soc. S Afr., Spec. Publ. 33, 315-326.

Navaya, S., Lenhardt, N., Eriksson, P.G., 2011. Geochemistry of the volcanic Machadodorp Member, Silverton Formation, South Africa. 23rd Colloquium of African Geology, Johannesburg, South Africa, 08.-14.01.11., Abstracts volume, p. 299.

Oberholzer, J.D., 1995. Die geologie van die piroklastiese gesteentes in die Hekpoort Formasie, Transvaal Opeenvolging. MSc thesis, University of Pretoria.

Oberholzer, J.D., Eriksson, P.G., 2000. Subaerial volcanism in the Palaeoproterozoic Hekpoort Formation (Transvaal Supergroup), Kaapvaal craton. Precambr. Res. 101, 193-210.

Orth, K., McPhie, J., 2003. Textures formed during emplacement and cooling of a Palaeoproterozoic, small-volume rhyolitic sill. J. Volcanol. Geotherm. Res. 128, 341-362.

Oshima, S., Tsuchide, M., Kato, S., Watanabe, K., Kudo, K., Ossaka, J., 1991. Birth of a submarine volcano Teisi Knoll. J. Phys. Earth 39, 1-19.

Perfit, M.R., Chadwick, Jr., W.W., 1998. Magmatism at midocean ridges: Constraints from volcanological and geochemical investigations, in: Buck, W.R. Delaney, P.T., Karson, J.A., Lagabrielle, Y. (Eds.), Faulting and Magmatism at Mid-Ocean Ridges. Geophysical Monograph 106, American Geophysical Union, Washington, DC, pp. 59-115.

Philpotts, A., Lewis, C., 1987. Pipe vesicles – an alternative model for their origin. Geology 15, 971-974.

Ransom, B., Byerly, G.R., Lowe, D.R., 1999. Subaqeous to subaerial Archean ultramafic phreatomagmatic volcanism, Kromberg Formation, Barberton Greenstone Belt, South Africa, in: Lowe, D.R., Byerly, G.R. (Eds.), Geologic evolution of the Barberton Greenstone Belt, South Africa. Geol. Soc. Am., Spec. Pap. 329, 151-166.

Reczko, B.F.F., 1994. The geochemistry of the sedimentary rocks of the Pretoria Group, Transvaal Sequence. PhD thesis, University of Pretoria.

Reczko, B.F.F., Eriksson, P.G., Snyman, C.P., 1995a, Some evidence for the base-metal potential of the Pretoria Group: stratigraphic targets, tectonic setting and REE patterns. Mineral. Depos. 30, 162-167.

Reczko, B.F.F., Oberholzer, J.D., Res, M., Eriksson, P.G., Schreiber, U.M., 1995b. A re-evaluation of the volcanism of the Palaeoproterozoic Pretoria Group (Kaapvaal craton) and a hypothesis on basin development. J. Afr. Earth Sci. 21, 505-519.

Reinson, G.E., 1984. Barrier island systems, in: Walker, R.G. (Ed.), Facies models, second edition. Geoscience Canada, Reprint Series 1, 119-140.

Res, M., 1993. Petrology and Geochemistry of the Hekpoort Andesite Formation between Frederikstad and Fochville. MSc thesis, Potchefstroom University.

Robins, B., Sandstå, N.R., Furnes, H., de Wit, M., 2010. Flow banding in basaltic pillow lavas from the Early Archean Hooggenoeg Formation, Barberton Greenstone Belt, South Africa. Bull. Volcanol. 72, 579-592.

Ross, C.S., Smith, R.L., 1961. Ash-flow tuffs. Their origin, Geological Relations and Identification. U.S. Geol. Survey Prof. Pap. 366, 81 pp.

Saucedo, R., Macías, J.L., Bursik, M., Mora, J.C., Gavilanes, J.C., Cortes, A., 2002. Emplacement of pyroclastic flows during the 1998–1999 eruption of Volcán de Colima, Mexico. J. Volcanol. Geotherm. Res. 117, 129-153.

Schmidt, R., Schmincke, H.-U., 2000. Seamounts and island building, in: Sigurdsson, H., Houghton, B., McNutt, S.R., Rymer, H., Stix, J. (Eds.), Encyclopedia of Volcanoes. Academic Press, San Diego, CA, pp. 383-402.

Schmincke, H.-U., Sunkel, G., 1987. Carboniferous submarine volcanism at Herbornseelbach (Lahn-Dill area, Germany). Geol. Rundsch. 76, 709-734.

Schreiber, U.M., 1991. A palaeoenvironmental study of the Pretoria Group in the eastern Transvaal. PhD thesis, University of Pretoria.

Schreiber, U.M., Eriksson, P.G., Snyman, C.P., 1991. A provenance study of the sandstones of the Pretoria Group, Transvaal Sequence (South Africa): petrography, geochemistry and palaeocurrent directions. S. Afr. J. Geol. 94, 288-298.

Schumacher, R., Schmincke, H.-U., 1991. Internal structure and occurrence of accretionary lapilli – a case study at Laacher See Volcano. Bull. Volcanol. 53, 612-634.

Schumacher, R., Schmincke, H.-U., 1995. Models for the origin of accretionary lapilli. Bull. Volcanol. 56, 626-639.

Schütte, S.S., 1992. Ongeluk volcanism in relation to the Kalahari manganese deposits. Unpubl. PhD thesis, University of Natal.

Scott, C.R., Mueller, W.U., Pilote, P., 2002. Physical volcanology, stratigraphy, and lithogeochemistry of an Archean volcanic arc: evolution from plume-related volcanism to arc rifting of SE Abitibi Greenstone Belt, Val d'Or, Canada. Precambr. Res. 115, 223-260.

Seaman, S.J., Dyar, M.D., Marinkovic, N., 2009. The effects of heterogeneity in magma water concentration on the development of flow banding and spherulites in rhyolitic lava. J. Volcanol. Geotherm. Res. 183, 157-169.

Sharpe, M.R., Brits, R., Engelbrecht, J.P., 1983. Rare earth and trace element evidence pertaining to the petrogenesis of 2.3 Ga old continental andesites and other volcanic rocks from the Transvaal Sequence, South Africa. University of Pretoria Institute for Geological Research on the Bushveld Complex Research Report 40, 63 p.

Simpson, K., McPhie, J., 2001. Fluidal-clast breccia generated by submarine fire fountaining, Trooper Creek Formation, Queensland, Australia. J. Volcanol. Geotherm. Res. 109, 339-355.

Smith, T.L., Batiza, R., 1989. New field and laboratory evidence for the origin of hyaloclastite flows on seamount summits. Bull. Volcanol. 51, 96-114.

Smith, G.A., Lowe, D.R., 1991. Lahars: volcano-hydrologic events and deposition in the debris flow – hyperconcentrated flow continuum, in: Fisher, R.V., Smith, G.A. (Eds.), Sedimentation in Volcanic Settings. Soc. Econ. Paleontol. Mineral., Spec. Publ. 45, 59-70.

Sparks, R.S.J., 1976. Grain size variations in ignimbrites and implications for the transport of pyroclastic flows. Sedimentology 23, 147-188.

Tankard, A.J., Jackson, M.P.A., Eriksson, K.A., Hobday, D.K., Hunter, D.R., Minter, W.E.L., 1982. Crustal evolution of southern Africa. Springer-Verlag, New York.

Trendall, A.F., Compston, W., Williams, I.S., Armstrong, R.A., Arndt, N.T., McNaughton, N.J., Nelson, D.R., Barley, M.E., Beukes, N.J., De Laeter, J.R., Retief, E.A., Thorne, A.M., 1990. Chronological comparison of the volcano-sedimentary sequences of the Kaapvaal and Pilbara cratons between 3.2 and 2.4 Ga. Abstract, Third Archaean Symposium, Perth, Australia, p. 81-83.

Van der Neut, M., 1990. Afsettingstoestande van die Pretoria Groep gesteentes in die Pretoria-Bronkhorstspruit-Delmas gebied. MSc thesis, University of Pretoria.

Visser, J.N.J., 1969. 'n Sedimentologiese studie van die Serie Pretoria in Transvaal. PhD thesis, University of Orange Free State.

Visser, J.N.J., 1971. The deposition of the Griquatown Glacial Member in the Transvaal Supergroup. Trans. Geol. Soc. S. Afr. 74, 187-199.

Von Gruenewald, G., Harmer, R.E., 1993. Tectonic setting of Proterozoic layered intrusions with special reference to the Bushveld Complex, in: Condie, K.C. (ed.), Proterozoic Crustal Evolution. Elsevier, Amsterdam, pp. 181-213.

Walker, G.P.L., 1992. Morphometric study of pillow-size spectrum among pillow lavas. Bull. Volcanol. 54, 459-474.

Walraven, F., 1997. Geochronology of the Rooiberg Group, Transvaal Supergroup, South Africa. Information Circular 316, Economic Geology Research Unit, University of the Witwatersrand, Johannesburg, South Africa.

Walraven, F., Grobler, D.F., 1996. Age equivalence of the Plantation Porphyry and the Kanye Volcanic Formation, south east Botswana. S. Afr. J. Geol. 99, 23-31.

Walraven, F., Martini, J., 1995. Zircon Pb-evaporation age determinations for the Oak Tree Formation, Chuniespoort Group. Transvaal Sequence; implications for Transvaal-Griqualand West basin correlations. S. Afr. J. Geol 98, 58-67.

Walton, A.W., Palmer, B.A., 1988. Lahar facies of the Mount Dutton Formation (Oligocene-Miocene) in the Marysvale volcanic field, Utah. Geol. Soc. Am. Bull 100, 1078-1091

Waters, A.C., 1960. Determining direction of flow in basalts. Am. J. Sci. 258A, 350-366.

Wessell, P., Lyons, S., 1997. Distribution of large pacific seamounts from Geosat/ERS-1: Implications for the history of intraplate volcanism. J. Geophys. Res. 102, 22459-22475.

White, J.D.L., 1996. Pre-emergent construction of a lacustrine basaltic volcano, Pahvant Butte, Utah (USA). Bull. Volcanol. 58, 249-262.

White, J.D.L., 2000. Subaqueous eruption-fed density currents and their deposits. Precambr. Res. 26, 87-109.

White, J.D.L., Houghton, B.F., 2000. Surtseyan and related eruptions, in: Sigurdsson, H., Houghton, B., McNutt, S., Rymer, H., Stix, J. (Eds.), Encyclopedia of Volcanoes. Academic Press, New York, pp. 495-512.

Willemse, J., 1959. The floor of the Bushveld Igneous Complex and its relationships with special reference to the eastern Transvaal. Trans. Geol. Soc. S. Afr. 82, 21-80.

Williams, G.E., 2004. The paradox of Proterozoic glaciomarine deposition, open seas and strong seasonality near the Palaeo-Equator: global implications, in: Eriksson, P.G., Altermann, W., Nelson, D.R., Mueller, W.U., Catuneanu, O. (Eds.), The Precambrian Earth: tempos and events. Elsevier, Amsterdam, pp. 448-459.

Wilmoth, R.A., Walker, G.P.L., 1993. P-type and S-type pahoehoe: A study of vesicle distribution patterns in Hawaiian lava flows. J. Volcanol. Geotherm. Res. 55, 129-142.

Wilson, L., Head, J.W., 1981. Ascent and eruption of basaltic magma on the Earth and Moon. J. Geophys. Res. 86, 2971-3001.

Wilson, C.J.N., Walker, G.P.L., 1982. Ignimbrite depositional facies: the anatomy of a pyroclastic flow. J. Geol. Soc. Lond. 139, 581-592.

Wohletz, K.H., 1983. Mechanisms of hydrovolcanic pyroclast formation: Grain-size, scanning electron microscopy, and experimental results. J. Volcanol. Geotherm. Res. 17, 31-63.

Wright, I.C., 1996. Volcaniclastic processes on modern submarine arc stratovolcanoes: Sidescan and photographic evidence from the Rumble IV and V volcanoes, southern Kermadec Arc (SW Pacific). Mar. Geol. 136, 21-39.

Wright, I.C., 1999. Southern Kermadec submarine caldera arc volcanoes (SW Pacific): Caldera formation by e!usive and pyroclastic eruption. Mar. Geol. 161, 207-227.

Wright, J.V., Smith, A.L., Self, S., 1980. A working terminology of pyroclastic deposits. J. Volcanol. Geotherm. Res. 8, 315-336.

Yamagishi, H., 1987. Studies on the Neogene subaqeous lavas and hyaloclastite in southwest Hokkaido. Rep. Geol. Surv. Hokkaido 59, 55-117.

Yamagishi, H., 1991. Morphological and sedimentological characteristics of the Neogene submarine coherent lavas and hyaloclastites in Southwest Hokkaido, Japan. Sed. Geol. 74, 5-23.

Yamagishi, H., Sakamoto, I., Ishii, J., 1989. Internal structures of pillow lobes of Cenozoic and Mesozoic pillow lavas in and around Hokkaido. Proceedings Hokaido Tokai Univ. Science and Engineering 2, 107-118.

Young, G.M., 2004. Earth's two great Precambrian glaciations: aftermath of the "Snowball Earth" hypothesis, in: Eriksson, P.G., Altermann, W., Nelson, D.R., Mueller, W.U., Catuneanu, O. (Eds.), The Precambrian Earth: tempos and events. Elsevier, Amsterdam, pp. 440-448.