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TECTONIC STYLE AND MECHANISM OF EARLY PROTEROZOIC SUCCESSOR BASIN DEVELOPMENT, SOUTHERN AFRICA

by

C.W. CLENDENIN, E.G. CHARLESWORTH and S. MASKE

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*(Department of Geology, University of the Witwatersrand
Wits 2050, Republic of South Africa)*

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September, 1987

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ABSTRACT

A series of early Proterozoic successor basins developed on the Kaapvaal Craton from 3,0 to pre-2,2 Ga and this succession includes the Dominion-Witwatersrand, Ventersdorp, and Chuniespoort/Ghaap depositional systems. These successor basins developed along lines of pre-existing tectonic weakness and resulted from a simple tectonic style of deformation in which neither extension nor compression dominated. As changes occurred within the simple tectonic style, these changes were imposed on the existing thermomechanical properties of the Kaapvaal Craton. Even though the basins are interrelated, they are dissimilar because stress changes resulted in combinations of renewable and non-renewable lithospheric stress that further modified individual basin evolution. Intraplate deformation due to these modifications has been used to identify plate boundary forces which developed the simple tectonic style. Changes in both intraplate deformation and transmitted intraplate stress were related to the angle of subduction between the Kaapvaal and Rhodesian cratons. The evolution of subduction geometries due to the kinematics of plate interactions resulted in successor basin development and genetically links them together via a causal mechanism.

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1. INTRODUCTION

In a recent review, Clendenin *et al.*, (in press) proposed an early Proterozoic, three-stage rift system in which a number of superimposed basins developed on the Kaapvaal Craton from Dominion Group (3,0 Ga) to Chuniespoort/Ghaap Group (pre-2,2 Ga) time. This system includes the Dominion-Witwatersrand, Ventersdorp, and Chuniespoort/Ghaap depositional basins which were related to the pre-graben, graben, and post-graben stages of a fully evolved rift system (Clendenin *et al.*, in press) (Fig. 1). A unique sequence of crustal responses was defined that led to these interrelated Proterozoic basins, and the defined responses were compared to a number of Phanerozoic examples. However, the more fundamental questions of causation and mechanism have remained relatively unaddressed.

Even though the basins are superimposed and interrelated, they are dissimilar because each is characterized by particular lithologies; each is subjected to various periods of evolution; and each is dominated by specific styles of deformation. Due to these dissimilarities, the three-stage rift system is made up of successor basins as defined by Klemme (1975) who recognized that changing tectonic styles develop a succession of basin types along zones of pre-existing tectonic weakness. Tectonic style is the result of a combination of two major types of superimposed lithospheric stress which have been referred to as renewable and non-renewable (Bott, 1982). Renewable-type stresses are those generated by plate boundary forces or loading and are fed from the re-application boundary or body forces, even though the strain energy is continually relieved by tectonic activity (Bott and Kusznir, 1984). Renewable-type stresses due to plate boundary forces can be either compressional or extensional in nature, and changes in plate boundary interactions generating the stress fields probably have the most effect on intraplate deformation. In contrast, non-renewable-type stresses are those either due to lithospheric flexure or resulting from thermal stresses and are stress systems that are dissipated by release of the strain energy present (Bott and Kusznir, 1984).

Non-renewable stresses also differ from renewable types in that non-renewable-types are strongly time dependent and any combination of renewable and non-renewable stress systems can be generally distinguished by tectonic analysis because of the time-dependent nature of non-renewable systems. Non-renewable stress may be of some importance in creating fundamental zones of weakness within the lithosphere that may be subsequently exploited by renewable stresses (Bott and Kusznir, 1984). In addition, combinations of renewable and non-renewable stress may amplify the effectiveness of non-renewable stress in creating weakness in those zones of weakness. However, the effect that combinations of non-renewable and renewable stresses have on one another depends on which type of stress system is producing the tectonic style at any given time. Renewable stress systems are particularly susceptible to stress amplification because stress decay in the lower lithosphere results in amplification of upper lithosphere stress (Kusznir, 1982; Mithen, 1982) and subsequent modification of structural style.

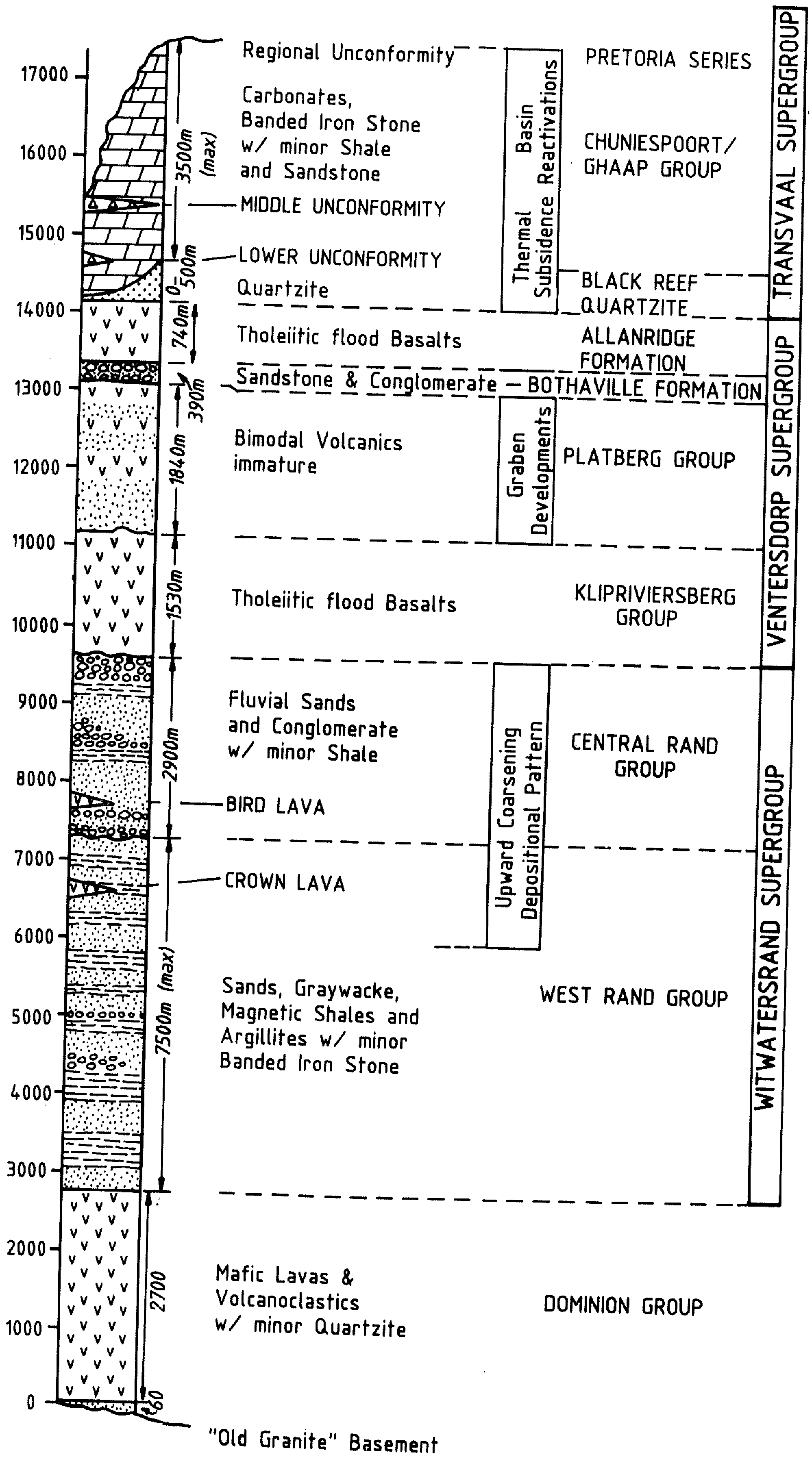

Figure 1: General stratigraphic section of successor basin sequence developed on the Kaapvaal Craton 3,0 to pre-2,2 Ga.

If the lithosphere contains a pre-existing tectonic weakness (i.e. local thermal anomaly, sedimentary basin, or large wrench fault zone), the perturbation will localize the combination of stress systems. The perturbation will be amplified and amplification will perpetuate the perturbation, as well as allow its growth with time (Lambeck, 1983). It should be emphasized that changing tectonic styles during the early Proterozoic developed the successor basins on the Kaapvaal Craton, while zones of pre-existing weakness influenced their location. It is the intent of this paper to propose a causal mechanism that genetically and chronologically links these events.

11. GEOLOGIC CHARACTERISTICS OF THE SUCCESSOR BASINS

In recent years, a number of papers have reviewed and discussed the geologic relationships between the early Proterozoic Kaapvaal Craton successor basins, with a particular interest given to the Witwatersrand and Ventersdorp Supergroups (Buck, 1980; Button *et al.*, 1981; Tankard *et al.*, 1982; Matthews, 1982; Burke *et al.*, 1985, 1986; Winter, 1986a; Stanistreet *et al.*, 1986; McCarthy *et al.*, 1987; Clendenin *et al.*, in press). Most of these studies were either specific about some stage of successor basin development or gave detailed descriptions of the sedimentology and stratigraphy; but it has only been the more recent authors that have tried to define and model the tectonic relationships (i.e. Burke *et al.*, 1985, 1986; Winter, 1986a,b; Clendenin *et al.*, in press). In earlier models, Pretorius (1966) proposed that the successor basins were developed under extension where vertical tectonics prevailed. Hunter (1974) expanded Pretorius' (1966) model and proposed a model for the successor basins that can be compared with the widely accepted McKenzie (1978) two-stage basin model which involves mechanical, followed by thermal subsidence. Probably the most important conclusion Hunter (1974) made was that the Limpopo Belt is fundamental in the evolution of the Kaapvaal Craton and is an interpretation that has only recently been re-proposed (Burke *et al.*, 1985, 1986; Winter, 1976; Clendenin *et al.*, in press). (Fig. 2).

A complete review of the geologic relationships of the successor basins is beyond the scope of this paper, and the reader is referred to the previously mentioned papers for specific detail. However, a brief overview of the stratigraphy together with several of the newer structural interpretations will be presented. The reason for this overview is that the best record of vertical movements in 'continental' interiors is stored in the stratigraphic record (Beaumont, 1978) and that changing intraplate stress fields influencing the vertical movements can be identified without confusion of specific stratigraphic detail. In this study, the writers have assumed that the initial topography of the basement on which the successor basins developed was an imperfectly peneplaned surface and any vertical movements of this surface would be recorded in the stratigraphic record. This assumption is based on Vlaar's (1986a) suggestions that Archaean continental crust had a moderate surface relief and stood only slightly above sea level.

The Dominion Group is the oldest in the successor basin sequence which was deposited nonconformably on the 'Old Granite' basement (S.A.C.S., 1980) and is made up of mixed volcano-sedimentary lithologies (Fig. 1). Volcanics and volcanoclastics make up the bulk of the lithologies within the Group and overlie a thin basal quartzite (Button *et al.*, 1981). The Dominion Group is overlain by clastics of the Witwatersrand Supergroup.

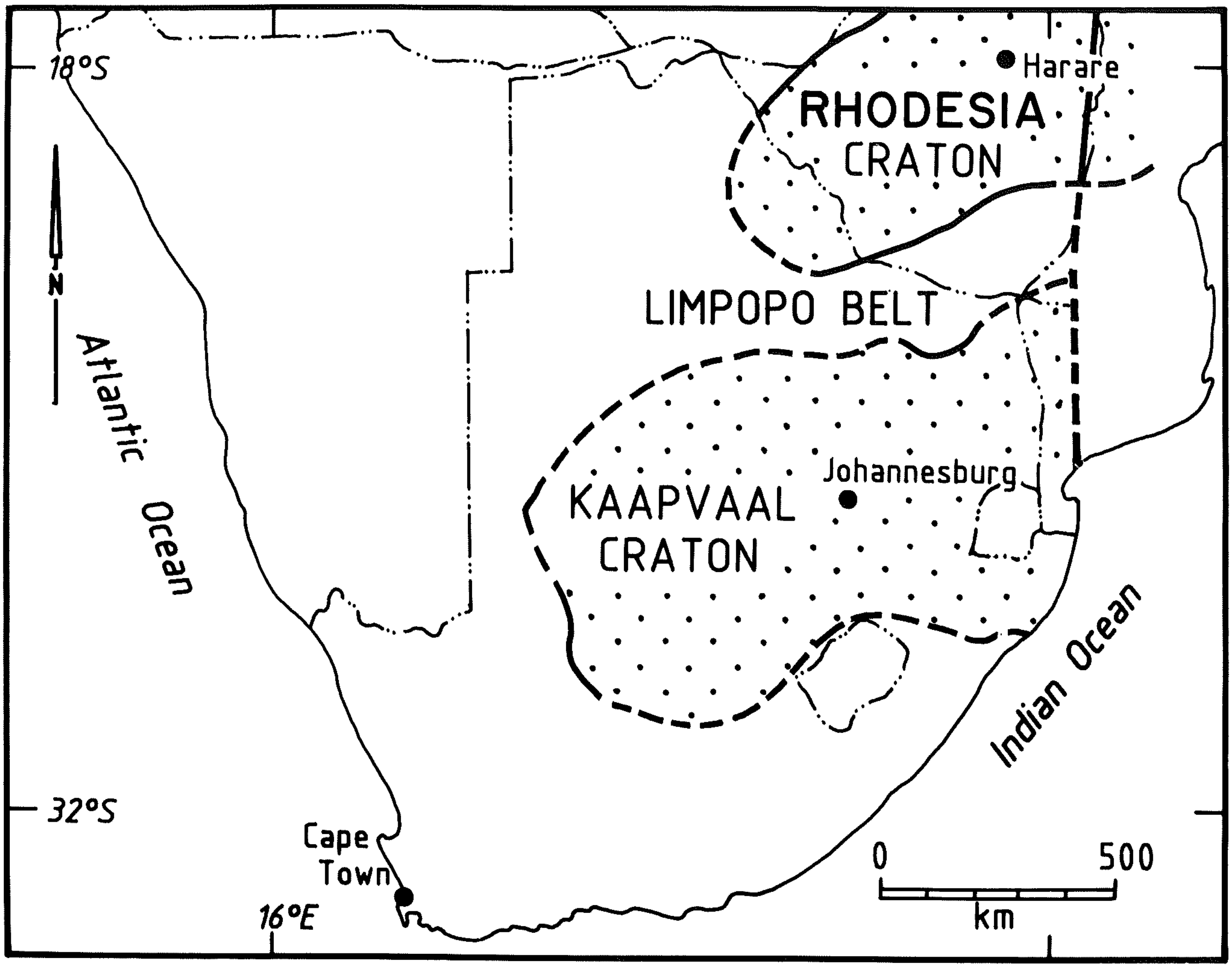

Figure 2. Regional relationships of the Kaapvaal and Rhodesian cratons of southern Africa.

The Witwatersrand Supergroup is divided into the lower West Rand Group and upper Central Rand Group (S.A.C.S., 1980) (Fig.1). The sediments of the West Rand Group (re-worked orthoquartzite, ferruginous shale, and minor iron-formation) have been interpreted as representing a more distal, lower energy, depositional environment than those of the overlying Central Rand Group (coarse siliciclastics with minor shale) that indicate a more proximal environment (Tankard *et al.*, 1982) (Fig. 1). Overall progressive regression resulted in a general coarsening upward of lithologies through the two Groups as the basin became restricted to the outer seas during middle and upper Central Rand time. During this progressive restriction, the basin appears to have been increasingly modified with fluvial and lacustrine sedimentation becoming more prominent in late West Rand into Central Rand time (Button *et al.*, 1981). The Central Rand Group contains numerous upward-fining packages separated by shallow disconformities with the most proximal, higher energy clastics deposited in upper Central Rand time (Button *et al.*, 1981; Tankard *et al.*, 1982) (Fig.1). Periods of folding, and both compressive block and strike-slip faulting, have been interpreted as having occurred episodically through Central Rand time with a period of extensional block faulting initiated during late Central Rand time (Button *et al.*, 1981; Tankard *et al.*, 1982; Charlesworth *et al.*, 1986; Stanistreet *et al.*, 1986; McCarthy *et al.*, 1987). Of particular note are two mafic lava flows that lie within the Witwatersrand Supergroup; the Crown lava of the upper West Rand Group and the Bird tuffs and amygdaloid of the middle Central Rand Group (Fig. 1). The Crown lava is the thicker of the two and locally attains thicknesses of 250m along the northern margin (Burke *et al.*, 1986). However, the upper lava unit of the Bird sequence may be 170m thick and both mafic lava flows are absent in the south-southeastern portion of the preserved basin (Tankard *et al.*, 1982).

The volcanic and sedimentary rocks of the Ventersdorp Supergroup lie both conformably and unconformably on Witwatersrand strata depending on what stage of successor basin development is being considered (Fig. 1). Winter (1976) subdivided the Ventersdorp Supergroup into four subdivisions: the Klipriviersberg Group, the Platberg Group, the Bothaville Formation, and the Allanridge Formation (Fig. 1). The Klipriviersberg Group consists of a repetitive sequence of alkali-rich, continental tholeiitic flood basalts (Tankard *et al.*, 1982); but locally, at the base, high-Mg, mafic komatiitic basalt flows are present (McIver *et al.*, 1981). Klipriviersberg lava flows of variable thickness filled irregularities on the underlying Witwatersrand Supergroup palaeosurface (Tankard *et al.*, 1982) and these lower flows do not have the areal extent of some of the upper ones (Palmer, pers. comm., 1987). The Platberg Group is made up of an interfingering sequence of bimodal volcanics, immature clastics, and playa lake sediments (Buck, 1980) and overlies the Klipriviersberg Group with a pronounced unconformity. Faulting was penecontemporaneous with extrusion of Klipriviersberg Group lavas (Tyler, 1979a) and field relationships indicate that faulting increased with time leading to periods of major extensional block faulting during Platberg Group time (Fig. 1). Penecontemporaneous erosion of faulted blocks during graben development supplied coarse clastic material to the Platberg group (Buck, 1980). The Platberg Group graben appears to be a late-stage development of extensional deformation that was initiated during late Central Rand time (Clendenin *et al.*, in press) and graben development was virtually complete by the end of Platberg time (Buck, 1980). The Bothaville Formation is unconformable with the underlying Platberg Group and consists of clastic sedimentation that is cyclic from conglomerate to impure sand and shale to conglomerate (S.A.C.S., 1980). The Bothaville Formation clastics were buried by the flood basalts of the Allanridge Formation which records a final episode of Ventersdorp volcanism. The Allanridge Formation usually has a structurally conformable contact with the underlying Bothaville Formation; but in places, it is known to unconformably overlie the Bothaville Formation and an older diverse stratigraphy (S.A.C.S.,1980).

The Black Reef Quartzite Formation is a time transgressive unit that unconformably overlies the volcanic and sedimentary units of older successor basins (Fig. 1). In addition, it oversteps these units along structural basin margins and rests nonconformably on older granites and greenstones (Tankard *et al.*, 1982). The Black Reef Quartzite Formation has both an early fluvial and a later marine component; and, being time transgressive, the formation has a gradational contact with the overlying carbonates of the Chuniespoort/Ghaap Group, Transvaal Sequence (Fig. 1). This gradational relationship between the two is interpreted as representing an episodic southwest to north-northeast transgression that resulted in marine sedimentation replacing fluvial conditions (Clendenin, in prep.) Within the Chuniespoort/Ghaap Group, shallow platform carbonates interfinger with ferruginous carbonates, carbonaceous shale, and iron-formation in a south-southwesterly direction (Beukes, 1977): while to the north-northeast, platform carbonates interfinger with quartzarenite and mixed siliciclastic-carbonate sediments. Episodic transgressions to the north-northeast progressively enlarged the basin, but there were periods of uplift and erosion. One such hiatus in the lower Chuniespoort/Ghaap Group seems to define an arch-type structure in the approximate area of the Central Rand Group compressive block faults and appears to have divided the Chuniespoort/Ghaap depositional basin into two discreet segments (Fig. 1). Pronounced basin enlargement followed the downwarping of this arch-type structure (Clendenin, in prep.) A second hiatus in middle Chuniespoort/Ghaap time allowed a thin sheet of quartzarenite to cover the entire south-southwest portion of the carbonate basin while sand-filled channels eroded the northern portion prior to a rapid transgression that both re-established carbonate sedimentation and enlarged the basin (Clendenin, in prep.) (Fig. 1). Later periods of uplift and erosion appear to be short-lived and are marked by marginal disconformities defined by chert-in-shale breccias and thin quartzite lenses (Clendenin and Maske, 1986). Marginal disconformities pass laterally into conformable relationships in the west-southwest Transvaal near the centre of the basin

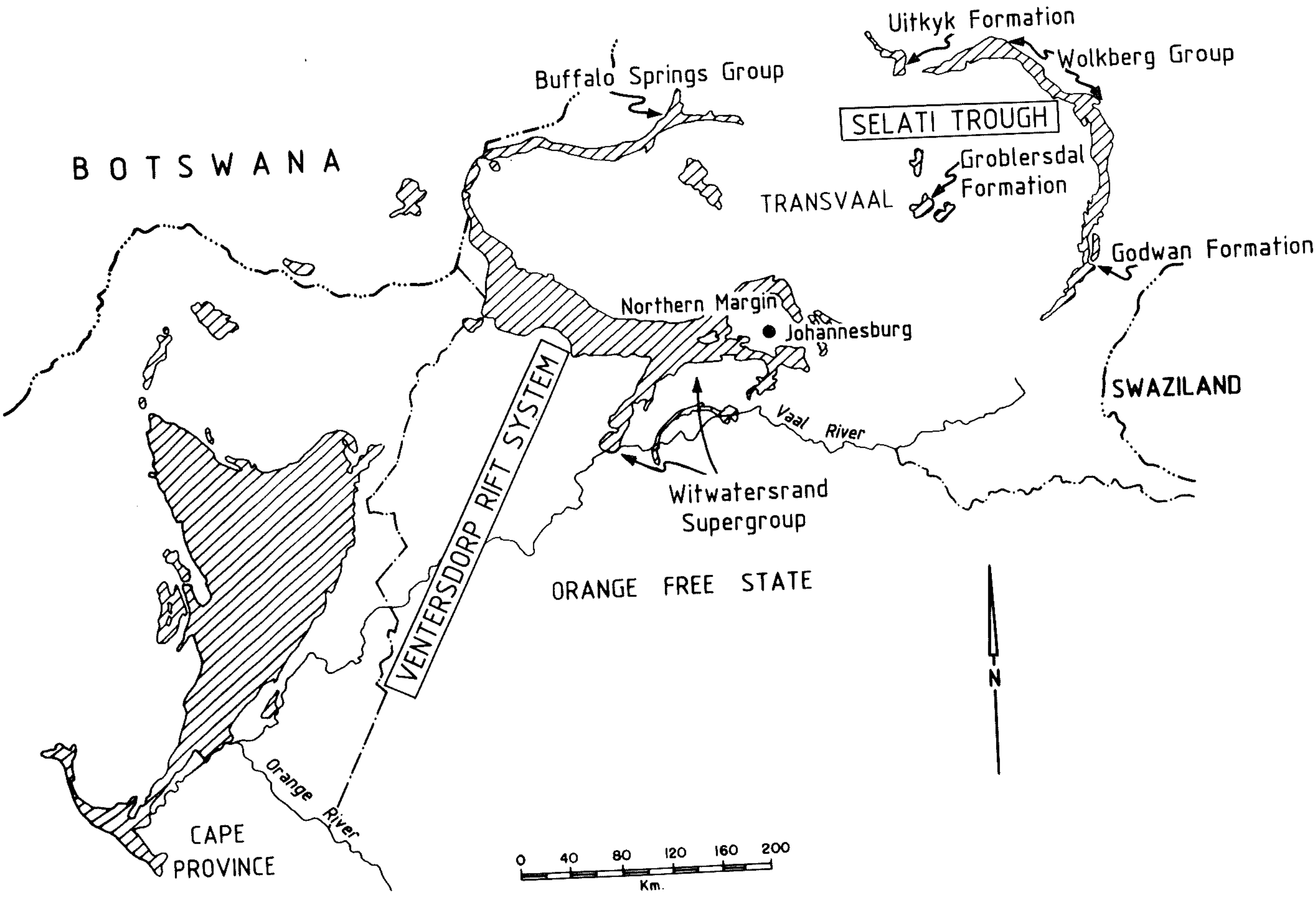

Figure 3. General location map showing major structural elements of the successor basin sequence in relationship to preserved Chuniespoort/Ghaap Group (striped areas) and various other units that are correlatable to successor basin sequence.

that overlies the Ventersdorp Supergroup (Clendenin and Maske, 1986; Clendenin *et al.*, in press) (Fig. 3). Transgressions following each hiatus were rapid and deposited subtidal carbonate lithofacies over marginal disconformities (Clendenin and Maske, 1986). These rapid transgressions progressively advanced distal carbonaceous shale and iron-formation to the north-northeast over both subtidal and shallow basin carbonate lithofacies as the basin continued to expand with time. The distal lithofacies indicate that associated subsidence could have been as much as 100m with these transgressions, but it is believed that subsidence was generally in the 10-15m range when the entire basin is considered. Episodic basin expansions associated with these rapid transgressions allowed the Chuniespoort/Ghaap sea to cover most of the Kaapvaal Craton during late Chuniespoort/Ghaap time. Following this transgressive event, uplift produced a rapid regression and led to a regional unconformity that abruptly terminated the continuance of the successor basin sequence prior to the deposition of the Pretoria Group, Transvaal Sequence (Fig. 1).

Other stratigraphic units that lie nonconformably on the Archaean 'Old Granite' basement and that either have an unconformable or questionable conformable relationship with the Black Reef Quartzite Formation are: the Uitkyk Formation, the Godwan Formation, the Groblersdal Group, the Buffalo Springs Group, and the Wolkberg Group (Fig. 4). These units have been correlated with either the Dominion-Witwatersrand or Ventersdorp Supergroup and their geologic relationships are briefly described. Van Rooyen (1947) suggested that the Uitkyk Formation was correlatable with the Witwatersrand Supergroup and Button (1973a) tended to support this interpretation. Button (1977) also suggested that the Godwan Formation was correlatable with the Witwatersrand Supergroup, and the Godwan Formation is unconformable with the younger Black Reef Quartzite Formation. Little is known about the Groblersdal Group, other than it overlies the 'Old Granite' and has an apparent conformable (?) contact with the Black Reef Quartzite Formation (S.A.C.S., 1980). The present correlation between the Wolkberg Group and the Buffalo Springs Group is questionable, as well as the correlation of both Groups to stages of the successor basin sequence. Tyler (1979b) proposed that the Buffalo Springs Group and the Black Reef Quartzite Formation were conformable and he correlated the Buffalo Springs Group with the Wolkberg Group based on this relationship. Field observations by the writers indicate that an unconformable relationship exists between the Buffalo Springs Group and the Black Reef Quartzite Formation. Based on these new observations, any correlation between the Buffalo Springs Group and the Wolkberg Group that is based on an apparent conformable relationship proposed by Button (1973a) between the Wolkberg Group and the Black Reef Quartzite Formation is not justifiable. Button (1973a) also defined an erosional contact between the Wolkberg Group and Black Reef Quartzite Formation and then ignored it as a regional feature because the Wolkberg Group is found only in the rift-like Salati Trough. The presence of a defined discontinuity makes the relationship between the Wolkberg Group and Black Reef Quartzite Formation also suspect. Based on Tyler's (1979b) stratigraphic descriptions, the writers suggest that the lower Buffalo Springs Group, with its overall upward coarsening siliciclastics and low-angle discordances, is correlatable with the Witwatersrand Supergroup while the upper Buffalo Springs Group bimodal volcanics (basaltic lavas followed by dominantly acid volcanics) is correlatable with the Ventersdorp Supergroup (Fig. 4). The correlations of the Wolkberg Group with the successor basin sequence will be discussed later in the paper. Regardless of their exact correlation within the successor basin sequence, the five preserved fragments define the

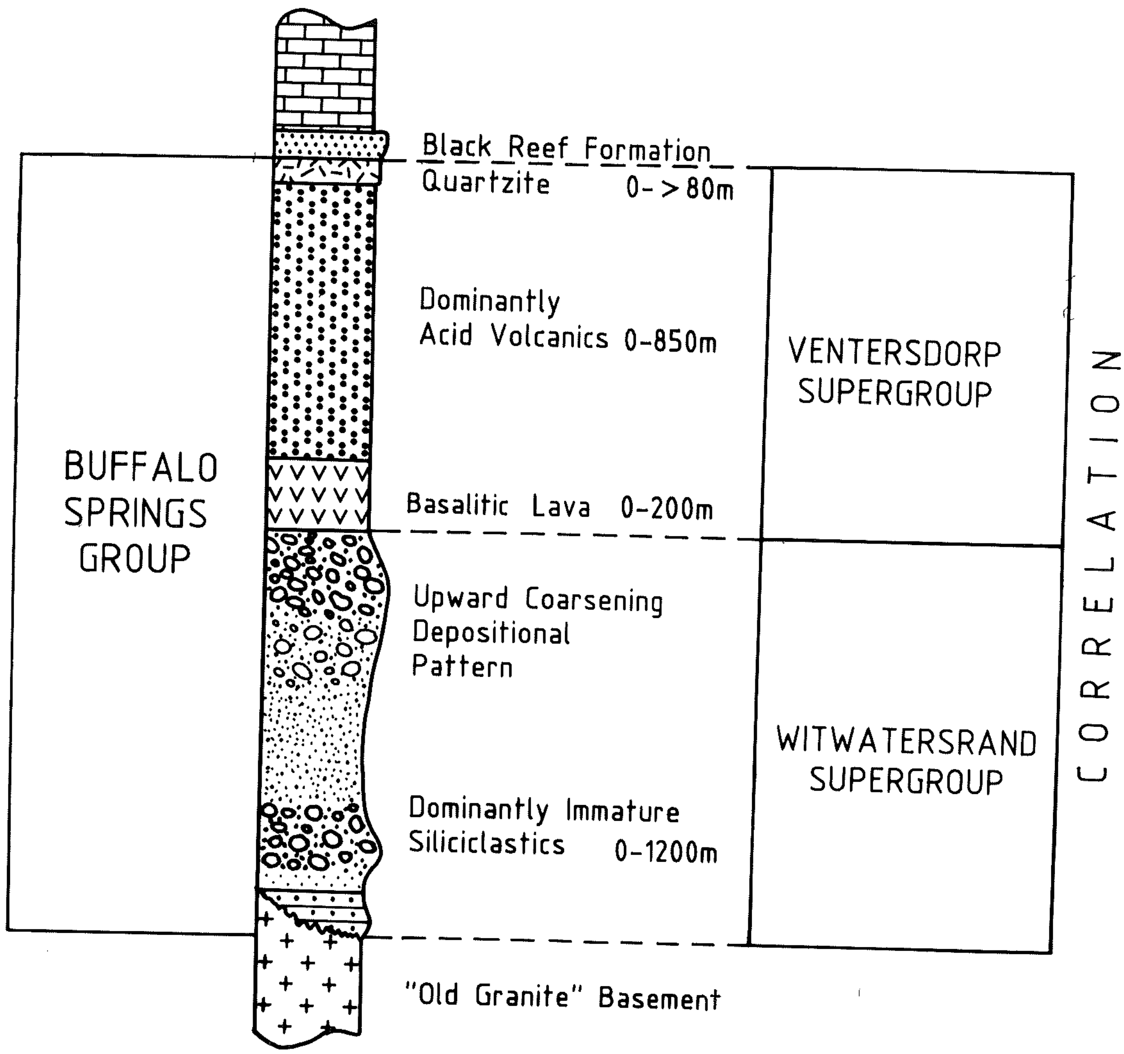

Figure 4. General stratigraphic section of the Buffalo Springs Group and its correlation with the successor basin sequence,

implication suggested by Mathews (1982) that the successor basins extended well beyond their present structural limits.

111. IDENTIFICATION OF TECTONIC STYLE

Tectonic style results from changing intraplate stress systems that are imposed on the thermomechanical properties of the crust (Cross and Pilger, 1982) and the identification of changes in the stress systems is somewhat dependent on the identification of tectonic style. Molnar and Atwater (1978) have proposed that tectonic styles can be divided into three broad categories: extensional, compressional, and simple. The category of simple tectonic style is not self-explanatory and is a tectonic style in which neither extension nor compression dominates. Simple tectonic style should be expected to characterize successor basin evolution because differing basin types developing along zones of pre-existing tectonic weakness would be influenced by changing stress systems. However, a particular successor basin may be dominated by specific intraplate stresses, as well as reflect changes within the stress systems through time.

Having briefly reviewed the stratigraphy of the successor basins, a number of specific changes within the sequence can be directly related to local and regional intraplate stress systems while others have to be inferred. Volcanism is generally associated with thermal uplift that is followed by time-dependent thermal subsidence. Thermally induced vertical movements would represent changes in the thermomechanical properties of the crust and define periods of non-renewable thermal stress. From the stratigraphic record, periods of non-renewable thermal stress can be related to the major periods of volcanism during Dominion and Ventersdorp time and to brief periods of thermal subsidence during Witwatersrand and Chuniespoort/Ghaap times.

Matthews (1982) has pointed out that periods of alternating volcanism and non-volcanism can be related to the evolution of an overall expanding basin and has proposed that periods of volcanism can be related to extensional stress systems while periods of non-volcanism are due to compressional ones. Dewey (1980) has also pointed out that under a compressional stress system volcanism will be inhibited because feeders will close. Even though it might be a rather simplistic approach to renewable stress systems, alternating periods of volcanism within the stratigraphic record would define a chronology of changing intraplate stresses due to distant plate interactions, as well as the tectonic style. As previously mentioned, major periods of volcanism occurred during Dominion and Ventersdorp times and these periods alternated with extended periods of non-volcanism during Witwatersrand and Chuniespoort/Ghaap times (Fig.1). Due to these alternating periods, the writers suggest that a simple tectonic style developed the early Proterozoic Kaapvaal successor basins. However, the reader is reminded that this definition of tectonic style, based on volcanism, is an overview of the changing stress systems and that the initiation of changes within the stress systems cannot be simply correlated with the first appearance of volcanism.

1V. IDENTIFICATION OF GEODYNAMIC PROCESSES

As previously mentioned, Hunter (1974) has pointed out that the Limpopo Belt is fundamental in the evolution of the Kaapvaal Craton and this linear belt can be readily attributed to convergence and collision between the Kaapvaal and Rhodesian cratons (Light, 1982; Burke *et al.*, 1985, 1986; Winter, 1986a, b; Clendenin *et al.*, in press) (Fig. 2). The driving and resisting forces due to a converging margin would cause the cratons to be stressed and transmission of these plate boundary forces into intraplate regions would generate the different renewable stress systems (Bott and Kusznir, 1984). A number of authors have interpreted the Kaapvaal Craton as the overriding plate (Light, 1982; Burke *et al.*, 1985, 1986; Winter, 1986a, b)even if it has only been implied in the graphics associated with these studies. Accepting this interpretation, renewable inplane stress would be transmitted into the Kaapvaal Craton due to convergence and collision. This transmitted stress would affect both tectonic style and resulting intraplate deformation. Changes in renewable stress systems within the Kaapvaal Craton can thus be related to the history of the Limpopo Belt and the effect the Limpopo Belt has had on the Kaapvaal Craton can be interpreted.

Light (1982) has proposed the following chronology of events for the Limpopo Belt within the 3,0 Ga to pre-2,2 Ga time frame of this study. The zone was static around 2,9 Ga, collision occurred about 2,6 Ga, and culmination of convergence resulted between 2,5 Ga and 2,3 Ga when the Kaapvaal and Rhodesian cratons formed a stable unit. On a broad scale, each of these events can be related to the defined, changing renewable stress systems within the Kaapvaal Craton. With the cessation of compressional forces about 2,9 Ga, the static nature of the zone can be interpreted as a period of extension within the Kaapvaal Craton because external extension, or equivalently relaxation of compression, would lead to such stress systems. Periods of extension at this time would be supported by Light's (1982) interpretation that a series of igneous centres intruded along a major zone of release (extensional) fractures in the old cratonic crust at about 2,9 Ga. This zone of fractures can be related to a resurgent Precambrian taphrogenic lineament extending from the Afar triangle to the Orange River, South Africa (McConnell, 1980) and possibly represents the oldest of the pre-existing zones of weakness that influenced the successor basin sequence. This period of extension at about 3,0 Ga localized the Dominion Group thermal event and weakening of the crustal lithosphere due to changes in the thermal structure perpetuated and expanded the pre-existing crustal flaw.

Light (1982) defined a recommencement of convergence between the Kaapvaal and Rhodesian cratons following the 2,9 Ga period of static or extensional stress. The understanding and interpretation of this period of convergence is critical in the identification of the geodynamic processes which were operating and subsequent tectonic styles defined in this review. Even though Light (1982) and others have readily used terms like convergence and collision, invariably they have implied or suggested subduction. Recently, Shackleton (1986) challenged this implied subduction between the Kaapvaal and Rhodesian cratons even though he supported the interpretation of collision; but as he pointed out, one must tread lightly in the minefields of Archaean-early Proterozoic geodynamics. As with most geodynamics, this challenge and warning may be due to a preoccupation with marginal processes at the expense of the identification of intraplate deformation that may well assist in the identification of the operating boundary processes, and in particular, subduction.

Four relationships within the Witwatersrand successor basin provide such intraplate evidence for not only subduction, but for an initial shallow angle of subduction. These relationships are: the cessation of volcanism at the end of Dominion time; the development of basement-involved compressive block fault during lower Central Rand time; the basin fill is composed of continental clastics; and the depositional basin was larger than the presently preserved structural limits. It is pertinent to note that no fold-thrust belt or magmatic arc has been positively identified even though these two features, like subduction, have been implied (Winter, 1986a, b; Burke *et al.*, 1986). However, the writers believe that these two seemingly negative points can also be used as evidence for proposed initial shallow-angle subduction.

The first two points define that the extensional stress system that operated in Dominion time changed to one of compression at the beginning of assigned Witwatersrand time. It has been previously discussed that under a compressional stress system volcanism will be inhibited and that non-volcanic periods can be equated with compressional stress. The transmission of compressional stress through the crust can be directly related to the angle of subduction because low-angle subduction results in a more effective coupling between the subducting and overriding plates than does steeper subduction (Cross and Pilger 1978a, b). Low-angle subduction is induced by rapid absolute motion of the upper plate toward the trench, rapid relative convergence, and subduction of young lithosphere (Cross and Pilger, 1982). Low-angle subduction and effective transmission of compressive stress has been invoked to explain Laramide-style, compressive block faults in the western United States some 1000km from the trench (Cross and Pilger, 1978a. 1982; Dickinson and Snyder, 1978) (Fig. 5a). A similar style of compressive block faulting has been interpreted as having developed in lower Central Rand time (McCarthy *et al.*, 1987) (Fig. 5b) and this faulting along the northern margin of the preserved Witwatersrand structural basin is more than 300km from the Limpopo Belt. Fault-bounded uplifts provided a flood of coarse clastics that filled adjoining sites of deposition and maintained shallow water or emergent conditions, and a similar style of sedimentation has been defined by Kluth and Coney (1981) for block uplifts associated with the Ancestral Rocky Mountains. Low angle subduction would also induce regional subsidence due to isostatic subsidence by subcrustal loading and thermal subsidence by subcrustal cooling (Cross and Pilger, 1978a), and such regional subsidence would develop a depositional basin larger than any particular site of deposition, i.e. the preserved Witwatersrand structural basin. Subcrustal loading would result in rapid subsidence associated with compressive stress transmission soon after the establishment of low-angle subduction (Cross and Pilger, 1978a). However, with thinner crusts, the effects of compressive shortening may be delayed due to flexural rigidity being a function of thermal structure (Allen *et al.*, 1986). Subcrustal cooling produces a slower subsidence pattern that follows rapid subsidence due to subcrustal loading and thus modifies the thermal structure of the region (Cross and Pilger 1978a, b). Modifications of the thermal structure would increase the density of the continental crust (Cross and Pilger, 1978a) and produce a relatively cold, elastic region a great distance from the trench that was susceptible to Laramide-style, compressive block faulting (Cross and Pilger, 1982). Disruption of the geothermal state by low-angle subduction can also be defined by a large drop in temperature indicated by the retrogression of metamorphic grade in the Limpopo Belt just prior to collision about 2,6 Ga (Light, 1982). As for the two seemingly negative points mentioned above, Dickinson and Snyder (1978) pointed out that the Laramide structural style, due to low angle subduction, results in a suppression of arc magmatism and the lack of an integrated system of nappes being developed. Both have been strong points against suggested subduction. The interpretative evidence for subduction presented here is also supported by Vlaar's (1986a, b) interpretations that lithospheric doubling (low-angle subduction) was common in the Archaean-early Proterozoic and by Nisbet and Fowler's (1983) suggestions that Archaean subduction not only occurred, but was easy and rapid. The writers have interpreted that Nisbet and Fowler's (1983) easy and rapid Archaean subduction implies low-angle subduction associated with both rapid absolute motion and rapid relative convergence.

Modification of the geothermal state and lithospheric rheology due to low-angle subduction resulted in the wholesale brittle failure of the Kaapvaal Craton about 2, 6 Ga that Light (1982) has used to define collision. Collision at about 2,6 Ga would be penecontemporaneous with the Ventersdorp Supergroup extensional stage and its associated volcanism. The writers believe that this period of intraplate deformation can again define marginal processes. As long as low-angle subduction continued, the Ventersdorp extensional stage could not develop; but with the relaxation of compressive stress, extension in the upper plate would be facilitated and would be associated with basaltic volcanism (Cross and Pilger, 1978b). Periods of extensional block faulting occurred in late Central Rand time and were followed by the extrusion of Klipriviersberg Group mafic lavas associated with faulting. With a collision event occurring at about this time, the subducted lithosphere would have been acted upon less by

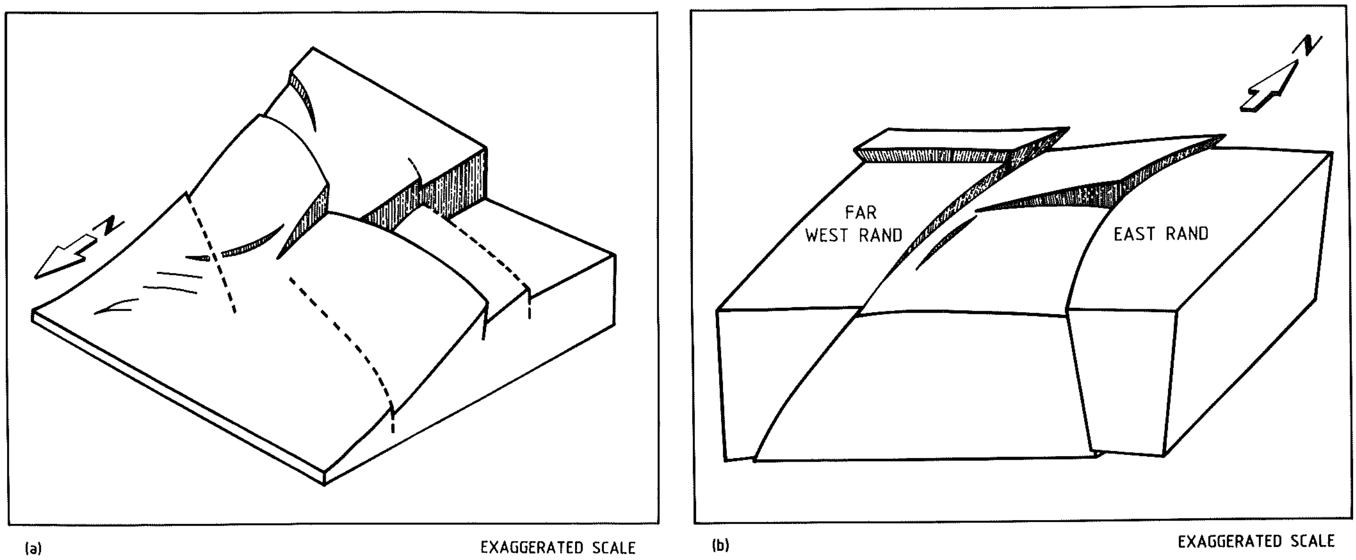

Figure 5a. Laramide-type compressive block faulting, Beartooth Range, Montana (Modified after Foose et al., 1961).
5b. Character of compressive block faulting, Central Rand Group (Modified after McCarthy et al., 1987; with permission of T.S. McCarthy, 1987).

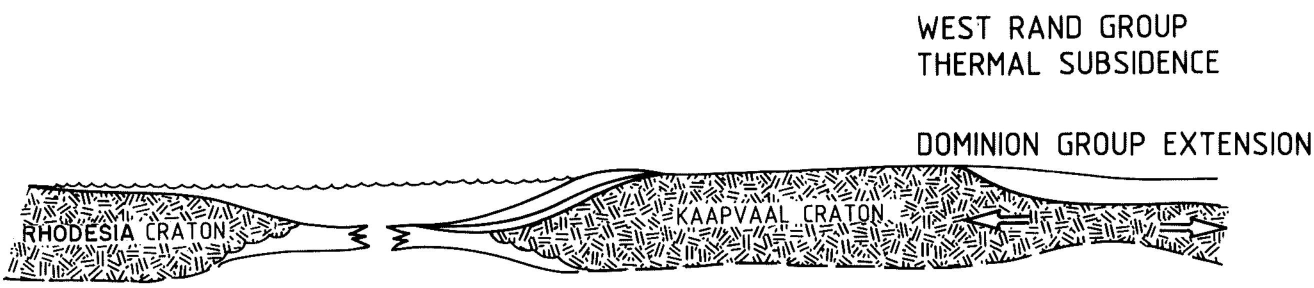

Figure 6. Cartoon of intraplate deformation during Dominion and West Rand Group time.

horizontal forces and more by vertical gravity forces so that it will roll back increasingly toward a vertical orientation producing the extensional stress systems. Models for this type of geodynamic process have been proposed by Royden *et al.*, (1983) and Murrell (1986) and these models are supported by a well-defined, near-vertical zone of intermediate-focus earthquakes that are associated with subducted, but not totally assimilated, remnants of Tethys lithosphere along the Carpathian Arc in Romania and the Hindu Kush region of India (Royden *et al.*, 1983; Gupta and Bhatia, 1986). With the roll-back of the subducting plate, the asthenosphere would rise to maintain isostatic conditions; and by maintaining isostatic conditions, large, relative, and absolute vertical displacements of the continental crust are possible (Vlaar, 1986a, b). It should be noted that initiation of extension could be before the cessation of subduction (Cross and Pilger, 1978b) and such extension may have been amplified by collision due to the relatively rapid roll-back of subducted lithosphere combined with slowdown of absolute upper plate motion. With the development of such a regional extensional stress system, the resulting deformation would be confined to pre-existing zones of tectonic weakness (Cross and Pilger, 1978b) and the Platberg Group graben appears to be controlled by such zones (Clendenin *et al.*, in press). With the development of the graben, a simple overview of the geodynamics during Platberg and Chuniespoort/Ghaap time can be defined by the two-stage McKenzie (1978) model of rapid mechanical subsidence followed by longer thermal subsidence. However, this two-stage model would represent only the graben and post-graben stages of a fully evolved three-stage rift system. The responses to changing stress systems are also not as straightforward as the McKenzie model predicts because a period of tectonic subsidence is an overlooked event between the mechanical and thermal subsidence stages. The recognition of a tectonic subsidence stage or readjustment during Bothaville-Allanridge time is important in that it is due to the episodic removal of asthenopheric upwelling prior to the thermal subsidence stage (Houseman and England, 1986).

Continued convergence between the Kaapvaal and Rhodesian cratons is defined by Light (1982) to have occurred between 2,7 and 2,5 Ga (Fig. 2). Following collision, subduction should be expected to continue as long as the force pulling the continental crust down is not balanced by the buoyancy effect (Molnar and Gray), 1979). However, resistance to subduction by young lithosphere may have generated the regional compressive stress system that during late Ventersdorp time replaced the extensional one. During early Ventersdorp time, a smaller angle of subduction, a faster convergence rate, and a thinner plate would cause a smaller resistive force at the convergence zone (Nisbet and Fowler, 1983; Vlaar, 1986b); but with a near vertical angle during late Ventersdorp time, resistive forces and friction should have increased considerably. Episodic uplift followed by basin expansion during Chuniespoort/Ghaap time defines that renewable compressive stress existed and was being relieved. Inplane compressive stress due to transmitted boundary forces will induce marginal uplift (Cloetingh *et al.*, 1985; Karner, 1986) while rapid stress relaxation will occur as the lithosphere undergoes another period of rapid thinning that causes basin expansion (DeRito *et al.*, 1983). The episodic nature of compressive stress transmission also suggests that subduction resistance was a stick-slip process and that periodic readjustments between the Kaapvaal and Rhodesian cratons resulted in stress relaxation, as well as basin expansion.

As the buoyancy effect overcomes downward pull, convergence/subsidence would stop, but gravity would continue to act on the vertical excess lithosphere of the downgoing slab (Molnar and Gray, 1979). Under this continued force, the excess lithosphere would detach itself and such detachment will be followed by rapid isostatic uplift (Murrell, 1986). Detachment is interpreted to have occurred and appears to be the geodynamic process that led to the pre-Pretoria Group regional unconformity. Detachment could also explain why crustal thickening is only preserved in the Limpopo Belt (Shackleton, 1986).

V. OPENING OF THE SELATI TROUGH

The Selati Trough is probably one of the most ignored structural features on the Kaapvaal Craton (Fig. 3). The Wolkberg Group is found in the Selati Trough; but as previously mentioned, its correlation with the overlying Chuniespoort/Ghaap Group is speculative due to an unconformable relationship with the Black Reef Quartzite Formation. The Selati Trough is defined as an east-west trending, rift-like structure that has a pronounced coincidence with the Murchison Lineament (Button, 1973a). Descriptions and sections indicate that the Selati Trough was a rather passive structure and that it was filled by middle Chuniespoort/Ghaap time (Button, 1973a, b).

Even though stratigraphic correlation is speculative, the described characteristics of the Selati Trough suggest interpretations that would allow tectonic correlation with the successor basin sequence. The passive, rift-like nature of the structure and its coincidence with the Murchison Lineament suggest that the Selati Trough opened along pre-existing zones of weakness in the 'Old Granite' basement. These zones of weakness could have been inherited from the Murchison Lineament, or due to a proto-fault system developing penecontemporaneously as a result of oblique collision, or both. The development of a proto-fault system would take up part of the oblique relative motion between plates and tend to cut through the weakened magmatic arc. Convergence between the Kaapvaal and Rhodesian craton has been defined as oblique (Clendenin *et al.*, in press) and relatively rapid, but a magmatic arc is not clearly defined. The east-west trend of the Selati Trough parallels both the Limpopo Belt and a line of granitic bodies that have been interpreted as a remnant magmatic arc (Burke *et al.*, 1986). A closer examination of these granitic bodies reveals that they form several parallel lines, that the exposed size of bodies decrease, and that they progressively young to the north of the Selati Trough toward the Limpopo Belt from about 2,6 Ga (Barton *et al.*, 1983). The writers suggest that the magmatic arc proposed by Burke *et al.*, (1986) existed to the north of the Selati Trough and evolved during Ventersdorp time. The development of this magmatic arc can be equated with steepening of the angle of subduction, while the migration of the arc further supports the interpretation of flexural roll-back following collision at about 2,7 Ga. The size of the granitic bodies would also support roll-back since arc volcanism subsides as the continental plates involved in collision are welded together (Dickinson, 1977).

Based on these descriptions and relationships, the writers suggest that the Selati Trough developed due to interarc spreading as regional extension began to exploit a weak, ductile area previously heated by arc magmatism. Molnar and Atwater (1978) indicate that interarc spreading results in largely passive openings along lines of (pre-existing) weakness within the overriding plate in response to divergent motion. Slowing of absolute plate motion combined with slower convergence should create extensional stresses or divergent motion in the overriding plate and both are associated with steeper angles of subduction (Cross and Pilger, 1982). The Selati Trough interarc basin probably began to open during early

Ventersdorp time following the expansion of volcanism due to developing regional extensional stresses in the Kaapvaal Craton. Following spreading, the Selati Trough was bounded to the north by an active, but migrating, magmatic arc. The Wolkberg Group began to accumulate in this rift-like interarc basin that was coupled to the magmatic arc. Spreading expanded and deepened the Wolkberg Group interarc basin throughout the Ventersdorp extensional stage, but spreading probably stopped as crustal re-organizations, due to tectonic subsidence, developed an erosional surface over the area. The depth of the Selati Trough due to continued spreading may explain several features defined by Button (1973a), i.e. why the Wolkberg Group is confined to the Selati Trough, why the upper marginal unconformity appears to pass laterally into conformable sequences with it, and why the region continued to be a passive structural downwarp until filled in middle Chuniespoort/Ghaap time.

1V. SUMMARY OF REGIONAL RELATIONSHIPS

The Dominion Group was deposited on the 'Old Granite' basement under a regional extensional stress system and represents a local thermal anomaly. It should be noted that a granite basement would have a lower density than a metamorphic one and form a significant mass deficiency which could be compensated by magmatic upwelling (Logatchev and Zorin, 1984). Regardless, a thinner and thermally weakened crust, which probably existed in the Archaean-early Proterozoic, would have lower cohesion and less resistance to deformation under a regional extensional stress system. With the replacement or relaxation of extensional stress in late Dominion time, compression would have aborted the progressive rift process while preserving an expanded linear zone of weakness that would have amplified and perpetuated for as much as 500 Ma (Fig.6).

The onset of regional compression or the relaxation of regional extensional stress terminated volcanism and suggests that re-initiation of convergence and subduction had modified the boundary forces influencing the transmitted in-plane stresses. The general character of the West Rand sediments indicate a subdued landscape and the absence of 'high relief' and the character of the basin appears to be simply flexural (Fig. 6). With changes in the thermal structure due to Dominion Group volcanism, thermal stress would have exerted a greater influence on the development of the early Witwatersrand successor basin than did the regional compressive stress system by which it was initiated and may explain why the effects of in-plane compression were seemingly delayed till late West Rand time. Changes in sedimentation in late West Rand time suggest that the earlier subdued landscape had been tectonically modified and imply that regional compressive stress possibly began to influence tectonic style even before the end of thermal subsidence. Cross (1986) has recently proposed that stress transmitted into the overriding plate would be relieved along the zone of least strength and this zone would be coincident with the junction between a double and a single thickness of lithosphere. If this proposal is correct, the position of greatest contrast of mechanical properties and zone of least strength would have been along the northern margin of the preserved Witwatersrand structural basin (Fig. 7).

Localization of deformation within this narrow zone would have been due to the modifications of thermal structure induced by the subducted slab (Cross, 1986) and compressive block faulting might represent a form of supracrustal loading (Fig. 8).

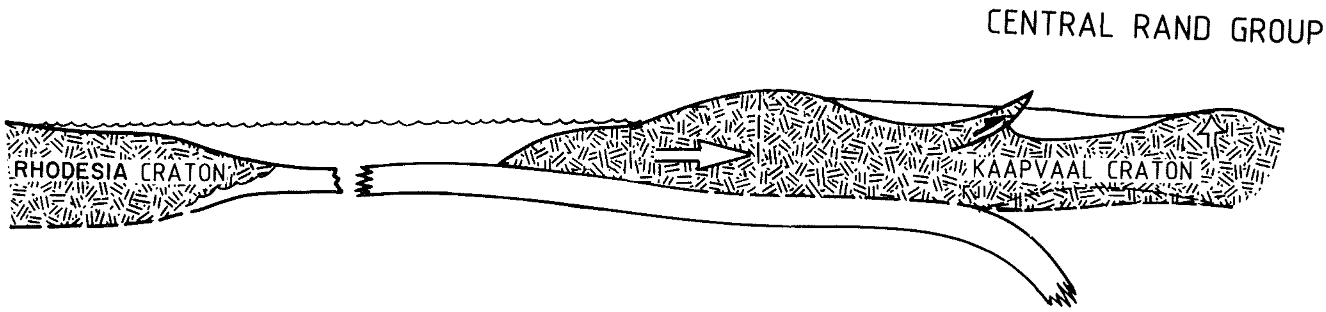

Figure 7: Cartoon of low-angle subduction and intraplate deformation during Central Rand Group time.

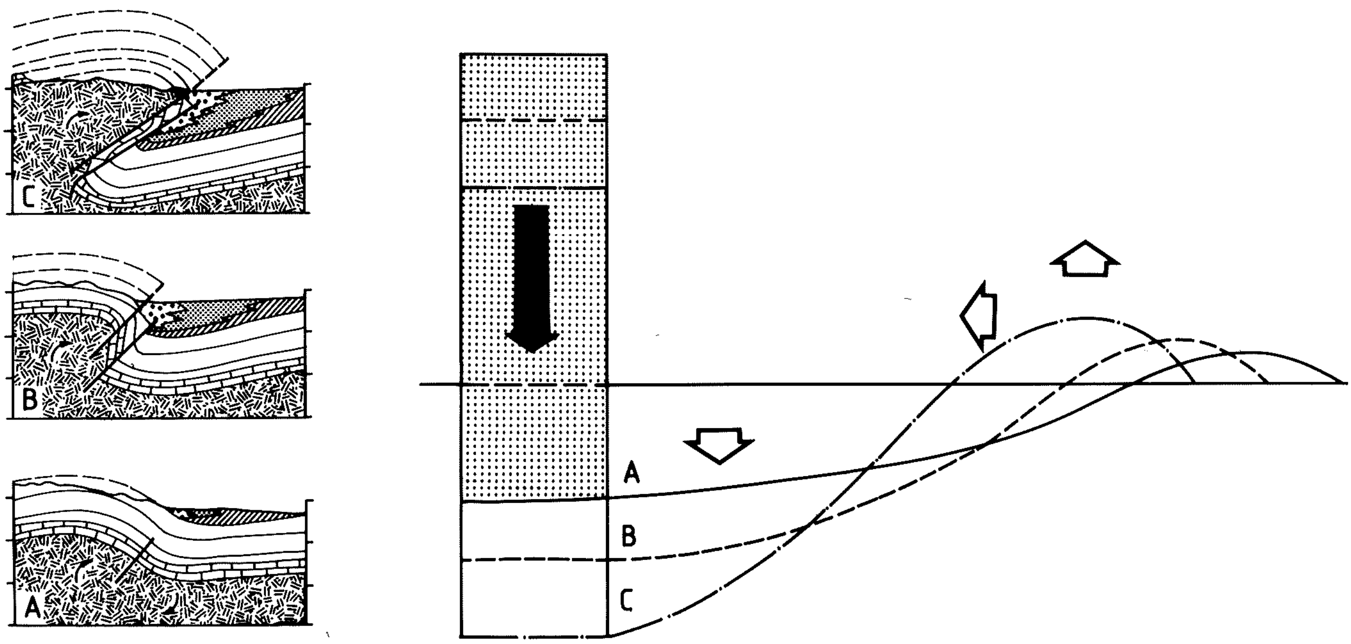

Figure 8: Evolution of compressive block fault and possible relationship to development of peripheral bulge (Modified after Quinlan and Beaumont, 1984).

Restriction of the basin in middle Central Rand time indicates that time-dependent thermal subsidence had ended and regional compressive stress dominated. The restriction of the basin could have been due to a rising peripheral bulge along the south-southwest margin of the presently preserved Witwatersrand structural basin (Fig. 7, 8) and the development of foreland-type tectonics as suggested by Winter (1986a, b) and Burke *et al* (1986). The numerous disconformities in the Central Rand Group would support the suggestion of foreland-type tectonics for the evolving basin since unconformities should be developed throughout foreland basin development (Karner, 1986). In contrast, it should be pointed out that in the case of intracratonic basins, unconformities should only be developed in the middle stages of development due to time dependent thermal subsidence (Karner, 1986). Quinlan and Beaumont (1984) have reported that loads on a viscoelastic lithosphere gradually evolve toward local isostatic equilibrium with the result that the evolving basin becomes deeper and narrower as the peripheral bulge is uplifted and broadened (Fig. 8). However, no potential supracrustal loads of the right age, i.e. penecontemporaneous magmatic arc or fold-thrust belt, other than speculative compressive block faulting, have been identified during this stage of successor basin development and these are critical assumptions that Winter (1986a, b) and Burke *et al.*, (1986) have made. If a basin pigeon-hole classification is necessary, the writers propose that the characteristics of the Witwatersrand Basin are similar to the Chinese-type basins defined by Bally and Snelson (1980). Chinese-type basins are bounded by either distal compressive block faults or wrench faults and develop without an associated A-subduction margin or fold-thrust belt (Bally and Snelson, 1980). Their origin appears to be similar to both Laramide- and Ancestral Rocky Mountains-type basins of the western United States and they are filled mostly with continental clastics (Bally and Snelson, 1980). These relationships would certainly support the interpretation that the Witwatersrand Basin was a type of foreland basin developed by low-angle subduction, but it is not the type of foreland basin proposed by either Winter (1986a, b) or Burke *et al.*, (1986). Karner and Watts (1983) have also suggested that an extra driving force is necessary for foreland flexure and such an extra driving force could be due to transmitted compressive stress (Allen *et al.*, 1986). The transmission of in-plane compressional or extensional stress would either amplify or decrease the height of the peripheral bulge (Cloetingh, 1986). The Central Rand Group was being actively influenced by in-plane compression as suggested by compressive block faulting; and with amplified uplift, the peripheral bulge should have broadened and begun to restrict the basin. However, periods of thermal subsidence, following the extrusion of the Crown and Bird lavas and defined by the deposition of shale or very fine-grained quartzite (subdued landscape sedimentation), punctuated the regional compressive stress system. Volcanic extrusions under a regional compressive stress field could only be associated with transtensional zones along penecontemporaneous wrench faults because other feeders would be closed (Dewey, 1980) and periods of superimposed thermal stress suggest periods of amplified downwarping. The period following the extrusion of the Bird lavas may have downwarped the peripheral bulge and supports the interpretation that the basin was merely restricted, not closed, to the outer southwestern seas (Clendenin *et al.*, in press).

Major changes in the sedimentation and the initiation of extensional block faulting during upper Central Rand time point to a change in the renewable stress systems before the extrusion of the Klipriviersberg Group lavas associated with a regional extensional stress field (Fig. 9). The gradual, widening areal distribution of the Klipriviersberg Group lavas and

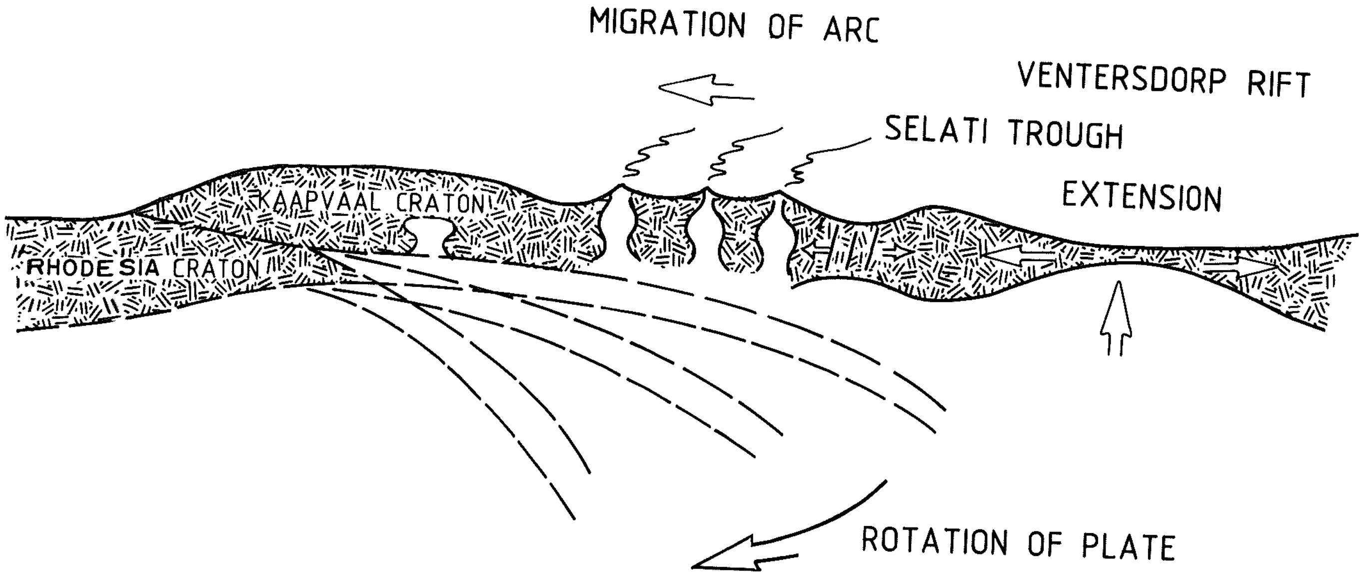

Figure 9. Cartoon of roll back of subducted slab following collision of Kaapvaal and Rhodesian cratons and resulting intraplate deformation during Ventersdorp Supergroup time.

the acceleration of faulting with time indicate that, with the rising geotherm, stress amplification was resulting in the propagation of rift-type tectonics under the regional extensional stress field. Stress amplification should be expected since areas of high geotherm and thinner lithosphere are more susceptible to faulting (Kusznir, 1982; Kusznir and Karner, 1985; Mithen, 1982). Since major graben formation is generally a late stage event and is often preceded by uplift and volcanism, stress amplification would also lead to graben development. The reason for this is that once initiated the process would accelerate as further amplification would arise due to the release of stress by brittle failure, and re-stressing of the lower lithosphere would result in creep and stress transmission again to the upper lithosphere (Kusznir and Park, 1982, 1984; Bott and Kusznir, 1984). With the lack of sediments defining tectonics in Platberg time, the bimodal volcanics in the Platberg and Buffalo Springs Groups support roll-back generated extensional stresses (Fig. 9). Intermediate and silicic volcanism following mafics have been attributed to a slowing of absolute upper plate motion and reduced convergence rates (Cross and Pilger, 1978b). Such plate motion has been interpreted to have occurred following collision in Ventersdorp time.

The recognition of a tectonic subsidence stage or readjustment in Bothaville-Allanridge time is important in that it is due to the episodic removal of asthenospheric upwelling prior to the thermal subsidence stage (Houseman and England, 1986). It may also represent a change from a regional extensional stress system to a compressional one in the Kaapvaal successor basin sequence as resistance to convergence transmitted compressive stress into the Kaapvaal Craton. The Platberg Group graben is a failed rift; and with slightly elevated mantle temperature during the Archaean-early Proterozoic, a regional compressive stress system would resist fragmentation of the cratonic mass (Vlaar, 1986b). Compressive stress systems seem to have prevailed during the Proterozoic (Vlaar, 1986a, b) and may be the reason why the Proterozoic is characterized by large intracratonic basins.

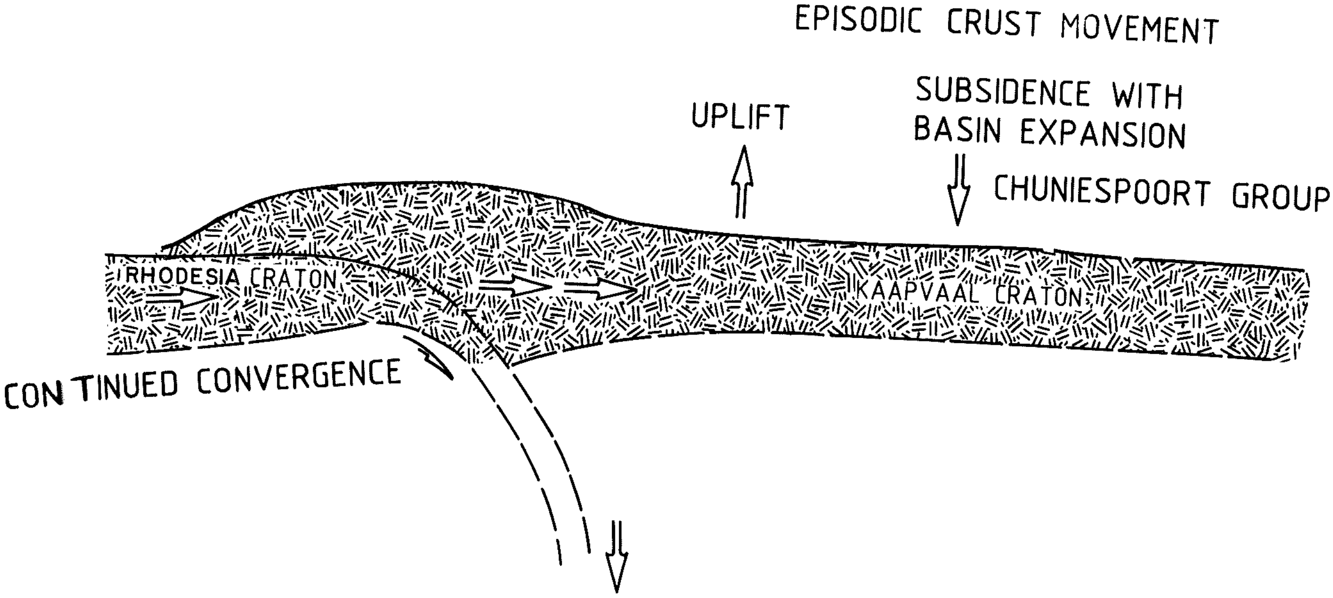

Figure 10: Cartoon of continued convergence between Rhodesian and Kaapvaal cratons and effects of transmitted in-plane stress during Chuniespoort/Ghaap Group time.

Following tectonic subsidence, thermal subsidence began to dominate over any regional renewable stress systems and the Chuniespoort/Ghaap Group intracratonic basin began to subside (Fig. 10). Thermal subsidence, as previously mentioned, is time dependent and episodic, as lateral heat flow from the cooling rift can terminate the pattern of marginal flexure and overstepping (Watts, 1982; Vlaar, 1986a). However, flexure soon overcomes the effects of lateral heat flow and younger sediments progressive onlap, due to flexural rigidity with age (Watts, 1982). Arching and the development of local unconformities during these periods of early subsidence would suggest regional compressive stress was also operating and developed a peripheral bulge-like structure that divided the basin. Stresses during this period did not exceed the strength of the upper crust and flexure resulted, instead of faulting, as is evident in Central Rand time. Such an arch structure might be expected at this time since the effectiveness of compressive in-plane stress in producing deflections is greatest on young margins subject to rapid loading (Cloetingh *et al.*, 1985) and might represent perpetuation of a crustal flaw developed during Central Rand time. Enlargement of the basin following arching suggests that relaxation of renewable stress amplified the effects of non-renewable thermal stress. It is assumed that in a hotter earth, thermal subsidence and equilibrium is reached within the same or a shorter time frame (Vlaar, 1986b). The recognized hiatus during middle Chuniespoort/Ghaap time would represent a major crustal reorganization as siliciclastics replaced carbonates and is interpreted as representing the end of thermal subsidence (Clendenin and Maske, 1986). Continued convergence following the end of thermal subsidence episodically transmitted compressive stress into the Kaapvaal Craton and stress relaxation resulted in the reactivation and progressive enlargement of the basin following brief periods of marginal uplift (Fig. 10). The maximum stress difference occurs within the basement below the basin centre and relaxation of stress would amplify the perturbation allowing the basin to expand (DeRito *et al.*, 1983; Lambeck, 1983). The filling of the Selati Trough during lower Chuniespoort/Ghaap

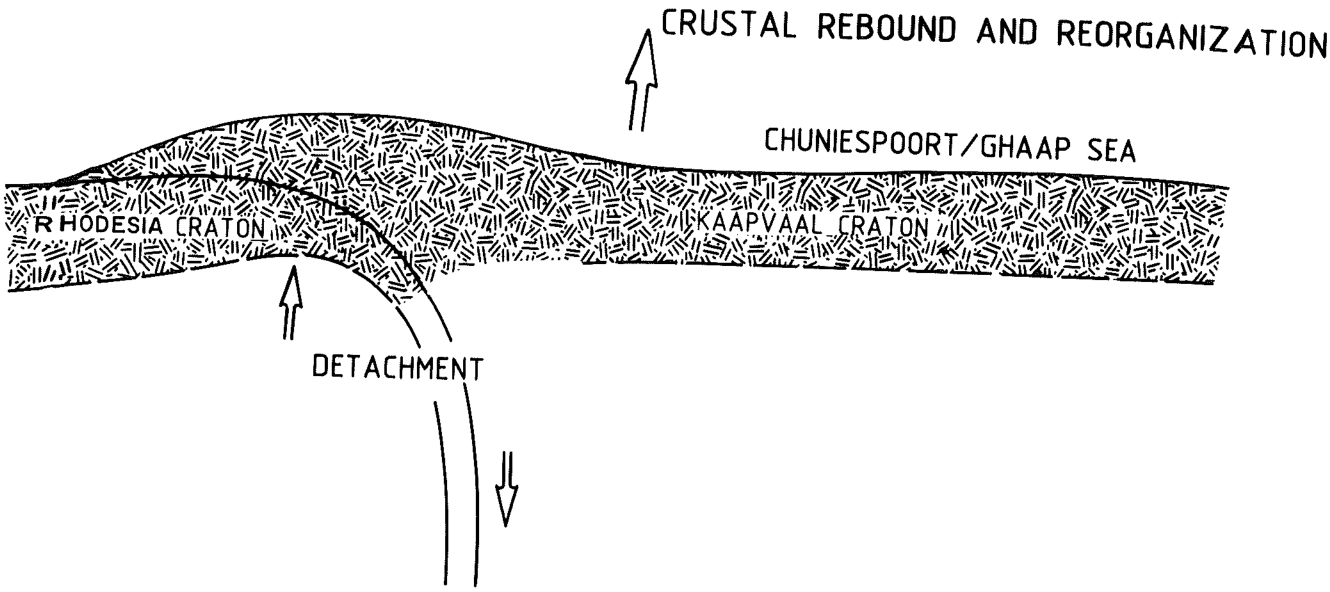

Figure 11: Cartoon of crustal rebound and reorganization leading to pre-Pretoria Group unconformity following detachment of subducted slab.

time implies that it was a temporary arm of the basin and had no long term effects as did the Platberg Group graben. Following four major periods of reactivation and expansion, the successor basin sequence was terminated by rapid uplift and development of the pre-Pretoria Group unconformity (Clendenin and Maske, 1986) (Fig. 11). A major crustal reorganization due to uplift and termination of a successor basin sequence may be a modification of the Wilson Cycle when failed rifts are involved in basin development.

V11. CONCLUSIONS

The successor basins, developed on the Kaapvaal Craton from Dominion Group (3,0 Ga) to Chuniespoort/Ghaap Group time (pre-2,2 Ga), seem to be the result of a balance of various, transmitted in-plane stress systems (renewable) and thermal stress (non-renewable), and the forces generated by a subducting slab. Particular stress systems either dominated the tectonic style, combined with others to modify it, or simply had no effect at certain times. However, it is the evolution of subduction geometries due to the kinematics of plate interactions that genetically links the events together.

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