ELSEVIER

Precambrian Research 97 (1999) 215–251

Precambrian Research

www.elsevier.com/locate/precamres

The sequence stratigraphic concept and the Precambrian rock record: an example from the 2.7–2.1 Ga Transvaal Supergroup, Kaapvaal craton

Octavian Catuneanu [a,*], Patrick G. Eriksson [b]

[a] *Department of Geology, Rhodes University, Grahamstown 6140, South Africa*
[b] *Department of Geology, University of Pretoria, Pretoria 0002, South Africa*

Accepted 21 April 1999

Abstract

Sequence stratigraphic concepts are applied to the 2.7–2.1 Ga Transvaal Supergroup, which constitutes the sedimentary floor to the Bushveld igneous complex. The Transvaal succeeds the foreland-fill of the Witwatersrand Supergroup and, in its lowermost portions, is partly synchronous with the Ventersdorp Supergroup lavas that followed upon the Witwatersrand sediments in the stratigraphic record. The unconformable contacts at the base and the top of the Transvaal Supergroup mark significant changes in the overall tectonic setting, which qualifies them as first-order sequence boundaries. Sedimentation within the Transvaal basin was controlled by cycles of extensional and/or thermal subsidence resulting in the accumulation of sequences, separated by stages of base-level fall generating sequence-bounding subaerial unconformities. The Transvaal first-order sequence is subdivided into five second-order unconformity-bounded depositional sequences, i.e. the Protobasinal, Black Reef, Chuniespoort, Rooihoogte–Timeball Hill and Boshoek–Houtenbek sequences. The span of time of these second-order sequences, averaging 130 Ma, is much longer than the duration of the second-order Phanerozoic cycles, which is explained by the less evolved and thus slower plate tectonic processes that dominated the Late Archaean–Early Proterozoic times.

Each Transvaal second-order sequence preserves a variable number of systems tracts, mainly as a function of the strength of the erosional processes associated with the ravinement surfaces and subaerial unconformities. In a complete succession, a sequence would include a basal lowstand systems tract (LST), followed by transgressive (TST), highstand (HST) and falling stage (FSST) systems tracts. The Protobasinal and Rooihoogte–Timeball Hill sequences only preserve the LST and TST, due to the strong post-depositional erosion. The Black Reef sequence is composed of LST, TST and HST. The Chuniespoort sequence preserves the TST, HST and FSST, but lacks the LST due to ravinement erosion. The Boshoek–Houtenbek sequence develops the complete succession of systems tracts. Common features may be emphasized between the same types of systems tract developed within different sequences. The LST includes high-energy coarse prograding facies, such as alluvial fans and fan–delta systems, that fill the irregularities of the pre-existing topography and lead to the peneplanation of the depositional areas. The TST may be recognized from retrogradational stacking patterns that are associated with the transgression of marine environments. The HST marks the normal regression of the previous marine environments, which assumes coeval aggradation of fluvial and basinal facies within an overall progradational framework. The FSST would normally include only basinal facies age-equivalent with the correlative subaerial unconformities. In our case study, the stages of base-level fall resulted

* Corresponding author. Tel.: +27-46-603-8313; fax: +27-46-622-9715.
E-mail address: oc@rock.ru.ac.za (O. Catuneanu)

PII: S0301-9268(99)00033-9

invariably in fully nonmarine environments within the Transvaal area, generally leading to the development of subaerial unconformities. In particular circumstances, the nonmarine environment may still accumulate and preserve fluvial, alluvial fan or lacustrine sediments as a function of the relative position between topography and local equilibrium profiles and base-levels.

The interpretations of base-level rise and fall resulting from the application of the sequence stratigraphic concept to the 2.7–2.1 Ga Transvaal Supergroup enable estimation of sea level changes during deposition of this succession. With an understanding of local tectonic conditions on the Kaapvaal craton and of coeval sedimentation elsewhere in Africa, eustatic and relative sea level changes and variation in continental freeboard conditions may be suggested for the Transvaal Supergroup.

Keywords: Base-level changes; Sea-level changes; Sequence stratigraphic concept; Transvaal Supergroup.

1. Introduction

Sequence stratigraphy is a subdiscipline of stratigraphy dealing with the genetic interpretation of lateral and vertical facies changes of a basin-fill. Sequence stratigraphic analysis assumes a subdivision of the sedimentary pile into sequences, which are stratigraphic units related to cyclic changes in the sedimentation regime through time. They include relatively conformable successions of genetically related strata displaying repetitive vertical profiles and stacking patterns. Sequences are subdivided into systems tracts, which refer to linkages of correlative depositional systems commonly associated with significant lateral changes of facies. Sequences and systems tracts are bounded by event-significant stratigraphic surfaces that mark sharp changes in the sedimentation regime or in the stratal stacking patterns.

In the simplest way, sequence stratigraphy is the study of facies changes within a time framework. It follows that the two main prerequisites for a sequence stratigraphic interpretation are to obtain, firstly, facies control and, secondly, time control. Whereas the former may be achieved by means of sedimentology and facies analysis, the latter requires independent research methods, such as biostratigraphy, radiometric dating or magnetostratigraphy. The application of sequence stratigraphic analysis to the rock record is increasingly difficult as older strata are investigated, as for example studying the Precambrian relative to the Phanerozoic, because both facies analysis and age determinations encounter practical problems when applied to such successions. Facies analysis and palaeoenvironmental reconstructions for Precambrian depositional systems may be limited by poor preservation, post-depositional tectonics, diagenetic transformations and metamorphism. The constraint of Precambrian rocks' ages, based essentially on radiometric dating, is also less precise relative to younger sequences, with error margins of at least 1 Ma. For this reason, no time control can be provided for the correlation of high frequency Precambrian sequences, with durations less than 10^0–10^1 Ma. It thus follows, that sequence stratigraphic interpretations of early Precambrian successions are not common in the literature [for an example see Beukes and Cairncross (1991)].

The purpose of this paper is to illustrate the application of sequence stratigraphic concepts to the Precambrian, using an example represented by the 2.7–2.1 Ga Transvaal Supergroup, the floor to the Bushveld igneous complex. Owing to the relatively poor chronology available for this ~650 Ma succession, we only attempt a low resolution preliminary interpretation of the high-order (low frequency) Transvaal sequence stratigraphy. We base our interpretation on an extensive data base of the sedimentology, lithostratigraphy and stratal stacking patterns, as well as on models proposed for the evolution and tectonic settings affecting the Transvaal basin [see syntheses by Eriksson and coworkers (Eriksson and Reczko, 1995; Eriksson et al., 1995)]. Changes in sea level, one of the main thrusts of this special volume and related to the continental freeboard concept (Eriksson, 1999), are implicit in the sequence stratigraphic subdiscipline. The sequence stratigraphic analysis of the Transvaal Supergroup reveals major events of base-level rises and falls, which is relevant for the record of sea-level changes that occurred during the studied interval of the Precambrian rock record.

2. Concepts of sequence stratigraphy

Sequences are defined as relatively conformable successions of genetically related strata bounded by unconformities or their correlative conformities (Mitchum, 1977). Strata within sequences display predictable changes in their stacking patterns through time, which relates to the subdivision of sequences into systems tracts. Three main types of sequence have been defined, i.e. the depositional sequence (Jervey, 1988; Posamentier et al., 1988; Van Wagoner et al., 1990; Haq, 1991; Vail et al., 1991; Hunt and Tucker, 1992), the genetic stratigraphic sequence (Galloway, 1989), and the transgressive–regressive (T–R) sequence (Embry and Johannessen, 1992; Embry, 1993, 1995). Each uses a unique combination of stratigraphic surfaces to delineate the sequence boundary and each is defined in relationship to the timing of its bounding surfaces relative to the T–R or base-level curves.

The central elements in defining sequence stratigraphic concepts (i.e. sequences, systems tracts and bounding surfaces) are the T–R and base-level curves. A brief review of the factors controlling transgressions, regressions and base-level changes is therefore necessary before exploring the various sequence models.

2.1. Factors controlling transgressions, regressions, and base-level changes

A transgression is defined as a landward migration of the marine shoreline, preserved in the stratigraphic record as a retrogradational shift of facies. In contrast, marine regression results in progradational stacking patterns related to the seaward shift of facies. Base-level is the downward limit of subaerial erosion, usually approximated with the sea-level (Schumm, 1993). The rise and fall in the base-level are independent from sedimentation, which does not allow the establishment of a direct relationship between base-level changes and movements of the shoreline.

Three main factors control the manifestation of marine transgressions and regressions (Fig. 1): eustatic oscillations; vertical tectonism (subsidence, uplift); and sedimentation. Sediment compaction may also interfere in this process, but it has essentially the same effect as tectonic subsidence. The combined effect of eustasy and tectonics provides the relative sea-level changes, with a relative increase dominated either by subsidence or eustatic rise, and a relative fall controlled either by tectonic uplift or eustatic fall. A relative sea-level rise is accompanied by a corresponding base-level rise and creation of accommodation space for sedimentation. A relative sea-level fall generates an equal amount of base-level fall, with destruction of accommodation space (Fig. 1). Independent of the direction and magnitude of base-level change, the space available for sediments to accumulate (i.e. accommodation space) may be partially or totally consumed by sedimentation.

The interaction between base-level changes and sedimentation controls the manifestation of transgressions and regressions (Fig. 1). A rise in the base-level that outpaces the sedimentation rates results in a marine transgression and retrogradation of facies. On the other hand, progradation of facies takes place during a stage of base-level rise when the sedimentation rate exceeds the rate of base-level rise. In this case, the newly created accommodation space is totally consumed by sedimentation, aggradation is accompanied by sediment bypass, and a marine regression occurs. A fall in the base-level also results in shoreline regres-

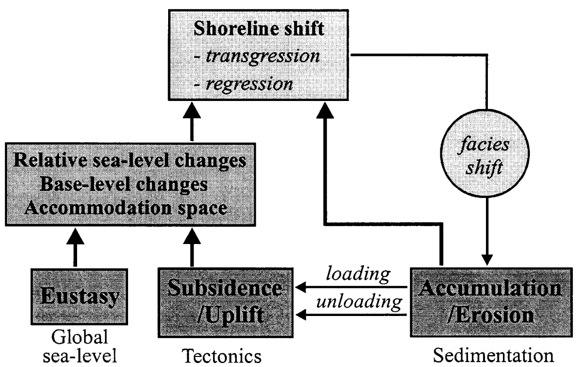


Fig. 1. Diagrammatic illustration of the controlling factors on transgressions, regressions and base-level (relative sea-level) changes. Modified from Catuneanu et al. (1998).

sion, but usually accompanied by subaerial erosion in the nonmarine portion of the basin.

In addition to the primary effects of eustasy, tectonics and sedimentation, crustal loading through sediment accumulation or unloading as a result of erosion, as well as the feedback effect of facies shift on the sedimentation rates, may also control to some extent changes in the base-level or the migration trends of the marine shoreline (Fig. 1).

2.2. T–R and base-level curves

Both the pattern of shoreline migration and the variations in base-level undergo cyclic changes through time represented by the two sinusoidal curves in Fig. 2, which are offset relative to each other. Taking the base-level curve (dashed line in Fig. 2) as a reference, the point where regression changes to transgression will always be delayed relative to the point where base-level fall is replaced by subsequent rise, as the initial rates of base-level rise are very low and therefore likely to be outpaced by the sedimentation rates. Thus, the regression that accompanies a stage of base-level fall also continues in the early stages of base-level rise. Similarly, the end of a transgressive stage will always occur before the end of base-level rise, as the latest stage of base-level rise is characterized by rates lower than the sedimentation rates. In this way, the interaction between sedimentation and an oscillating base-level will always generate strongly asymmetric T–R curves, with the regressive portions much longer than the transgressive ones, assuming that the reference base-level changes are represented by a symmetric sine curve (Fig. 2). The regression taking place during base-level fall is defined as 'forced regression' (driven by eustasy and/or tectonics), whereas the regression taking place during base-level rise is called 'normal regression' (driven by sedimentation) (Posamentier et al., 1992).

2.3. Sequence-bounding surfaces

The main types of bounding surface used in sequence stratigraphy are the subaerial unconformity (SU) and its correlative conformity (c.c.), the basal surface of forced regression (BSFR), the conformable transgressive surface (CTS), the maximum flooding surface (MFS) and the ravinement surface (RS) (Fig. 2).

The SU is an erosional surface created during times of base-level fall by subaerial processes such as fluvial or wind degradation. It gradually extends basinward during the forced regression of the shoreline and reaches its maximum extent at the end of base-level fall. Criteria for recognition of SU in the field have been reviewed by Shanmugam (1988). The SU has a marine c.c. whose timing corresponds to the end of base-level fall in the shoreline area (Fig. 2). The c.c. develops within a continuous regressive succession, which poses practical problems in its field recognition (Embry, 1995). At a regional scale, the c.c. separates rapidly prograding and offlapping forced regressive strata from the overlying aggradational normal regressive strata.

The BSFR develops within the marine portion of the basin, with a timing corresponding to the start of base-level fall in the shoreline area (onset of forced regression; Fig. 2). It separates aggrading regressive strata from the overlying rapidly prograding and offlapping forced regressive deposits. The BSFR is an unconformity cut by waves in the shallower portion of the basin [the regressive surface of marine erosion of Plint (1988)], or generated as a scour surface associated with gravity flows in the deeper basinal areas and mapped at the base of the submarine fans (Hunt and Tucker, 1992).

The CTS forms within a marine succession at the limit between underlying coarsening-upward regressive strata and overlying fining-upward transgressive strata (Embry, 1995). As the name implies, it is a nonerosional surface and it is generated during base-level rise when the rate of rise begins to exceed the sedimentation rate. The key to the identification of this surface is reliable facies analysis that allows regressive and transgressive trends in strata to be recognized and a boundary between the two to be drawn. The nonmarine correlative of the CTS ('CTS-c' in Fig. 2) is a conformable surface that may be indicated by an abrupt decrease in fluvial energy, e.g. a change from braided to overlying meandering styles

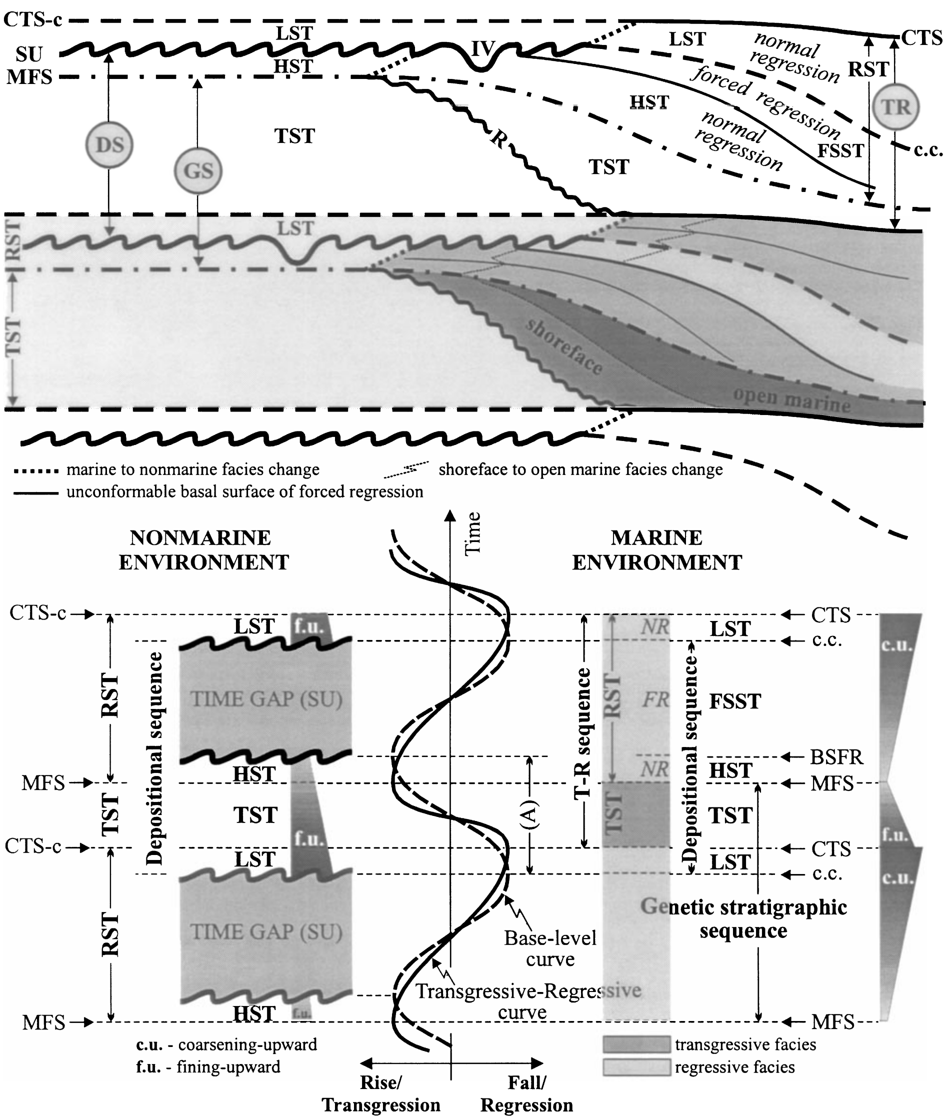


Fig. 2. Types of sequence, systems tract and bounding surface, defined in relationship to the base-level and T–R curves. Abbreviations: DS = depositional sequence; GS = genetic stratigraphic sequence; T–R = transgressive–regressive sequence; LST = lowstand systems tract; TST = transgressive systems tract; HST = highstand systems tract; FSST = falling stage systems tract; RST = regressive systems tract; SU = subaerial unconformity; c.c. = correlative conformity; BSFR = basal surface of forced regression; CTS = conformable transgressive surface; CTS-c = CTS-correlative (i.e. the nonmarine correlative of the marine CTS); MFS = maximum flooding surface; R = ravinement surface; IV = incised valley; (A) = creation of accommodation space (base-level rise); NR = normal (sediment supply-driven) regression; FR = forced (base-level fall-driven) regression. Modified from Catuneanu et al. (1998).

(Shanley et al., 1992), corresponding to the timing of the seaward-most position of the shoreline. In coastal settings, the CTS underlies the oldest estuarine facies. In offshore regions, the CTS may be onlapped by the overlying transgressive strata (marine onlap). A synonymous term for the CTS is ‘maximum regressive surface’ (Catuneanu, 1996; Helland-Hansen and Martinsen, 1996; Catuneanu et al., 1998).

The MFS (in Fig. 2) separates underlying transgressive strata (retrogradational stacking pattern) from the overlying regressive strata (progradational stacking pattern). The MFS is generally conformable, excepting for the outer shelf and upper slope regions where the lack of sediment supply may leave the seafloor exposed to erosional processes (Galloway, 1989). Within a marine succession that preserves transgressive strata, the MFS can be mapped at the limit between fining-upward and overlying coarsening-upward profiles. Owing to sediment trapping within the shoreline systems during the landward shift of facies, the offshore transgressive strata underlying the MFS may be reduced to a condensed section (Loutit et al., 1988). Criteria for the recognition of MFS in the nonmarine portion of the basin-fill have been provided by Shanley et al. (1992), mainly based on tidal influences in fluvial sandstones. The position of the nonmarine MFS may also be indicated by an abrupt increase in fluvial energy, e.g. a change from meandering to overlying braided styles (Shanley et al., 1992), or by extensive coal seams (Hamilton and Tadros, 1994). The latter criterion would obviously not be applicable to Precambrian successions. In coastal settings, the MFS overlies the youngest estuarine facies. The MFS is generated late in the base-level rise portion of a cycle when the rates of sedimentation begin to exceed the rates of rise, i.e. at the turnaround point from transgression to subsequent regression. A synonymous term for the MFS is 'maximum transgressive surface' (Catuneanu, 1996; Helland-Hansen and Martinsen, 1996; Catuneanu et al., 1998). In seismic stratigraphic terms, the marine MFS is a 'downlap surface', being overlain by the downlap-type of foreset terminations.

The RS is an erosional (scour) surface that is formed by the erosive action of waves and currents in the upper shoreface area. This surface is generated during the portion of base-level rise when the shoreline moves landward (transgression). It is a highly diachronous surface, varying with the rate of shoreline transgression. The RS qualifies as an 'onlap surface' in seismic stratigraphic terms, as it is onlapped during the retrogradational shift of facies by marine transgressive strata (coastal onlap).

2.4. Systems tracts

A systems tract forms a subdivision of a sequence; it refers to a linkage of correlative depositional systems, usually associated with significant lateral changes of facies. Systems tracts are interpreted on the basis of stratal stacking patterns and types of bounding surface. The following types of systems tract are used in sequence stratigraphic analysis (Fig. 2).

The lowstand systems tract (LST) is bounded by the SU and its marine c.c., at the base, and by the CTS and its nonmarine correlative at the top. It forms during the early stage of base-level rise when the rate of rise is outpaced by the sedimentation rate (case of normal regression). The LST includes the coarsest sediment fraction of both marine and nonmarine sections, i.e. the upper part of an upward-coarsening profile in a marine succession, and the lower part of a fining-upward profile in nonmarine strata, including lag deposits and reworked debris in the rejuvenated nonmarine sedimentation area. Coastal aggradation decreases the slope gradient in the downstream portion of the fluvial systems, which induces a lowering in fluvial energy, fluvial aggradation, and an overall upwards-decrease in grain size. The increase in the rate of base-level rise also contributes to the overall fining-upward fluvial profile, as it creates more accommodation space for floodplain deposition and increases the ratio between floodplain and channel sedimentation. The nonmarine portion of the LST is equivalent to the coarse-grained 'Low Accommodation Systems Tract' of Dahle et al. (1997), with the coarse clastic input related to the rejuvenated sediment source areas. Typical examples of LST deposits include incised valley fills and amalgamated fluvial channels in nonmarine successions, as well as low-rate aggradational and progradational (normal regressive) marine strata, including sharp-based shoreface deposits overlying the regressive surface of marine erosion.

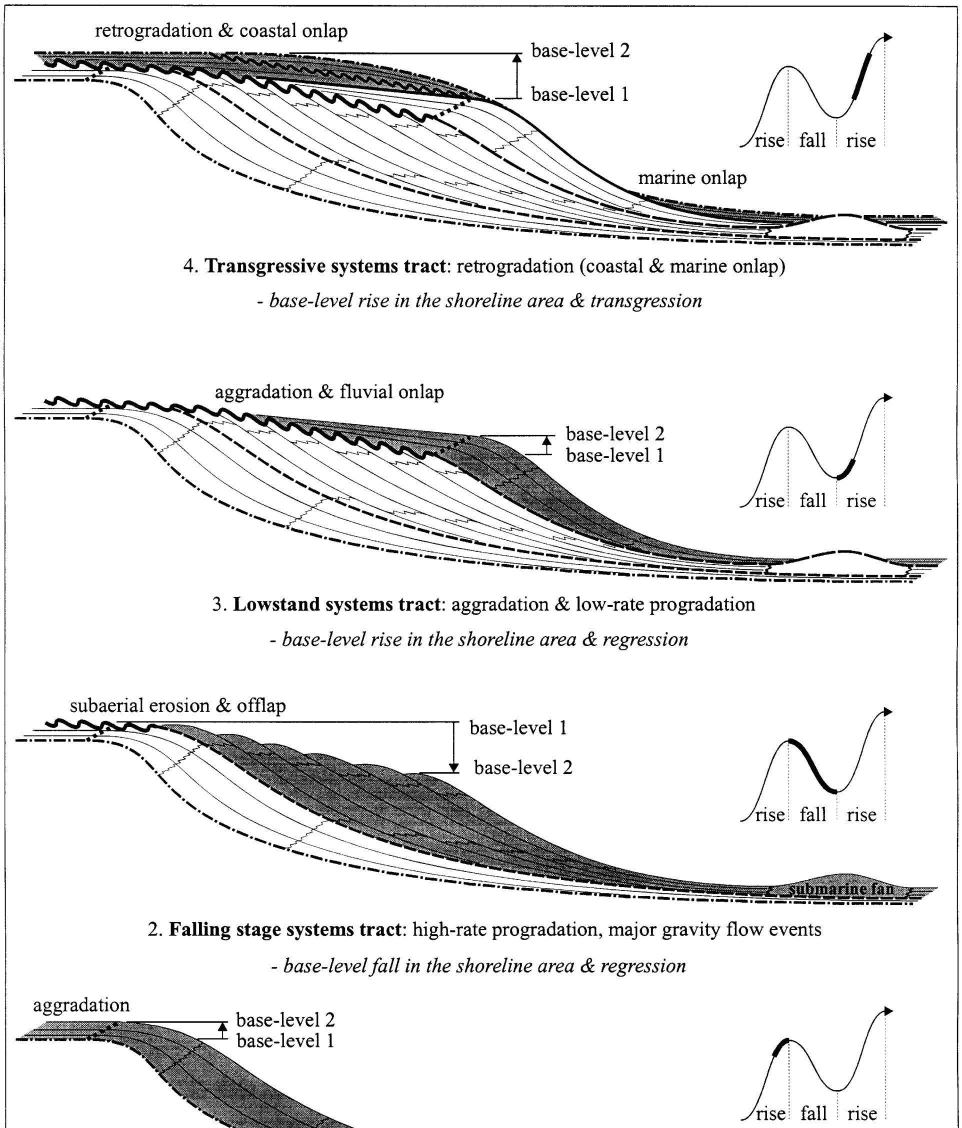
The transgressive systems tract (TST) is bounded by the marine CTS and its nonmarine correlative, at the base, and by the MFS at the top. The TST forms during the portion of base-level rise when the rates of rise outpace the sedimentation rates. It can be recognized, firstly, from retrogradational stacking patterns, and, secondly, from overall fining-upward profiles within both marine and nonmarine strata in relationship to the landward facies shift and the gradual denudation of the sediment source areas. The marine portion of the TST mainly develops in shallow areas adjacent to the shoreline, with correlative condensed sections, unconformities or onlapping gravity flow deposits offshore (Galloway, 1989). In coastal settings, the TST includes diagnostic estuarine facies. The nonmarine portion of the TST displays a lowering of fluvial energy with increasing stratigraphic height, which may lead to a change in fluvial style from braided to meandering, in response to the landward shift of the shoreline and the low-energy estuarine system (Shanley et al., 1992). As braided fluvial styles predominate within Precambrian successions (e.g. Cotter, 1978; Eriksson et al., 1998), this criterion will be difficult to apply in the pre-Phanerozoic rock record.

The highstand systems tract (HST) is bounded by the MFS at the base, and by the SU and the BSFR at the top (Fig. 2). It corresponds to the late stage of base-level rise during which the rates of rise drop below the sedimentation rates, resulting in the normal regression of the shoreline. Concomitantly with differentiated fluvial aggradation and decrease in topographic slope, the nonmarine portion of the HST records an upward lowering in fluvial energy that may result in a change in fluvial style from braided to meandering (Shanley et al., 1992). This trend, superimposed on continued denudation of the sediment source areas, tends to generate an upward-fining profile that continues the overall upwards decrease in grain size recorded by the underlying LST and TST. However, the late HST may be characterised by laterally interconnected, amalgamated channel and meander belt systems with poorly preserved floodplain deposits, due to the lack of floodplain accommodation space once the rate of base-level rise decreases, approaching the stillstand (Shanley and McCabe, 1993). Overall, the nonmarine portions of the TST and overlying HST are equivalent to the fine-grained 'High Accommodation Systems Tract' of Dahle et al. (1997). During the subsequent base-level fall, the top of the nonmarine HST may be affected by erosion or pedogenic processes. The marine portion of the HST displays a coarsening-upward profile related to the basinward facies shift, and includes low-rate prograding and aggrading normal regressive strata. Within the overall regressive marine succession, the HST occupies the lower part of the coarsening-upward profile (Fig. 2).

The falling stage systems tract (FSST) develops during base-level fall. Diagnostic for this systems tract is the marine forced regressive deposits with rapidly prograding and offlapping stacking patterns [the 'slope component' of Hunt and Tucker (1992), comprising autochthonous wedges developed on the continental slope]. The forced regressive prograding lobes are age-equivalent with deep submarine fans, as well as the stratigraphic hiatuses corresponding to the SU and the regressive surface of marine erosion that reshapes the continental shelf during the forced regression. Although subaerial processes such as fluvial and wind degradation tend to be more active during stages of base-level fall, the nonmarine portion of the basin may still accumulate and preserve fluvial, alluvial fan or lacustrine sediments as a function of the relative positions of topography and local equilibrium profiles and base-levels. The marine portion of the FSST is bounded by the BSFR, at the base, and the c.c. at the top. Owing to the gradual seaward migration of the SU during base-level fall, the older marine forced regressive strata may be capped by the basinward termination of the SU (Fig. 2; Hunt and Tucker, 1992).

The regressive systems tract (RST) is the counterpart of the TST, and includes all strata accumulated during shoreline regression, i.e. the succession of HST–FSST–LST systems tracts. It is bounded by the MFS at the base, and by the CTS and its nonmarine correlative at the top. The RST is defined by progradational stacking patterns in both marine and nonmarine strata. Within a nonmarine succession, the RST includes the time gap corresponding to the SU that separates coarser facies above (LST) from finer facies below (HST). Within marine strata, the RST displays a coarsening-upward (upward-shallowing) profile, which also includes an unconformity corresponding to the basal surface of forced regression.

Diagnostic stacking patterns for the four inde-



4. **Transgressive systems tract**: retrogradation (coastal & marine onlap)
- base-level rise in the shoreline area & transgression

3. **Lowstand systems tract**: aggradation & low-rate progradation
- base-level rise in the shoreline area & regression

2. **Falling stage systems tract**: high-rate progradation, major gravity flow events
- base-level fall in the shoreline area & regression

1. **Highstand systems tract**: aggradation & low-rate progradation
- base-level rise in the shoreline area & regression

subaerial unconformity
ravinement surface
marine to nonmarine facies change
lateral changes within the marine facies
correlative conformity
basal surface of forced regression
maximum flooding surface
conformable transgressive surface

Fig. 3. Diagnostic stacking patterns for the HST, FSST, LST and TST. The sinusoidal curves illustrate base-level changes in the shoreline area, and the timing of each systems tract relative to the base-level curve. Modified from Catuneanu et al. (1998).

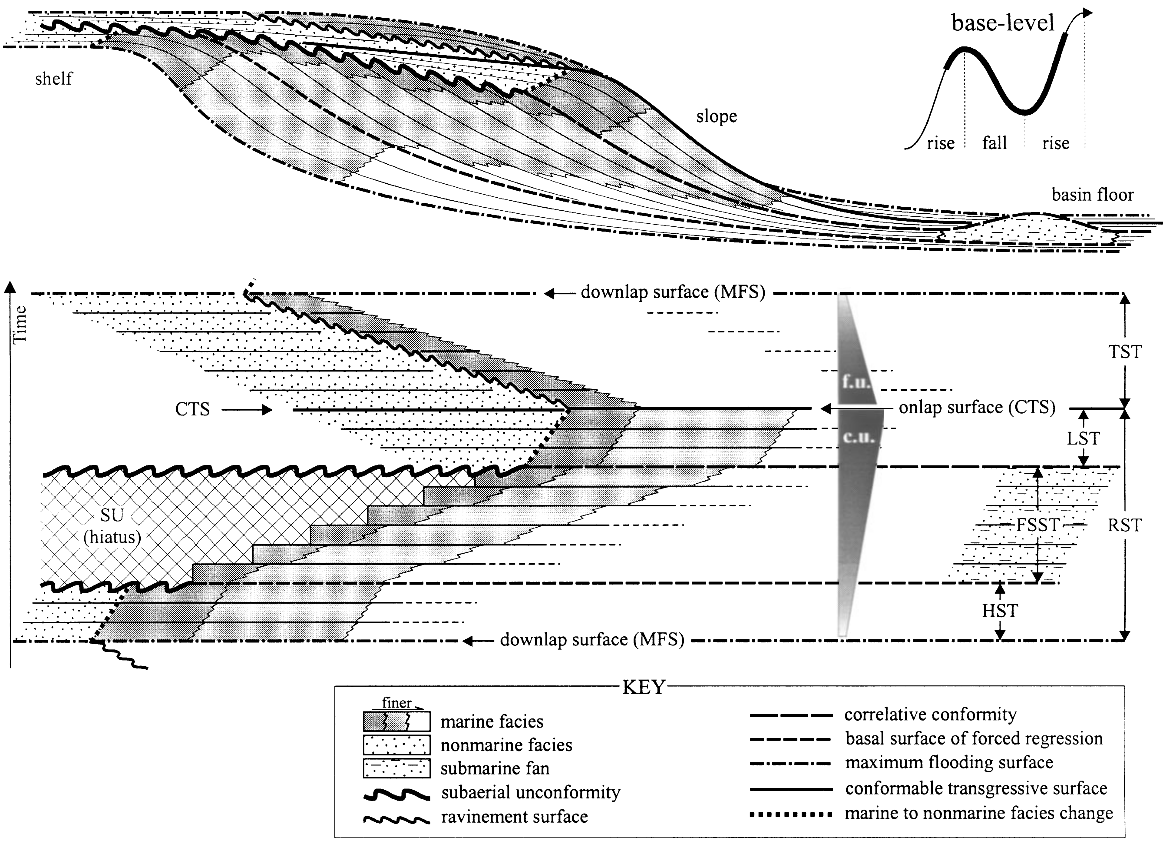


Fig. 4. Cross-sectional view and Wheeler diagram illustrating the internal architecture and evolution of a genetic stratigraphic sequence. Abbreviations: MFS = maximum flooding surface; CTS = conformable transgressive surface; SU = subaerial unconformity; HST = highstand systems tract; FSST = falling stage systems tract; LST = lowstand systems tract; RST = regressive systems tract; TST = transgressive systems tract; c.u. = coarsening-upward; f.u. = fining-upward. Modified from Catuneanu et al. (1998).

pendent systems tracts (HST, FSST, LST and TST) are illustrated in Figs. 3 and 4.

2.5. Sequence types

Three types of sequence (described below) are currently used in sequence stratigraphic analysis (Fig. 2).

The depositional sequence (Posamentier et al., 1988) is bounded by the SU and its marine c.c. Vertical profiles normally display a fining-upward trend in nonmarine strata, and fining-upward followed by coarsening-upward trends separated by an MFS in a marine succession. The depositional sequence is defined, relative to the base-level curve, as the timing of the SU and its c.c. depends on base-level changes and not on the migration trends of the shoreline depicted by the T–R curve. The depositional sequence is built by a LST–TST–HST–FSST succession of systems tracts.

The genetic stratigraphic sequence (Galloway, 1989) uses the MFS as bounding surface, and it is divided by the CTS and its nonmarine correlative into an RST followed by a TST. This sequence type is defined relative to the T–R curve. As the RST is composed of an HST–FSST–LST succession of systems tracts, Figs. 3 and 4 suggest the evolution and internal architecture of a genetic stratigraphic sequence.

The T–R (Embry and Johannessen, 1992) is half a wavelength out of phase relative to the previous type, being composed of a TST followed by an RST. Within a marine succession, the T–R sequence boundary is represented by the CTS (top of regressive facies). Within a nonmarine succession, the T–R sequence boundary is not taken at the top of the RST, owing to the difficulties in field recognition of the CTS-nonmarine correlative, but is chosen to coincide with the depositional sequence boundary (i.e. the SU). As the SU and the CTS are temporally offset (corresponding to the end of base-level fall and the end of regression respectively), a linking surface is necessary to make the physical connection between the SU and the CTS. The RS was selected to make the linkage, based on the assumption that the erosion associated with this scour surface would entirely remove the nonmarine LST in the area adjacent to the shoreline, placing in this way the CTS in direct continuation with the SU. Thus, despite the fact that the SU is not a regressive–transgressive boundary, being overlain by LST strata which are still regressive, the T–R sequence is taken by definition as being "bounded by subaerial unconformities and/or ravinement surfaces and their correlative transgressive surfaces" (Embry, 1995).

3. Tectonic settings and lithostratigraphy of the Transvaal Supergroup

The Late Archaean–Palaeoproterozoic Transvaal Supergroup overlies the Witwatersrand Supergroup in the stratigraphic record and constitutes the sedimentary floor to the Bushveld igneous complex (Fig. 5). The lavas of the Ventersdorp Supergroup overlie the Witwatersrand basin as well as being relatively widespread on the Kaapvaal craton. Available age data suggest that the Ventersdorp volcanism, the main event of the Limpopo collisional orogen, and the lowermost Transvaal rocks are approximately coeval (e.g. Eriksson and Reczko, 1995). The Transvaal Supergroup is preserved in three separate basins, i.e. the Transvaal, Kanye and Griqualand West basins. We only deal in this paper with the sedimentary fill of the Transvaal basin, situated on the Kaapvaal craton and extending from South Africa into easternmost Botswana (Fig. 6).

The age of the Transvaal succession is constrained between 2714 Ma for the basal Ventersdorp lavas (Armstrong et al., 1991) and 2050 Ma for the Bushveld complex (Harmer and von Gruenewaldt, 1991). Within these limits, the Transvaal chronology lacks a sufficient degree of detail for high-resolution stratigraphic correlation. The Transvaal Supergroup is subdivided into four main lithostratigraphic units, i.e. the Protobasinal rocks, Black Reef Formation, Chuniespoort Group and Pretoria Group (Eriksson and Reczko, 1995; Fig. 5). Traditionally, the Rooiberg Group has also been included within the Transvaal Supergroup. Recent zircon dating of the Rooiberg lavas indicates a synchronous genesis of the Bushveld and Rooiberg magmas (Hatton and Schweitzer, 1995), for which reason the Rooiberg Group could be considered as part of the Bushveld complex (Eriksson and Reczko, 1995; Fig. 6). Accordingly, we do not include the Rooiberg succession in our analysis.

3.1. *Protobasinal rocks*

The Protobasinal rocks accumulated within an overall tectonic escape-related extensional setting that developed in response to the collision between the Zimbabwe and Kaapvaal cratons during the 2.71–2.64 Ga Limpopo (Ventersdorp age) orogeny (Burke et al., 1985; Stanistreet and McCarthy, 1991; McCourt, 1995; Eriksson and Reczko, 1995). Sedimentation took place simultaneously in discrete fault-related basins identified as either small and deep pull-apart basins associated with strike-slip faults (e.g. Buffelsfontein, Tshwene-Tshwene and Godwan depositories), or larger rift basins associated with extensional tectonism (e.g. Wolkberg, Wachteenbeetje, Bloempoort and Mogobane depositories; Figs. 6 and 7). These basins underwent active subsidence and creation of accommodation space during Ventersdorp times, illustrated as base-level rise in Fig. 5.

The basin-fill successions of the Protobasinal rocks are preserved in isolated areas, which together with the stratal geometry would suggest separate depocentres for sediment accumulation (Eriksson and Reczko, 1995). However, it is possible that these depositories were interconnected during the syn-depositional time, as the individual

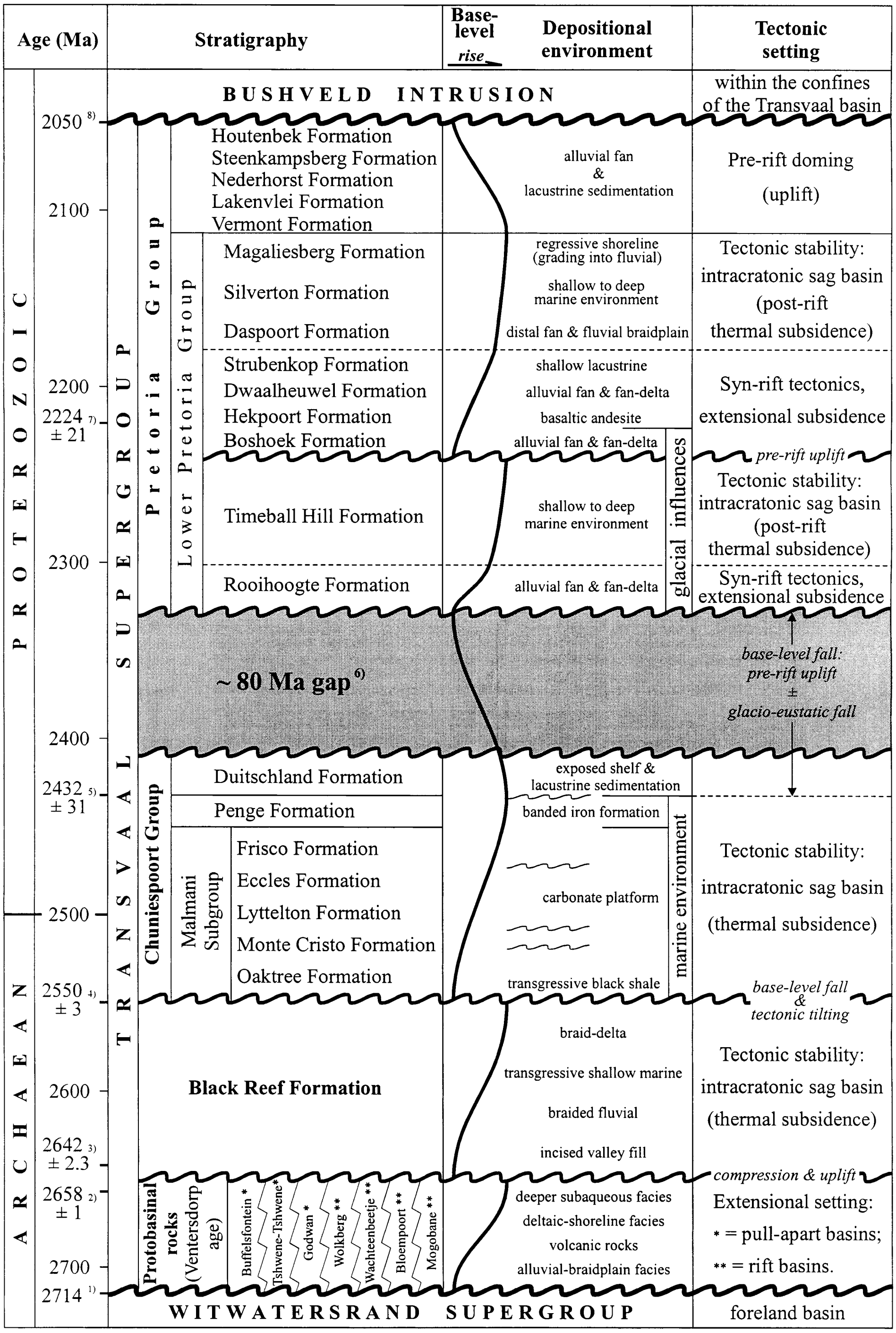


Fig. 5. Lithostratigraphy, chronology, tectonic settings, palaeoenvironments and inferred base-level changes for the Transvaal Supergroup. (1) Armstrong et al. (1991); (2) and (6) Eriksson and Reczko (1995); (3)–(5) and (7) Walraven and Martini (1995); (8) Harmer and von Gruenewaldt (1991). Wavy lines suggest unconformable contacts.

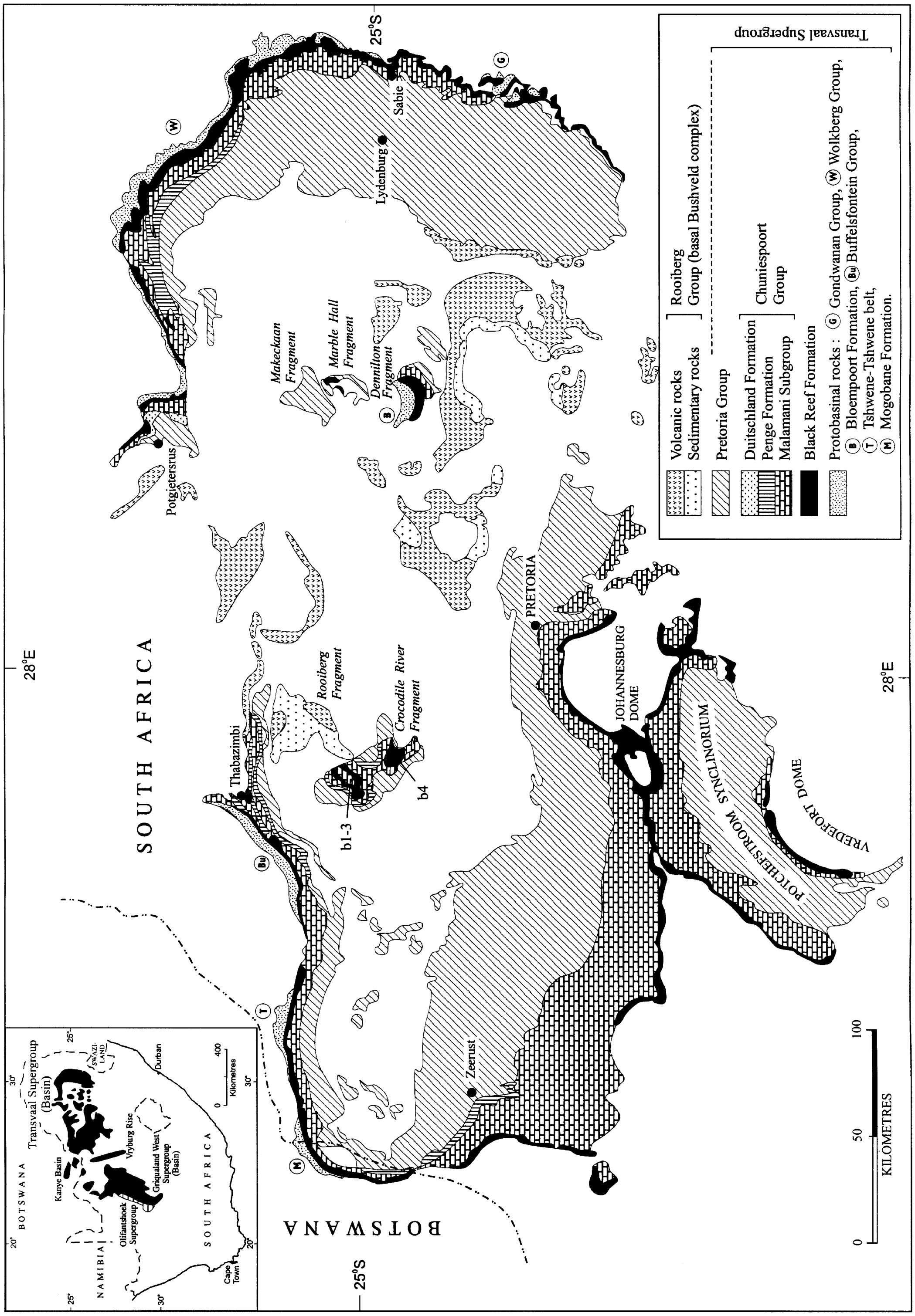


Fig. 6. Outcrop distribution of the Transvaal lithostratigraphic units within the confines of the Transvaal basin. Note that the Wachteenbeetje Formation, not represented on this map, is only known from boreholes (locations of four boreholes given by b1–3, b4, Crocodile River fragment). Modified from Eriksson and Reczko (1995).

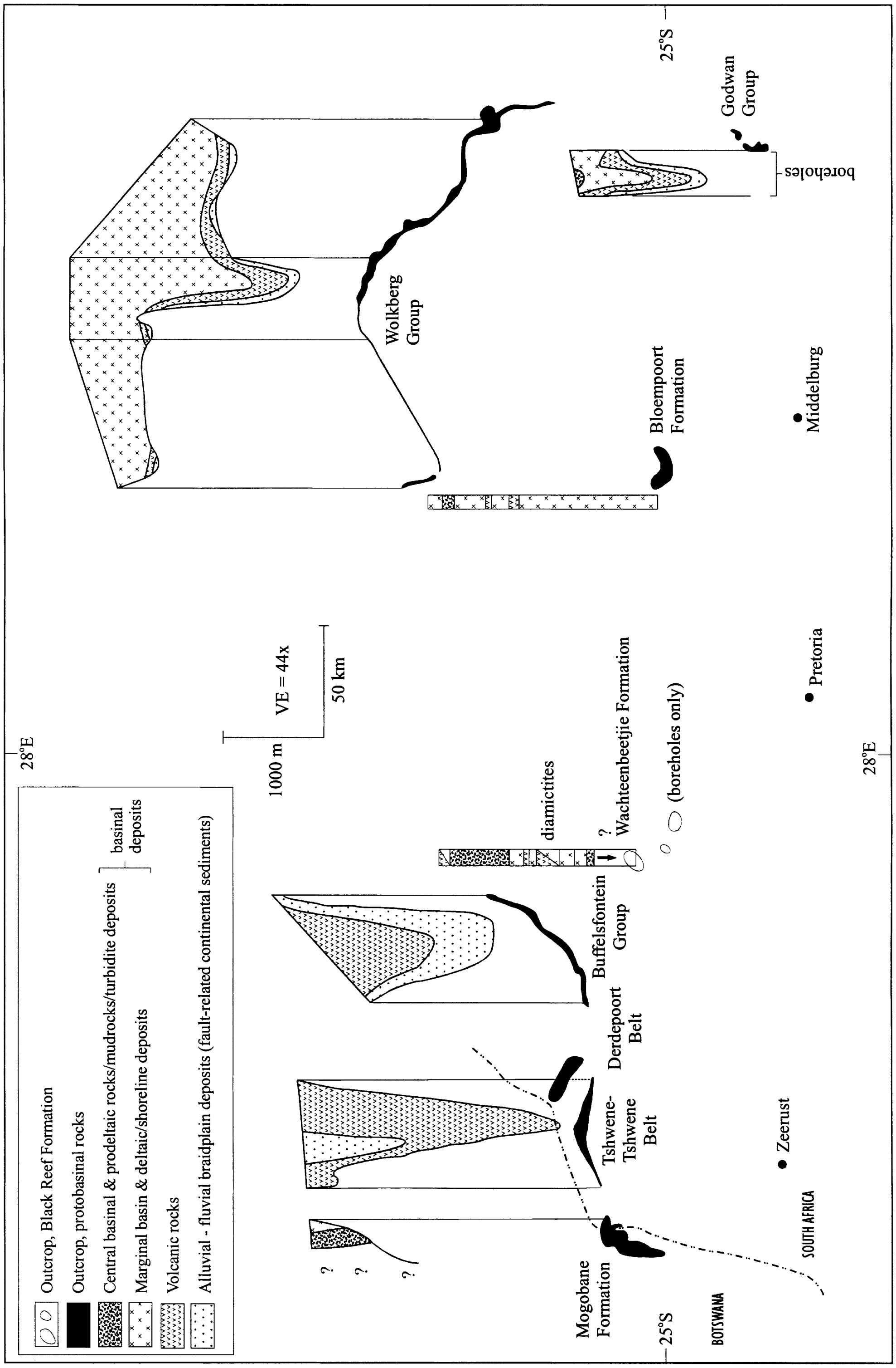


Fig. 7. Fence diagram illustrating the geometry and facies of the Protobasinal rocks of the Transvaal Supergroup [from Eriksson and Reczko (1995)].

basins tend to display common fill patterns and similar successions of facies (Fig. 8). The oldest lithofacies assemblage is generally represented by coarse nonmarine beds, ranging from boulder conglomerates to sandstones, which are compositionally and texturally immature and associated with gravity-flow deposits (Eriksson and Reczko, 1995). This lithofacies is interpreted to be the product of proximal, fault-related alluvial fans and braided fluvial systems (Button, 1973; Tyler, 1979a,b; Myers, 1990; Bosch, 1992; Eriksson and Reczko, 1995), most likely associated with the early syn-rift opening of the extensional basins. Apart from this first pulse of nonmarine sedimentation that took place before and/or coeval with the volcanic activity, fluvial systems also succeeded the stage of emplacement of volcanic rocks (e.g. within the Tshwene-Tshwene basin, Fig. 8).

A second lithofacies assemblage is represented by bimodal, basaltic and acid lavas, indicative of both pull-apart and rift basins. The volcanic rocks may have formed a laterally continuous blanket that covered the entire Transvaal basin, separating the two distinct generations of fluvial strata identified earlier as a first lithofacies assemblage (Fig. 8).

A third lithofacies assemblage, best exposed in the Wolkberg and Godwan basins where it overlies the volcanic rocks, is represented by basinal deposits. The basinal succession starts with basin margin facies, overlain by deeper water facies. The basin margin facies assemblage is dominantly arenaceous with intercalations of stromatolitic carbonates, and associated with shallow water sedimentary structures such as ripples, mudcracks and cross-bedding. A shoreline-deltaic interpretation is attributed to this basin margin assemblage (Button, 1973; Key, 1986; Myers, 1990; Hartzer, 1994; Eriksson and Reczko, 1995). The overlying deeper water facies include argillaceous lithologies, commonly calcareous or carbonaceous, with

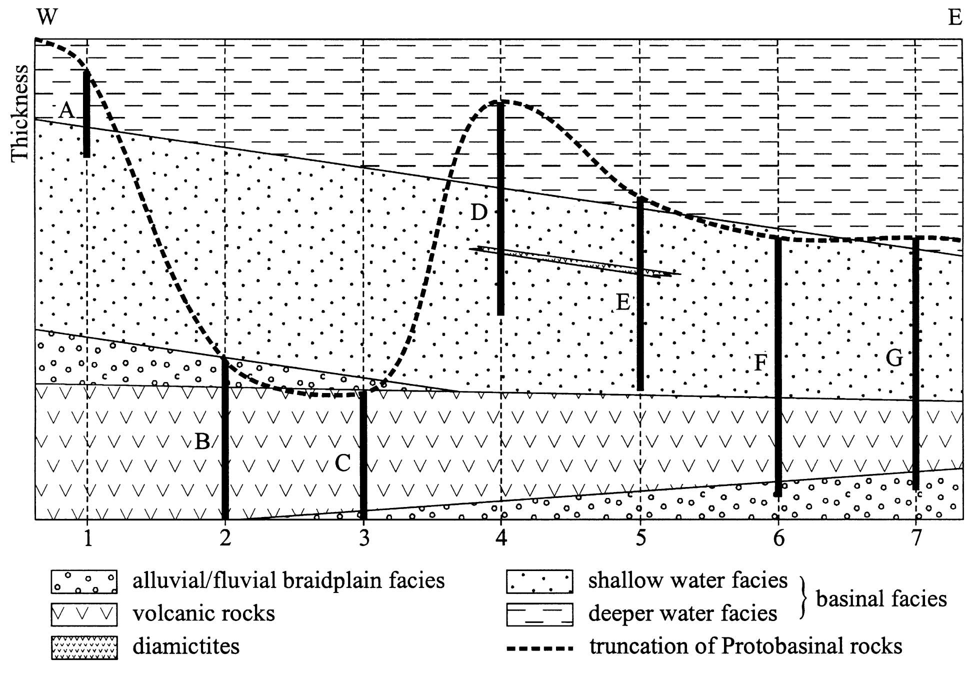


Fig. 8. Generalized correlation of facies of the discrete Protobasinal basins. Not to scale. Total thickness represented ~5 km. For a more detailed description of the Protobasinal lithostratigraphies, see synthesis by Eriksson and Reczko (1995). Abbreviations: 1 = Mogobane basin; 2 = Tshwene-Tshwene basin; 3 = Buffelsfontein basin; 4 = Wachteenbeetje basin; 5 = Bloempoort basin; 6 = Wolkberg basin; 7 = Godwan basin; A = Mogobane Formation; B = Tshwene-Tshwene belt; C = Buffelsfontein Group; D = Wachteenbeetje Formation; E = Bloempoort Formation; F = Wolkberg Group; G = Godwan Group.

inferred turbidite cycles (Hartzer, 1994; Eriksson and Reczko, 1995), suggesting water depths of a few hundreds of metres (Eriksson et al., 1994). The basinal succession is attributed to closed basin settings rather than to a fully marine basin, based on the rapid fluctuations in both inferred sedimentation energy levels and clastic/carbonate depositional systems (Eriksson and Reczko, 1995). The shallow water facies of the basinal succession in the eastern part of the Transvaal basin (e.g. Wolkberg and Godwan areas) directly overlie the volcanic rocks, and may be partly age-equivalent with the western fluvial facies identified in the Tshwene-Tshwene belt (Fig. 8).

The deposition of the Protobasinal rocks was terminated by a stage of compression and tectonic uplift (Myers, 1990; Hilliard, 1994; Els et al., 1995; Eriksson and Reczko, 1995) that generated the erosional contact (subaerial unconformity) with the overlying formation (truncation line in Fig. 8).

The base of the Protobasinal rocks may be approximated as 2714 Ma old, which is the age of the basal Ventersdorp lavas (Armstrong et al., 1991). The top contact of the Protobasinal rocks is placed between 2658 ± 1 Ma (age determined on the upper lavas of the Buffelsfontein Group and quoted in a report of the South African Committee of Stratigraphy; Eriksson and Reczko, 1995) and 2642 ± 2.3 Ma (Vryburg Formation, a likely equivalent of the overlying Black Reef Formation in the Griqualand West basin; Walraven and Martini, 1995; Fig. 5). However, until an age is obtained on the Black Reef Formation itself, the often assumed correlation with the Vryburg Formation may be open to revision (e.g. Altermann, 1997).

3.2. Black Reef Formation

Following the pre-Black Reef compressive deformation and uplift, particularly strong along the southern margin of the Transvaal basin (Myers, 1990; Eriksson and Reczko, 1995), the deposition of the Black Reef sediments took place within the tectonic setting of an intracratonic basin characterized by thermal subsidence (stage of base-level rise, Fig. 5). The isopach map of the Black Reef Formation (Eriksson and Reczko, 1995) indicates a steady increase in thickness from the margins of the Transvaal basin towards the centre, where thicknesses in excess of 200 m are recorded in the Crocodile River and Dennilton areas. This supports the hypothesis of the Transvaal basin being a unitary subsiding intracratonic sag basin during Black Reef time, subject to differential thermal subsidence with higher rates towards the centre of the basin.

The lithology of the Black Reef Formation is dominated by clastic rocks, ranging from conglomerates to sandstones and mudstones. The formation may be subdivided into a lower fining-upward succession (basal conglomerates grading into mature quartz arenites and mudstones) followed by an upper coarsening-upward sandy succession (Button, 1973; Key, 1983; Henry et al., 1990; Eriksson and Reczko, 1995; Fig. 9). The lower succession displays irregular thicknesses and is interpreted as an alluvial fan-braided fluvial facies that fills up and levels the erosional topography inherited from the top of the underlying Protobasinal rocks (Button, 1973; Henry et al., 1990; Els et al., 1995; Eriksson and Reczko, 1995). Following the alluvial-fluvial peneplanation of the initial depositional surface, the upper succession is characterized by a sheet-like geometry, alternatively interpreted to be the product of braided fluvial, transgressive shallow marine or braid-delta (*sensu* Nemec and Steel, 1988) sedimentation (Henry et al., 1990; Hartzer, 1994), or a combination of these environments (Eriksson et al., 1993a; Eriksson and Reczko, 1995). Palaeocurrent directions from the fluvial systems indicate a northward

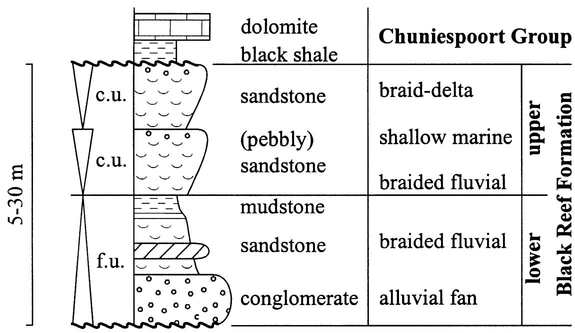


Fig. 9. Generalized vertical profiles and environments of the Black Reef Formation [modified from Henry et al. (1990)]. Abbreviations: f.u. = fining-upward; c.u. = coarsening-upward.

tilting of the syn-depositional surface (Clendenin et al., 1991; Els et al., 1995; Eriksson and Reczko, 1995), which is in agreement with the compressive deformation and uplift that affected the southern margin of the Transvaal basin prior to the Black Reef sedimentation (Myers, 1990; Hilliard, 1994). The top of the Black Reef Formation is marked by an unconformity that separates the uppermost Black Reef sandstones from the overlying transgressive black shales of the Chuniespoort marine succession (Clendenin, 1989; Figs. 5 and 9). The architecture of the Black Reef facies preserved in the eastern Transvaal is shown in Fig. 10.

3.3. Chuniespoort Group

The Chuniespoort Group accumulated within the tectonic setting of an intracratonic sag basin dominated by thermal subsidence (Altermann and Nelson, 1998; Fig. 5), which continues the structural style of the preceding Black Reef depository. However, the Black Reef and Chuniespoort basins are separated by a stage of tectonic tilting which changes the tilt of the depositional surface from northwards (Black Reef time: Myers, 1990; Clendenin et al., 1991; Els et al., 1995; Eriksson and Reczko, 1995) to southwestwards (Chuniespoort time: Clendenin, 1989; Altermann and Herbig, 1991; Hälbich et al., 1993). Most of the Chuniespoort Group strata (the lower six out of a total number of seven formations) are the product of continuous aggradation within a marine environment, which indicates a stage of base-level rise probably related to the intracratonic thermal subsidence (Fig. 5). Within the overall setting of a rapidly regressed shelf, the youngest formation includes palaeosol levels and subaerial unconformities, indicating negative accommodation and base-level fall (Fig. 5).

Overlying a thin basal layer of transgressive black shales, the lithostratigraphy of the Chuniespoort Group comprises seven formations with an overall sheet-like geometry (Figs. 5 and 11–13). The Malmani Subgroup includes the older five formations, i.e. Oaktree, Monte Cristo, Lyttelton, Eccles and Frisco (Figs. 5 and 12), which accumulated within the depositional setting of a carbonate platform, now preserved as dolomites. The five formations of dolomites are differentiated on the basis of chert content and types of stromatolite, as well as by interbedded subordinate carbonaceous mudstones and rare

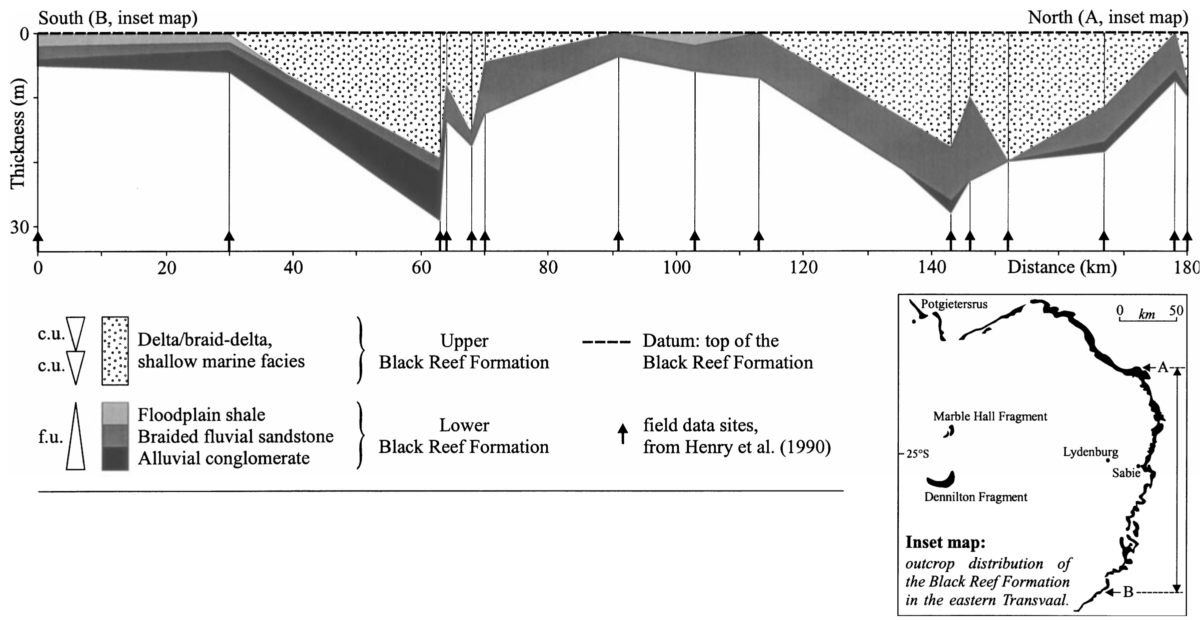


Fig. 10. Fence diagram illustrating the geometry and facies of the Black Reef Formation in the eastern part of the Transvaal basin.

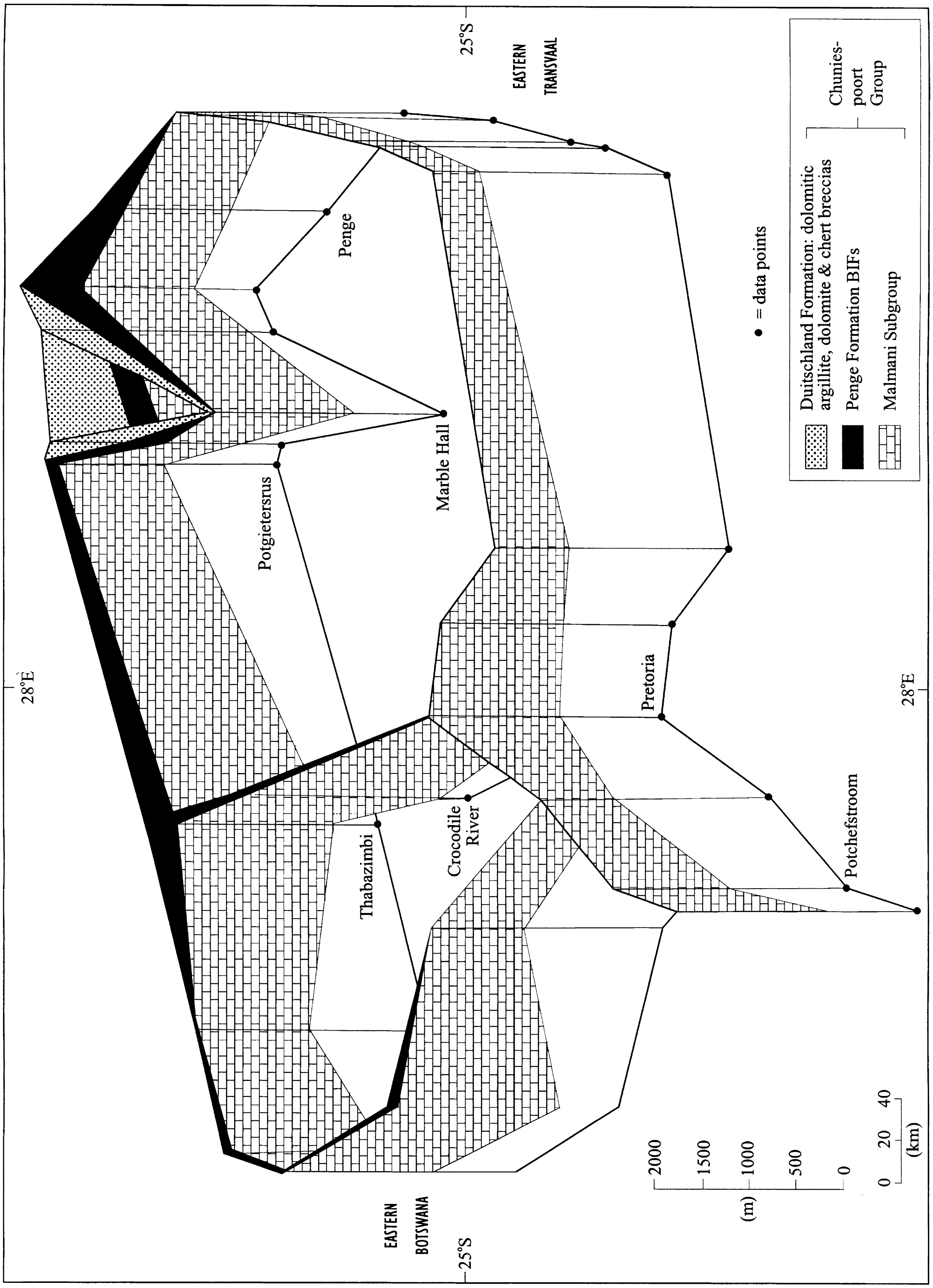


Fig. 11. Fence diagram illustrating the sheet-like geometry of the Chuniespoort facies within the Transvaal basin [from Eriksson and Reczko (1995)].

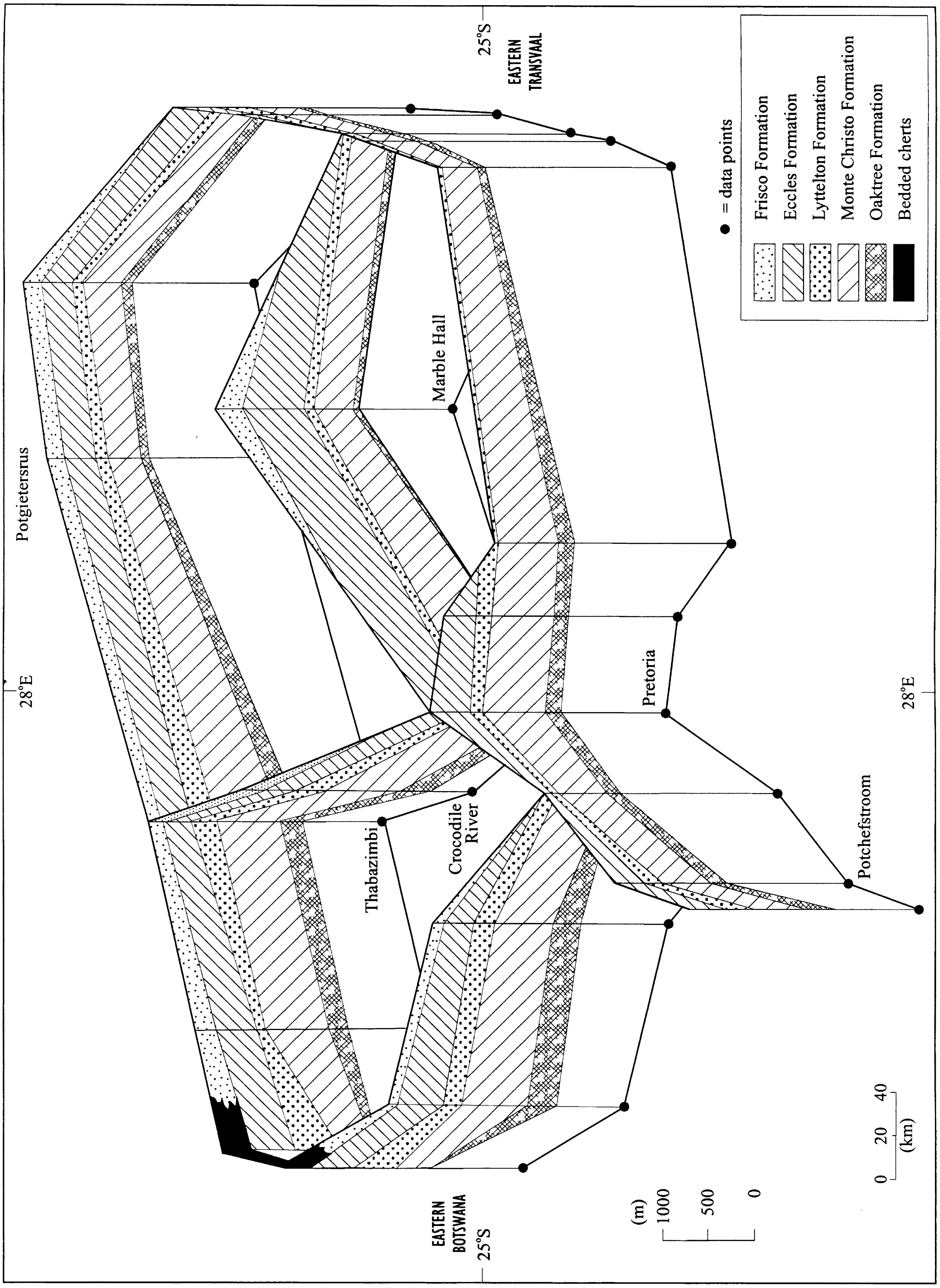


Fig. 12. Fence diagram illustrating the lithostratigraphy of the Malmani Subgroup [from Eriksson and Reczko (1995)].

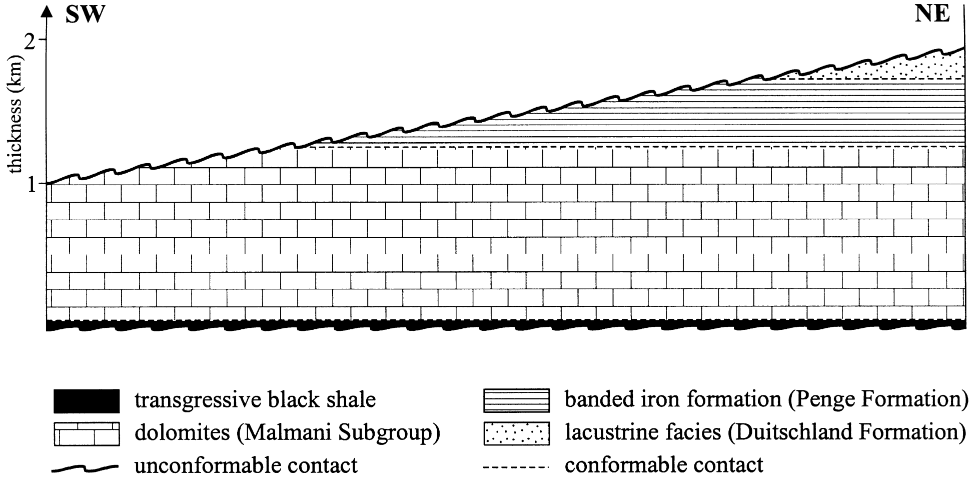


Fig. 13. Generalized stratigraphic cross-section through the Chuniespoort Group, showing the main lithofacies and the effect of the post-depositional erosional truncation. The approximate length of the section: 350 km.

quartzites (Button, 1973; Eriksson and Altermann, 1998). The interbedded clastic lithologies are interpreted to reflect proximity to the palaeo-shoreline and basin-margin unconformities (Clendenin, 1989; Fig. 5). The Malmani dolomites are gradationally replaced by the overlying iron-rich facies of the Penge Formation (Fig. 11) and its correlative in Botswana, the Ramotswa Formation (Key, 1983; Eriksson et al., 1995). The Penge Formation consists of micro- to macro-banded iron formations (BIFs) with shard structures and subordinate interbeds of carbonaceous mudstones and intraclastic iron formation breccias (Beukes, 1978; Hälbich et al., 1992). Although the origin of BIFs is still vigorously debated, some of the most recent interpretations of the Penge Formation infer a shallow marine to shelf (regressive) marine palaeoenvironment, with the possibility of a fluvial input of Fe from weathered Ventersdorp lavas (Beukes, 1980; Hälbich et al., 1992, 1993; Klemm, 1991; Altermann, 1998). Alternatively, deep marine conditions and fumarolic sources of both iron and silica may also apply to certain of the Transvaal BIF (e.g. Beukes and Klein, 1990). The Ramotswa Formation includes ferruginous mudstones interpreted as regressive, back-reef mudflat deposits (Key, 1983). The uppermost Duitschland Formation of the Chuniespoort Group (Fig. 11) represents the product of deposition during the final southwestwards withdrawal of the sea from the Kaapvaal craton (Clendenin, 1989). The Duitschland Formation comprises predominant dolomitic mudstones with interbedded dolomites and quartzites, a thick erosionally based chert breccia body, three palaeosol levels and two thin lava beds. The contact with the underlying Penge lithologies is gradational, suggesting an initial continuation of the marine sedimentation. As the sea regressed, the depositional environment changed upwards into a continental setting, as supported by the palaeosol and lava horizons (Clendenin, 1989). A lacustrine facies is inferred for most of the preserved Duitschland Formation, including the 20 to 300 m thick chert breccia body bounded at the base by a subaerial unconformity that incises locally into the Penge lithologies (Eriksson and Reczko, 1995).

The two lava beds of the Duitschland Formation mark the end of an extended period of time of tectonic stability in which intracratonic sag basins dominated by thermal subsidence-driven base-level rise constituted the tectonic setting for the accumulation of the Black Reef Formation and most of the Chuniespoort Group (Fig. 5). The base-level fall that followed the deposition of the Penge Formation may be primarily related to tectonic uplift, possibly with some additional contribution from a glacio-eustatic fall, considering the glacial influences recognized within the overlying Pretoria Group (Fig. 5). However, a glacio-eustatic fall alone cannot explain the pattern of post-depositional erosion of the Chuniespoort lithologies, stronger to the south, as the syn-depositional dip was from northeast to southwest.

The Chuniespoort Group is followed by a significant hiatus (~80 Ma, Fig. 5) in the stratigraphic record corresponding to a combination of base-level fall, tilting and differential uplift and erosion, stronger to the south, leading to the truncation illustrated in Fig. 13. The age of the Chuniespoort succession is relatively well constrained, between 2550 ± 3 Ma for the basal dolomites, and 2432 ± 31 Ma for the upper Penge Formation (Walraven and Martini, 1995). No information is available on the age of the top of the Duitschland Formation, nor on the age of the base of the overlying Pretoria Group, and implicitly also not on the precise position of the 80 Ma stratigraphic hiatus (Fig. 5).

3.4. *Pretoria Group*

The tectonic setting of the Transvaal basin during the accumulation of the Pretoria Group underwent cyclic changes involving active extension (syn-rift tectonics), post-rift stability and pre-rift uplift (Fig. 5). This indicates an overall less stable tectonic environment relative to the previous Chuniespoort and Black Reef times, marking the transition towards the significant tectonic event leading to the emplacement of the Bushveld igneous complex (Fig. 5). The succession of sedimentary facies of Pretoria age suggests that the syn-rift, post-rift and pre-rift stages alternated in a specific order, with cyclic repetition in time. We recognize two such cycles during the deposition of the Pretoria Group (Fig. 5). The syn-rift tectonic regime encompassed strong extensional subsidence, and generally allowed the accumulation of coarse-grained alluvial fan facies with a wedge-like geometry. The subsequent post-rift stage was associated with a more stable tectonic environment, lower-rates of thermal subsidence, and finer-grained sedimentation within an intracratonic-type basin, generating sheet-like geometries. The syn-rift and post-rift stages together generate subsidence-driven base-level rise, as well as accommodation space for sediment accumulation. The differences between the rates of base-level rise for the syn-rift and post-rift stages are suggested in Fig. 5. Separating periods of base-level rise, stages of base-level fall were most likely related to pre-rift tectonic uplift (crustal doming). During such doming stages, subaerial erosion occurred along with the destruction of accommodation space, thereby generating sharp facies contacts at the top of the base-level rise-related sedimentary successions, as for example the contact between the Timeball Hill and Boshoek formations (Fig. 5). Within this tectonic context, it is conceivable that the unconformity underlying the Pretoria Group (80 Ma time gap; Fig. 5) was also related to a stage of pre-rift doming, as it preceded inferred syn-rift tectonism and was associated with crustal tilting.

The Pretoria Group includes 14 formations dominated by clastic and volcanic lithologies (Figs. 5, 14 and 15). The older nine formations, known as the lower Pretoria Group, are well preserved throughout the Transvaal basin. The younger five formations are only preserved in the southeast of the Transvaal basin, being affected by post-Pretoria time erosion, as well as by the subsequent intrusion of the Bushveld complex (Eriksson and Reczko, 1995). The lithologies that define the Pretoria formations, together with the palaeoenvironmental interpretations, are illustrated in Fig. 14. Only limited age data are available for the Pretoria Group, the Hekpoort lavas being 2224 ± 21 Ma (Walraven and Martini, 1995; Fig. 5).

4. Sequence stratigraphy of the Transvaal Supergroup

4.1. *Hierarchy of the Transvaal sequences*

Assuming that the lowermost Protobasinal Transvaal rocks are approximately coeval with the Ventersdorp Supergroup volcanism, the 2714 Ma boundary between the Witwatersrand and Transvaal supergroups (Fig. 5) marks a significant change in the structural style of the receiving

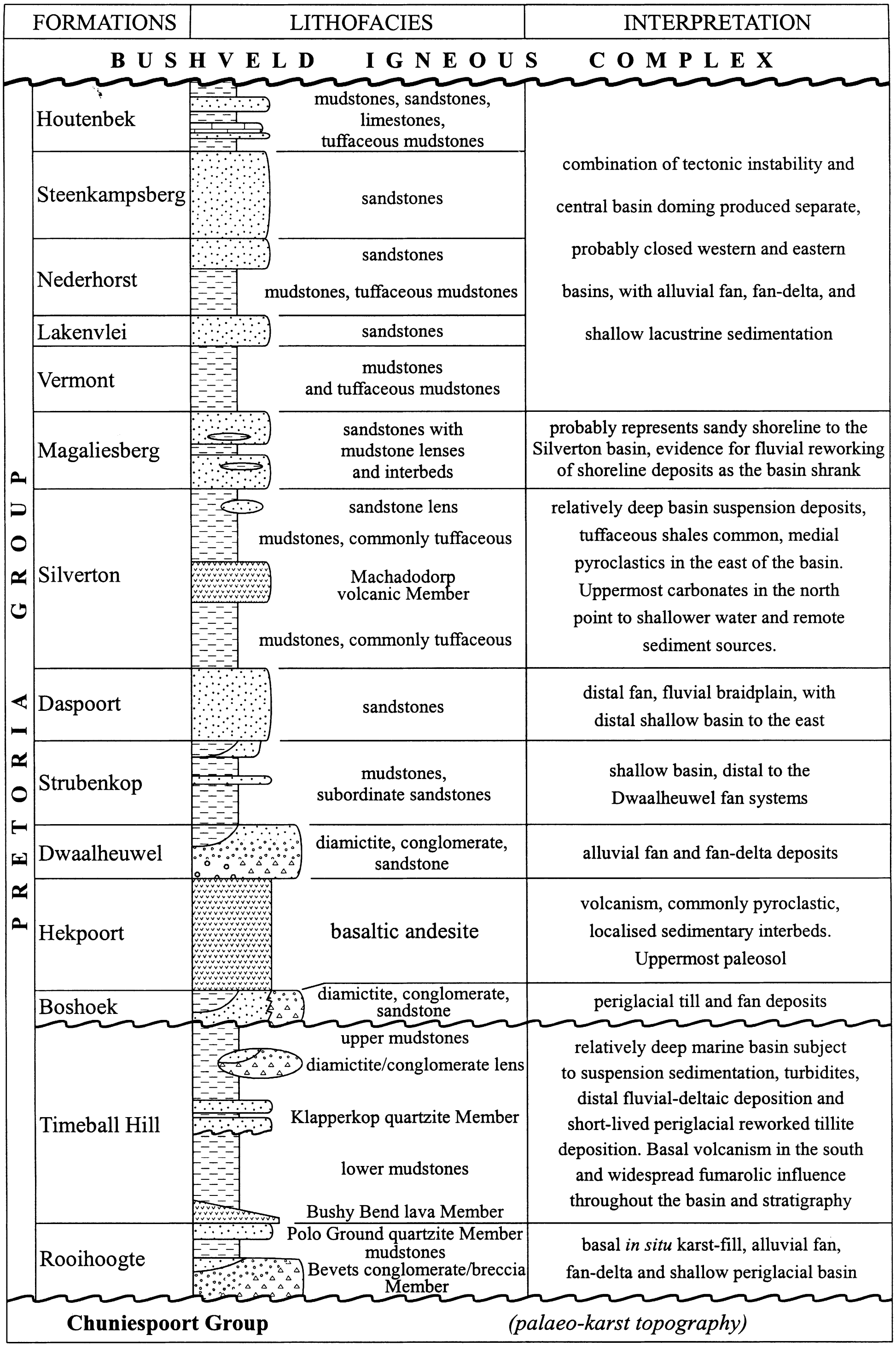


Fig. 14. Lithostratigraphy of the Pretoria Group [modified from Eriksson and Reczko (1995)]. Wavy lines indicate unconformable contacts. In relationship to the Pretoria tectonic settings (Fig. 5), the main unconformities underlie the syn-rift stage-related coarse alluvial facies. For more detailed information regarding thicknesses and facies distribution within the preserved Transvaal basin see synthesis by Eriksson and Reczko (1995).

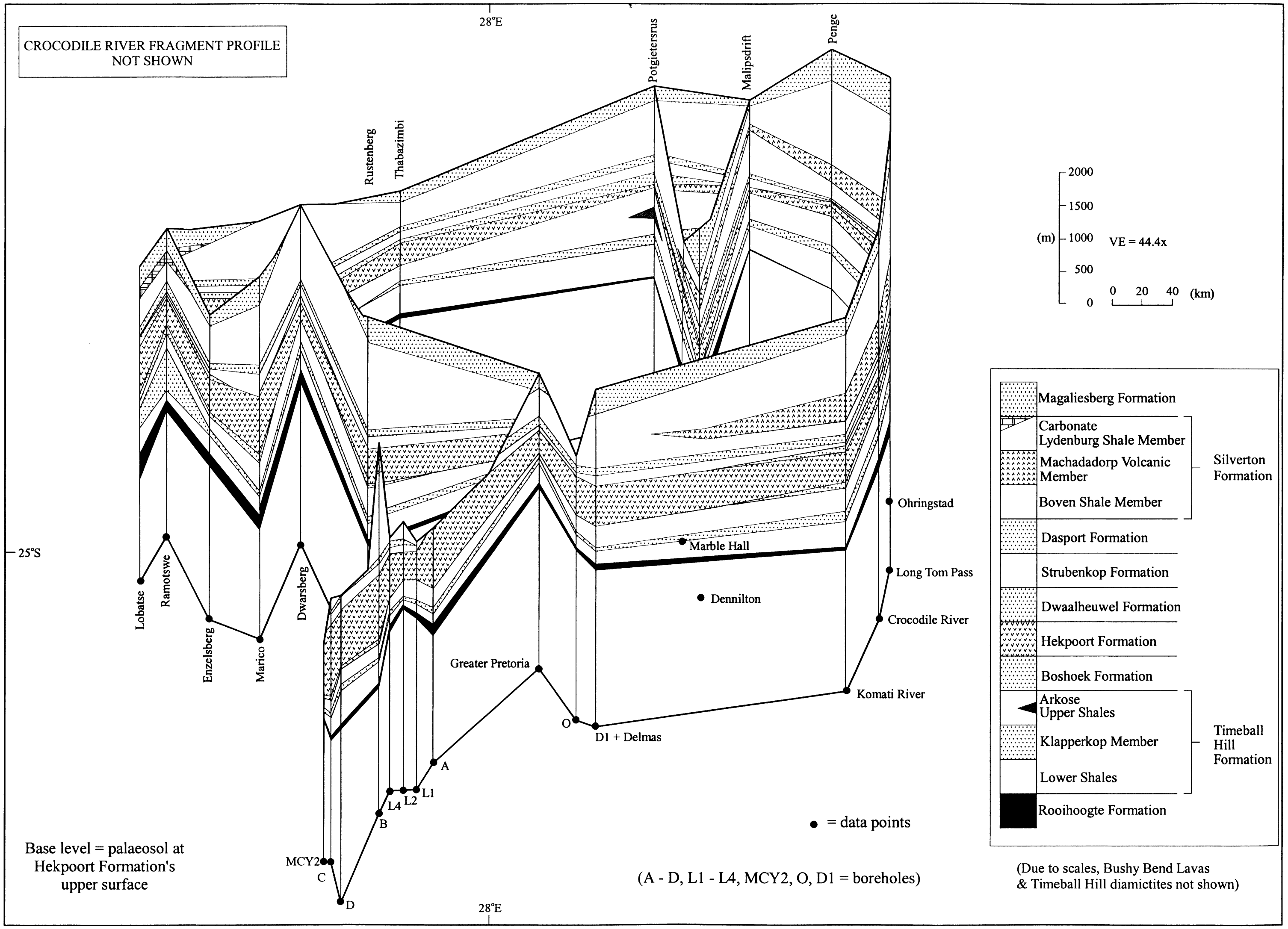


Fig. 15. Fence diagram illustrating the geometry and lithostratigraphy of the Lower Pretoria Group [from Eriksson and Reczko (1995)].

sedimentary basins. The Witwatersrand succession accumulated within the tectonic setting of a foreland basin developed in relationship to the supracrustal loading associated with the Limpopo orogeny (Stanistreet and McCarthy, 1991). In contrast, sedimentation within the Transvaal basin was controlled by cycles of extensional and/or thermal subsidence separated by stages of uplift or glacio-eustatic base-level fall. According to the sequence hierarchy system based on the magnitude of base-level changes that resulted in the boundary generation (Embry, 1995), the 2714 Ma boundary qualifies as a first-order sequence boundary (subaerial unconformity), as it marks a major shift in the tectonic setting. The upper boundary of the Transvaal Supergroup, i.e. the contact with the Bushveld complex, falls in the same category of magnitude of change, so we also take it as a first-order sequence boundary. Bounded by these two 2050 and 2714 Ma contacts (Fig. 5), the Transvaal Supergroup may be interpreted as a first-order sequence related to the sediment accumulation within the tectonic setting of the Transvaal and correlative basins.

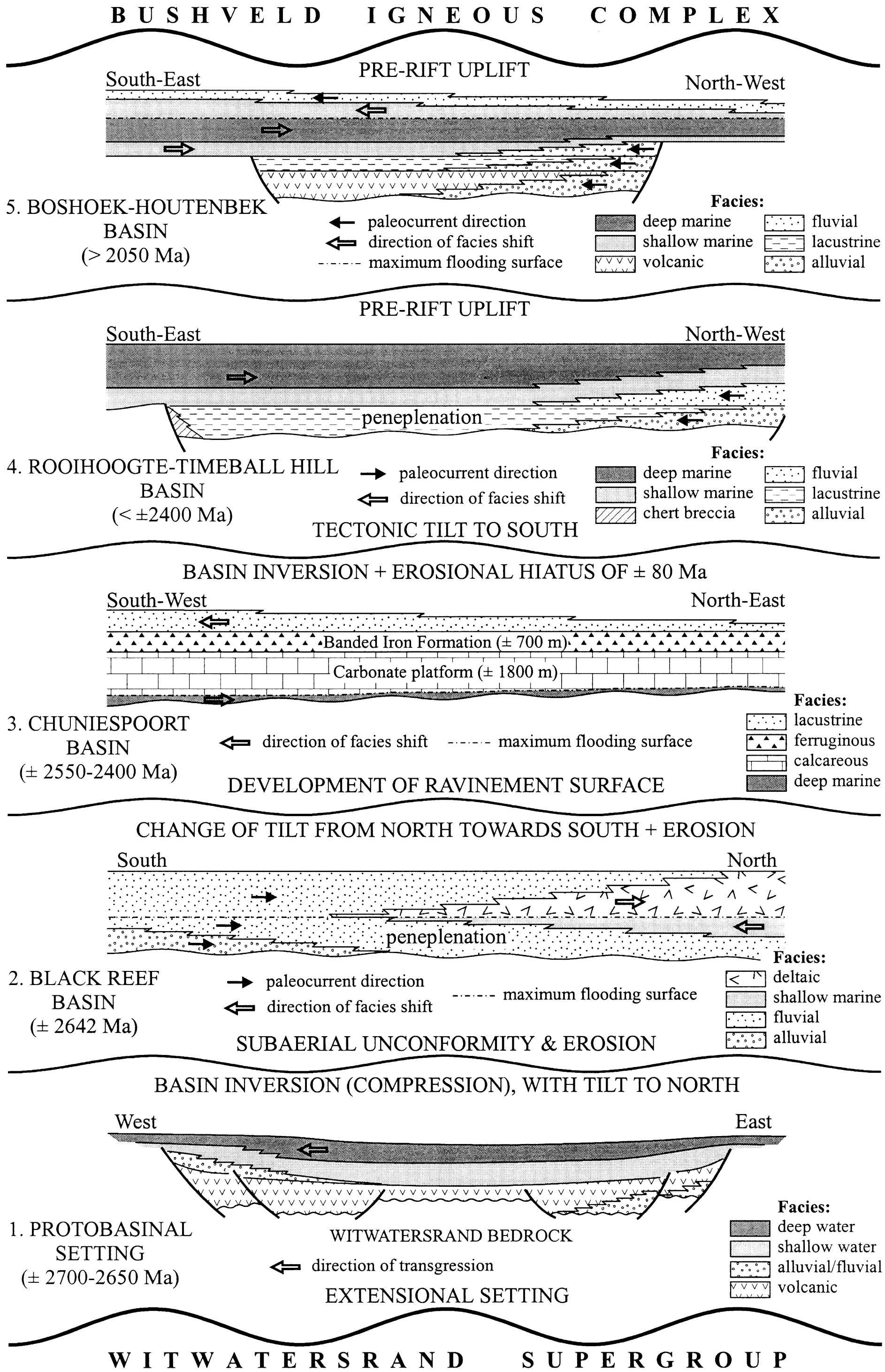
The inferred base-level curve for the Transvaal basin (Fig. 5) suggests the subdivision of the Transvaal first-order sequence into five second-order sequences, i.e. the Protobasinal, Black Reef, Chuniespoort, Rooihoogte–Timeball Hill and Boshoek–Houtenbek sequences. These sequences correspond to distinct cycles of extensional and/or thermal subsidence, and are separated by second-order boundaries related to stages of base-level fall. Important to note is the cyclic repetition of tectonic settings within the succession of second-order sequences, indicating no major shifts in structural styles during the evolution of the Transvaal basin. Our sequence stratigraphic interpretation, discussed below, thus deals with the five second-order depositional sequences and their subdivision into second-order systems tracts.

Fig. 16 provides a series of structural cartoons illustrating the relationship of the Transvaal sedimentary facies to the basins in which they occur. The five major stages of sediment accumulation correspond to periods of second-order base-level rise, and are separated by significant falls in the base-level associated with the development of subaerial unconformities (second-order sequence boundaries represented as wavy lines within the Transvaal succession; Fig. 16). Each second-order sequence is thus related to a particular setting marking a distinct stage in the evolution of the Transvaal basin.

4.2. Protobasinal sequence

The accumulation of the Protobasinal sequence is related to a stage of subsidence-driven base-level rise within Ventersdorp-age extensional basins (Fig. 5). Field data suggest that sedimentation of nonmarine facies (i.e. alluvial fan and high-energy fluvial systems) in the east, coeval with volcanism in the west, marked the debut of Transvaal deposition within the newly created extensional basins (Figs. 8 and 17). The volcanic facies gradually became dominant and replaced the nonmarine sedimentary facies to the east [diachronous facies boundary in Fig. 17(A), younger to the east], the two lithofacies together filling up the pre-existing topography associated with the underlying first-order subaerial unconformity. We interpret this assembly of sedimentary and volcanic facies as the LST of the Protobasinal sequence.

Succeeding the accumulation of the volcanic lithofacies, the younger fluvial and basinal deposits display a typical retrogradational shift of facies, upwards deepening, which allows us to interpret them as a second-order TST. The transgression of basinal facies took place from east to west, making the lower basinal deposits in the east age-equivalent with the western fluvial facies that overlie the volcanic rocks [Fig. 17(A)]. Within the TST, the limits between the fluvial and shallow water facies, as well as between the shallow and deeper water facies, are diachronous, and are younger to the west, reflecting the rate of shoreline transgression. The former limit of the fluvial facies is a ravinement surface, whereas the latter limit between shallow and deeper water facies is a conformable shift of facies. The erosion associated with the ravinement surface removed the entire transgressive fluvial facies in the east, which is why the basinal deposits directly overlie the volcanic facies there. In the west, the thicker transgressive fluvial pile allowed


BUSHVELD IGNEOUS COMPLEX
PRE-RIFT UPLIFT
South-East
North-West
5. BOSHOEK-HOUTENBEK BASIN (> 2050 Ma)
paleocurrent direction
direction of facies shift
maximum flooding surface
Facies:
deep marine
shallow marine
volcanic
fluvial
lacustrine
alluvial
PRE-RIFT UPLIFT
South-East
North-West
peneplenation
4. ROOIHOOGTE-TIMEBALL HILL BASIN (< ±2400 Ma)
paleocurrent direction
direction of facies shift
TECTONIC TILT TO SOUTH
Facies:
deep marine
shallow marine
chert breccia
fluvial
lacustrine
alluvial
BASIN INVERSION + EROSIONAL HIATUS OF ± 80 Ma
South-West
North-East
Banded Iron Formation (± 700 m)
Carbonate platform (± 1800 m)
3. CHUNIESPOORT BASIN (± 2550-2400 Ma)
direction of facies shift
maximum flooding surface
DEVELOPMENT OF RAVINEMENT SURFACE
Facies:
lacustrine
ferruginous
calcareous
deep marine
CHANGE OF TILT FROM NORTH TOWARDS SOUTH + EROSION
South
North
peneplenation
2. BLACK REEF BASIN (± 2642 Ma)
paleocurrent direction
direction of facies shift
maximum flooding surface
SUBAERIAL UNCONFORMITY & EROSION
Facies:
deltaic
shallow marine
fluvial
alluvial
BASIN INVERSION (COMPRESSION), WITH TILT TO NORTH
West
East
1. PROTOBASINAL SETTING (± 2700-2650 Ma)
WITWATERSRAND BEDROCK
direction of transgression
EXTENSIONAL SETTING
Facies:
deep water
shallow water
alluvial/fluvial
volcanic
WITWATERSRAND SUPERGROUP

the preservation of fluvial facies in the Tshwene-Tshwene basin (Figs. 8 and 17).

The upper boundary of the Protobasinal depositional sequence is a subaerial unconformity that was probably associated with strong erosional processes, as suggested by the truncation of Protobasinal facies illustrated in Fig. 8. This interpretation is also supported by the absence of HST and FSST systems tracts from the Protobasinal sequence (Fig. 17). The preserved Protobasinal LST and TST are separated by a second-order conformable transgressive surface that we place at the contact between the volcanic rocks and the overlying fluvial facies. Prior to the transgression of the basinal facies, the conformable transgressive surface developed across the entire Transvaal basin [Fig. 17(A)], but it is currently preserved only in the west due to the subsequent erosion associated with the ravinement surface [Fig. 17(B)].

4.3. Black Reef sequence

The key element in the interpretation of the Black Reef sequence is the degree of constraint on the palaeoenvironmental reconstructions. Although the lower Black Reef Formation is well understood as the product of alluvial fan and braided fluvial sedimentation, the upper part of the formation is either interpreted using a fluvial–transgressive shallow marine model (Hartzer, 1994), or a fluvial–deltaic model (Henry et al., 1990). In fact, the two models do not exclude each other and could be used in conjunction to explain the observed succession of facies (Fig. 9). The fluvial–deltaic model, which seems to explain at least the uppermost part of the formation (Eriksson and Reczko, 1995), implies the existence of a regressive shoreline within the Transvaal basin. In turn, this requires a preceding transgression of that shoreline, also during the upper Black Reef time, as the lower Black Reef environments were dominated by the northward progradation of alluvial and fluvial facies into the basin (Fig. 18). We therefore use the following evolutionary scenario for the interpretation of the Black Reef facies.

1. The lower Black Reef Formation developed from the northward progradation of alluvial fan and braided fluvial facies. The coarse alluvial sediments (proximal facies) are replaced by finer-grained fluvial sediments both laterally (distal, to the north) and upwards as well, in response to the gradual peneplanation of the topographic profile and denudation of the source areas. This explains the fining-upward profile that characterizes the lower Black Reef Formation (Fig. 9).
2. The upper Black Reef Formation is probably the result of a T–R, with an initial transgression of beach-shoreline facies over fluvial facies (coarsening-upward succession) followed by fluvial progradation and the development of regressive deltaic facies (also displaying a coarsening-upward profile). We therefore assume a change in time of the controls on the coarsening-upward profiles observed within the upper Black Reef Formation (Figs. 9 and 18).

The Black Reef depositional sequence is bounded by two second-order subaerial unconformities (Figs. 9 and 18). The upper unconformable boundary suggests the complete retreat of the upper Black Reef marine environment, followed by subaerial exposure and erosional degradation of the surface profile across the entire Transvaal basin. The base-level fall associated with these erosional processes may be eustatic in origin because (i) no compressional activity and tectonic uplift are known to have manifested during that time, and (ii) there is no change in the tectonic

Fig. 16. Relationship of the Transvaal sedimentary facies to the basins in which they occur. Each structural cartoon corresponds to a second-order sequence, reflecting the structural style for that particular stage of basin development. Stage 1: the Protobasinal sequence accumulated within an extensional setting dominated by pull-apart and rift basins. Stages 2 and 3: the Black Reef and Chuniespoort sequences accumulated within intracratonic sag basins. Stages 4 and 5: the two Pretoria sequences display similar features, starting with syn-rift deposits overlain by post-rift strata; both sequences are bounded by subaerial unconformities related to stages of pre-rift uplift. The two first-order sequence boundaries are represented with thicker and higher magnitude wavy lines. The four second-order sequence boundaries are suggested by relatively thinner and lower magnitude wavy lines. Not to scale.

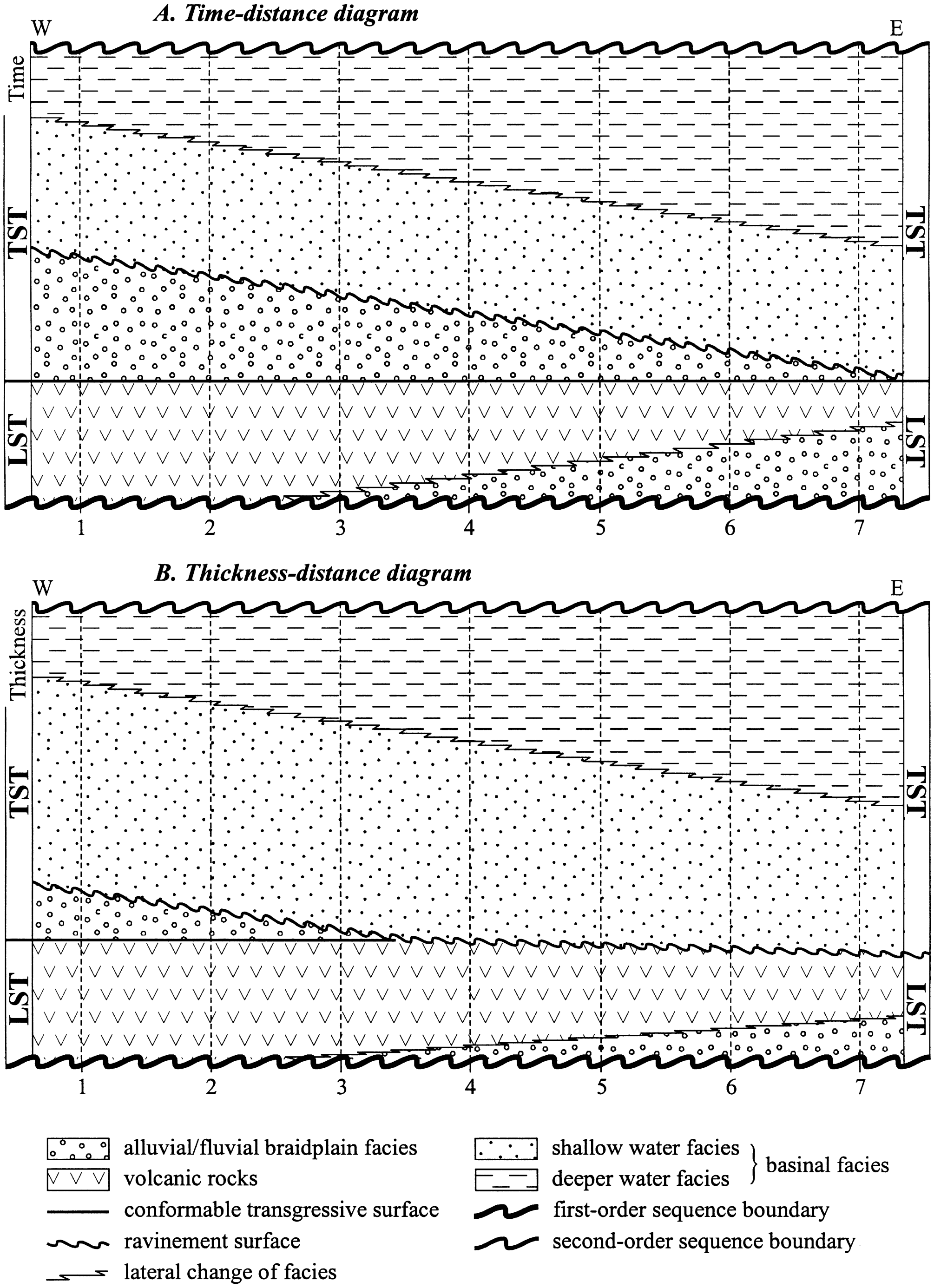


Fig. 17. Sequence stratigraphic interpretation of the Protobasinal succession, considering the individual pull-apart and rift basins as being interconnected during the syn-depositional stage. Not to scale. Abbreviations: 1 = Mogobane basin; 2 = Tshwene-Tshwene basin; 3 = Buffelsfontein basin; 4 = Wachteenbeetje basin; 5 = Bloempoort basin; 6 = Wolkberg basin; 7 = Godwan basin; LST = second-order lowstand systems tract; TST = second-order transgressive systems tract.

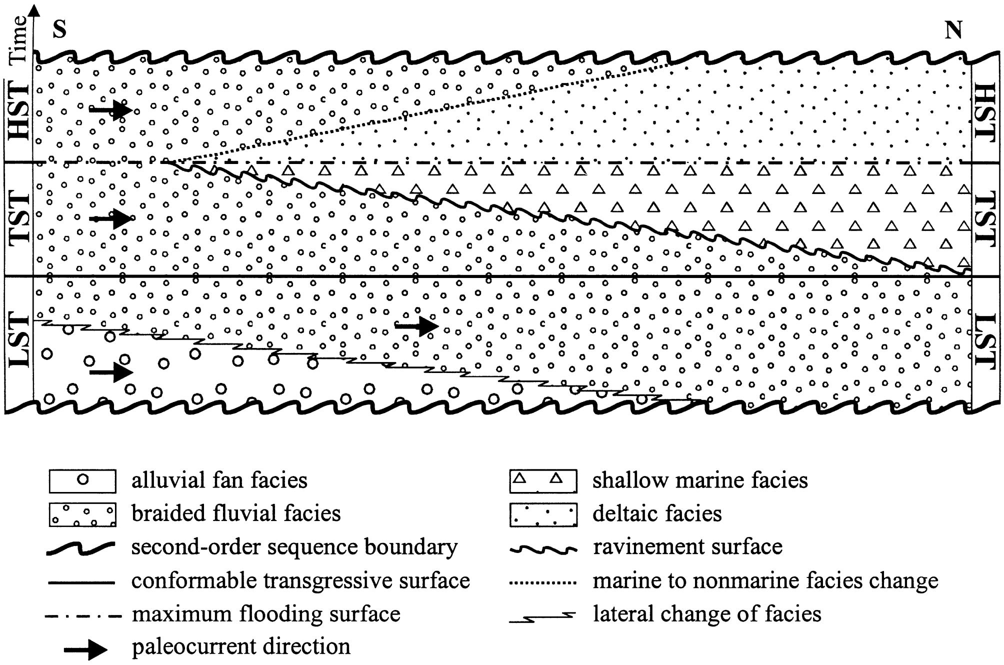


Fig. 18. Sequence stratigraphic interpretation of the Black Reef Formation. Not to scale. The LST corresponds to the lower Black Reef Formation, and the TST and HST to the upper Black Reef Formation. Abbreviations: LST = second-order lowstand systems tract; TST = second-order transgressive systems tract; HST = second-order highstand systems tract.

setting across the boundary (Fig. 5). Nevertheless, the possibility of a tectonic control on this stage of base-level fall should still be kept in mind, as there is a distinct change in the tilt of the depositional surface between the Black Reef (tilting from south to north: Myers, 1990; Clendenin et al., 1991; Els et al., 1995; Eriksson and Reczko, 1995) and Chuniespoort (tilting from NE to SW: Clendenin, 1989; Altermann and Herbig, 1991; Hälbich et al., 1993) times.

The lower succession of the Black Reef Formation, with its alluvial-fluvial prograding facies and irregular thicknesses, forms a typical 'incised valley fill', which we interpret as the LST of the Black Reef sequence (Fig. 18). This facies is similar to the alluvial fan–braidplain succession at the base of the Protobasinal sequence, also interpreted as LST deposits.

The transgressive phase of the upper Black Reef Formation corresponds to the fluvial–transgressive shallow marine model of Hartzer (1994), and is interpreted as the TST of the Black Reef sequence (Fig. 18). Although fining-upward profiles would generally be the norm within both nonmarine and marine portions of the TST, the succession of fluvial facies overlain by coarser beach-shoreline facies within the marine to nonmarine transitional setting may generate the observed coarsening-upward profile.

The regressive phase of the upper Black Reef Formation assumes coeval deposition of fluvial and deltaic facies [the fluvial–deltaic model of Henry et al. (1990)], meaning that accommodation space was available throughout the Transvaal basin. This is a case of normal regression (developed during base-level rise), so we interpret this succession as an HST (Fig. 18). Based on the assumption that the upper unconformable boundary of the Black Reef sequence develops across the entire basin, it is apparent that no FSST deposits are currently preserved.

Two systems tract boundaries may be identified within the Black Reef sequence: a conformable transgressive surface, separating the lower from upper successions of the Black Reef Formation (LST–TST boundary), and a maximum flooding surface at the limit between TST and HST (Fig. 18).

4.4. Chuniespoort sequence

The Chuniespoort sequence starts directly with transgressive marine shales (Fig. 13), indicating that the younger strata that accumulated prior to the shoreline transgression, i.e. the LST and the nonmarine portion of the TST, were removed by the erosional processes associated with the ravinement surface (Fig. 19). In this case, the ravinement surface is superimposed on the second-order lower boundary of the Chuniespoort sequence, and the preserved TST is entirely represented by the transgressive black shale (Fig. 19). This interpretation of the TST (i.e. with the maximum flooding surface placed at the top of the shale) is also supported by the thick overlying carbonate facies, which displays a highstand-type of regressive character as explained below. The small thickness of this marine portion of the original TST, in a range of metres, suggests a distal setting relative to the maximum extent of the marine environment, which in turn indicates that the Chuniespoort marine transgression took place far beyond the limits of the preserved Transvaal basin (Fig. 19). As a product of low-rate offshore sedimentation, the black shale at the base of the Chuniespoort sequence corresponds to a condensed section in the TST model of Galloway (1989).

The change in environmental conditions, stratigraphically upwards, leading to the Malmani carbonate platform and subsequent deposition of the banded iron formation, requires continuous creation of accommodation space, little clastic input to the marine basin, tectonic stability, peneplanation in the depositional area and denudation of the extrabasinal source areas, which all point towards a late stage of base-level rise characteristic of normal regressive 'highstand' conditions (Fig. 2). The thick carbonate facies, locally in excess of 2 km, indicates base-level rise with a rate matched, if not slightly outpaced, by the rate of aggradation of the carbonate platform (i.e. a case of normal regression). If the rate of base-level rise would have exceeded the rate of vertical carbonate growth (i.e. a case of transgression), this would have led to a drowned carbonate platform with the development of a drowning unconformity onlapped by clastic facies (Schlager, 1989), which has not been observed for the Malmani platform. This places the maximum flooding surface of the Chuniespoort sequence at the limit between the transgressive black shale and the overlying dolomites (Fig. 19). The Penge BIF and its correlative ferruginous mudstones in Botswana, recognized by many workers as a product of a regressive environment (Klemm, 1979, 1991; Key, 1983; Hälbich et al., 1992, 1993; Altermann, 1998), seem to continue the Malmani normal regression as opposed to a falling stage forced regression, as suggested by the low-energy environment and the low clastic sediment supply. We therefore interpret the Malmani dolomites and the overlying Penge BIF as forming together the HST of the Chuniespoort sequence. These two lithostratigraphic units are separated by an interval of gradual transition, which further supports their grouping within one systems tract. If an alternative deeper water model is adopted for BIF sedimentation (e.g. Beukes et al., 1990), then drowning of the Malmani carbonate platform as a result of base-level rise exceeding the carbonate sedimentation rate would presumably have occurred. However, the proponents of this deeper water BIF model interpret the BIF succession as being upward-shallowing [Beukes and Klein, 1990; see also discussion by Altermann (1998)], which supports the normal regressive model suggested above.

The uppermost Duitschland Formation is associated with the final withdrawal of the Chuniespoort sea from the Kaapvaal craton (Clendenin, 1989). As most of the Duitschland facies are interpreted as continental, mainly lacustrine (Clendenin, 1989; Eriksson and Reczko, 1995), there is an indication that the shoreline regression was rapid, which is compatible with a base-level fall-driven forced regression. The difference between the rates of normal and forced regressions are suggested in Fig. 19. The fall in the base-level during the deposition of the Duitschland Formation is also supported by the presence of

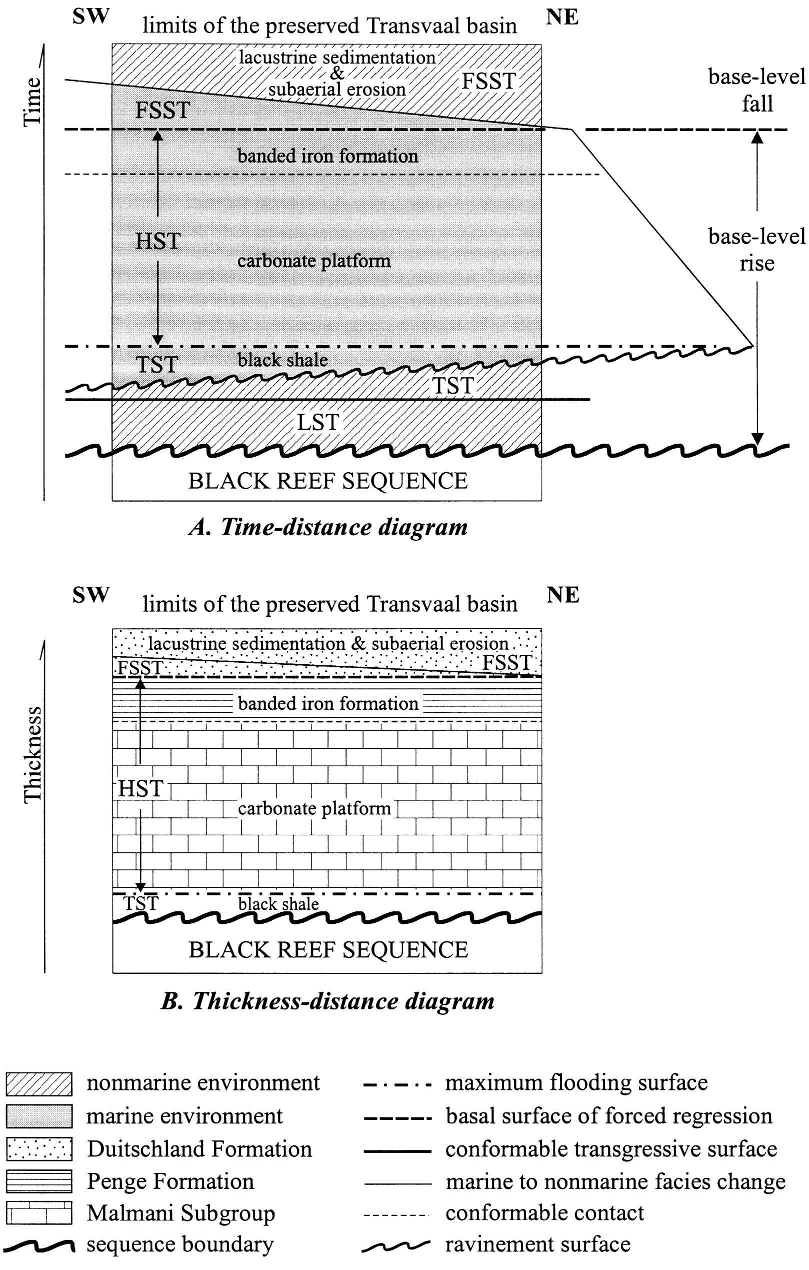


Fig. 19. Sequence stratigraphic interpretation of the Chuniespoort Group. Not to scale. The post-depositional erosional truncation of the Chuniespoort facies (illustrated in Fig. 13) is not represented here. Abbreviations: LST = second-order lowstand systems tract; TST = second-order transgressive systems tract; HST = second-order highstand systems tract; FSST = second-order falling stage systems tract.

palaeosol levels and intraformational subaerial unconformities. Within this scenario, it is conceivable that the base of the Duitschland Formation preserves marine strata, which is supported by the gradual transition with the underlying Penge Formation. As a product of deposition during base-level fall, we interpret the Duitschland Formation as an FSST. The limit between the HST and FSST is represented by the basal surface of forced regression (Fig. 19).

Within this overall second-order sequence stratigraphic framework (Fig. 19), the marginal unconformities of Clendenin (1989; Fig. 5) may provide a first indication towards the differentiation of higher-frequency (third-order) sequences. Such sequences in the Transvaal carbonate–BIF platform rocks are also discussed by Altermann and Nelson (1998) and by Altermann (1998).

4.5. Rooihoogte–Timeball Hill sequence

The Rooihoogte–Timeball Hill sequence is the product of deposition during the first Pretoria cycle of syn-rift and subsequent post-rift base-level rise, and it is bounded by base-level fall-related subaerial unconformities (Figs. 5, 14 and 20). The sequence displays an overall fining-upward profile, grading from coarse incised valley fills into shallow lacustrine and then deep marine facies (Fig. 14). The initial stage in the deposition of the Rooihoogte–Timeball Hill sequence was of peneplanation of the pre-existing karst topography via the progradation of alluvial-fan and fan–delta systems into the basin, generally from the north, coeval with periglacial lacustrine sedimentation in the western part of the basin. These deposits correspond to the overall fining-upward succession of the Rooihoogte Formation (Fig. 14), which may be interpreted as a second-order 'incised valley fill'. This initial stage of peneplanation was followed by subsequent transgressive shallow to deep marine facies, dominated by aggradation and the accumulation of low-energy fine sediments (Timeball Hill Formation). Aggradation of correlative fluvial (braid-delta) facies took place at the same time with the transgression of the marine environment, to the north of the coeval shoreline (Eriksson and Reczko, 1995), which is in agreement with an overall southerly dipping topographic slope. We interpret the second-order incised valley fill as the LST of the Rooihoogte–Timeball Hill sequence, and the overlying transgressive marine facies with their fluvial correlatives as the second-order TST. The boundary between the two systems tracts, which is age-equivalent with the onset of transgression, is taken as a conformable transgressive surface developed at the top of the Rooihoogte Formation (Figs. 14 and 20).

Detailed facies analysis may provide the basis for further subdivision of the second-order systems tracts into higher frequency sequence stratigraphic elements. The widespread marine black shales at the base of the Timeball Hill Formation (Eriksson and Reczko, 1998) are inferred to have transgressed approximately from east to west (Eriksson and Reczko, 1998), as is supported by the apparently retrograding geometries (Fig. 20). Above the transgressive black shales at the base of the Timeball Hill Formation, a shallowing upward succession of pelagic, distal delta-fed turbidites ['lower mudstones' in Fig. 14, interpreted by Eriksson and Reczko (1998) as being deposited under highstand conditions] and erosively based tidally reworked braid-delta deposits [arenaceous Klapperkop Member, Fig. 14, interpreted as lowstand facies by Eriksson and Reczko (1998)] developed, followed by a deepening upward succession of suspension deposits and delta-fed turbidite fan systems ['upper mudstones' in Fig. 14, interpreted as transgressive facies by Eriksson and Reczko (1998)]. Based on this succession of facies, the Timeball Hill second-order TST may be subdivided into two third-order depositional sequences separated by the erosional surface underlying the Klapperkop Member (third-order sequence boundary): a lower sequence, comprising a third-order TST (the basal transgressive black shales) and a third-order HST ('lower mudstones'), and an upper sequence that includes a third-order LST (Klapperkop Member) followed by a third-order TST ('upper mudstones').

The uppermost deep marine facies of the Rooihoogte–Timeball Hill sequence are sharply overlain by the high-energy alluvial fan deposits of the Boshoek Formation (Figs. 14 and 20), indicating an abrupt change in the sedimentation regime across the sequence boundary. This may be related to a stage of pre-rift uplift-driven base-level fall that was probably associated with erosional processes, as suggested by the absence of a

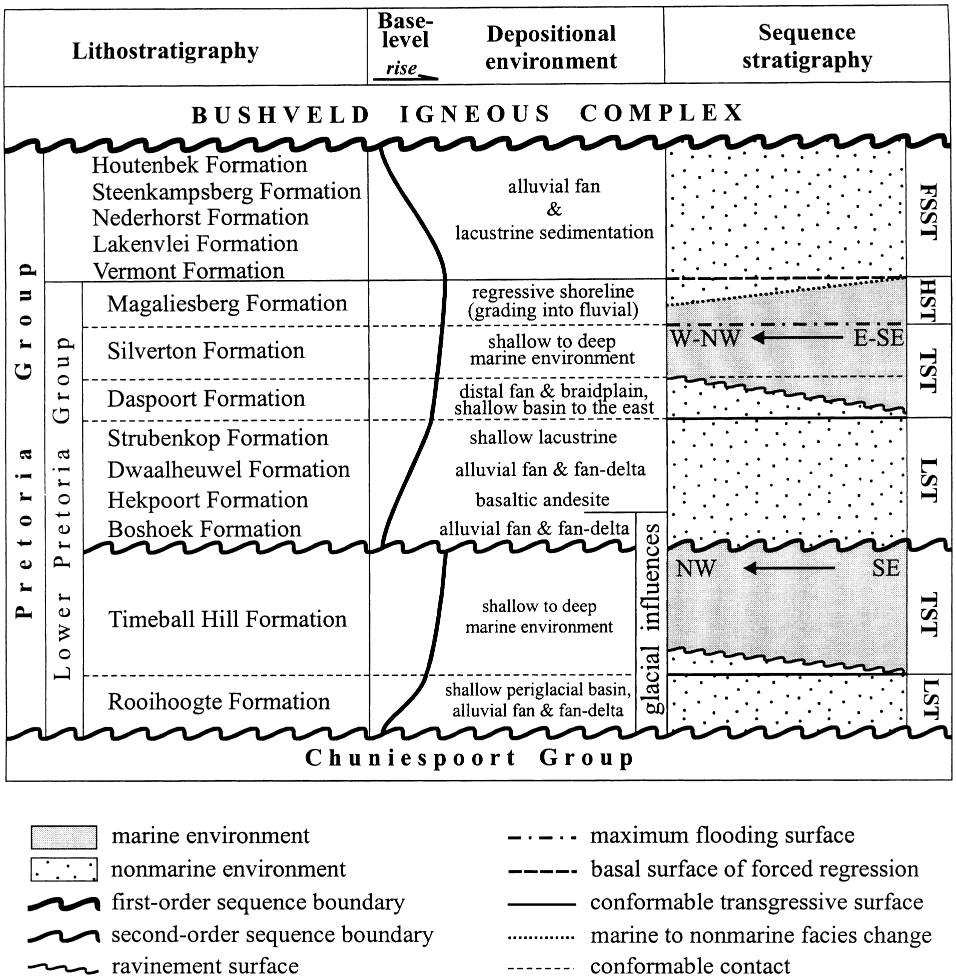


Fig. 20. Sequence stratigraphic interpretation of the Pretoria Group. Not to scale. Vertical axis suggests both time and thickness. The arrows in the 'sequence stratigraphy' column indicate the directions of shoreline transgression. Abbreviations: LST = second-order lowstand systems tract; TST = second-order transgressive systems tract; HST = second-order highstand systems tract; FSST = second-order falling stage systems tract.

second-order HST as well as by the truncation of the preserved TST.

4.6. Boshoek–Houtenbek sequence

The Boshoek–Houtenbek sequence corresponds to a second cycle of renewed syn- and post-rift base-level rise followed by uplift-related base-level fall (Figs. 5, 14 and 20).

As in the case of the previous sequence, sedimentation started with the southwards progradation of coarse alluvial fan and fan–delta facies (Boshoek and Dwaalheuwel formations), temporarily interrupted by the volcanic event leading to the emplacement of the Hekpoort basaltic andesites (Figs. 14 and 20). With the gradual lowering in depositional energy as a result of progressive topographic peneplanation and denudation of the source areas, the upper alluvial facies were replaced distally by coeval low-energy lacustrine sedimentation, reflected by the mudrocks of the Strubenkop Formation, interpreted as a shallow lacustrine facies distal to the Dwaalheuwel fans (Schreiber and Eriksson, 1992; Eriksson et al., 1993a, 1995).

The Boshoek–Dwaalheuwel/Strubenkop succession, corresponding to a stage of progradation of high-energy coarse alluvial facies into a rejuvenated depositional area, may be interpreted as the second-order LST of the Boshoek–Houtenbek sequence (Fig. 20).

The LST progradational facies is followed by retrograding geometries associated with the west–northwestwards transgression of the Silverton sea (Fig. 20). The marine transgression started during the deposition of the Daspoort Formation, the nonmarine facies of this unit being coeval with shallow marine deposits to the east (Eriksson et al., 1993b; Eriksson and Reczko, 1995; Fig. 20). The maximum extent of the Silverton sea is estimated to correspond to the top of the Silverton Formation (Eriksson and Reczko, 1995), which we therefore take as the maximum flooding surface of the Boshoek–Houtenbek sequence (Fig. 20). In this interpretation, the Daspoort and Silverton formations together build a second-order TST, with the conformable transgressive surface at the limit between the Daspoort Formation and the underlying Dwaalheuwel and correlative Strubenkop formations (Fig. 20).

The retrograding facies of the TST are overlain by the fluvial–deltaic–regressive shoreline deposits of the Magaliesberg Formation (Eriksson et al., 1995). These deposits, displaying progradational stacking patterns with coeval marine and nonmarine aggradation, accumulated during the final stages of base-level rise as normal regressive strata. They are interpreted as a second-order HST (Fig. 20).

The post-Magaliesberg formations accumulated during the stage of base-level fall (pre-rift tectonic doming/uplift) that preceded the intrusion of the Bushveld complex, which allows their interpretation as the FSST of the Boshoek–Houtenbek sequence (Fig. 20). The FSST is separated from the underlying HST by the basal surface of forced regression, which may be placed at the top of the Magaliesberg Formation. This surface is only preserved in the area of occurrence of the post-Magaliesberg formations (mainly in the southeast of the Transvaal basin, with lesser occurrences northeast of Pretoria and in eastern Botswana), as in the rest of the basin the top of the HST was affected by erosional truncation as well as by the intrusion of the Bushveld complex.

4.7. Duration of first- and second-order sequences in the Transvaal Supergroup

The whole question of the existence of Late Archaean–Early Proterozoic supercontinents is a vexed one. Reconstructions of Neoproterozoic supercontinents, namely Kanatia and Rodinia, are widely accepted and are becoming ever more sophisticated (e.g. Young, 1995; Li et al., 1996; Pelechaty, 1996; Dalziel, 1997). For older Precambrian supercontinents, the supporting data are often equivocal. Rogers (1996) proposes Ur, a supercontinent postulated to have stabilised at 3.0 Ga and centred on India, and different Late Archaean supercontinent reconstructions are proposed by a number of researchers (e.g. Button, 1976; Piper, 1983; Gaal, 1992; Stanistreet, 1993; Cheney, 1996). Most recently, Aspler and Chiarenzelli (1998) provide supporting evidence for the existence of a 'northern' (present-day frame of reference) Kenorland (Williams et al., 1991) supercontinent, encompassing the established Archaean crustal fragments of North America and the Baltic and Siberian shields. This postulated Late Archaean supercontinent is thought to have undergone a protracted attenuation from approximately 2.5 to 2.1 Ga, with dispersal occurring from about 2.1–2.0 Ga (Aspler and Chiarenzelli, 1998). Aspler and Chiarenzelli (1998) also debate a 'southern' supercontinental cycle, which may have extended from about 2.65 Ga until approximately 2.0 Ga, although evidence for this is less convincing than for Kenorland. For the period of ca 2.0–1.8 Ga, a stronger case can be made for supercontinental assembly events, for example that of Laurentia, including most of the fragments postulated for Kenorland (e.g. Hoffman, 1988), or a supercontinent encompassing Late Archaean–Early Proterozoic crustal provinces from Africa and South America (e.g. Eriksson et al., 1999).

The possibility of early Precambrian supercontinents and the merits of different postulates really rests intrinsically on a second debate, whether plate tectonics operated during the Archaean and, if not, when such processes first became important.

This debate is addressed elsewhere in this special issue (e.g. Arndt, 1999; Eriksson et al., 1999). Nelson and coworkers (Nelson, 1998; Nelson et al., 1999) discuss the possibility of episodic global magmatism as a primary influence on Early and Middle Archaean continental crustal growth, and suggest that a gradual transition, diachronous on a global scale, to a plate tectonic regime occurred near the Archaean–Proterozoic boundary. This supports the existence of Late Archaean supercontinents dominated by still-developing, and thus slower, plate tectonic processes. In such a scenario, the temporal duration of the earliest supercontinental assembly–disassembly events, or first-order cycles, would not be comparable with much shorter Phanerozoic cycles developed within a lithosphere almost completely dominated by evolved plate tectonic processes. The apparent length of the proposed first-order cycle for the Transvaal Supergroup, in excess of 650 Ma ($\pm$2714–2050 Ma), when compared to well-studied Phanerozoic examples, should be seen within the light of the above discussion. Similarly, second-order cycles of relative sea level change discussed here for the Transvaal, with an average duration of 130 Ma, were also much longer than the younger (Phanerozoic) equivalents. The difference in the temporal duration between same-order Precambrian and Phanerozoic cycles indicates the boundary frequency as a less relevant criterion in establishing a hierarchy system for stratigraphic sequences and their bounding surfaces. Instead, the hierarchy system based on the magnitude of base-level changes that resulted in boundary generation could be used for both Precambrian and Phanerozoic successions regardless of the time span of the various cycles.

5. Conclusions

The 2.7–2.1 Ga Transvaal Supergroup is recognized here as a first-order depositional sequence, as being bounded by unconformable contacts that mark major shifts in the geodynamic trends of the Kaapvaal craton. The lower first-order sequence boundary ($\sim$2714 Ma) punctuates the change from the foreland setting of the Witwatersrand basin to the alternating extensional/intracratonic settings of the Ventersdorp Supergroup and partly coeval Transvaal basin, whereas the upper first-order boundary ($\sim$2050 Ma) represents the limit with the overlying Bushveld igneous complex. Between these boundaries, sedimentation within the Transvaal basin was cyclic, with periods of time of sediment aggradation associated with stages of base-level rise (sequences), separated by stages of base-level fall mainly leading to the development of subaerial unconformities (sequence boundaries). The inferred base-level changes are related to cycles of extensional and/or thermal subsidence followed by compressional or pre-rift tectonic uplift, with secondary overprinting from glacio-eustatic changes. The creation of accommodation space during stages of base-level rise allowed the accumulation of lowstand, transgressive and highstand systems tracts. During base-level fall, nonmarine facies may still accumulate in restricted areas as a function of the relative position between the topography and the local fluvial equilibrium profiles or lacustrine base-levels, although their preservation potential is low due to subsequent subaerial erosion. When preserved, the falling stage facies are grouped within the falling stage systems tract.

Four second-order subaerial unconformities are recognized within the Transvaal succession, subdividing the Transvaal first-order sequence into five second-order unconformity-bounded depositional sequences. The Protobasinal sequence, at the base of the Transvaal succession, is currently preserved within several extensional basins, which were very likely interconnected during the syn-depositional time, as suggested by the regional correlation of facies. Only the LST and the overlying TST are preserved to date, due to post-depositional erosional truncation. The Black Reef sequence accumulated during intracratonic subsidence-related base-level rise, and preserves the LST, TST and HST. The overlying Chuniespoort sequence also accumulated during intracratonic subsidence base-level rise. It does not preserve the LST, due to ravinement erosion, but has distinct TST, HST and FSST deposits. The upper two second-order sequences of the Transvaal succession, i.e. Rooihoogte–Timeball Hill and overlying Boshoek–Houtenbek, are related to distinct syn-rift/post-rift cycles of subsidence-driven base-level rise, preceded and succeeded by pre-rift uplift base-level fall. The Rooihoogte–Timeball Hill sequence only preserves the LST and part of the overlying TST, whereas the Boshoek–Houtenbek sequence includes all the four systems tracts characteristic of a depositional sequence. The five Transvaal second-order sequences record an average duration of 130 Ma, much longer than the Phanerozoic second-order sequences. This is explained by the still-developing, and thus slower, plate tectonic processes that operated near the Archaean–Proterozoic boundary relative to the evolved processes that dominated the Phanerozoic. The difference in time scales between similar tectonic processes-driven Precambrian and Phanerozoic cycles favours the use of the hierarchy system based on the magnitude of base-level changes that resulted in boundary generation (Embry, 1995), rather than the hierarchy system based on boundary frequency (Vail et al., 1977, 1991; Mitchum and Van Wagoner, 1991; etc.).

The base-level rise inferred for the Protobasinal sequence most probably reflects the influence of tectonism in the Kaapvaal craton on relative sea level. Within the Limpopo collision-related extensional setting envisaged for the Transvaal Protobasinal rocks, it is not possible to make meaningful inferences on continental freeboard conditions. The subsequent Black Reef sequence represents the thermal subsidence following upon the extensional fault-related Protobasinal rocks, thereby suggesting once again a local tectonic explanation of the upper Black Reef relative sea level rise. Base-level rise inferred for the Chuniespoort sequence (Fig. 5) appears to be related to a much more widespread event, possibly of global proportions (see Eriksson et al., 1999), thereby probably reflecting eustatic sea level changes, most likely related to enhanced continental crustal growth rates at the Archaean–Proterozoic boundary (e.g. Eriksson, 1999). In view of the tectonic stability and lack of clastic sedimentation generally associated with the development of large carbonate–BIF platform successions such as that of the Transvaal, continental freeboard of the Kaapvaal craton was probably reduced at this time. Local tectonism as a cause of the Timeball Hill transgression is proposed, especially in view of the apparent association of these sedimentary rocks with periglacial deposits; sea level would logically have been lowered during this apparently global glacigenic event (e.g. Eriksson et al., 1999). The relative sea level rise interpreted for the Daspoort–Silverton–Magaliesberg formations within the Boshoek–Houtenbek sequence (Fig. 5) appears to form part of a wider development of epeiric marine sedimentation within Africa (Eriksson et al., 1999). The rise in sea level implicit in these widespread epeiric marine deposits on many of the cratons of Africa may reflect enhanced mid-ocean ridge activity prior to supercontinent assembly during the approximately 2.0 Ga Eburnean orogeny of Africa (Eriksson et al., 1999). The collisions during the Eburnean would have increased continental freeboard significantly after its lowering due to aggressive early Precambrian weathering (Corcoran et al., 1999) during the epeiric marine transgressions preceding the orogeny.

Acknowledgements

OC thanks Rhodes University for financial support during the completion of this study. PGE acknowledges research support from the University of Pretoria and the Foundation for Research Development, South Africa. PGE is also indebted to graduate students who have stimulated debate and research on Transvaal rocks over the years: Markus van der Neut, Pieter Bosch, Ute Schreiber, Boris Reczko and Adam Bumby. We thank Andrew Miall and John Hancox for excellent reviews and Ken Eriksson for guidance and encouragement.

References

Altermann, W., 1997. Sedimentological evaluation of Pb–Zn exploration potential of the Precambrian Griquatown Fault Zone in the Northern Cape Province, South Africa. Mineral. Deposita 32, 382–391.

Altermann, W., 1998. Die sedimentäre Entwicklung der Karbonat–Schieferton–BIF-Sequenz Des Kaapvaal Kratons, Südafrika: eine Folge von Transgressionen, erkennbar auch auf dem Pilbara Kraton, Australien. Münchener Geol. Hefte A 23, 5–25.

Altermann, W., Herbig, H.G., 1991. Tidal flat deposits of the Lower Proterozoic Campbell Group along the southwestern margin of the Kaapvaal Craton, Northern Cape Province, South Africa. J. Afr. Earth Sci. 13, 415–435.

Altermann, W., Nelson, D., 1998. Sedimentation rates, basin analysis, and regional correlations of three Neoarchean and Paleoprotoerozoic subbasins of the Kaapvaal Craton as inferred from precise U-Pb zircon ages from volcaniclastic sediments. Sediment. Geol., in press.

Armstrong, R.A., Compston, W., Retief, E.A., Williams, I.S., Welke, H.J., 1991. Zircon ion microprobe studies bearing on the age and evolution of the Witwatersrand Triad. Precambrian Res. 53, 243–266.

Arndt, N., 1999. Why was flood volcanism on submerged continental platforms so common in the Precambrian? Precambrian Res. 97, 155–164.

Aspler, L.B., Chiarenzelli, J.R., 1998. Two Neoarchean supercontinents? Evidence from the Paleoproterozoic. Sediment Geol. 120, 75–104.

Beukes, N.J., 1978. Die karbonaatgesteentes en ysterformasies van die Ghaap–Groep van die Transvaal Supergroep in Noord Kaapland. Ph.D. dissertation, Rand Afrikaans University, Johannesburg, South Africa.

Beukes, N.J., 1980. Lithofacies and stratigraphy of the Kuruman and Griquatown Iron-Formations, Northern Cape Province, South Africa. Trans. Geol. Soc. S. Afr. 83, 69–86.

Beukes, N.J., Cairncross, B., 1991. A lithostratigraphic–sedimentological reference profile for the Late Archaean Mozaan Group, Pongola Sequence: application to sequence stratigraphy and correlation with the Witwatersrand Supergroup. S. Afr. J. Geol. 94, 44–69.

Beukes, N.J., Klein, C., 1990. Geochemistry and sedimentology of facies transition from microbanded to granular iron-formation in the early Proterozoic Transvaal Supergroup, South Africa. Precambrian Res. 47, 99–139.

Beukes, N.J., Klein, C., Kaufmann, A.J., Hayes, J.M., 1990. Carbonate petrography, kerogene distribution, and carbon and oxygene isotope variations in Early Proterozoic transition from limestone to iron formation deposition, Transvaal Supergroup, South Africa. Econ. Geol. 85, 663–690.

Bosch, P.J.A., 1992. Die geologie van die Wolkberg Groep tussen die Abel Erasmuspas en Graskop, Oos-Transvaal. M.Sc. dissertation, Pretoria University, Pretoria, South Africa.

Burke, K., Kidd, W.S.F., Kusky, T.M., 1985. Is the Ventersdorp Rift System of southern Africa related to a continental collision between the Kaapvaal and Zimbabwe cratons at 2.64 Ga ago? Tectonophysics 115, 1–24.

Button, A., 1973. A regional study of the stratigraphy and development of the Transvaal Basin in the eastern and northeastern Transvaal. Ph.D. dissertation, University of the Witwatersrand, Johannesburg, South Africa.

Button, A., 1976. Transvaal and Hamersley basins – review of basin development and mineral deposits. Miner. Sci. Eng. 8, 262–292.

Catuneanu, O., 1996. Reciprocal architecture of Bearpaw and post-Bearpaw sequences, Late Cretaceous–Early Tertiary, Western Canada Basin. Ph.D. dissertation, University of Toronto, Toronto, Canada.

Catuneanu, O., Willis, A.J., Miall, A.D., 1998. Temporal significance of sequence boundaries. Sediment. Geol. 121, 157–178.

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

Clendenin, C.W., 1989. Tectonic influence on the evolution of the early Proterozoic Transvaal Sea, southern Africa. Ph.D. dissertation, University of the Witwatersrand, Johannesburg, South Africa.

Clendenin, C.W., Henry, G., Charlesworth, E.G., 1991. Characteristics of and influences on the Black Reef depositional sequence in the eastern Transvaal. S. Afr. J. Geol. 94, 321–327.

Corcoran, P.L., Mueller, W.U., Chown, E.H., 1998. Climatic and tectonic influences on Fan deltas and wave- to tide-controlled shoreface deposits: evidence from the Archean Keskarrah Formation, Slave Province, Canada. Sediment. Geol. 120, 125–152.

Cotter, E., 1978. The evolution of fluvial style, with special reference to the central Appalachian Paleozoic. In: Miall, A.D. (Ed.), Fluvial Sedimentology, Can. Soc. Pet. Geol. Mem. 5., 361–383.

Dahle, K., Flesja, K., Talbot, M.R., Dreyer, T., 1997. Correlation of fluvial deposits by the use of Sm-Nd isotope analysis and mapping of sedimentary architecture in the Escanilla Formation (Ainsa Basin, Spain) and the Statfjord Formation (Norwegian North Sea), Abstracts 6th International Conference on Fluvial Sedimentology, Cape Town, 22–26 September, 46.

Dalziel, I.W.D., 1997. Neoproterozoic-Paleozoic geography and tectonics: review, hypothesis, environmental speculation. Geol. Soc. Am. Bull. 109, 16–42.

Els, B.G., van den Berg, W.A., Mayer, J.J., 1995. The Black Reef Quartzite Formation in the western Transvaal: sedimentological and economic aspects, and significance for basin evolution. Mineral. Deposita 30, 112–123.

Embry, A.F., 1993. Transgressive–regressive (T–R) sequence analysis of the Jurassic succession of the Sverdrup Basin, Canadian Arctic Archipelago. Can. J. Earth Sci. 30, 301–320.

Embry, A.F., 1995. Sequence boundaries and sequence hierarchies: problems and proposals. In: Steel, R.J., Felt, V.L., Johannessen, E.P., Mathieu, C. (Eds.), Sequence Stratigraphy on the Northwest European Margin, Norwegian Petroleum Society special publication 5., 1–11.

Embry, A.F., Johannessen, E.P., 1992. T–R sequence stratigraphy, facies analysis and reservoir distribution in the uppermost Triassic–Lower Jurassic succession, western Sverdrup Basin, Arctic Canada. In: Vorren, T.O., Bergsager, E., Dahl-Stamnes, O.A. Holter, E., Johansen, B., Lie, E., Lund, T.B. (Eds.), Arctic Geology and Petroleum Potential, Norwegian Petroleum Society spec. publ. 2., 121–146.

Eriksson, P.G., 1999. Sea level changes and the continental freeboard concept: general principles and application to the Precambrian. Precambrian Res., this issue.

Eriksson, P.G., Altermann, W., 1998. An overview of the geology of the Transvaal Supergroup dolomites (South Africa). Environment. Geol. 36, 179–188.

Eriksson, P.G., Reczko, B.F.F., 1995. The sedimentary and tectonic setting of the Transvaal Supergroup floor rocks to the Bushveld complex. J. Afr. Earth Sci. 21, 487–504.

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

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

Eriksson, P.G., Schreiber, U.M., van der Neut, M., Labuschagne, H., van der Schyff, W., Potgieter, G., 1993b. Alternative marine and fluvial models for the non-fossiliferous quartzitic sandstones of the Early Proterozoic Daspoort Formation, Transvaal Sequence of southern Africa. J. Afr. Earth Sci. 16, 355–366.

Eriksson, P.G., Reczko, B.F.F., Merkle, R.K.W., Schreiber, U.M., Engelbrecht, J.P., Res, M., Snyman, C.P., 1994. Early Proterozoic black shales of the Timeball Hill Formation, South Africa: volcanogenic and palaeoenvironmental influences. J. Afr. Earth Sci. 18, 325–337.

Eriksson, P.G., Hattingh, P.J., Altermann, W., 1995. An overview of the geology of the Transvaal Sequence and Bushveld Complex, South Africa. Mineral. Deposita 30, 98–111.

Eriksson, P.G., Condie, K.C., Tirsgaard, H., Mueller, W.U., Altermann, W., Miall, A.D., Aspler, L.B., Catuneanu, O., Chiareznelli, J.R., 1998. Precambrian clastic sedimentation systems. Sediment. Geol. 120, 5–53.

Eriksson, P.G., Mazumder, R., Sarkar, S., Bose, P.K., Altermann, W., Van der Merwe, R., 1999. The 2.7–2.0 Ga volcano-sedimentary record of Africa, India and Australia: evidence for global and local changes in sea level and continental freeboard. Precambrian Res. 97, 269–302.

Gaal, G., 1992. Global Proterozoic tectonic cycles and Early Proterozoic metalogeny. S. Afr. J. Geol. 95, 80–87.

Galloway, W.E., 1989. Genetic stratigraphic sequences in basin analysis I: architecture and genesis of flooding-surface bounded depositional units. A. A. P. G. Bull. 73, 125–142.

Hälbich, I.W., Lamprecht, D.F., Altermann, W., Horstmann, U.E., 1992. A carbonate-banded iron formation transition in the early Proterozoikum of South Africa. J. Afr. Earth Sci. 15, 217–236.

Hälbich, I.W., Scheepers, R.S., Lamprecht, D., van Deventer, J.N., de Kock, N.J., 1993. The Transvaal–Griqualand West banded iron formation: geology, genesis, iron exploitation. J. Afr. Earth Sci. 16, 63–120.

Hamilton, D.S., Tadros, N.Z., 1994. Utility of coal seams as genetic stratigraphic sequence boundaries in nonmarine basins: an example from the Gunnedah Basin, Australia. A. A. P. G. Bull. 78, 267–286.

Haq, B.U., 1991. Sequence stratigraphy, sea-level change and significance for the deep sea. In: Macdonald, D.I.M. (Ed.), Sedimentation, Tectonics and Eustasy, Int. Assoc. Sedimentol. spec. publ. 12., 3–39.

Harmer, R.E., von Gruenewaldt, G., 1991. A review of magmatism associated with the Transvaal Basin — implications for its tectonic setting. S. Afr. J. Geol. 94, 104–122.

Hartzer, F.J., 1994. Transvaal Inliers: geology and relationship with the Bushveld Complex. Ph.D. dissertation, Rand Afrikaans University, Johannesburg, South Africa.

Hatton, C.J., Schweitzer, J.K., 1995. Evidence for synchronous extrusive and intrusive Bushveld magmatism. J. Afr. Earth Sci. 21, 579–594.

Helland-Hansen, W., Martinsen, O.J., 1996. Shoreline trajectories and sequences: description of variable depositional-dip scenarios. J. Sediment. Res. 66, 670–688.

Henry, G., Clendenin, C.W., Charlesworth, E.G., 1990. Depositional facies of the Black Reef Quartzite Formation in the eastern Transvaal, Abstracts 23rd Geocongress (Geocongress '90) of the Geological Society of South Africa, Cape Town, 234–237.

Hilliard, P. 1994. The structural evolution of the Johannesburg Dome, Kaapvaal Craton, South Africa. M.Sc. dissertation, Pretoria University, Pretoria, South Africa.

Hoffman, P.F., 1988. United plates of America, the birth of a craton: Early Proterozoic assembly and growth of Laurentia. Annu. Rev. Earth Planet. Sci. 16, 543–603.

Hunt, D., Tucker, M.E., 1992. Stranded parasequences and the forced regressive wedge systems tract: deposition during base-level fall. Sediment. Geol. 81, 1–9.

Jervey, M.T., 1988. Quantitative geological modeling of siliciclastic rock sequences and their seismic expression. In: Wilgus, C.K., Hastings, B.S., Kendall, C.G.St.C., Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (Eds.), Sea-Level Changes: An Integrated Approach, Soc. Econ. Paleontol. spec. publ. 42., 47–70.

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

Key, R.M., 1986. Sedimentation along the eastern margin of the Bushveld Basin, SE Botswana, Abstracts 21st Geocongress (Geocongress '86) of the Geological Society of South Africa, Johannesburg, 527–530.

Klemm, D.D., 1979. A biogenic model for the formation of the Banded Iron Formation in the Transvaal Supergroup/South Africa. Mineral. Deposita 14, 381–385.

Klemm, D.D., 1991. The formation of highly concentrated iron ore bodies within the BIF. The Sishen case. In: Pagel, M., Leory, J.L. (Eds.), Source, Transport and Deposition of Metals. Balkema, Rotterdam, pp. 63–64.

Li, Z.-X., Zhang, L., Powell, C.McA., 1996. Positions of the East Asian cratons in the Neoproterozoic supercontinent Rodinia. Aust. J. Earth Sci. 43, 593–604.

Loutit, T.S., Hardenbol, J., Vail, P.R., Baum, G.R., 1988. Condensed sections: the key to age dating and correlation of continental margin sequences. In: Wilgus, C.K., Hastings, B.S.Kendall, C.G.St.C., Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (Eds.), Sea-level Changes: An Integrated Approach, Soc. Econ. Paleontol. Mineral., Tulsa spec. publ. 42., 183–213.

McCourt, S., 1995. The crustal architecture of the Kaapvaal crustal block South Africa, between 3.5 and 2.0 Ga. Mineral. Deposita 30, 89–97.

Mitchum, R.M., 1977. Seismic stratigraphy and global changes of sea-level, part 11: glossary of terms used in seismic stratigraphy. In: Payton, C.E. (Ed.), Seismic Stratigraphy: Applications to Hydrocarbon Exploration, A. A. P. G. Memoir 26, 205–212.

Mitchum, R.M., Van Wagoner, J.C., 1991. High-frequency sequences and their stacking patterns: sequence stratigraphic evidence of high-frequency eustatic cycles. Sediment Geol. 70, 131–160.

Myers, R.E., 1990. The geology of the Godwan Basin, eastern Transvaal. Ph.D. dissertation, University of the Witwatersrand, Johannesburg, South Africa.

Nelson, D.R., 1998. Granite-greenstone crust formation on the Archaean Earth: a consequence of two superimposed processes. Earth Planet. Sci. Lett. 158, 109–119.

Nelson, D.R., Trendall, A.F., Altermann, W., 1999. Chronological correlations between the Pibara and Kaapvaal cratons. Precambrian Res. 97, 165–189.

Nemec, W., Steel, R.J., 1988. What is a fan delta and how do we recognize it? In: Nemec, W., Steel, R.J. (Eds.), Fan Deltas. Blackie, Glasgow, pp. 3–13.

Pelechaty, S.M., Kaufman, A.J., Grotzinger, J.P., 1996. Evaluation of $\delta^{13}C$ chemostratigraphy for intrabasinal correlation: Vendian strata of northeast Siberia. Geol. Soc. Am. Bull. 108, 992–1003.

Piper, J.D.A., 1983. Dynamics of continental crust in Proterozoic times. In: Medaris Jr., L.G., Byers, C.W., Nickelson, D.M., Shanks, W.C. (Eds.), Proterozoic Geology, Geol. Soc. Am. Mem. 161, 11–34.

Plint, A.G., 1988. Sharp-based shoreface sequences and "offshore bars" in the Cardium formation of Alberta: their relationship to relative changes in sea level. In: Wilgus, C.K., Hastings, B.S., Kendall, C.G.St.C., Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (Eds.), Sea-Level Changes: An Integrated Approach, Soc. Econ. Paleontol. Mineral., Tulsa spec. publ. 42., 357–370.

Posamentier, H.W., Allen, G.P., James, D.P., Tesson, M., 1992. Forced regressions in a sequence stratigraphic framework: concepts, examples and exploration significance. A. A. P. G. Bull. 76, 1687–1709.

Posamentier, H.W., Jervey, M.T., Vail, P.R., 1988. Eustatic controls on clastic deposition 1, conceptual framework. In: Wilgus, C.K., Hastings, B.S., Kendall, C.G.St.C., Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (Eds.), Sea-Level Changes: An Integrated Approach, Soc. Econ. Paleontol. Mineral., Tulsa spec. publ. 42., 109–124.

Rogers, J.J.W., 1996. A history of continents in the past three billion years. J. Geol. 104, 91–107.

Schlager, W., 1989. Drowning unconformities on carbonate platforms. In: Crevello, P.D., Wilson, J.L., Sarg, J.F., Read, J.F. (Eds.), Controls on Carbonate Platforms and Basin Development, Soc. Econ. Paleontol. Mineral., Tulsa spec. publ. 44., 15–25.

Schreiber, U.M., Eriksson, P.G., 1992. An Early Proterozoic braid–delta system in the Pretoria Group, Transvaal Sequence, South Africa. J. Afr. Earth Sci. 15, 111–125.

Schumm, S.A., 1993. River response to base-level change: implications for sequence stratigraphy. J. Geol. 101, 279–294.

Shanley, K.W., McCabe, P.J., Hettinger, R.D., 1992. Tidal influence in Cretaceous fluvial strata from Utah, USA: a key to sequence stratigraphic interpretation. Sedimentology 39, 905–930.

Shanley, K.W., McCabe, P.J., 1993. Alluvial architecture in a sequence stratigraphic framework: a case history from the Upper Cretaceous of southern Utah, USA. In: Flint, S., Bryant, I.D. (Eds.), International Association of Sedimentologists spec. publ. 15., 21–56.

Shanmugam, G., 1988. Origin, recognition and imprudence of erosional unconformities in sedimentary basins. In: Kleinspehn, K.L., Paola, C. (Eds.), New Perspectives in Basin Analysis. Springer, New York, pp. 83–108.

Stanistreet, I.G., 1993. Ancient and modern examples of tectonic escape basins: the Archaean Witwatersrand Basin compared with the Cenozoic Maracaibo Basin. In: Frostick, L.E., Steel, R.J. (Eds.), Tectonic Controls and Signatures in Sedimentary Successions, Int. Assoc. Sedimentol. Spec. Publ. 20, 363–376.

Stanistreet, I.G., McCarthy, T.S., 1991. Changing tectono-sedimentary scenarios relevant to the development of the Late Archaean Witwatersrand Basin. J. Afr. Earth Sci. 13, 65–81.

Tyler, N., 1979a. The stratigraphy of the early Proterozoic Buffalo Springs Group in the Thabazimbi area, west-central Transvaal. Trans. Geol. Soc. S. Afr. 82, 215–226.

Tyler, N., 1979b. Stratigraphy, geochemistry and correlation of the Ventersdorp Supergroup in the Derdepoort area, west-central Transvaal. Trans. Geol. Soc. S. Afr. 82, 133–147.

Vail, P.R., Mitchum Jr., R.M., Thompson III, S., 1977. Seismic stratigraphy and global changes of sea level, part four: global cycles of relative changes of sea level. A. A. P. G. Memoir 26, 83–98.

Vail, P.R., Audemard, F., Bowman, S.A., Eisner, P.N., Perez-Cruz, C., 1991. The stratigraphic signatures of tectonism, eustasy and sedimentology — an overview. In: Einsele, G., Ricken, W., Seilacher, A. (Eds.), Cycles and Events in Stratigraphy. Springer, New York, pp. 611–659.

Van Wagoner, J.C., Mitchum, R.M., Campion, K.M., Rahmanian, V.D., 1990. Siliciclastic sequence stratigraphy in well logs, cores, and outcrops: concepts for high-resolution correlation of time and facies, A. A. P. G. Methods in Exploration Series vol. 7.

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

Williams, H., Hoffman, P.F., Lewry, J.F., Monger, J.W.H., Rivers, T., 1991. Anatomy of North America: thematic portrayals of the continent. Tectonophysics 187, 117–134.

Young, G.M., 1995. Glacial environments of pre-Pleistocene age. In: Menzies, J. (Ed.), Past Glacial Environments, Sediments, Forms and Techniques. Butterworth-Heinemann, Oxford, pp. 239–252.