The Kaapvaal craton (South Africa): no evidence for a supercontinental affinity prior to 2.0 Ga?

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We briefly examine the possible antiquity of the supercontinental cycle while noting the likely unreliability of paleomagnetic data >~1.8 Ga. An assumption of a gradual change from a magmatically dominated Hadean Earth to a plate tectonically dominated Neoarchaean system is made. A brief review of one of Earth's oldest cratons, Kaapvaal, where accent is placed on the lithostratigraphic and geodynamic-chronological history of its cover rocks from ~3.1 – 2.05 Ga, forms the factual basis for this paper. The ~3.1-2.8 Ga Witwatersrand-Pongola (Supergroups) complex retroarc flexural foreland basin developed while growth and stabilization of the craton were still underway. Accretion of relatively small composite granite-gneiss-greenstone (island arc complex) terranes from both N and W, does not support formation of a Neoarchean supercontinent, but may well have been related to a mantle plume which enhanced primary gold sources in the accreted terranes and possibly controlled the timing and rate of craton growth through plate convergent processes. Subsequent deformation of the Witwatersrand basin-fill with concomitant loss of ≤1.5 km of stratigraphy must logically have been due to far-field tectonic effects, but no known mobile belt or even greenstone belts can be related to this. At ~2714-2709 Ma, a large mantle plume impinged beneath the thinned crust underlying the Witwatersrand basin, forming thick, locally komatiitic flood basalts at the base of the Ventersdorp Supergroup, with subsequent thermal updoming leading to graben basins within which medial bimodal volcanics and immature sediments accumulated. Finally (possibly at ~2.66-2.68 Ga), thermal subsidence enabled deposition of uppermost Ventersdorp sheetlike lavas and sediments, with minor komatiites still present. Ongoing plume-related influences are thus inferred, and an analogous cause is ascribed to a c. 2.66-2.68 Ga dyke swarm to the north of the Ventersdorp, where associated rifting allowed formation of discrete "protobasinal" depositories of the Transvaal (a ~2.6 – 2.05 Ga Supergroup, preserved in three basins). Thin fluvial sheet sandstones (Black Reef Formation, undated) above these lowermost rift-fills show an association with localized compressive deformation (along the paleo-Rand anticline, N of Johannesburg), but again with no evidence of any major terrane amalgamations with the Kaapvaal. From ~2642-2432 Ma, the craton was drowned with a long-lived epeiric marine carbonate-banded iron formation platform covering much of it, and preserved in all three Transvaal basins. During this general period, at ~2691-2610 Ma, Kaapvaal collided with a small exotic terrane (the Central Zone [CZ], Limpopo Belt) in the north. Although far-field tectonic effects are likely implicit in Transvaal basin geodynamics, again there is no case to be made for supercontinent formation. Following a hiatus (80-200 My?), with localized deformation and removal of large thicknesses of chemical sediments (along the paleo-Rand anticline), the uppermost Pretoria Group of the Transvaal Supergroup was deposited. This reflects two episodes of rifting (associated with volcanism) and subsequent thermal subsidence within a sag basin setting; an association of the second such event with flood basalts supports a plume affinity. At ~2050 Ma the Bushveld

Complex intruded the northern Kaapvaal craton and reflects a major plume, following which Kaapvaal-CZ collided with the Zimbabwe craton, when for the first time, strong evidence exists for a small supercontinent assembly, at ~2.0 Ga. We postulate that the long-lived evidence in favour of active mantle (cf. plume) influences with subordinate and localized tectonic shortening, implicit within the review of the ~3.1-2.05 Ga geological history of Kaapvaal, might reflect an ongoing influence of earlier Precambrian mantle-dominated thermal systems, at least for this craton.

Keywords: Kaapvaal craton; mantle plumes; supercontinent cycle; Limpopo Belt

Introduction

The antiquity of the supercontinental cycle is a subject of ongoing debate (e.g., Unrug 1992; Rogers 1996; Aspler and Chiarenzelli 1998). Such debate is predicated on views of the antiquity of a Phanerozoic-type plate tectonic regime (e.g., de Wit 1998). Here, we favour a gradual change from a magmatically dominated Hadean regime to a plate tectonically dominated one (possibly somewhere close to the Neoarchean-Paleoproterozoic boundary), with Trendall's (2002) "plughole model" providing a conceivable hypothesis for achieving the envisaged transition (Eriksson and Catuneanu 2004, for discussion). The inherent link between the supercontinent cycle and the mantle plume (definition of Condie 2004a, b; Condie *et al.* 2001) concept (e.g., Zhong *et al.* 2007; Santosh *et al.* 2009), suggests that the cycle may only have begun once the interaction of mantle thermal and plate tectonic regimes became the norm on Earth. This debate is exacerbated by the recent postulate of Condie *et al.* (2009) for global tectonic-mantle thermal stagnation at c. 2.45-2.2 Ga (similar to the quiescence of Eriksson *et al.* (2004) at c. 2.7-2.2 Ga), based on global distribution of U-Pb age data. Somewhat analogously, Aspler and Chiarenzelli (1998) propose a very protracted breakup of the Neoarchean "Kenorland" supercontinent from c. 2.45-2.1 Ga.

On the one side of the supercontinent antiquity debate is a somewhat confusing number of relatively poorly constrained early Precambrian supercontinents: e.g., "Ur" (c. 3.0 Ga); "Kenorland" and "expanded Kenorland" (Neoarchean); "Superia" (Neoarchean-Paleoproterozoic); "Arctica" (c. 2.5 Ga); "Atlantica" (c. 2.0 Ga); "Columbia" (c. 1.9 Ga); "Laurentia" (c. 2.0-1.8 Ga); "expanded Ur" and "Nena" of c. 1.5 Ga antiquity; each is proposed to comprise a unique set of ancient cratonic nuclei (e.g., Piper 1976; Hoffman 1988; Williams *et al.* 1991; Aspler and Chiarenzelli 1998; Meert 2002; Rogers and Santosh 2002; Bleeker and Ernst 2006). Breakup and re-assembly of Columbia is thought to have led to the supercontinent Rodinia (e.g., Dalziel 1997), which is recognized widely, although its component cratons and configuration remain uncertain (e.g., Weil *et al.* 1998). Postulated supercontinents >c. 1.8 Ga come up against the other side of this debate: the reliability of the underlying paleomagnetic data (e.g., Meert 2002; Pesonen *et al.* 2003). Meert (2002) provides a recent summary of the challenges and problems implicit in interpreting paleomagnetic data >c. 1.8 Ga and argues against its application. Attempts to reconstruct a reliable apparent polar wander path for the Kaapvaal craton for the 3.0-1.9 Ga period have not met with success (Strik *et al.* 2007; see however, de Kock *et al.* 2009, for a counter-view).

Correlating widespread impact ejecta/fallout units (e.g., Glikson 2008) or mobile belts may provide an alternative to magnetic data. Regional lithostratigraphy (including a glacigenic horizon) across the cratonic blocks of North America, the Baltic and Siberian shields, as well as good geochronological data including that on precisely dated dyke swarms with matching geometries on postulated cratons within a Superia/Kenorland supercontinent, make a relatively compelling case for this amalgamation (e.g., Aspler and Chiarenzelli 1998; Bleeker and Ernst 2006). However, in contrast, support for a "southern" (modern reference) equivalent, including Kaapvaal is parlous (e.g., Eriksson *et al.* 2009) and forms the subject of this paper.

2. Kaapvaal craton – brief overview of evolution

2.1. Formation of the craton

Assuming some form of plate tectonics to have been operative, formation of the nucleus of the Kaapvaal craton (Figure 1) by approximately 3.1 Ga can be ascribed to initial (c. 3.6-3.4 Ga) thin-skin thrusting within ocean and arc settings and subsequent (c. 3.3-3.2 Ga) amalgamation of displaced oceanic and arc terranes, accompanied by significant granitoid magmatism (de Wit *et al.* 1992). It is suggested that the bulk of the terrane accretion which formed the Kaapvaal craton, occurred along two prominent ENE-WSW suture zones, the Barberton lineament (BL) and the Thabazimbi-Murchison lineament (TML) between 3.23 and 2.9 Ga (Poujol *et al.* 2003; Anhaeusser 2006; Robb *et al.* 2006) (Figure 1). Recent U-Pb and Lu-Hf isotope data from zircons indicate that the Kaapvaal craton is composed of at least four distinct terranes (Barberton-North [BN] and Barberton-South [BS] either side of the BL; Murchison-Northern Kaapvaal [MNK] north of the TML, and Limpopo Central Zone [LCZ] – see Figure 1) that underwent different crustal evolutions, and were successively accreted at c.3.23 (BN and BS), 2.9 (assembled BN-BS and MNK) and 2.65-2.7 Ga (three existing terranes and LCZ) (Zeh *et al.* 2009).

The Murchison greenstone belt, which accreted from approximately N to S along the NE margin of the Kaapvaal nucleus, was characterized by c. 3.1 – 2.9 Ga Ga mafic and granitic magmatism within an arc-subduction system (Poujol and Robb 1999; Poujol *et al.* 2003; Robb *et al.* 2006; Zeh *et al.* 2009), and formed part of the MNK composite terrane (Figure 1), along with the Pietersburg greenstone belt. Between c. 2.7 and 2.6 Ga, further accretion took place from the north, with juxtaposition of an exotic terrane, the Central zone (LCZ in Figure 1) of the Limpopo mobile belt (LMB), along a ENE-WSW trending, inward dipping, strike-slip shear zone, the Palala-Zoetfontain shear zone. Rocks of the Southern Marginal zone (SMZ) of this mobile belt, which represent high-grade equivalents of the granite-greenstone cratonic successions, were thrust onto the Kaapvaal craton along the Hout River shear zone at 2691-2620 Ma (Barton and Van Reenen 1992; Barton *et al.* 1992; Kreissig *et al.* 2001). The western accretion onto the Kaapvaal nucleus took place at c. 2-8-2.72 Ga, along a suture zone now preserved as the Colesburg magnetic lineament (Tinker *et al.* 2002) (Figure 1).

2.2. Witwatersrand basin: transition from granite-greenstone to stable craton platform

These Neoarchean cratonisation processes were largely contemporaneous with the evolution of the c. 3.1-2.8 Ga Witwatersrand basin, the oldest large sedimentary depository known (e.g., Frimmel 2005), which derived detritus from both >3.1 Ga nuclear crust and <3.1 Ga juvenile granitoid-gneissic-greenstone accreted terranes; pulses of <3.1 Ga granites were likely related to sedimentation episodes in the evolving depository (Robb and Meyer 1995) (Figure 2a). Application of the widely accepted flexural foreland basin model to the inferred syn-Witwatersrand accretion from both north and west (Figure 1), results in a complex or double foreland depository, with two convergent stress fields at about 100^{o} to each other (Catuneanu 2001 and references therein). Flexural foreland basins typically are characterized by foredeep and back-bulge sub-basins, separated by a flexural forebulge; in this case, the former sub-basin accommodated the Witwatersrand Supergroup succession and the back-bulge sub-basin was filled by deposits of the partly correlated Pongola Supergroup (Beukes and Cairncross 1991; Catuneanu 2001) (Figure 1). An emergent forebulge is postulated to have persisted throughout, due to low rates of subduction along the accretionary margins of the Kaapvaal nucleus (Catuneanu 2001; Eriksson *et al.* 2009). Initial largely volcanic deposits characterized both sub-basins, with subsequent thermal subsidence allowing deposition of starved foredeep sediments within the Witwatersrand sub-basin and shallow back-bulge sub-basin sedimentation in the Pongola Supergroup sub-basin (respectively, West Rand and Mozaan Groups; c. 2970-<2914 Ma; Eriksson *et al.* 1981; Stanistreet and McCarthy 1991; Robb and Meyer 1995; Catuneanu 2001; Eriksson *et al.* 2005). Braided fluvial sandstones and subordinate auriferous conglomerates subsequently formed the Central Rand Group (Figure 2a) of the Witwatersrand Supergroup within the foredeep sub-basin, which acquired a stuffed basin character, with no equivalent deposits in the back-bulge part of the depository (Robb and Meyer 1995; Els 1998a and b; Catuneanu 2001).

2.3. Ventersdorp superplume

The 2714-<2709 Ma Ventersdorp Supergroup overlies the largely sedimentary strata of the Witwatersrand foredeep sub-basin, and the former predominantly volcanic and lesser sedimentary rocks were also deposited widely onto surrounding cratonic lithologies (e.g., Eriksson *et al.* 2002). The c. 100 My lacuna (Maphalala and Kröner 1993; Beukes and Nelson 1995) separating these two supergroups was characterized by tectonic shortening and erosion resulting in the loss of ≤1.5 km of stratigraphy of the earlier unit (Hall 1996). Applying a plume model to the largely volcanic Ventersdorp succession (cf., Hatton 1995) is compatible with the lower, c. 2 km thick Klipriviersberg Group (2714±8 Ma; Armstrong *et al.* 1991) (Figure 2b) flood basalts of the Ventersdorp Supergroup, which also include basal komatiites (van der Westhuizen *et al.* 1991); ponding of mafic magma beneath thinned lithosphere underlying the earlier Witwatersrand foredeep basin, related to a plume head which was possibly marginal to the Kaapvaal craton, has been suggested (Eriksson *et al.* 2002). Subsequent crustal extension concomitant with this envisaged geodynamic setting formed a set of graben/half-graben basins within this volcanic floor, within which immature clastic sedimentary and bimodal volcanic rocks accumulated to form the medial unconformity-based Platberg Group (c. 2709±4 Ma; Armstrong *et al.* 1991) of the Ventersdorp succession (van der Westhuizen *et al.* 1991, and references therein). The uppermost two units of the supergroup which succeed the Platberg Group, the widespread and sheet-like Bothaville and Allanridge Formations, point to an overall regime of thermal subsidence, with continued plume (minor komatiites in the latter formation) and graben influences (e.g., van der Westhuizen *et al.* 1991; Eriksson *et al.* 2002).

2.4. Transvaal basin: dominant epicratonic shallow seas

The Transvaal Supergroup unconformably succeeds the Ventersdorp, and also transgresses onto older basement rocks; it is preserved within three basins on the Kaapvaal craton: Transvaal itself (= TB) in the north, Kanye (KB; Botswana, in the NW of the craton) and Griqualand West (GW; SW of craton). Descriptions of the relevant geology, as well as basin and sequence stratigraphic models are given in many sources (e.g., Catuneanu and Eriksson 1999, 2002; Eriksson *et al.* 2001, 2006; references therein) (Figure 2c). The supergroup comprises four main parts: basal "protobasinal" (a purely descriptive term) rocks (TB only); Black Reef Formation (TB and KB) and its presumed equivalent in Griqualand West, the Vryburg Formation; the largely dolomite-banded iron formation (BIF) succession of the Chuniespoort-Ghaap-Taupone Groups (respectively, TB, GB, KB); the uppermost essentially clastic sedimentary-lesser volcanic Pretoria-Postmasburg-Segwagwa Groups (again, TB, GB, KB) (e.g., Eriksson *et al.* 2006).

The protobasinal successions comprise discrete, fault-bounded relatively small basins, with fills varying from predominantly volcanic (bimodal) to largely clastic sedimentary; the basal parts of the latter tend to be more immature and mainly alluvial, and to become more mature basin-marginal and basin-central deposits upwards (e.g., Hartzer 1994, 1995; Eriksson *et al.* 2001). An overall tectonic setting of either strike-slip or extensional basins across a wide zone of rifting is envisaged, possibly related to the Ventersdorp event in both style and age. Olsson *et al.* (in press) have dated a dyke swarm in the eastern Kaapvaal craton to 2.66-2.68 Ga which they relate to both protobasinal depository evolution and the uppermost Allanridge Formation of the Ventersdorp Supergroup. Only one of the protobasinal successions has any published age data, that of the Buffelsfontein Group (NW of the TB), varying from 2657-9 to 2664 Ma (unpublished report, South African Committee for Stratigraphy; Barton *et al.* 1995, respectively). Thin (mostly ~30-60 m) sheet sandstones of the undated Black Reef Formation unconformably overlie the protobasinal successions as well as surrounding older rocks in the TB and KB, and are ascribed to initial fluvial sedimentation, passing up into transgressive epeiric marine deposits (e.g., Button 1973; Key 1983; Henry *et al.* 1990; Els *et al.* 1995) (Figure 2c). Northward-directed tectonic shortening affected some of the protobasinal rocks, was synchronous with and post-dated Black Reef deposition (e.g., Eriksson *et al.* 2006 and references therein). The Vryburg Formation of the GB unconformably succeeds the Ventersdorp, and is often correlated with the Black Reef; it is mostly between 100 and c. 300 m thick, comprising clastic sedimentary and lesser carbonate sedimentary lithologies and basaltic-andesitic lavas, the latter dated at 2642±3 Ma (Walraven and Martini 1995). Deposition is inferred to have been through a spectrum of settings from fluvial to marginal marine (Beukes 1979) or deeper marine environments (Altermann and Siegfried 1997).

A regional unconformity related to tilting and base level fall, ushered in the transgressive epeiric sea which covered much of the Kaapvaal, and in which a thick package of stromatolitic carbonate rocks (~1200 m in TB and >2.5 km in GB), succeeding BIF (~640 m in TB) and uppermost mixed clastic and chemical sediments (Duitschland Fm., ~1100 m in TB; Koegas Subgroup in GB) was laid down (e.g., Altermann and Siegfried 1997; Eriksson *et al.* 2001, 2006) (Figure 2c). Available age data suggest over 200 My of chemical sedimentation, from <2642±3 Ma till at least 2432±31 Ma (Trendall *et al.* 1990; c. 2.65-2.40 Ga based on Knoll and Beukes 2009). Initial carbonate deposition began in the SW of the GB, with a later major transgression at c. 2550 Ma of the carbonate platform over the rest of the GB basin and those now preserved in the KB and TB; depositional realms varied from exposed peri-tidal flats to deep carbonate platform conditions (e.g., Eriksson and Altermann 1998). A second major transgression at c. 2500 Ma that was accompanied by deeper epeiric marine conditions, drowned the carbonate platform, and led to the deposition of BIF across all three preserved basins (Altermann and Nelson 1998). The uppermost mixed clastic-chemical sediments are related to final withdrawal of the epi-continental sea from NE to SW (e.g., Eriksson *et al.* 2005). Deposition of the localised Duitschland Formation (far NE of the TB only) probably occurred within a depositional hiatus of possibly 80 My (possibly up to c. 200 My according to Mapeo *et al.* 2006) separating Chuniespoort and Pretoria Groups (and equivalents in the other two basins), during which the chemical sedimentary succession was uplifted and extensively eroded; this erosion was largely along the southern part of the TB succession, essentially coincident with the paleo-Rand anticline (e.g., Eriksson *et al.* 2001; their Figure 8).

The Pretoria Group of the TB and closely correlated equivalents in the KB comprise a 6-7 km thick succession of dominantly argillaceous rocks, lesser yet prominent sandstones, with two major volcanic intervals; overall geometry is sheetlike and depositional conditions are inferred to have varied between two major epeiric marine intervals interspersed with thinner fluvial deposits, with minor inferred glacigenic sediments (e.g., Eriksson *et al.* 2006) (Figure 2c). Two episodes of rifting and subsequent thermal subsidence are thought to have accommodated the two second-order unconformity-bounded depositional sequences identified within this group (Catuneanu and Eriksson 1999). A major flood basalt (Hekpoort-Tsatsu-Ongeluk Fm's., respectively in TB, KB, GB) is dated in the GB at 2222±13 Ma (Pb-Pb; Cornell *et al.* 1996). The base of the Pretoria Group is dated at 2316±7 Ma (Re-Os; Hannah *et al.* 2004); detrital zircon dating within successively higher sandstone units varies from maximum sedimentation ages of 2250±14/15 Ma near the base, to 2236±13 Ma in the medial Daspoort sandstones to 2193±20 Ma in the Magaliesberg Formation, in a stratigraphically high position (Figure 1) (Mapeo *et al.* 2006; similar detrital age data are given by Dorland *et al.* 2004). Although the emplacement age for the Bushveld Complex reflects a minimum age for the Pretoria Group of 2058±0.8 Ma (Buick *et al.* 2001), a time gap between the end of Transvaal sedimentation and Bushveld magmatism is indicated by regional compressive deformation of the sedimentary succession (Bumby *et al.* 1998; Eriksson *et al.* 1998). A less complete succession in the GB incorporates the major c. 2.2 Ga flood basalt (Ongeluk Fm.) and a well developed glacial deposit, but available age data have led to divergent opinions, varying from good correlation with the thicker Pretoria-Segwagwa Groups of TB/KB, to a placement of the equivalent GB succession essentially within the ~ 80 My time gap below the Pretoria (see Moore *et al.* 2001 for discussion).

3. Kaapvaal as part of a >c 2.0 Ga supercontinent – evidence for and against

The "Vaalbara" concept of a Kaapvaal-Pilbara amalgamation (encompassing also the Zimbabwe craton and the Grunehogna province, Antarctica) was first mooted by Button (1976), and later taken up in more detail by Cheney (1996) largely on the basis of inferred correlations of unconformity-bound volcano-sedimentary units on the two cratons. De Kock *et al.* (2009) detail a single paleomagnetic pole established for the Ventersdorp Supergroup and argue for a validation of the Vaalbara reconstruction on that basis; this view stands in contrast to widely held skepticism towards applying paleomagnetism prior to about c. 1.8 Ga (discussed in section 1 above). Other paleomagnetic work carried out by Wingate (1998) and more recently and more exhaustively by Strik *et al.* (2007) on Kaapvaal, covering the time period relevant to the Witwatersrand-Transvaal successions, also does not support the Vaalbara amalgamation. Similarly, detailed precise chronological data examined by Nelson *et al.* (1999; see also Nelson, 2008) contradict any such reconstruction in the 3650-2200 Ma time period. The alternative argument, that an overall analogous geological character observed for Kaapvaal and Pilbara, reflects global events such as superplumes, eustasy and glaciation has been stated by several workers (e.g., Nelson *et al.* 1999; Eriksson *et al.* 2005). A recent study by Eriksson *et al.* (2009) reviewed the basic geology of possible cratons that might potentially have formed part of the postulated "southern" supercontinent of Aspler and Chiarenzelli (1998; possibly comprising "Zimvaalbara" [Stanistreet, 1993], the São Francisco, as well as Indian cratons), and found no support for this hypothesis.

The amalgamation of the Kaapvaal and Zimbabwe cratons along the Limpopo belt, implicit in the "Zimvaalbara" reconstruction, has been the subject of much debate. Historically, the metamorphic, magmatic and deformational characteristics of the Limpopo mobile belt (LMB) have been attributed to an Alpine-Himalayan style collision event between the Zimbabwe and Kaapvaal cratons at c. 2.7-2.6 Ga (Rigby *et al.* 2008a and references therein). However, in the late 1990's the first reports of the now ubiquitous Paleoprotorozoic age for metamorphism in the Central zone (CZ) started to emerge (Barton and Sergeev 1997; Jaeckel *et al.* 1997; Holzer *et al.* 1998; Kröner *et al.* 1999), and this ultimately led Holzer *et al.* (1998) to conclude that the "Limpopo orogeny" formed as a result of the oblique collision between the Kaapvaal and Zimbabwe cratons at c. 2.0 Ga. An Archean-only versus a Paleoproterozoic-only collision is, however, an oversimplication that is inconsistent with recent studies, which have unequivocally demonstrated that the LMB has a long and protracted evolution spanning over 700 My of Earth history (Barton *et al.* 2006; Boshoff *et al.* 2006; Zeh *et al.* 2007; Zeh *et al.* 2008; Millonig *et al.* 2008; Van Reenen *et al.* 2008; Perchuk *et al.* 2008; Chudy *et al.* 2008; Gerdes and Zeh 2009). The CZ is characterized by discrete metamorphic and magmatic activity which is attributable, in part, to a Neoarchean event (Millonig *et al.* 2008; Van Reenen *et al.* 2008; Gerdes and Zeh 2009), and a final c. 2.0 Ga overprint (Zeh *et al.* 2004; Zeh *et al.* 2005; Zeh *et al.* 2007; Perchuk *et al.* 2008; Rigby *et al.* 2008b; Van Reenen *et al.* 2008; Chudy *et al.* 2008; Rigby 2009). Conversely, the granulite-facies metamorphism developed in the Southern marginal zone (SMZ) is undisputedly characterized by a single monometamorphic P-T path, indicative of crustal thickening (Stevens and Van Reenen 1992) and constrained by U-Pb dating of monazite and zircon dating of melt leucosomes to be 2691+/-7 Ma and 2643+/-1 Ma, respectively (Kreissig *et al.* 2001). Furthermore, the thrusting of the granulite facies rocks of the SMZ onto the adjacent Kaapvaal Craton along the mylonitic oblique-dip slip Hout River shear zone (HRSZ) (e.g., Smit *et al.* 1992) is constrained by zircon dates from the syn-kinematic Matok intrusive complex (MIC) to be between c. 2671 and 2664 Ma (Barton and Van Reenen 1992; Barton *et al.* 1992) and by Ar-Ar dating of amphiboles from the HRSZ, which yield maximum ages ranging from 2650-2620 Ma (Kreissig *et al.* 2001). Collectively, these data favour a Central zone-Kaapvaal amalgamation during the Neoarchean, which is consistent with recent U-Pb and Lu-Hf data from zircons that indicate the exotic CZ accreted onto the Kaapvaal craton at 2.67-2.61 Ga (Zeh *et al.* 2009).

The Northern Marginal Zone (NMZ) also displays evidence of polymetamorphism, however, the Archean-aged metamorphism is not associated with accretion or collisional –style orogenesis but is commonly attributed to prolonged and widespread charno-enderbitic magmatism (Rollinson and Blenkinsop 1995; Kamber and Biino 1995;) in a northward dipping subduction zone setting (Kramers *et al.* 2001). A regional c. 2.0 Ga tectono-metamorphic event, similar to that recorded in parts of the CZ (Zeh *et al.* 2005; Rigby *et al.* 2008b), is reported for the southern half the NMZ (Kamber *et al.* 1995), which in addition to contemporaneous transpressional deformation in the CZ and NMZ suggests that a Kaapvaal-Zimbabwe amalgamation was a Paleoproterozoic phenomena (Kamber *et al.* 1995; Holzer *et al.* 1998; Rigby *et al.* 2008a). This assertion is also supported by applying high quality new data from the global LIP (large igneous province, cf., plume/superplume) record (e.g., Ernst *et al.* 2005); ongoing such work suggests that Zimbabwe may have belonged to the Superia supercontinent at about 2.7-2.6 Ga, with a Kaapvaal union only at about 2.0 Ga (R.E. Ernst, pers. comm. 2009).

4. Discussion: what was unique to Kaapvaal from c. 3.0 – 2.0 Ga: mantle plumes?

The brief lithostratigraphic and geodynamic-chronological history of the c. 3.1 – 2.05 Ga interval on Kaapvaal (Figure 2; section 2) provides the factual basis for postulates of possible Neoarchean-Paleoproterozoic supercontinents which may have included this craton. During evolution of the c. 3.1-2.8 Ga Witwatersrand basin, the Kaapvaal craton was still in its early development, undergoing amalgamation of its older nucleus with younger and smaller composite (greenstone-granitoid-gneissic) terranes (Figure 1); supercontinent reconstructions apart, during Witwatersrand times, the craton itself was still being formed (e.g., Robb and Meyer 1995). Examination of the c. 3.1 – 2.8 Ga preserved geological history of Neoarchean cratons (Pilbara, Zimbabwe, Dharwar, São Francisco, Amazon and Congo) which might potentially have been amalgamated with Kaapvaal (Eriksson *et al.* 2009) provides no support for the concept of a “southern” supercontinent as suggested by Aspler and Chiarenzelli (1998). On the contrary, a mantle superplume event at c. 3.0-2.9 Ga has been suggested (Abbott and Isley 2002); however, the pre-c. 2.7 Ga plume record can be seen as equivocal, although the LIP record appears rather continuous over time (Ernst *et al.* 2004, 2005). Eriksson *et al.* (2009) have postulated that such a possible global plume event may also have affected Kaapvaal, with a plume that impinged beneath the craton driving arc-subduction complexes “offshore” of the nucleus; these complexes would later have amalgamated with the nucleus, as the northern and western composite terranes discussed previously in this paper (see also,

Figure 1). These processes possibly also resulted in gold-rich source areas for the syn-craton amalgamation-aged Witwatersrand basin (Eriksson *et al.* 2009). The complex flexural foreland basin model applied to this basin is tied to amalgamation of relatively small composite terranes rather than any inherent building of an incipient supercontinent (e.g., Catuneanu 2001). Within the model proposed for Witwatersrand basin evolution by Eriksson *et al.* (2009) the role of an inferred plume was paramount over tectonic processes, both disturbed by and following upon plume impingement beneath Kaapvaal.

Following Witwatersrand sedimentation, Kaapvaal was affected by a much better defined c. 2.7 Ga superplume event of demonstrable global proportions, resulting in the Ventersdorp Supergroup. During the c. 100 My time gap separating these two supergroups, tectonic shortening of Witwatersrand sedimentary rocks occurred, with concomitant erosive loss of up to 1.5 km of stratigraphy (e.g., Hall 1996). The precise cause of this deformation is not known, but there are no preserved mobile belts or suitably aged greenstone belts marginal to the then-craton to support any significant collisions of the growing Kaapvaal craton with any other craton. In a widely accepted scenario, post-Witwatersrand deformation has been related by many to incipient collision along the LMB on the northern margin of Kaapvaal (e.g., Stanistreet and McCarthy 1991); however, recent dating does not support this, with only a collision of the exotic Central zone of the Limpopo with Kaapvaal at about 2691-2610 Ma (Barton and Van Reenen 1992; Barton *et al.* 1992; Kreissig *et al.* 2001; Zeh *et al.* 2009), well after formation of the Ventersdorp Supergroup. The latter was unconstrained by the existing tectonic architecture and grain of the Kaapvaal craton and can be interpreted within a fully mantle-thermally dominated plume model, resulting in 2714±8 Ma locally komatiitic mafic volcanics at the base, 2709±4 Ma medial graben-related bimodal volcanic and sedimentary deposits, with undated uppermost thermal subsidence related widespread sedimentary and volcanic lithologies (e.g., Armstrong *et al.* 1991; van der Westhuizen *et al.* 1991; Eriksson *et al.* 2002).

A plume event might also have been responsible for the set of discrete fault-bounded volcano-sedimentary basins of the Transvaal "protobasinal" phase; a c. 2.66-2.68 dyke swarm in the east of the craton might have been related to a plume responsible for both uppermost Ventersdorp volcanics and the protobasinal depositories (Olsson *et al.* in press). The protobasinal Godwan basin-fill was subsequently deformed, but there was no regional expression of this localised tectonic shortening event (e.g., Eriksson *et al.* 2001). This was followed by thermal relaxation and deposition of the thin sheet sandstones of the undated Black Reef Formation across the protobasinal rifting zone, with craton-marginal (SW of craton) passive margin deposits of the Vryburg Formation forming at 2642±3 Ma (Walraven and Martini 1995), which may be correlates (Eriksson *et al.* 2001). Syn- and post-Black Reef deformation (Hilliard and McCourt 1995) of these sandstones was restricted to the region of the Johannesburg dome (south of the TB) and along the paleo-Rand anticline E and W of it (Els *et al.* 1995; Eriksson *et al.* 2001). Subsequent drowning of much of the Kaapvaal craton during deposition of the thick carbonate-BIF platform succession, from <2642±3 Ma till at least 2432±31 Ma (Trendall *et al.* 1990) may have been related to a global crustal growth rate maximum and concomitant lowered freeboard (e.g., Eriksson *et al.* 2006). This resulted in an epeiric sea advancing onto the craton from the SW towards the NW, with two major drowning episodes, at c. 2.55 and 2.50 Ga (Altermann and Nelson 1998). The possible role of far-field tectonic forces related to the Kaapvaal – Limpopo Belt Central zone collision at c. 2691-2610 Ma (Barton and Van Reenen 1992; Barton *et al.* 1992; Kreissig *et al.* 2001; Zeh *et al.* 2009) for the Transvaal sedimentary basins has not been investigated, but an association would appear to be likely. However, once again, this does not constitute a supercontinental amalgamation, but rather the collision of a small crustal plate with shallow marine sediments covering much of its surface (cf., Central zone) with the Kaapvaal craton.

In the hiatus (~80-200 My?) separating chemical platform sediments from the Pretoria Group, weathering and erosive removal of BIF and carbonate lithologies was mainly along the southern preserved margin of the Transvaal basin (TB); a local downcutting event is thus inferred rather than widespread uplift and removal, which could be interpreted to reflect a major tectonic and thus possibly supercontinental-type event. With deposition of the uppermost clastic sedimentary (volcanic) Pretoria Group of the Transvaal Supergroup being interpreted as the result of two episodes of rifting and subsequent thermal subsidence which accommodated two major epicontinental seas advancing onto parts of the craton (Catuneanu and Eriksson 1999; Eriksson *et al.* 2001), a return can be motivated to the apparently thermally dominated history of Kaapvaal since c. 3.1 Ga. A plume influence is once again supported, with volcanics related to both rifting episodes, the second being tied to a major flood basalt at c. 2222±13 Ma (Cornell *et al.* 1996). The final major mantle plume related event on the craton was intrusion of the Bushveld Complex soon after Transvaal deposition ceased, at 2058±0.8 Ma (Buick *et al.* 2001). Subsequent collision of the Zimbabwe craton with the assembled Central zone-Kaapvaal crustal segment at c. 2.0 Ga (e.g. Kamber *et al.* 1995; Holzer *et al.* 1998; Kröner *et al.* 1999) finally resulted in Kaapvaal becoming amalgamated with another major craton to form its first undoubted “supercontinent”, albeit one of limited compass.

5. Conclusions

The brief synopsis of the geological evolution of the cover sequences on the Kaapvaal craton from c. 3.1 – 2.05 Ga presented in this paper suggests that mantle thermal processes, mainly in the form of plumes and superplume events might have been predominant over plate tectonic influences during this period, and that there is no unequivocal support for Kaapvaal having formed part of any supercontinental amalgamation prior to c. 2.0 Ga. Evidence for tectonic shortening, such as the deformation of the Godwan protobasinal fill, that related to the Black Reef Formation, or removal of chemical sediments along the southern TB, all appear to have been local events, and concomitantly all in the general area of the paleo-Rand anticline, a long-lived feature subject to repeated uplifts (e.g., Eriksson *et al.* 1991, 2001). None of these inferred tectonic shortening events lend themselves in either scale or intensity to an interpretation as far-field effects of a craton-marginal supercontinent assembly.

This raises the question of possible geodynamic settings for the craton during this long period of over a billion years, and here we have recourse to the possible conditions at the transition in Earth history from a thermally-dominated planet to one where a layered mantle and plate tectonics became pre-eminent, as discussed in the first paragraph of this paper. The Trendall (2002) model of such early crustal evolution, whereby the earliest cratonic nuclei developed above centres of convective descent on a fully molten early Earth, and as they grew and stabilized, were subject to lateral tectonic forces, underplating and overplating, as well as amalgamations with small composite granite-greenstone arc-subduction terranes, can possibly serve as at least a partial explanation of the supposedly dominant mantle-thermal character of Kaapvaal from c. 3.1 – 2.0 Ga. We thus suggest as the main thesis of this paper that the Kaapvaal craton during this time period might have been subject to an ongoing influence from mantle-thermal processes, beyond the transition of Earth from a mantle-dominated system into a fully plate tectonically driven system. Possibly, either Kaapvaal remained subject to the influence of a longer-lived mantle-thermal regime as an exception on a changing Earth already subject to dominant plate movements and a layered mantle, or, speculatively, the transition from mantle- to plate-dominated systems may have taken place over a much longer time period, and a later one at that, than is generally accepted or even implicit within the Trendall (2002) model itself.

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Figure 1. (a) Schematic sketch map of the Kaapvaal craton, showing early southeastern nucleus (made up of Barberton-S and –N terranes), accreted Murchison-North Kaapvaal (MNK) terrane, major Archean greenstone belts and the Witwatersrand-Pongola, Ventersdorp and Transvaal basins. Colesberg magnetic lineament represents the suture of the assembled B-S, B-N, MNK terranes with the Kimberley (=westerly accreted terrane) block. The Central Zone (LCZ terrane) of the Limpopo mobile belt subsequently accreted to the north of the assembled Kaapvaal craton. (Modified after de Wit *et al.* 1992; Cheney, 1996; Tinker *et al.* 2002; Zeh *et al.* 2009).

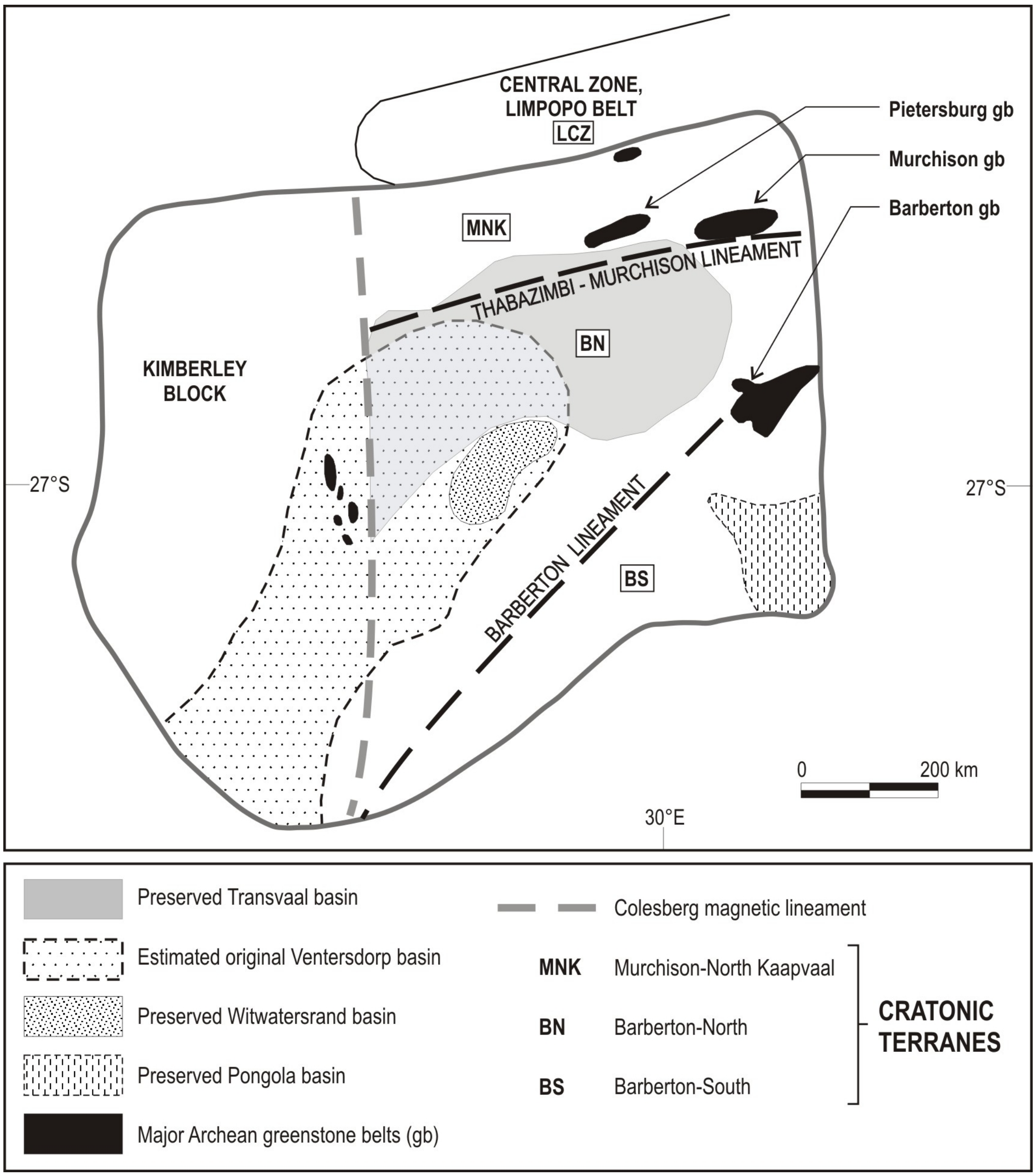
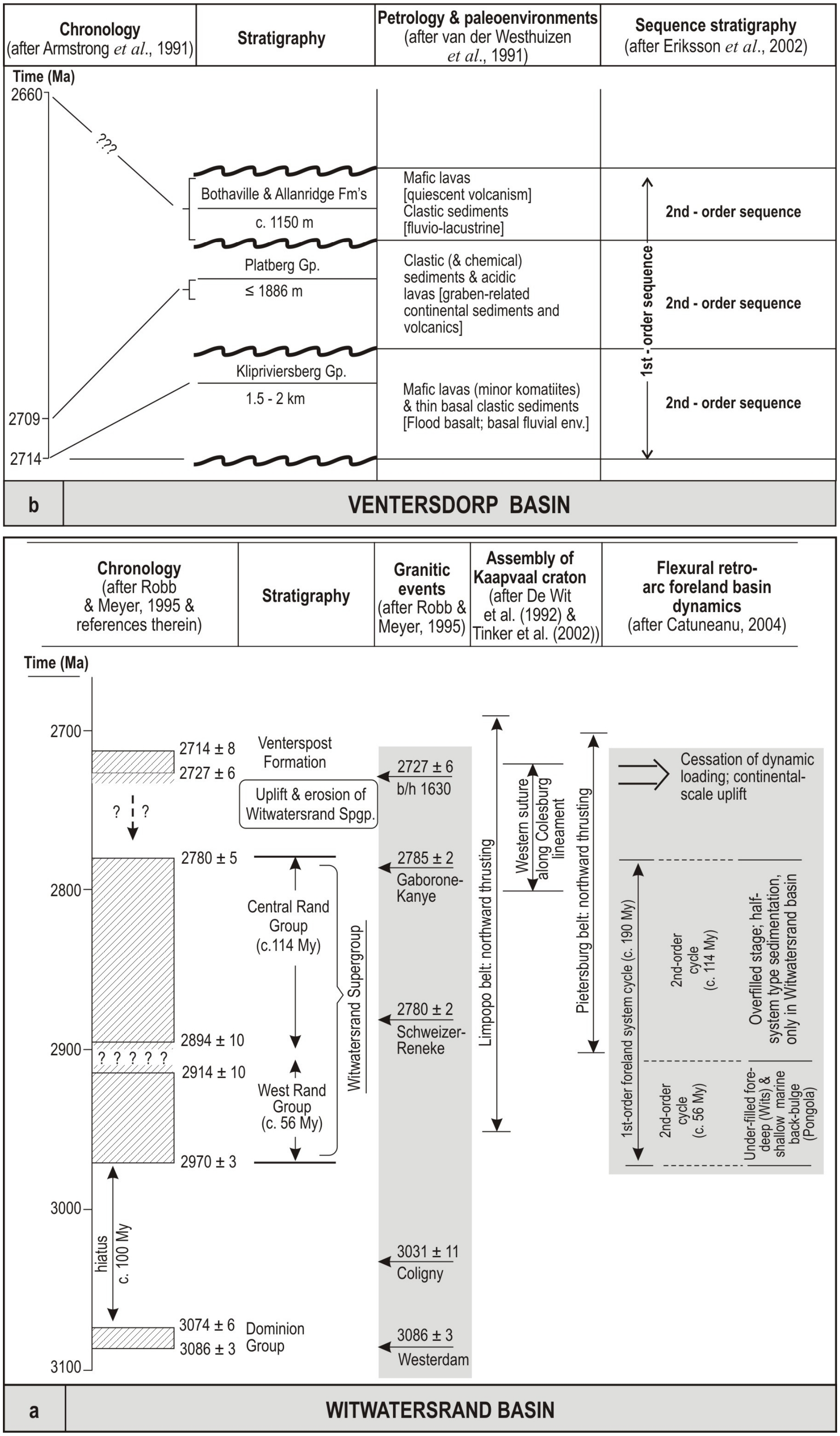
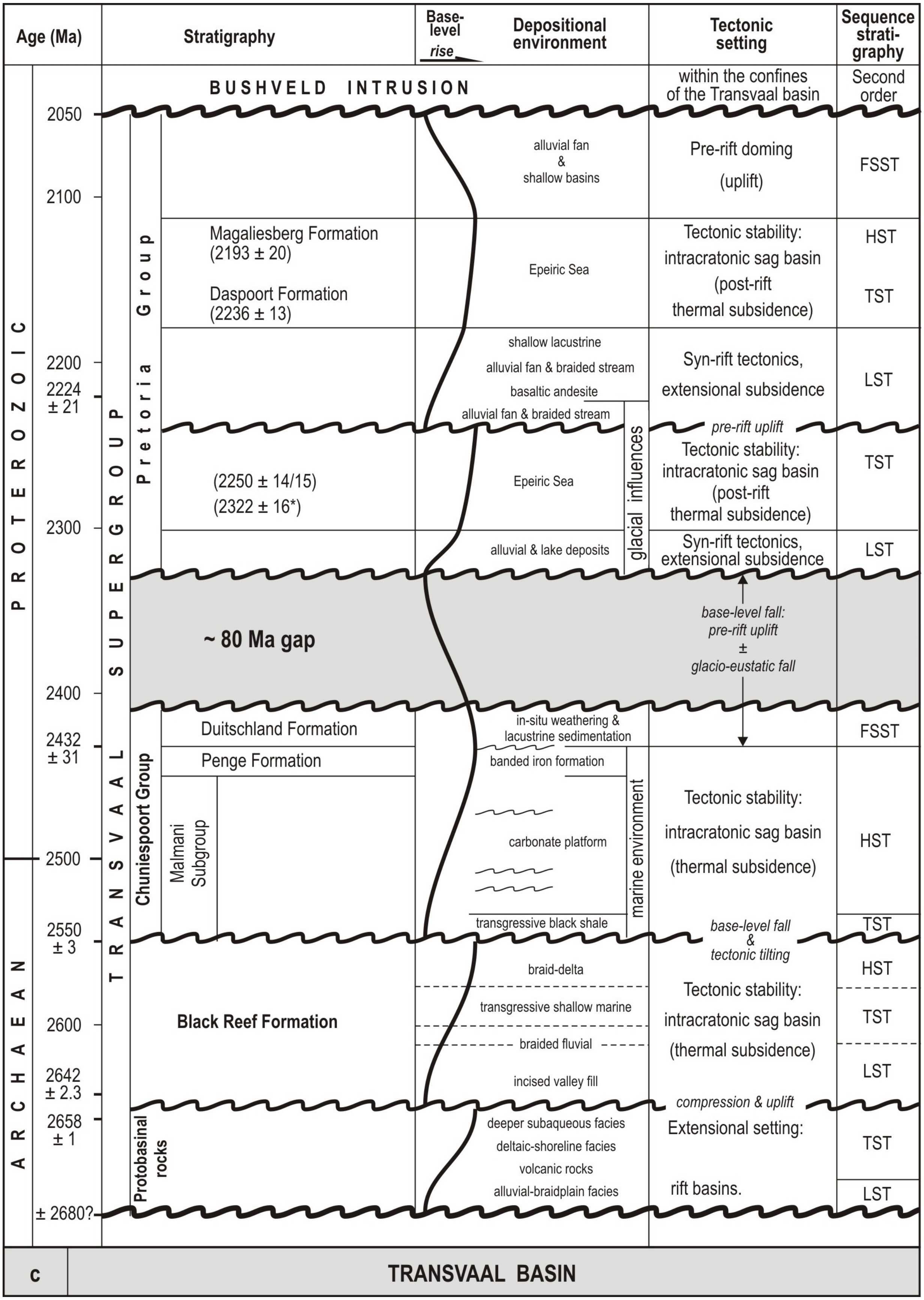


Figure 2. (a) Geodynamic history chart for the "greater Witwatersrand basin" (=Witwatersrand and correlated Pongola Supergroups), detailing chronology, stratigraphy and major granitic events affecting this basin and its hinterland, as well as major terrane accretion and amalgamation events affecting the Kaapvaal craton, and flexural retroarc foreland basin system stages for the greater Witwatersrand depository. (b) Geodynamic history chart for the Ventersdorp basin. Minimum age of c. 2.66 Ga (Olsson *et al.* in press) remains speculative. (c) Geodynamic history chart for the Transvaal Supergroup, shown only for the Transvaal basin, illustrates lithostratigraphy, chronology, inferred tectonic settings and depositional paleoenvironments, as well as interpreted base-level changes and sequence stratigraphy (modified after Catuneanu and Eriksson 1999). Maximum age for protobasinal rocks of c. 2.68 Ga (Olsson *et al.* in press) is speculative. Age data for Pretoria Group sandstones (shown in brackets; Mapeo *et al.* 2006) reflect detrital zircons and thus represent maximum depositional ages for sampled sandstones; age at base of Pretoria Group (shown with *) from Hannah *et al.* (2004); remaining age data from references in Eriksson *et al.* (2001).



Chronology (after Armstrong et al., 1991)
Stratigraphy
Petrology & paleoenvironments (after van der Westhuizen et al., 1991)
Sequence stratigraphy (after Eriksson et al., 2002)
Time (Ma)
2660
???
2709
2714
Bothaville & Allanridge Fm's
c. 1150 m
Platberg Gp.
≤ 1886 m
Klipriviersberg Gp.
1.5 - 2 km
Mafic lavas [quiescent volcanism] Clastic sediments [fluvio-lacustrine]
Clastic (& chemical) sediments & acidic lavas [graben-related continental sediments and volcanics]
Mafic lavas (minor komatiites) & thin basal clastic sediments [Flood basalt; basal fluvial env.]
1st - order sequence
2nd - order sequence
2nd - order sequence
2nd - order sequence
b
VENTERSDORP BASIN
Chronology (after Robb & Meyer, 1995 & references therein)
Stratigraphy
Granitic events (after Robb & Meyer, 1995)
Assembly of Kaapvaal craton (after De Wit et al. (1992) & Tinker et al. (2002))
Flexural retro-arc foreland basin dynamics (after Catuneanu, 2004)
Time (Ma)
2700
2800
2900
3000
3100
2714 ± 8
2727 ± 6
Venterspost Formation
Uplift & erosion of Witwatersrand Spgp.
2780 ± 5
2894 ± 10
2914 ± 10
2970 ± 3
3074 ± 6
3086 ± 3
Dominion Group
Central Rand Group (c.114 My)
West Rand Group (c. 56 My)
Witwatersrand Supergroup
hiatus c. 100 My
2727 ± 6 b/h 1630
2785 ± 2 Gaborone-Kanye
2780 ± 2 Schweizer-Reneke
3031 ± 11 Coligny
3086 ± 3 Westerdam
Limpopo belt: northward thrusting
Western suture along Colesburg lineament
Pietersburg belt: northward thrusting
Cessation of dynamic loading; continental-scale uplift
1st-order foreland system cycle (c. 190 My)
2nd-order cycle (c. 114 My)
2nd-order cycle (c. 56 My)
Overfilled stage; half-system type sedimentation, only in Witwatersrand basin
Under-filled fore-deep (Wits) & shallow marine back-bulge (Pongola)
a
WITWATERSRAND BASIN



Age (Ma)
Stratigraphy
Base-level rise
Depositional environment
Tectonic setting
Sequence strati-graphy
BUSHVELD INTRUSION
within the confines of the Transvaal basin
Second order
2050
2100
2200
2224 ± 21
2300
2400
2432 ± 31
2500
2550 ± 3
2600
2642 ± 2.3
2658 ± 1
± 2680?
PROTEROZOIC
ARCHAEAN
TRANSVAAL SUPERGROUP
Pretoria Group
Chuniespoort Group
Malmani Subgroup
Protobasinal rocks
Magaliesberg Formation (2193 ± 20)
Daspoort Formation (2236 ± 13)
(2250 ± 14/15)
(2322 ± 16*)
~ 80 Ma gap
Duitschland Formation
Penge Formation
Black Reef Formation
alluvial fan & shallow basins
Epeiric Sea
shallow lacustrine
alluvial fan & braided stream
basaltic andesite
alluvial fan & braided stream
Epeiric Sea
alluvial & lake deposits
glacial influences
in-situ weathering & lacustrine sedimentation
banded iron formation
carbonate platform
transgressive black shale
marine environment
braid-delta
transgressive shallow marine
braided fluvial
incised valley fill
deeper subaqueous facies
deltaic-shoreline facies
volcanic rocks
alluvial-braidplain facies
Pre-rift doming (uplift)
Tectonic stability: intracratonic sag basin (post-rift thermal subsidence)
Syn-rift tectonics, extensional subsidence
pre-rift uplift
Tectonic stability: intracratonic sag basin (post-rift thermal subsidence)
Syn-rift tectonics, extensional subsidence
base-level fall: pre-rift uplift ± glacio-eustatic fall
Tectonic stability: intracratonic sag basin (thermal subsidence)
base-level fall & tectonic tilting
Tectonic stability: intracratonic sag basin (thermal subsidence)
compression & uplift
Extensional setting:
rift basins.
FSST
HST
TST
LST
TST
LST
FSST
HST
TST
HST
TST
LST
TST
LST
c
TRANSVAAL BASIN