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Sequence Stratigraphic Development of the Neoarchean Transvaal carbonate platform, Kaapvaal Craton, South Africa

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ABSTRACT

The ~2.67 to ~2.46 Ga lower Transvaal Supergroup, South Africa, consists of a mixed siliciclastic-carbonate ramp that grades upward into an extensive carbonate platform, overlain by deep subtidal banded iron-formation. It is composed of 14 third-order sequences that develop from a mixed siliciclastic-carbonate ramp to a steepened margin followed by a rimmed margin that separated lagoonal environments from the open ocean. Drowning of the platform coincided with deposition of banded iron-formation across the Kaapvaal Craton. The geometry and stacking of these sequences are consistent with more recent patterns of carbonate accumulation, demonstrating that Neoarchean carbonate accumulation responded to subsidence, sea level change, and carbonate production similarly to Proterozoic and Phanerozoic platforms. The similarity of carbonate platform geometry through time, even with significant changes in dominant biota, demonstrates that rimmed margins are localized primarily by physiochemical conditions rather than growth dynamics of specific organisms.

Stratigraphic patterns during deposition of the Schmidtsdrift and Campbellrand-Malmani subgroups are most consistent with variable thinning of the Kaapvaal Craton during extrusion of the ~2.7 Ga Ventersdorp lavas. Although depositional patterns are consistent with rifting of the western margin of the Kaapvaal Craton during this time, a rift-to-drift transition is not required to explain subsidence. Heating and thinning during Ventersdorp time can produce the observed thermal subsidence from ~2.7 to ~2.45 Ga if the thermal diffusivity of the craton was moderately low.

Introduction

Carbonate platforms record the interplay of life, ocean chemistry, sea level, and subsidence through Earth history. Studies of Proterozoic and Phanerozoic microbial reefs demonstrate that platform geometry, response to changes in accommodation space, and facies distributions are similar even in the absence of skeletal organisms to build the reef framework (*e.g.*, Playford *et al.*, 1989; Sami and James, 1994; Elrick, 1996; Stephens and Sumner, 2003; Porta *et al.*, 2004). These similarities have been documented by detailed facies and sequence stratigraphic studies of platform architecture, allowing reconstruction of facies patterns, relative sea level change, and tectonic environment. Results demonstrate that carbonate accumulation is often predominantly driven by physical and chemical dynamics of platform growth rather than specific biological activity. These results predict similar behavior for carbonate platforms of all ages, including Archean carbonates.

Few Neoarchean carbonates are preserved extensively enough to allow sequence stratigraphic analysis of platform dynamics. The Campbellrand-Malmani carbonate platform of the Transvaal Supergroup, South Africa, is one such platform (Beukes, 1980; 1987; Sumner, 1995; Sumner and Grotzinger, 2004). It provides a case study of platform development during latest Archean time that demonstrates many similarities to younger platforms in terms of geometry and stratigraphic development even though the characteristics of many of the facies are distinctly different (Beukes, 1987; Grotzinger, 1989; Sumner and Grotzinger, 2004). The sequence stratigraphic development of this platform demonstrates that the dynamics of sedimentation, subsidence, and sea level change were similar from Neoarchean to Phanerozoic time even with dramatic evolutionary changes in reef-building organisms.

Geologic Setting

The lower Transvaal Supergroup consists of a mixed siliciclastic-carbonate ramp that grades upward into an extensive carbonate platform, overlain by banded iron-formation. It was deposited on the Kaapvaal Craton between ~2.67 to ~2.46 Ga (Table 1; Armstrong *et al.*, 1986; Barton *et al.*, 1994; Walraven and Martini, 1995; Pickard, 2003; Dorland, 2004). Lower Transvaal Supergroup strata are extremely well preserved. Structural disruption is limited to gentle warping over most of the craton with locally steeper dips around the ~2.06 Ga Bushveld Complex (Walraven *et al.*, 1990) and intense folding and faulting in the Kheis Belt and the

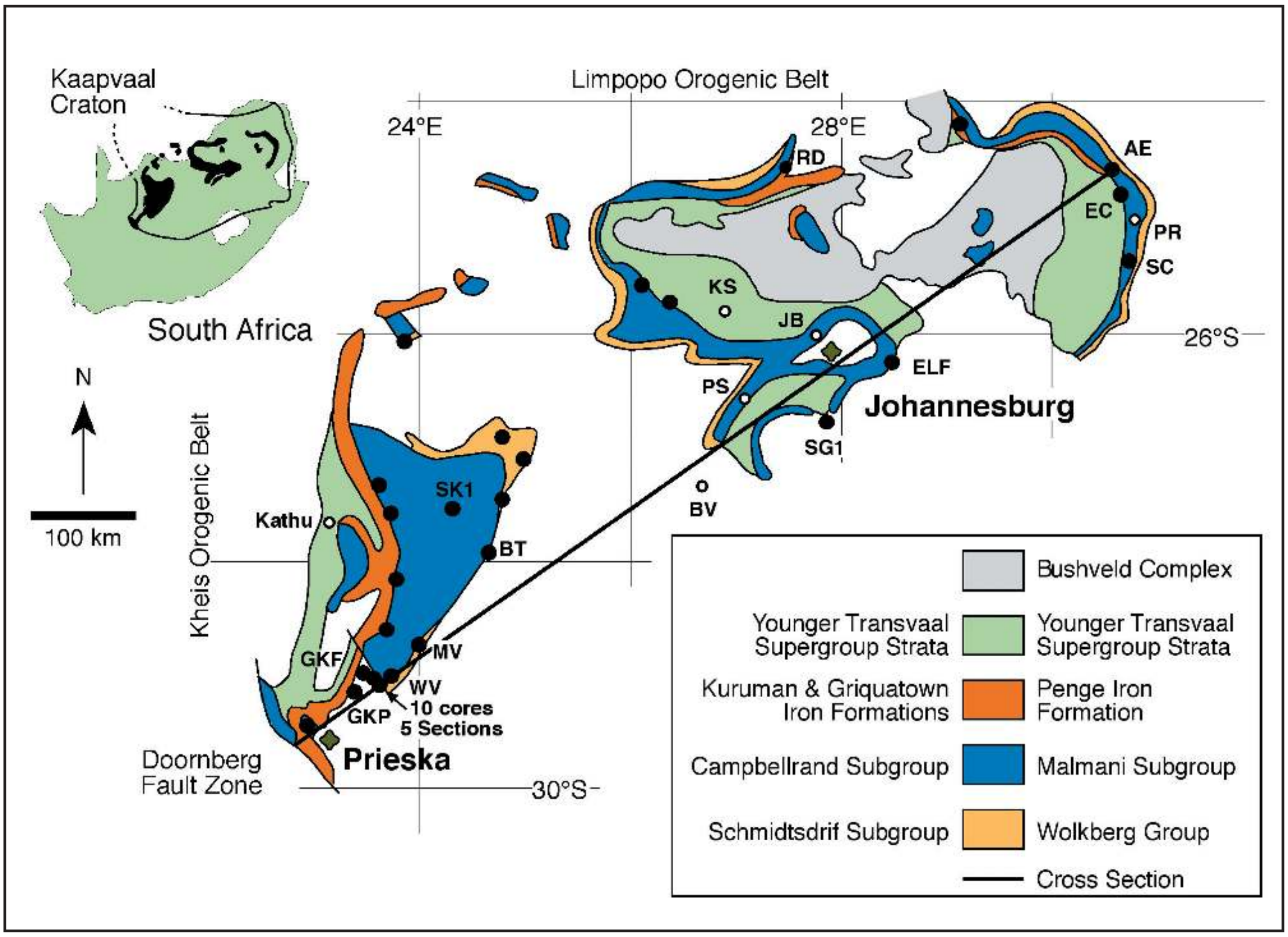

Figure 1. Map of exposures of the lower Transvaal Sequence, South Africa. Solid circles represent stratigraphic sections and core logs measured by the authors (Beukes, 1977; 1984; Sumner, 1995; Schröder *et al.*, 2006). Open circles represent stratigraphic data taken from the literature, including Clendenin (PR, KS, PS, and BV; Clendenin, 1989) and Eriksson and Truswell (JB; Eriksson and Truswell, 1974). Data from Kathu come from both Alterman and Siegfried (1997) and core logs by the authors.

Doornberg Fault Zone, which coincide with the western boundary of the Kaapvaal craton (Figure 1; Stowe, 1986; Beukes and Smit, 1987; Cornell *et al.*, 1998). Post-depositional erosion limits the distribution of preserved strata, which does not reflect the extent and shape of the original depositional basin. The eastern half of the platform is traditionally called the Transvaal Basin (TB), whereas the western half is called the Griqualand West Basin (GB).

Most outcrops experienced sub-greenschist facies metamorphism (Button, 1973b; Miyano and Beukes, 1984). However, amphibole is locally present due to Bushveld contact metamorphism in the Malmani Subgroup, and supergene alteration during late fluid flow produced local Pb-Zn, fluorite, and gold deposits in both the Malmani and Campbellrand subgroups (*e.g.*, Martini, 1976; Clay, 1986; Duane *et al.*, 1991; Tyler and Tyler, 1996; Duane *et al.*, 2004). Early, fabric-retentive dolomite replaced most of the Malmani carbonates, particularly peritidal facies, which are also associated with chert replacement (Button, 1973b; Eriksson et al., 1975; Eriksson *et al.*, 1976). However, significant amounts of the Campbellrand Subgroup still consist of limestone (Beukes, 1987; Sumner and Grotzinger, 2004).

The Kaapvaal craton forms basement for the Transvaal Supergroup. It consists of diverse Archean granitoids, gneisses, and greenstone belts (see review in Poujol *et al.*, 2003) with cratonic sediments as old as the 3.07 Ga Dominion Group, the >2837±5 Ma Mozaan Group, and the correlative Witwatersrand Supergroup (Armstrong *et al.*, 1991; Gutzmer *et al.*, 1999; Poujol *et al.*, 2003). These are overlain by the 2740-2690 Ma Ventersdorp Supergroup (Armstrong *et al.*, 1991; Schmitz and Bowring, 2003b), a succession of sedimentary and volcanic rocks including extensive flood basalts. Extrusion of the Ventersdorp Supergroup was associated with crustal extension and ultra high temperature metamorphism of the mantle lithosphere (Schmitz and Bowring, 2003a; b). Subsidence associated with slow cooling of the lithosphere likely provided accommodation space for the Transvaal Supergroup.

A series of narrow basins filled with volcanics and siliciclastic sediments overlie cratonic basement where the Ventersdorp Supergroup is absent. They include the Derdepoort Belt, Tshwene-Tshwene Belt, Mogobane Formation, Buffelsfontein Group, Wachteenbeetje Formation, Bloempoort Formation, Godwan Group, and Wolkberg Group (Button, 1973a; Tyler, 1979; Hartzer, 1989; Eriksson and Reczko, 1995; Catuneanu and Eriksson, 1999; Eriksson *et al*., 2001). These basins have been variously correlated to the Ventersdorp and lowermost Transvaal supergroups, and it is likely that they represent basins of various ages. The Derdepoort Belt contains felsic volcanics deposited at 2781±5 Ma (Wingate, 1998), which is older than the dated portions of the Ventersdorp Supergroup, which yields 2714±8 Ma zircons (Armstrong *et al*., 1991), although this age may not represent earliest Ventersdorp Supergroup volcanism (see discussion in Wingate, 1998). In contrast, volcanics in the Buffelsfontein Group are 2664±6 Ma, which is similar in age to the Schmidtsdrif Group, Transvaal Supergroup (Dorland, 2004). Based on the proportions of volcanic to non-volcanic deposits, it is reasonable to consider the Derdepoort and Tshwene-Tshwene belts equivalent to the Ventersdorp Supergroup, and the Buffelsfontein Group, Wachteenbeetje Formation, Bloempoort Formation, Godwan Group, and Wolkberg Group ("Wolkberg-equivalent units") as associated with lowermost Transvaal Supergroup deposition until sufficient stratigraphic and age data are available to refine correlations.

The Transvaal Supergroup unconformably overlies the Ventersdorp Supergroup, the Witwatersrand Supergroup, and granitic-greenstone basement (Figure 1; Button, 1973a; Beukes, 1977; 1979; Tyler, 1979; Clendenin *et al*., 1991; Els *et al*., 1995). The base of the Transvaal Supergroup in the TB is marked by 0-2 km of quartz arenite and shale forming the Wolkberg-equivalent units (Figures 1 and 2; Button, 1973a; Tyler, 1979; Eriksson and Reczko, 1995; Hartzer, 1995; Catuneanu and Eriksson, 1999). Isopach maps reveal large thickness variations in Wolkberg-equivalent units across the eastern Kaapvaal Craton related to paleotopography and differential subsidence (Button, 1973a; Tyler, 1979; Eriksson and Reczko, 1995; Hartzer, 1995), and deposition of at least the Buffelsfontein Group may overlap metamorphism in the Southern Marginal Zone of the Limpopo Belt (Barton and Van Reenen, 1992; Barton *et al*., 1995; Kreissig *et al*., 2001). However, the Limpopo Belt contains abundant ~2.7 to ~2.6 Ga zircons, whereas detrital zircons in Wolkberg sandstones are dominantly older than ~2.9 Ga, suggesting that uplift of the Limpopo Belt did not provide sediment to the Wolkberg basin (Dorland, 2004). Thus, deposition and subsequent deformation of Wolkberg-equivalent units are unlikely to be related to Limpopo Belt uplift; rather deformation and subsidence may be due to residual extension after extrusion of the Ventersdorp Supergroup. Wolkberg-equivalent sediments are unconformably overlain by the fluvial to shallow marine Black Reef Quartzite, which was broadly deposited over much of the craton with a sheet-like geometry (Eriksson *et al*., 2005). The Black Reef Quartzite grades upward into the Malmani carbonate platform (Button, 1973b; Clendenin *et al*., 1991; Els *et al*., 1995; Eriksson *et al*., 2005). In the GB, the base of the Transvaal Supergroup consists of mixed siliciclastic and carbonate rocks of the ~2.67 to ~2.65 Ga Schmidtsdrif Subgroup (Beukes, 1979; Dorland, 2004). The Schmidtsdrif Subgroup forms a sedimentary ramp that deepens southwestward and grades conformably upward into the carbonate platform of the Campbellrand Subgroup (Figure 2; Beukes, 1977; 1979; 1983a).

The 2650-2500 Ma Campbellrand and Malmani subgroups are correlative and form an extensive carbonate platform (*e.g*., Button, 1976; Cheney, 1996; Martin *et al*., 1998). Preserved outcrops cover 190,000 km^2 (Figure 1) and probably originally covered the entire Kaapvaal Craton, >600,000 km^2 (Beukes, 1980; 1987). The platform is 1.5 to 2 km thick, with the predominantly peritidal facies in the north and east and deeper facies to the south and west (Figure 2; Beukes, 1987; Sumner and Grotzinger, 2004). Platform slope and basinal sediments are preserved near Prieska and are only about 500 m thick (Figures 1 and 2; Beukes, 1987; Sumner and Grotzinger, 2004). The ~2.5 to ~2.46 Ga Kuruman and Penge iron-formations conformably overlie the Campbellrand and Malmani subgroups, respectively (Figures 1 and 2). The Kuruman Iron Formation consists of deep water, microbanded iron formation (Beukes, 1984; Klein and Beukes, 1989), and the correlative lower Penge Iron Formation is lithologically similar. Both formed on a stable marine shelf below wave base and then shallowed to sea level (Beukes, 1983b; Miyano and Beukes, 1997).

Overall, Wolkberg-Schmidtsdrif to lower Penge-Kuruman deposition represents a genetically related package of sediments, whose accommodation space reflects evolution of the Kaapvaal craton after heating and thinning during Ventersdorp extrusion. They record the fluvial to marine transition during flooding of the craton, initiation of a long-lived, stable carbonate platform, drowning of that platform just prior to substantial iron-formation deposition, and subsequent shallowing back to sea level *(e.g*., Cheney, 1996; Martin *et al*., 1998; Eriksson *et al*., 2001).

Data and Methods

Age Constraints

Several U-Pb and Pb-Pb ages from zircons in tuffaceous sediments constrain the timing of lower Transvaal sedimentation (Table 1). Common inherited zircons produce ages older than sediment deposition indicating significant zircon inheritance (*e.g*. Altermann and Nelson, 1998; Pickard, 2003) and raising questions about the reliability of many age constraints (Table 1). In addition, analytical data for some age constraints have not been published; rather, they have been reported as "personal communication" or papers "in press". Many of these ages are not traceable to specific stratigraphic locations, and many have remained unpublished due to analytical complications. In the absence of sufficient data to evaluate the quality of these ages, we avoid using ages for which analytical data have not be published to evaluate sequence stratigraphic

Table 1. Published U-Pb and Pb-Pb Zircon Ages

Geological Unit	Reported Ages (Ma)	Interpretation
Kuruman Iron Formation, Riries Member	2460 ± 5 (Pickard, 2003)	Depositional age
Gamohaan Formation	2516 ± 4 (Altermann and Nelson, 1998)	
	2521 ± 3 (Sumner and Bowring, 1996)	Depositional ages
Nauga Formation		
480 m above base	2549 ± 7 (Altermann and Nelson, 1998)	May be depositional ages, but
450 m above base	2552 ± 11 (Barton *et al*., 1994a)	inheritance may also be an issue
Oaktree Formation	2550 ± 3 (Walraven and Martini, 1995)	Poor analytical technique for 2550 age;
	2588 ± 6 (Martin *et al*., 1998)	depositional age for 2588 age
Monteville Formation	2717 ± 26, 2637 ± 30, 2555 ± 19, 2455 ± 32	Inherited zircons and lead-loss or
	(Altermann and Nelson, 1998)	contaminants; no clear depositional ages

correlations. Ages listed as probable depositional ages in Table 1 have been used.

Stratigraphic Data

Facies descriptions and stratigraphic data from Beukes (1977; 1979; 1983a) form the basis of the Schmidtsdrif Subgroup sequence stratigraphic interpretation. Data for the correlative Buffelsfontein Group, Wachteenbeetje Formation, Wolkberg Group, and other correlative units show significant lateral variability (Button, 1973a; Tyler, 1979; Hartzer, 1989; Clendenin *et al*., 1991; Barton *et al*., 1995), and insufficient stratigraphic data have been published to allow detailed stratigraphic correlations among these Wolkberg-equivalent units. Stratigraphic data for the Campbellrand Subgroup are based on measured sections of Beukes (1980; 1987); Sumner (1995; 1997a), Altermann and Siegfried (1997), and Schröder *et al*. (2006), supplemented with additional sections and cores logged for this project. The stratigraphy of the Malmani Subgroup is based on a compilation of sections and core logs from Eriksson (1972), Button (1973b), Eriksson and Truswell (1974), Clendenin (1989), and Sumner (1995). Substantial variations in reported rock thicknesses in the Johannesburg area may be due to the absence of stratigraphic dip corrections for some logged cores from the literature and the composite nature of several field sections (see discussion in Sumner, 1995). However, Eriksson (1977) proposed uplift in the Johannesburg area during carbonate deposition based on facies distributions, which may also account for variable thicknesses. Data presented here are compiled from numerous publications and are not detailed enough to resolve this question. Thus, thickness variations have been smoothed to fit thicknesses reported by Sumner (1995), which are consistent among field sections and core logs that are corrected for stratigraphic dip. Stratigraphic data from Beukes (1984) and Miyano and Beukes (1997) were used for the thicknesses of the Kuruman-Griquatown and Penge iron-formations.

When projected into a cross section trend of N55°E from Prieska to section AE at Abel Erasmus Pass, stratigraphic data show across-strike depositional variations from below wave base to supratidal depositional environments (Figure 2). Sections within 100 km of the section line were used to constrain stratigraphic thicknesses within sequences, whereas, stratigraphic data collected more than 100 km from the section line were used to constrain facies and define sequence boundaries. However, thicknesses of individual units from distant sections were adjusted to provide consistency with sections closer to the projected cross section.

Facies interpretations provide the basis for identifying sequence boundaries and flooding intervals (Table 2; Eriksson, 1972; Button, 1973b; Eriksson and Truswell, 1974; Eriksson *et al*., 1975; Eriksson *et al*., 1976; 1977; Beukes, 1979; 1980; 1983a; 1987; Sumner, 1995; 1997a; b; Sumner and Grotzinger, 2004). Local to regional unconformities correspond to significant shallowing of facies, demonstrating relative sea level fall. These were correlated to lateral conformities using facies patterns that reflect significant shallowing of relative sea level. Maximum water depths, as determined by facies distributions, allow identification of flooding intervals. Volcanic tuff and accretionary lapilli beds provide regional time line correlations, as do two impact spherule layers (Simonson *et al*., 2000a; Hassler and Simonson, 2001; Simonson and Sumner, 2004). Stratigraphic data from the Northern Cape Province were used to initially define sequences because the subtidal to intertidal depositional environments preserve the best evidence of relative sea level change. Sequences were then correlated to Transvaal Basin strata using the best available age constraints, the magnitude of basinward shifts in facies, and the extent of interpreted relative sea level change. Uncertainties in correlations are discussed below.

Sequence Stratigraphy

Schmidtsdrif Sequences

Two sequences can be identified in the Schmidtsdrif Subgroup, a lower sequence (SW1) immediately overlying the unconformity above the Ventersdorp Supergroup, and an upper sequence (SW2) that ends at the top of the Monteville Formation (Figure 2). SW1 is composed of transgressive facies of the Vryburg Formation followed by the high stand Boomplaas Formation. It represents the first marine transgression in the GB after Ventersdorp flood basalt extrusion. Seismic

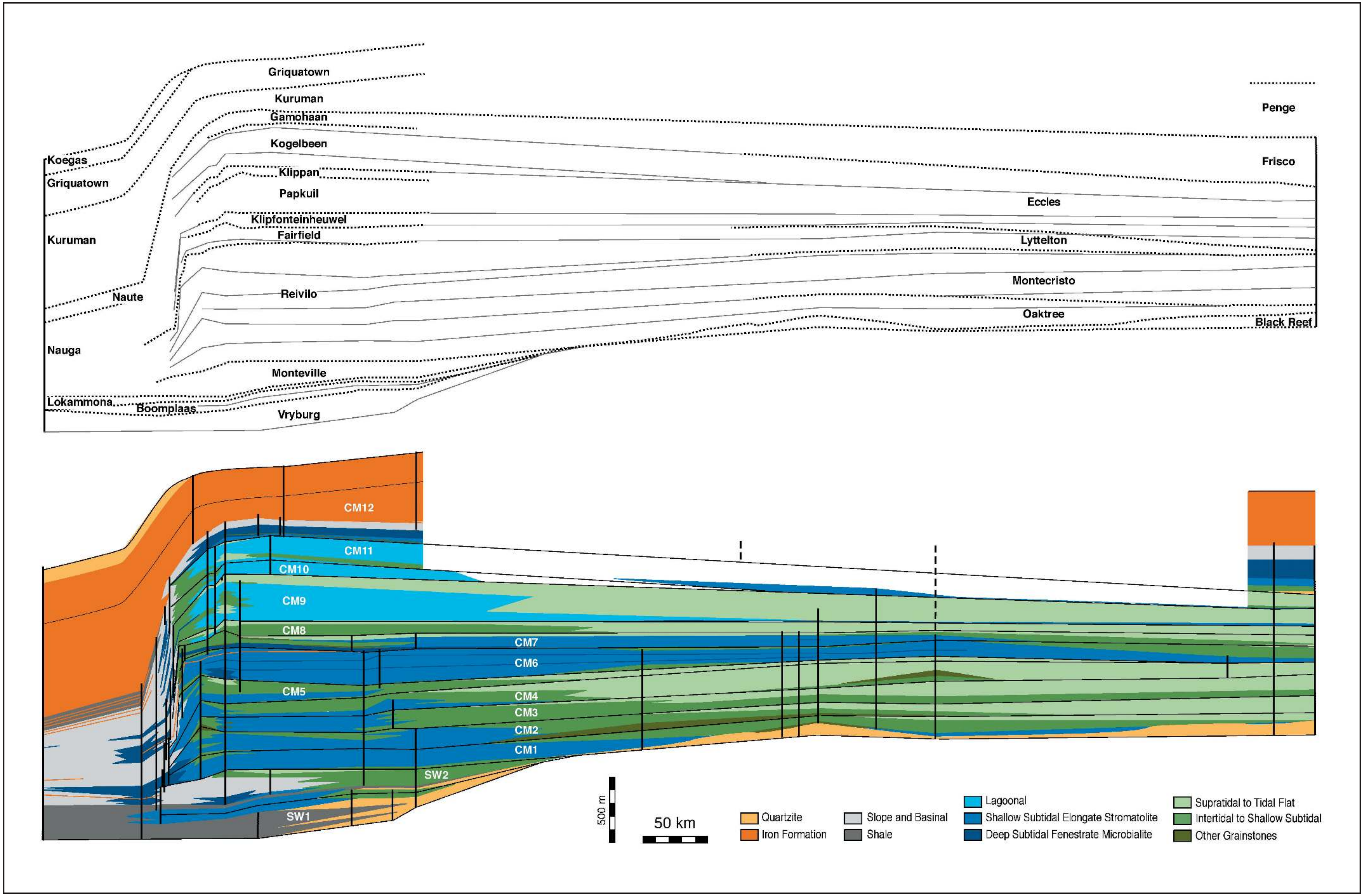

Figure 2. Cross section along line shown in Figure 1. (**A**) Formation boundaries are shown as dashed black lines and sequence boundaries are shown as gray lines. (**B**) Facies distributions and sequence boundaries show the development of the Campbellrand-Malmani carbonate platform. Vertical lines correspond to sections and cores marked in Figure 1. Sequence names are labeled in white letters).

Table 2. Lithofacies Assemblages (after Sumner and Grotzinger, 2004)

Lithofacies Assemblage	**Major Components**	**Environment Indicators**
Supratidal to Tidal Flat	Domal stromatolites, wavy-laminated dolomite, grainstones, micrite, breccia, aragonite pseudomorphs, halite pseudomorphs, herringbone calcite layers, quartz sand, shale	Erosional channels; wave and interference ripples imbricated intraclasts; mudcracks; tepee structures; rare evaporite pseudomorphs
Intertidal to Shallow Subtidal	Columnar and domal stromatolites; aragonite pseudomorphs; oolitic grainstones; other grainstones	Erosional unconformities; rare channeling; ripple, small dune, and low angle cross stratification; 1m synoptic relief of bioherms
Lagoonal	Fenestral, flat-laminated microbial mat; small, isopachously laminated domal stromatolites; rare breccias with <1 cm clasts; truncation surfaces	Small truncation surfaces; lack of cross stratification
Shallow Subtidal Elongate Stromatolite	Elongate 10's m diameter stromatolites; columnar stromatolites; aragonite pseudomorphs; stromatolite-herringbone calcite and grainstone-herringbone calcite cycles	Wave rippled and current rippled grainstones; bi-directional cross stratification; rare breccia
Deep Subtidal Fenestrate Microbialite	Fenestrate microbialites	Absence of cross stratification indicating strong currents; stratigraphic context
Slope and Basinal	Planar laminated dolomite; crinkly laminated dolomite; slump breccia and folds; nodular dolomitic limestone; shale; iron-formation; fenestrate microbialites	Turbidites; current ripple lamination; slump folds with downslope vergence
Iron-formation	Hematite, siderite, magnetite, greenalite, and chert	Various depositional environments
Other Grainstones	Massive beds; rare cross stratification (ripple, hummocky, trough cross stratification)	Various depositional environments
Quartzite	Quartz sand with lesser feldspar, pyrite, and other mafics; wave and current ripple lamination; trough, planar, & hummocky cross stratification	Various depositional environments
Shale	Fine-grained mudstone with abundant pyrite	Various depositional environments

data show that the Vryburg Formation is thicker in the west and progressively onlaps Ventersdorp deposits from west to east (Tinker *et al.*, 2002), consistent with the transgressive interpretation. Siliciclastic influx declined during the subsequent high stand, allowing accumulation of shallow subtidal carbonates of the Boomplaas Formation. The top of SW1 corresponds to an edgewise chert breccia and the return to transgressive facies patterns at the top of the Boomplaas Formation and the base of the Lokamona Formation, respectively.

Correlation of this sequence from the GB to the TB is uncertain. Carbonates and chertified stromatolites within the Buffelsfontein and Wachteenbeetje groups (Eriksson and Reczko, 1995; Hartzer, 1995) may correspond to development of carbonate facies in the Boomplaas Formation. However, Wolkberg-equivalent units do not show equivalent stratigraphic trends to SW1 in GB. These sediments are preserved in trough-like basins, pinching from greater than two kilometers thick to zero thickness across less than 50 km (Button, 1973a; Eriksson and Reczko, 1995; Hartzer, 1995). Unconformities within these predominantly fluvial coarse siliciclastic rocks demonstrate that differential subsidence and synsedimentary tectonism strongly influenced sedimentation patterns (Button, 1973a; Eriksson and Reczko, 1995; Hartzer, 1995). The lower fluvial sediments may correspond to deposition of SW1 with marine transgressive and near-shore deposits corresponding to transgression above the SW1 upper sequence boundary, but significantly more stratigraphic and chronological data are needed to establish a temporal correlation.

SW2 overlies the sequence boundary at the top of the Boomplaas Formation and consists of the Lokammona and the Monteville formations. It is composed of mixed siliciclastic and carbonate sediments, with an increasing proportion of carbonate facies upward and toward shallower subtidal depositional environments (Beukes, 1977; 1987; Bishop, 2004). The platform developed a steepened ramp geometry near the top of the Monteville Formation (Figure 2). The top of SW2 is defined as the unconformity and correlative conformity at the Monteville-Reivilo formation boundary. An impact spherule layer near the base of the Monteville Formation allows reliable correlation of this stratigraphic interval within the GB (Simonson *et al.*, 1999; Simonson *et al.*, 2000b; Hassler and Simonson, 2001).

SW2 temporally correlates to the lower Black Reef Formation. The lower Black Reef Formation consists of alluvial and fluvial facies that filled in topography on a landscape underlain by granite-greenstone basement, the Witwatersrand Supergroup, the Ventersdorp Supergroup, and Wolkberg-equivalent groups (Button, 1973a; b; Els *et al.*, 1995; Catuneanu and Eriksson, 1999; Eriksson *et al.*, 2001). A transgression resulted in a transition from fluvial to marine depositional environments and then a gradational transition into subtidal carbonate deposition of the Oaktree Formation (Figure 2; Button, 1973b; Clendenin *et al*., 1991; Eriksson *et al*., 2001; Eriksson *et al*., 2005).

The sequence boundary at the top of SW2 is firmly correlated to fluvial facies in the Black Reef Formation for several reasons. First, SW2 received a significant influx of siliciclastic sediment, which suggests correlative environments to the north and east consisted of erosional provinces or areas of siliciclastic deposition. Thus, SW2 cannot correlate to the carbonate-dominated Malmani Subgroup in the TB. Second, the unconformity at the top of SW2 is associated with extensive exposure of the Ventersdorp Supergroup as demonstrated by the merging of the sequence boundary with the top of the Ventersdorp between the GB and TB and zircons of Ventersdorp age in siliciclastics at the boundary (Figures 1 and 2, Table 1; Dorland, 2004). Both the Ventersdorp and Witwatersrand supergroups were exposed during deposition of the lower Black Reef Formation (Els *et al*., 1995; Dorland, 2004). Erosion of the Witwatersrand quartzites may have provided a source of mature quartz sand deposited in the Motiton Member at the top of the Monteville Formation. Third, both the Monteville and lower Black Reef formations represent the last units with strong regional to local topographic influences on sediment accumulation prior to flooding of the Kaapvaal Craton. After deposition of SW2, sequences extended across much or all of the preserved craton, and older rocks were not exposed until after deposition of the iron-formations at the top of the platform. Thus, the sequence boundary at the top of SW2 in the GB correlates to fluvial facies in the Black Reef Formation, in contrast to previous correlations which suggest that the Monteville and Oaktree formations are correlative (Beukes, 1986; Clendenin *et al*., 1991; Altermann and Wotherspoon, 1995; Altermann and Nelson, 1998).

Campbellrand-Malmani Sequences

The main Campbellrand-Malmani carbonate platform contains 12 sequences (CM), whose boundaries can be correlated across most of the preserved platform. Sequences generally thin to the northeast, and several pinch out entirely towards the platform interior. They reflect the flooding of the Kaapvaal Craton and initiation of widespread carbonate deposition (CM1), development of a platform with a steepened margin (CM2-CM5), reflooding of the platform (CM6), development of a rimmed platform with lagoonal to supratidal deposition dominating platform aggradation (CM7-CM11), and final drowning of the carbonate platform and initiation of regional iron-formation deposition (CM12).

CM1: First Transgression Across Entire Craton

The base of CM1 is defined as the Monteville-Reivilo formation boundary (GB) and within the fluvial facies of the Black Reef Formation (TB). It is composed of the lowermost Reivilo Formation (GB), the upper Black Reef Formation (TB), and the lower Oaktree Formation (TB). The upper boundary of CM1 consists of an unconformity (TB) and correlative conformity represented by deposition of intertidal facies (GB) (Figure 2).

Shales in the lower Reivilo Formation (GB) suggests that CM1 correlates to the upper Black Reef Formation (TB), the source of siliciclastic sediments; paucity of shales above the lower Reivilo Formation suggests correlation to carbonate deposition in the TB. Facies in the lower Reivilo Formation demonstrates progressive transgression with intertidal facies grading upward into subtidal facies. Similarly, the Black Reef Formation (TB) transitions from fluvial to shallow marine siliciclastic facies that grade upward into subtidal to intertidal carbonate facies, representing transgression and cessation of siliciclastic sediment influx. Correlation of shallowest facies in the GB with an erosional unconformity in the TB with minor quartzite in the far east (Button, 1973b) produces a consistent initial sequence for the CM platform.

The correlations from the GB to the TB for CM1 are the only ones consistent with available stratigraphic, compositional, and age constraints. The paucity of siliciclastic sediment makes it unlikely that the middle Reivilo Formation is correlative to the Black Reef Formation. Thus, CM1 is interpreted as a reliable correlation.

CM2-5: Development of Steep Platform Margin

Sequences CM2 to CM5 represent sea level fluctuations during development of a steep, sometimes rimmed margin for the carbonate platform. As a group, they represent an aggradational package with net migration of supratidal facies from northeast to southwest through time. CM2-CM5 sequences compose the uppermost Oaktree and entire Monte Christo formations (TB) and the middle Reivilo Formation (GB). They are defined by intertidal to supratidal (TB) and subtidal to intertidal (GB) facies shifts, separated by erosional unconformities with at least minor fluvial quartzite in the north (TB) and correlative conformities in the south (GB). For the most part, transgressive systems tracts are poorly developed, and flooding surfaces immediately overly sequence boundaries (Figure 2).

The base of CM2 is defined as a relative rise in sea level represented by deepening of facies. The presence of shallow subtidal to intertidal facies basinward of deeper subtidal facies documents development of a steep, partially rimmed margin during CM2 deposition (Figure 2). The margin shows two pulses of shallowing which may reflect a pause in relative sea level rise or variations in sediment accumulation rate at the platform margin. Similar facies changes have not been unambiguously identified elsewhere in the platform, suggesting that shallowing within CM2 reflects the dynamics of sedimentation at the platform margin rather than a significant change in sea level. The sequence boundary defining the top of CM2 resulted in exposure of the platform margin during a relative sea level fall that resulted in deposition of sandstones and mud cracked shales in the far east (Button, 1973b). Deposition of sequence CM3 continues the pattern of shallow sedimentation at the platform margin, with deeper subtidal environments within the platform.

During deposition of CM4, platform margin topographic relief flattened due to progradation of tidal flat to supratidal deposits from northeast to southwest, demonstrating that carbonate production was capable of filling all available accommodation space (Figure 2). The top of CM4 is correlated from the GB to the TB based on identifying the lowest sea level stands and subsequent transgressions between CM3 and CM5. However, the proposed correlation results in a substantial thickness increase in the central part of the platform. This thickness may be due to enhanced subsidence in the Johannesburg area, absence of significant erosion during exposure at the top of CM4, or an improper correlation between the GB and the TB (see below). Renewed shallowing of the platform margin during CM5 resulted in the first significant supratidal deposits in the GB and a substantial influx of grainstones and breccias to slope environments. The top of CM5 is defined as the most extensive erosional unconformity in the TB, which occurs at the Monte Cristo-Lyttelton formation boundary. It is marked by a chert breccia that is up to 2 m thick overlain by the fluvial 0-7 m thick Blyde River Quartzite in the far east and north (Button, 1973b; Sumner, 1995). The correlation of this SB to the GB is based on development of supratidal facies at the platform margin, evidence for significant erosion of the platform margin in slope deposits, and the extent of flooding at the base of CM6, which represents deepest flooding of the craton during continuous carbonate deposition in both the GB and TB. CM5, as correlated, shows the opposite geometry to CM4 and pinches out to the northeast, with the sequence boundaries above CM4 and CM5 coalescing in the eastern TB.

Correlations of CM2-CM5 from the GB to the TB are moderately reliable with the possible exception of the sequence boundary between CM4 and CM5. The significant number of stratigraphic columns with the same number of identified sequence boundaries in each area provides sufficient data to support these correlations. In addition, an impact spherule layer has been identified at the top of CM5 in three cores through slope deposits and one through platform deposits in the GB (Simonson and Sumner, 2004). Identification of this layer in the TB, particularly associated with or below the unconformity between the Monte Cristo and Lyttelton formations, will provide a test of proposed correlations.

CM6: Deepest Transgression within the Platform

Sequence CM6 represents the deepest transgression during growth of the carbonate platform. It consists of the uppermost Reivilo Formation (GB), the very lowermost Fairfield Formation near the platform margin (GB), the lower Lyttelton Formation (western TB), and the entire Lyttelton and lowermost Eccles formations (eastern TB). It overlies the significant erosional unconformity above CM5. CM6 consists predominantly of below to near wave base subtidal facies except in the far northeast, where accommodation space quickly filled after initial transgression, resulting in accumulation of shallow subtidal to intertidal facies (Figure 2). Two pulses of relative sea level rise led to deposition of subwave-base fenestrate microbialite facies that extend from the platform margin (GB) to core SG1 (TB), with one layer extending farther northeast to core ELF (TB). The continuity of these facies across a significant part of the platform suggests that depositional environments were extremely uniform across much of the flooded platform.

The Kamden Member of the Reivilo Formation consists of siderite and hematite facies iron-formation that was deposited near the top of CM6 in the GB. Kamden Member deposition was focused at the margin and reaches its greatest thickness in slope deposits, providing an excellent stratigraphic tie from platform to near-slope basinal sediments (Figure 2). However, the Kamden Member thins towards Prieska and is not present in fully basinal strata.

The top of CM6 is defined as a regional unconformity (TB) and correlative shallow subtidal to intertidal facies (GB). Collapse of the platform margin during this relative sea level fall led to erosion of the Kamden Member at the platform break, and renewed delivery of grainstones to the platform slope (Figure 2). In the Johannesburg area of the TB, the top of CM6 corresponds to an unconformity above a thin interval of intertidal facies within the upper Lyttelton Formation. This SB likely correlates to an unconformity above tidal flat to supratidal facies in the Eccles Formation (northeast TB), significantly above the top of the Lyttelton Formation as defined by lithostratigraphy (Figure 2; Button, 1973b).

The extensive relative sea level fall at the base of CM6 followed by a substantial rise makes the correlation of most of CM6 very reliable. However, correlation of the SB at the top of CM6 is less certain. Relative sea level fall in the GB and Johannesburg area of the TB is indicated by shallowing of facies to intertidal deposition, with an unconformity developed in the Johannesburg area. Correlation to the northeast TB is more tentative, with the top of CM6 correlated to the first identified unconformity above the CM6 flooding. However, a lower unconformity may be present within the tidal flat to supratidal facies below the top of CM6 as correlated.

CM7-11: Development of Lagoons and High Relief Margin

Sequences CM7 to CM11 represent development of a rimmed platform with lagoonal deposition and variable progradation of tidal flat to supratidal carbonates. The platform was primarily aggradational during this interval, and progradation of the reef margin was inhibited by steep inherited relief across the margin. Five sequence boundaries consisting of exposure surfaces and erosional unconformities have been identified in the GB. Fewer have been identified progressively to the northeast, where they likely merge into longer duration exposure surfaces and stratigraphic data are too sparse to resolve them.

CM7 and CM8 are thin, transitional sequences. CM7 consists of the Fairfield Formation (GB), the upper Lyttelton Formation (western TB), and lower Eccles Formation (eastern TB). CM8 consists of the Klipfonteinheuwel (GB) and lowermost Eccles (western TB) formations. Either CM7 or CM8 pinches out to the northeast, where the SB's above CM6 and CM7 or CM7 and CM8 merge (Figure 2). The favored interpretation is that CM7 pinches out, giving the platform a flat top. The development of supratidal to tidal flat facies across most of the platform demonstrates the absence of significant topographic relief. Minor quartzite was deposited immediately above the merged sequence boundaries in the northeast TB (Sumner, 1995).

After exposure on top of CM8, transgression was slow enough that the reef margin kept up with rising relative sea level, producing the lagoonal depositional pattern of CM9, represented by the Papkuil Formation (GB) and parts of the Eccles Formation (TB). Early in this transgression, sufficient currents were present in the northeast to elongate subtidal stromatolites, but accommodation space was rapidly filled as demonstrated by the predominance of supratidal to tidal flat deposition for the rest of the sequence. At least one shallowing and deepening episode occurred within CM9 based on the local development of supratidal facies overlain by intertidal facies within the lagoon.

This change in relative sea level has not been identified regionally, so it is not broken out as a separate sequence; rather, it is interpreted as the result of more localized dynamics in carbonate sedimentation and relative sea level rise. CM9 ends with progradation of supratidal facies (Klippan Formation in the GB) across the entire platform followed by subaerial exposure, with erosion in the northeast.

CM10, composed of part of the Kogelbeen Formation (GB), starts with reflooding of the platform behind a rimmed margin and ends with an exposure surface (Figure 2). Little progradation of facies occurred during deposition of this sequence, and it may be absent from the TB. CM11, which comprises the upper Kogelbeen (GB) and Eccles (TB) formations, contains a significant transgressive systems tract and substantial platform aggradation. It started with just enough sea level rise to reestablish lagoonal carbonate accumulation behind an aggrading reef margin in the GB, and facies aggraded to tidal flat and supratidal sediments (Figure 2). Continued sea level rise led to renewed growth of the rimmed margin and establishment of lagoonal (GB) and shallow subtidal to intertidal (TB) deposition. Supratidal depositional environments reestablished themselves in the TB. CM11 ends with exposure of the platform in the TB associated with deposition of shallow marine and fluvial siltstones and shales in the northeast TB. The correlative conformity in the GB is marked by upper intertidal to supratidal deposition in two closely spaced intervals (Sumner, 1997a) demonstrating higher order relative sea level changes.

Correlations of CM7 to CM11 are excellent within the shallower parts of the GB, but these sequences have not been correlated into basinal facies due to very rapid facies changes across the platform margin and an absence of time lines. Correlations of CM7-CM11 to the TB are tentative. Facies in the TB are dominated by tidal flat to supratidal deposition, and sequence boundaries are expected to merge within the shallowest parts of the platform. In addition, post-Archean erosion and intrusion of the Proterozoic Bushveld complex have reduced the extent of outcrop available to constrain the stratigraphy of the upper platform. The correlations proposed here are based on the extent of transgression and the relative thicknesses of sequences between the well constrained sequence boundaries below CM7 and above CM11. Correlations were chosen to provide a consistent stratigraphic model based on subsidence behavior lower in the platform.

CM12: Drowning of the Carbonate Platform

CM12 was deposited during a major transgression that led to the drowning of the Campbellrand-Malmani carbonate platform and the initiation of widespread deepwater iron-formation deposition (Klein and Beukes, 1989; Beukes and Klein, 1990b; Catuneanu and Eriksson, 1999). The lower part of this sequence consists of the Gamohaan (GB) and Frisco (TB) formations, whereas the upper part consists of the Kuruman (GB) and Penge (TB) iron-formations. The sequence starts with intertidal deposition and contains facies that progressively deepen upward into deep water iron-formation deposition. The top of the sequence corresponds to shallowing at the top of the Kuruman Iron Formation (GB) and within the Penge Formation (TB), but stratigraphic details are not documented here.

The extensive transgression and change in lithology across the platform, as well as more extensive exposure, make the proposed correlation of CM12 very reliable (Sumner, 1997a).

Discussion

Sequence Development and Platform Geometry

The average duration of a sequence for the platform can be calculated using age constraints from the top of CM1 (2588 ± 6 Ma) and the middle of CM12 (2521 ± 3 Ma). Eleven sequences were deposited in ~65 Ma, giving an average of ~6 Ma per sequence. This duration is consistent with 3rd order cycles (*e.g.*, Sarg, 1988) within a 2nd order sequence (Catuneanu and Eriksson, 1999), and sequence geometries show the same patterns of progradation and aggradation as those in younger platforms (*e.g.*, Sarg, 1988). Thus, the large scale development and stacking of these Neoarchean sequences are consistent with younger patterns of sea level change and subsidence. These similarities demonstrate that Neoarchean carbonate accumulation responded to subsidence, sea level change, and carbonate production in much the same way that carbonate platforms have throughout Proterozoic and Phanerozoic time; the Campbellrand-Malmani carbonate platform included a shallow water carbonate factory, rimmed margin development, and lower rates of carbonate sedimentation in deeper water. The similarity of carbonate platform geometry through time, even with significant changes in dominant biota, demonstrates that rimmed margins are localized primarily by physiochemical conditions rather than growth dynamics of specific organisms.

Implications for Neoarchean Kaapvaal Craton Tectonics

Subsidence associated with accumulation of ~2.6 to ~2.4 Ga sediments provides important constraints on the behavior of the Kaapvaal Craton after extrusion of the Ventersdorp Supergroup lavas. Thermal subsidence dominated after an ultra-high temperature event in the lower crust associated with the ~2.72 to ~2.7 Ga Ventersdorp Supergroup lava extrusion (Altermann and Nelson, 1998; Catuneanu and Eriksson, 1999). Temperatures peaked at >1000°C at ~2.72 to ~2.715 Ga (Schmidt and Bowring, 2003b). This event was associated with crustal thinning, as demonstrated by the presence of syn-eruption grabens. Thickening of the Ventersdorp Supergroup towards the western edge of the craton is consistent with a rifted margin in the west (Tinker *et al.*, 2002). However, typical rift-to-drift sediments are not preserved on the Kaapvaal Craton; the Schmidtsdrif Subgroup contains no stratigraphic evidence for continued rifting and records regional subsidence for 100 m.y. after extrusion of Ventersdorp lavas. Although the subgroup deepens and thickens to the southwest and west (Beukes, 1977; Tinker *et al.*, 2002), consistent with a southwestern margin, current stratigraphic constraints do not require a full continental rifting cycle. In fact, a sediment source from the southwest was also present (Beukes, 1977) suggesting incomplete rifting prior to deposition of the lower Schmidtsdrif Subgroup. Within the craton, variable subsidence and continued lower crustal heating resulted in accumulation of Wolkberg-equivalent sedimentary rocks in deep, localized basins. Lower crustal xenoliths show variable thermal histories from 2.7 to 2.6 Ga, suggesting that extension was laterally variable (Schmitz and Bowring, 2003b). Continued crustal thinning during Wolkberg deposition is likely (Button, 1973a; Barton *et al.*, 1995), and volcanics in both Wolkberg-equivalent units and the Schmidtsdrif Subgroup demonstrate the continuing presence of a heat source within the Kaapvaal Craton. Thus, stratigraphic patterns are consistent with heterogeneous crustal thinning, possibly including an unpreserved rift-to-drift transition to the west, but they do not require a rifted margin (Catuneanu and Eriksson, 1999).

The Black Reef Formation overlies these and older rocks with an angular unconformity (Clendenin *et al.*, 1991; Tinker *et al.*, 2002). The relatively uniform thickness of the Black Reef Formation demonstrates regional peneplanation and the establishment of a tectonically quiescent phase for the craton. The subsequent Campbellrand-Malmani carbonate platform was deposited on a uniformly subsiding craton 100-200 m.y. after Ventersdorp extension. Accommodation space may have been provided by regional subsidence due to slow cooling of the mantle lithosphere, enhanced by sedimentary loading. This interpretation is consistent with the fairly uniform depositional thickness of the Campbellrand-Malmani carbonate platform with the exception of the Prieska area (Figure 2; Button, 1973b; Beukes, 1980;

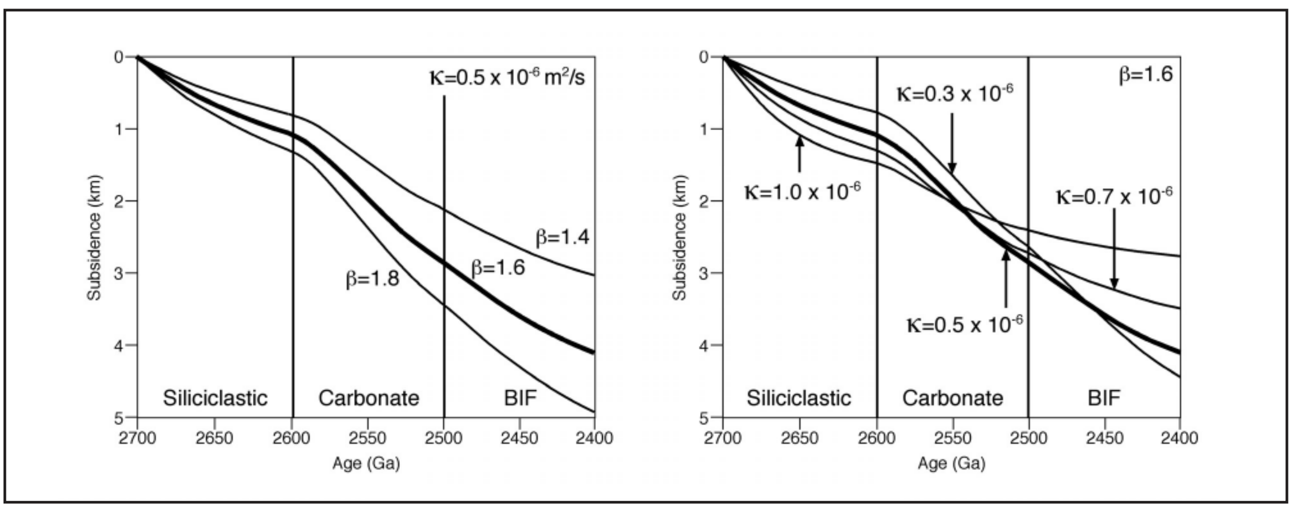

Figure 3. One dimensional thermal subsidence calculations due to thinning during extrusion of the Ventersdorp Supergroup. κ represents thermal diffusivity of the crust and β represents the uniform stretching factor. Green line segments represent loading by siliciclastic deposition, blue line segments represent loading by carbonate deposition, and orange line segments represent loading by iron-formation deposition. Thicker lines show subsidence through time for the preferred model with κ=0.5 x 10^{-6} m^2/s and β=1.6.

Table 3. Subsidence modeling results

κ^+	β^+	Subsidence (km)		
(km^2/s)		t=100 my	t=200* my	t=300* my
0.5	1.4	0.8	2.1 (1.3)	3.0 (0.9)
0.5	1.6	1.1	2.8 (1.7)	4.1 (1.3)
0.5	1.8	1.3	3.4 (2.1)	4.9 (1.5)
0.3	1.6	0.8	2.6 (1.8)	4.4 (1.8)
0.5	1.6	1.1	2.8 (1.7)	4.1 (1.3)
0.7	1.6	1.3	2.7 (1.4)	3.5 (0.8)
1.0	1.6	1.5	2.4 (0.9)	2.8 (0.4)

$^+\kappa$ is thermal diffusivity and b is the uniform crustal stretching factor.
*Total subsidence followed by subsidence since the previous time interval in parentheses.

1987; Tinker *et al.*, 2002), which had lower sedimentation rates due to its deeper depositional environment (Beukes, 1987; Sumner and Grotzinger, 2004).

Various crustal and mantle lithospheric thinning models can produce appropriate subsidence for accumulation of the Schmidtsdrif, Campbellrand-Malmani, and Kuruman-Penge sediments after the Ventersdorp thermal event. However, for thermal subsidence to extend from ~2.7 Ga to ~2.4 Ga, cooling rates must have been slow. Slow cooling rates may indicate a low thermal diffusivity. For example, a thermal diffusivity of 0.5 x 10^{-6} m^2/s rather than 1.0 x10^{-6} m^2/s (Allen and Allen, 1990, p. 58) produces subsidence appropriate for the thicknesses of observed sequences using a 1-D thermal subsidence model (Allen and Allen, 1990, p. 58) with loading of sediment with densities of 1,030 kg/m^3 for the first 100 m.y. of siliciclastic deposition, 2,800 kg/m^3 for 100 to 200 Ma during carbonate deposition, and 3,000 kg/m^3 for 200 to 300 Ma during BIF deposition (Figure 3; Table 3). A lithospheric mantle thinning factor of about 1.6 produces appropriate average subsidence (Figure 3a; Table 3) and is consistent with the temperatures proposed for 37 km depth during Ventersdorp extrusion (Schmitz and Bowring, 2003a; b). This model over estimates the depositional thickness of the iron-formation; the sea level fall between the Kuruman and Griquatown iron-formations (Beukes and Klein, 1990a) is consistent with a change in tectonic environment, which may be required to reduce subsidence after ~2.4 Ga.

Conclusions

The Neoarchean lower Transvaal Supergroup, South Africa, consists of 14 sequences that record development of a significant carbonate platform on the Kaapvaal Craton. This platform developed from a mixed siliciclastic-carbonate ramp to a steepened margin followed by a rimmed margin that separated lagoonal environments from the open ocean. Drowning of the platform resulted in deposition of banded iron-formation across the Kaapvaal Craton. The geometry and stacking of these sequences are consistent with younger patterns of carbonate accumulation, demonstrating that Neoarchean carbonate accumulation responded to subsidence, sea level change, and carbonate production similarly to Proterozoic and Phanerozoic platforms. The similarity of carbonate platform geometry through time, even with significant changes in dominant biota, demonstrates that rimmed margins are localized primarily by physiochemical conditions rather than growth dynamics of specific organisms.

Stratigraphic patterns during deposition of the Schmidtsdrift and Campbellrand-Malmani subgroups are most consistent with variable thinning of the Kaapvaal Craton during extrusion of the ~2.7 Ga Ventersdorp lavas. Although depositional patterns are consistent with rifting of the western margin of the Kaapvaal Craton during this time, a rift-to-drift transition is not required to explain subsidence. Heating and thinning during Ventersdorp time can produce the observed thermal subsidence from ~2.7 to ~2.45 Ga if the thermal diffusivity of the craton was moderately low.

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