An overview of the geology of the Transvaal Supergroup dolomites (South Africa)

P. G. Eriksson · W. Altermann

Abstract In the Neoarchaean intracratonic basin of the Kaapvaal craton, between approximately 2640 Ma and 2516 Ma, two successive stromatolitic carbonate platforms developed. Deposition started with the Schmidtsdrif Subgroup, which is probably oldest in the southwestern part of the basin, and which contains stromatolitic carbonates, siliciclastic sediments and minor lava flows. Subsequently, the Nauga formation carbonates were deposited on peritidal flats located to the southwest and were drowned during a transgression of the Transvaal Supergroup epeiric sea, around 2550 Ma ago. This transgression led to the development of a carbonate platform in the areas of the preserved Transvaal and Griqualand West basins, which persisted for 30–50 Ma. During this time, shales were deposited over the Nauga Formation carbonates in the southwestern portion of the epeiric sea. A subsequent period of basin subsidence led to drowning of the stromatolitic platform and to sedimentation of chemical, iron-rich silica precipitates of the banded iron formations (BIF) over the entire basin. Carbonate precipitation in the Archaean was largely due to chemical and lesser biogenic processes, with stromatolites and ocean water composition playing an important role. The stromatolitic carbonates in the preserved Griqualand West and Transvaal basins are subdivided into several formations, based on the depositional facies, reflected by stromatolite morphology, and on intraformational unconformities; interbedded tuffs and available radiometric age data do not yet permit detailed correlation of units from the two basins. Thorough dolomitisation of most formations took place at different post-depositional stages, but mainly during early diagenesis. Partial silicification was the result of diagenetic and weathering processes. Karstification of the carbonate rocks was related to periods of exposure to subaerial conditions and to percolation of groundwater. Such periods occurred locally at the time of carbonate and BIF deposition. Main karstification, however, probably took place during an erosional period between approximately 2430 Ma and 2320 Ma.

Key words Dolomite · Limestone · Karst · Stromatolites · Shallow marine · Archaean · Transvaal

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Introduction

The Neoarchaean to Palaeoproterozoic Transvaal Supergroup comprises lowermost protobasinal volcano-sedimentary deposits, followed by the carbonate-banded iron formation (BIF) succession of the Ghaap-Taupone-Chuniespoort Groups, overlying clastic sedimentary and volcanic rocks of the Postmasburg-Segwagwa-Pretoria Groups, and uppermost Rooiberg Group lavas (Eriksson and others 1995). These rocks are preserved within three structural basins: Griqualand West (Ghaap-Postmasburg Groups) in central South Africa, Kanye (Taupone-Segwagwa Groups) in eastern Botswana and Transvaal (protobasinal rocks–Chuniespoort-Pretoria-Rooiberg Groups) in northern South Africa (Fig. 1). The dolomites and subordinate limestones and shales of the Ghaap-Taupone-Chuniespoort Groups, often also referred to as the Malmani-Campbellrand (Subgroups) carbonates, are late Archaean in age, and represent one of the oldest preserved carbonate platform successions in the stratigraphic record (Altermann and Wotherspoon 1995; Altermann and Nelson 1996). It is very likely that this ancient carbonate platform extended across all three preserved basins, and that it covered an area in excess of 600 000 km^2 (Beukes 1987).

The purpose of this paper is to describe the geology of these dolomitic rocks in order to provide a background for the engineering geological papers in this issue. Al-

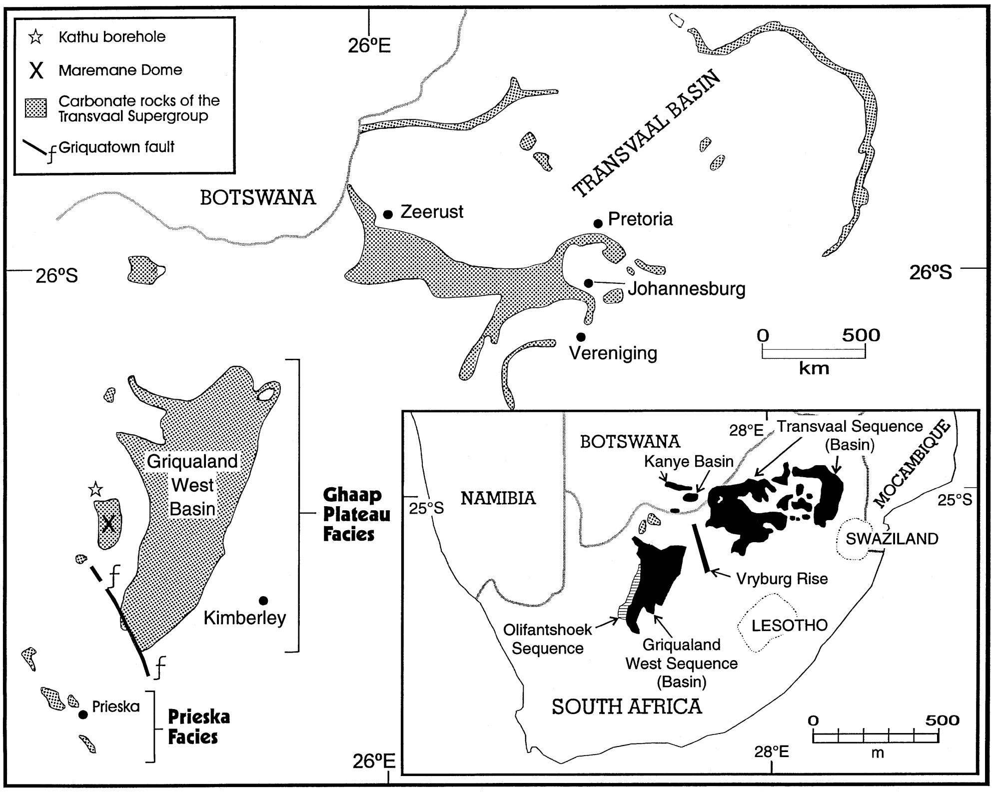


Fig. 1
Map showing the distribution of the carbonate rocks of the Transvaal Supergroup, within the Transvaal and Griqualand West basins. Note also the Kanye basin in Botswana (*inset map*) and the Maremane dome. The Griquatown fault separates the Prieska and Ghaap Plateau facies within the Griqualand West basin

though the geology of the Transvaal carbonate basin is well known, much of the detailed field work and interpretation on these rocks has been carried out in the more complex Griqualand West basin (Fig. 1) of the Northern Cape Province. The major region of urban and industrial development in South Africa, centred around the greater Johannesburg-Pretoria-Vereniging metropolitan area, lies within the southern part of the preserved Transvaal carbonate basin (Fig. 1), where major problems with karst-related features are encountered. As no large development has occurred on dolomitic rocks of the Kanye basin in Botswana, this paper will concentrate on the other two basins in South Africa.

Stratigraphic correlation and geochronology

Most previous workers have assumed a relatively simple lithostratigraphic correlation of the Transvaal carbonate units across the Transvaal and Griqualand West basins (Beukes 1987). This inferred simplicity has been offset by confusion arising from the use of different stratigraphic terminologies (Altermann and Wotherspoon 1995). In this paper, we will follow the widely accepted subdivision of Beukes (1980) for the Griqualand West basin, and that of Button (1973; as modified by Clendenin 1989) for the Transvaal basin.

The Griqualand West carbonate succession is cut by the major Griquatown fault in the southwestern portion of the basin (Fig. 1), which has led past workers to identify a Prieska carbonate facies (southwest of the fault) and a Ghaap Plateau carbonate facies (north-northeast of the fault). These two carbonate-shale successions located on either side of the fault were thought to reflect differences of water depth within the depositional palaeoenvironment (Beukes 1987; Altermann and Wotherspoon 1995). However, recent zircon dating (summed up by Altermann and Nelson 1996) of thin tuffs (Fig. 2) (Altermann 1996a) interbedded within the carbonates at Griqualand West indicates that carbonate deposition began in the Prieska region, southwest of the Griquatown fault zone, and that only the uppermost carbonates and shales from this area are age equivalents to the carbonates from the Ghaap Plateau region northeast of the fault (Fig. 2). The only age data available from the carbonates within the Transvaal basin indicates that deposition here began approximately simultaneously with that in the Ghaap Plateau region of the Griqualand West basin. These lateral relationships, between the carbonate successions within the two regions in the Griqualand West basin and in the Transvaal basin, reflect a major marine transgression from the southwest towards the northeast, as a large epeiric sea hosting widespread stromatolitic reefs developed on a tectonically stable Kaapvaal craton (Altermann and Nelson 1996).
The Griqualand West carbonates rest conformably on up to 650 m of mixed siliciclastic and chemical sediments and minor lava flows of the Schmidtsdrif Subgroup

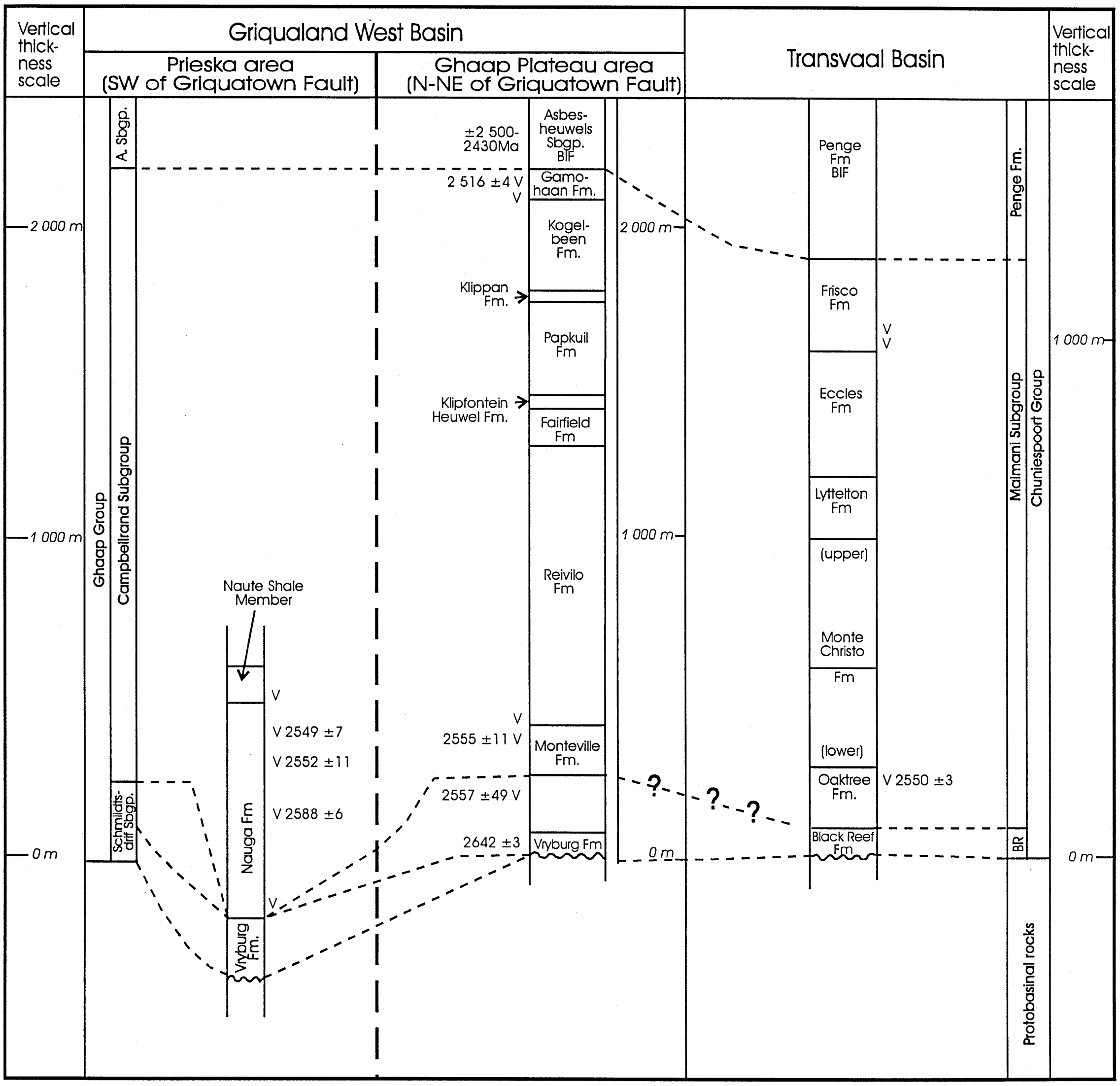


Fig. 2
Stratigraphic subdivision and correlation of the Transvaal Supergroup dolomites within the preserved Transvaal and Griqualand West basins. Note the location of thin tuffs (*v*) interbedded with the carbonate rocks and the radiometric ages determined from these volcanic rocks. Note also the separate columns for the Prieska and Ghaap Plateau facies of the Griqualand West succession

(Fig. 2) (Altermann and Siegfried 1997). It is possible that Schmidtsdrif rocks in the Prieska region are older than those in the Ghaap Plateau area northeast of the Griquatown fault. Initial Transvaal Supergroup carbonate deposition was restricted to the Prieska sub-basin, of which only a small part is preserved today. There, the Nauga Formation was deposited (Fig. 2), with an uppermost Naute Shale Member. A total of eight carbonate formations is identified in the Ghaap Plateau facies (Beukes 1980), with five formations being recognised in the Malmani Subgroup succession of the Transvaal basin (Fig. 2).

Carbonate lithostratigraphy in the Transvaal basin

The Malmani carbonate succession, almost 1200 m thick (Fig. 2), is characterised by predominant dolomites (a diagenetic product after primary limestones), subordinate limestones, lesser cherts and minor interbedded shales and chert-in-shale breccias (Button 1973; Beukes 1978; Clendenin 1989). Subdivision into the lowermost Oaktree, succeeding Monte Christo, Lyttelton, Eccles and uppermost Frisco Formations (Figs. 2 and 3) is based mainly on stromatolite types and interbedded cherts and shales (Button 1973). Clendenin (1989) added low angle unconformities as an additional stratigraphic discriminant between what he termed dolomite “packages”, this term being used for the alternating transgressive and regressive marine deposits comprising the Malmani succession. These packages are coincident with the formation names of Button (1973), and with an additional unconformity separating the lower and upper portion of the Monte

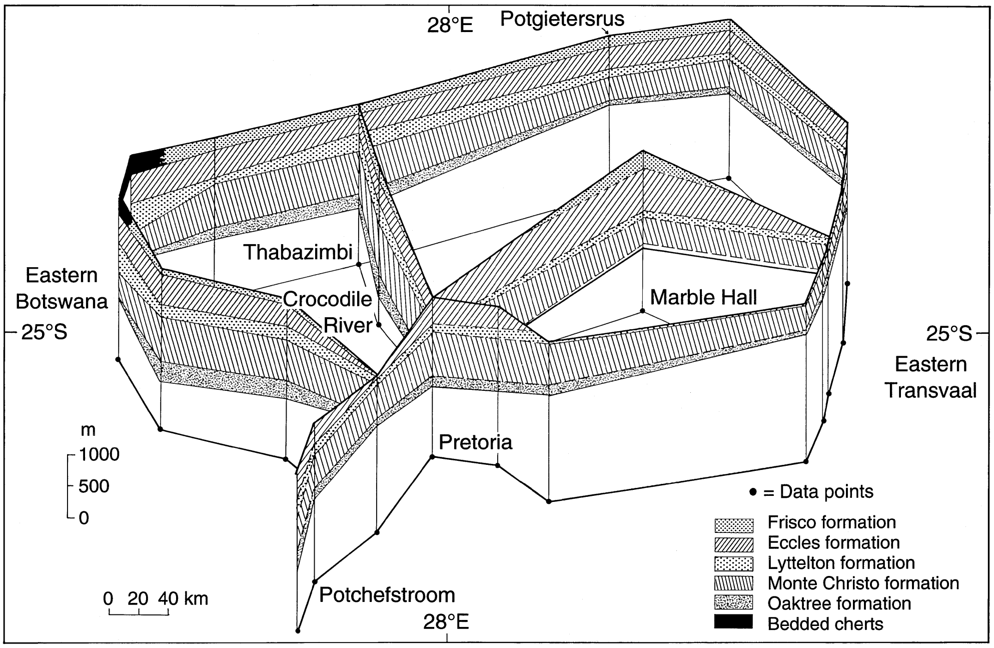


Fig. 3
Stratigraphic succession and sheet-like geometry of the carbonate succession of the Malmani Subgroup in the Transvaal basin

Christo Formation (Fig. 2). Clendenin's (1989) packages assumed a perfect correlation between the Griqualand West and Transvaal carbonate sequences, with most of the packages being common to both preserved basin successions, and the last and fifth package including the BIFs above the carbonates. The radiometric ages and correlation problems outlined above are at variance with this model.

The geometry of the preserved carbonate succession in the Transvaal basin indicates a sheet-like nature for the different formations, with little variation in thickness across the basin (Fig. 3) (Eriksson and Reczko 1995). The upper dolomite formations in the south and southeast of the preserved Transvaal basin have been removed by erosion (Fig. 3); some of this erosion was probably coeval with deposition of the post-BIF Duitschland Formation (Eriksson and Reczko 1995).

Carbonate lithostratigraphy in the Griqualand West basin

The eight formations making up the Ghaap Plateau facies of the Campbellrand Subgroup (Fig. 2) attain a maximum thickness of about 2500 m in the area of the Maremane Dome (Fig. 1), where the Kathu borehole provides a type section for the Ghaap Plateau carbonate succession (Altermann and Siegfried 1997). The basal Monteville Formation (Beukes 1980, 1987) is characterised by giant stromatolitic domes, fenestral microbial laminites, fine-grained clastic carbonates, shales and silts, reflecting a shelf carbonate association (Altermann and Wotherspoon 1995). The succeeding Reivilo Formation, up to 900 m thick, consists of carbonates with giant and columnar stromatolites, fenestral facies, thin intercalated oolitic beds and uppermost BIF-like lithologies (Beukes 1980, 1987). The cyclical Fairfield Formation comprises lower clastic laminated carbonates, passing up into rocks characterised by columnar stromatolites and microbial laminites. The Klipfontein Heuwel Formation exhibits domal stromatolitic dolomites and microbial laminites, which become increasingly siliceous upwards; sapropelic carbonates, oolite beds, laminated and columnar stromatolitic carbonates characterise the overlying Papkuil Formation. Domal stromatolites and microbial laminites of the 20-m-thick Klippan Formation are succeeded by over 300 m (Fig. 2) of dolomite, limestone and chert of the Kogelbeen Formation, with laminated mats, domal to columnar stromatolites and interbedded oolites (Beukes 1980, 1987). The uppermost Gamohaan Formation, with microbial mats, domal to columnar stromatolites, sapropelic carbonates and dolarenites, grades up into the succeeding BIFs of the Asbesheuwels Subgroup (Fig. 2) (Beukes 1980; Altermann and Wotherspoon 1995).

In the Prieska area, southwest of the Griquatown fault, only the Nauga Formation is recognised. Six hundred metres of carbonate at the base grade up into the Naute Shale Member (Fig. 2) (Beukes 1980; Altermann and Wotherspoon 1995). The Naute Shale Member of the Nauga Formation is the approximate age equivalent of the lower carbonates within the Ghaap Plateau and preserved Transvaal basin regions. Thin volcanic tuff interbeds within the Griqualand West and Transvaal carbonate platform successions indicate proximal-distal relationships, from southwest to northeast (Altermann 1996a).

Palaeoenvironmental models for the Transvaal Supergroup carbonates

Truswell and Eriksson (1973) proposed a tidal flat model for the Transvaal carbonates, characterised by giant domal subtidal stromatolites which decreased in size shorewards, to laminated microbial mats in the intertidal-supratidal settings (Fig. 4). Beukes (1978, 1980, 1987) developed this model further, identifying a deeper basinal facies in the south and west of the Griqualand West basin, which was separated from a more widespread tidally dominated shallower platform facies to the north and east (the Ghaap Plateau facies, which extended across to the area of the preserved Transvaal basin as well). The basinal facies was separated from the platform by the northwest–southeast striking Griquatown synsedimentary hinge fault, with stronger subsidence on its southwestern side. Basinal shales and turbidite-deposited carbonates were proposed by Beukes (1980) in the southwestern deeper portion (i.e. the Prieska facies) of this overall epeiric marine palaeoenvironment.
A more recent model (Clendenin 1989) envisaged three broad facies belts: distal shallow basin, subtidal periplatform and peritidal flats, which transgressed and regressed five times across a large portion of the Kaapvaal craton, under the influence of syndepositional extensional tectonics, with largely thermal subsidence. These transgressive-regressive phases formed five time-stratigraphic packages, including carbonates, overlying BIFs, and succeeding mixed siliciclastic and chemical sediments of the Duitschland Formation. This modified carbonate ramp model of Clendenin (1989) used the basic elements of the Truswell and Eriksson (1973) and Beukes (1978, 1987) hypotheses, and proposed a southwesterly inclination of the ramp, which passed into a more distal shallow basin in the southwest, the latter characterised by mixed carbonate-shale and BIF sedimentation.

The general agreement among these three models, with a deeper basinal facies in the southwest, is no longer tenable in the light of recent research. Klein and others (1987) discovered cyanobacterial microfossils (formerly referred to as "blue-green algae") in the inferred shelf-basinal transition zone from carbonate to BIF deposits, thereby implying shallower, photic depositional depths of about 45 m, with the more distal euxinic basin probably having had depths of about 80 m or less. Altermann and Herbig (1991) documented well-preserved intertidal to supratidal facies with definite desiccation features in the supposed basinal region southwest of the Griquatown fault. An overall modification of previous models, with the deepest part of the basin in the western portions of the Griqualand West region (western Ghaap Plateau facies), where both subsidence and carbonate accumulation were at a maximum (Altermann and Siegfried 1997), and mainly platform facies to the northeast into the Transvaal basin, is probably more acceptable. In the southwestern part of the Griqualand West basin (Prieska facies), predominantly shallow to emergent facies were formed. Upper shelf settings were subsequently developed in the entire region as drowning led to subsequent BIF deposition (Altermann and Wotherspoon 1995). Overall, the entire epeiric sea basin was most likely shallow throughout carbonate deposition, depths probably not exceeding 40–80 m in the shelf to "basinal" facies.
The recent zircon dating of tuffs within the Griqualand West carbonate succession has enabled a more accurate model of basin development to be proposed for the Transvaal carbonates (Altermann and Wotherspoon 1995; Altermann and Nelson 1996). An initial shallow peritidal flat carbonate basin developed in the region southwest of the Griquatown fault (Prieska facies). The extension of this basin to the south and west is not known, due to destruction by younger tectonic events. The basin possibly included a volcanic arc to the south of the Prieska area (Altermann 1996a), and was bounded by a palaeohigh to the northeast, near the Griquatown fault zone. Carbonate deposition north–northeast of the Griquatown fault zone, heralding the onset of sedimentation of the Ghaap Plateau facies and deposition within the preserved Transvaal basin area, reflects a major transgression of the epeiric sea at about 2550 Ma. As a result of the elevated sea level, the southwestern (Prieska facies) carbonates were drowned, and Naute Shale Member (Nauga Formation)

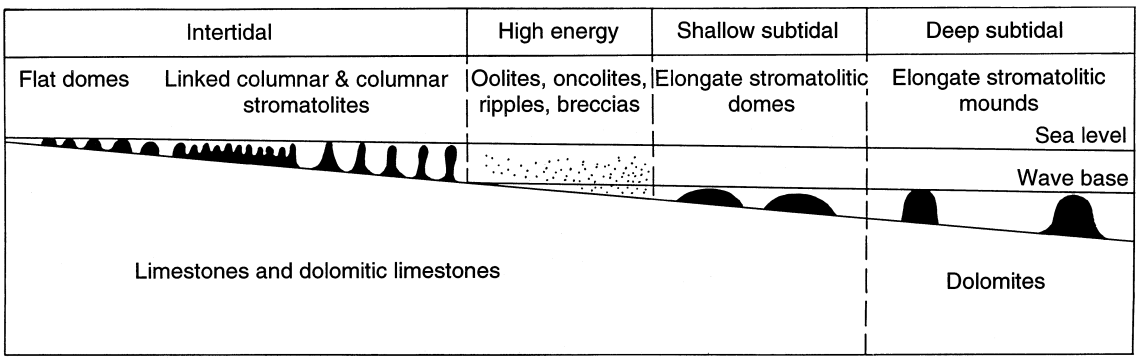


Fig. 4
Tidal-subtidal palaeoenvironmental model for carbonate rocks of the Transvaal Supergroup (after Truswell and Eriksson 1973). Lithologies shown reflect the role of early diagenesis due to adjacent waters (see text)

argillites were shed into this part of the basin from the inferred southernmost volcanic arc. The Naute Shales are thus the time equivalents of the initial shelf carbonate deposits of the Ghaap Plateau facies further to the north-northeast. In the Ghaap Plateau and Transvaal basin region, subsidence of the basin was matched by stromatolitic reef growth and carbonate sediment accumulation; shallow marine conditions thus persisted during deposition of the Reivilo to Kogelbeen Formations (and their Transvaal basin equivalent units) (Altermann, in press). The uppermost Gamohaan Formation reflects a return to shelf carbonate deposition as tectonically induced subsidence drowned the shallow water Ghaap Plateau-Transvaal basin carbonate platform, and deeper water BIF deposition gradually replaced shelf carbonate sediments (Altermann and Wotherspoon 1995).

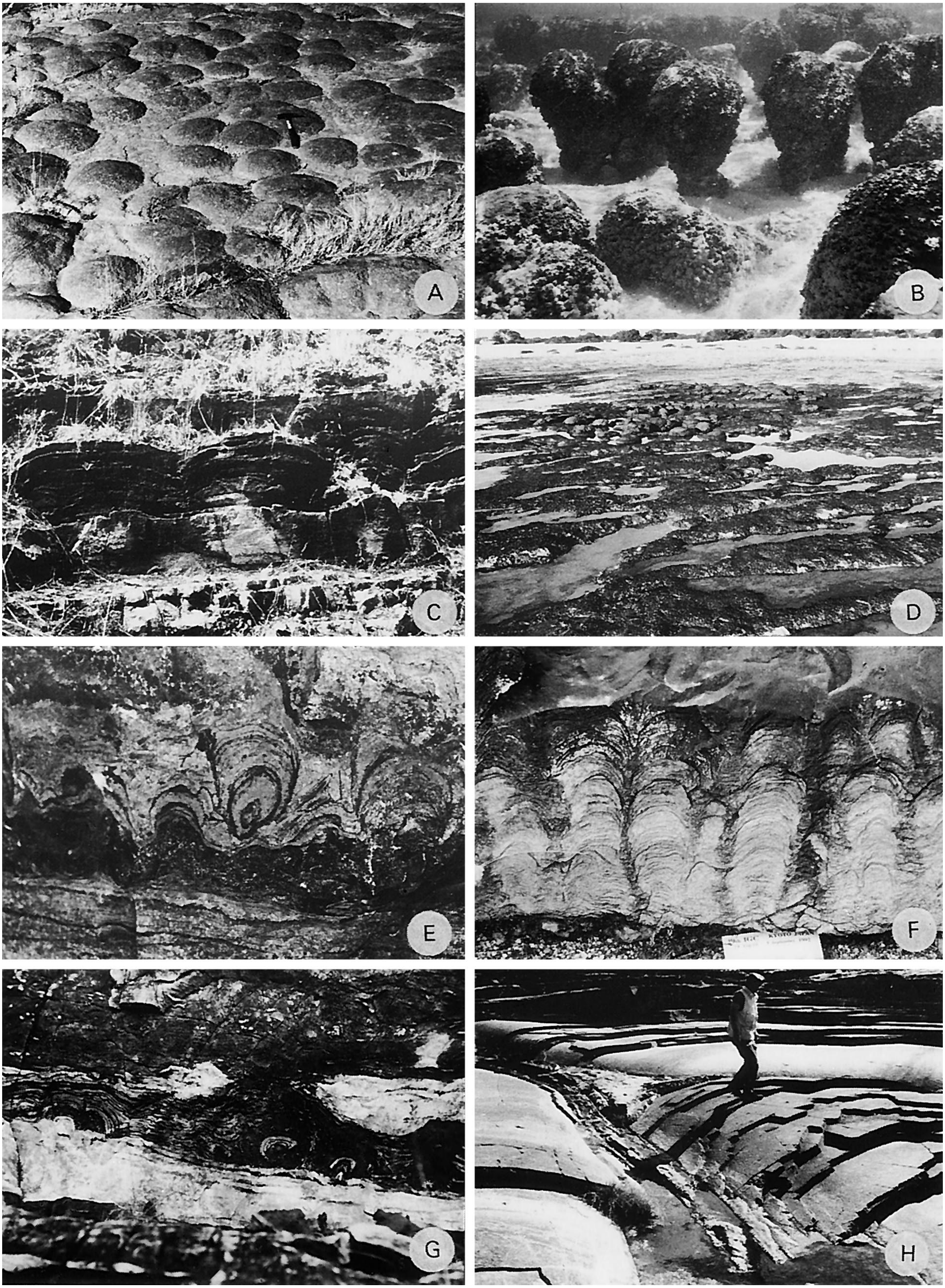
Sediment accumulation rates for the Ghaap Plateau facies of the Transvaal carbonates, derived from recent zircon dating, appear to have been approximately 60–80 m/Ma (Altermann 1996b). These rates are comparable to modern carbonate shelves, but are much lower than those for modern reefs. Much slower sedimentation rates are calculated for the carbonates of the Nauga Formation (Prieska facies, southwest of the Griquatown fault), below 10 m/Ma. These sediments, interpreted by earlier workers (Beukes 1987) to be deeper basinal deposits, were shown by Altermann and Herbig (1991) to be peritidal to supratidal flat carbonates. Within the preserved sedimentary pile from such a palaeoenvironmental setting, up to 90% of the time is represented by contacts between the rock layers (i.e. periods of non-sedimentation) rather than by the accumulated sediment itself (Osleger 1994), thereby explaining the low sedimentation rates for the Nauga Formation. Lanier (1988) estimated much greater depositional rates for the overall Transvaal carbonate succession, ca. 6000 m/Ma, based on calculations of biomass productivity of microstromatolites, which he found analogous to modern productivity. Such calculations, however, are not comparable with direct measurements of stromatolite growth rates, as they do not account for some important aspects like calcification, cementation and preservation potential. The best solution to this major discrepancy is provided by intraformational unconformities, representing considerable time gaps within the carbonate depositional system. Such time gaps would be expected within the inferred shallow water–tidal flat model generally accepted for the carbonates (Altermann and Wotherspoon 1995).

General discussion of Archaean carbonate precipitation

Carbonate precipitation in sea water results from biochemical and chemical processes. Lime solubility in water depends primarily on CO_2 pressure, and temperature. Any process which removes CO_2 from the normal (pH 8.4) water supports lime precipitation. Among the most effective processes in support of carbonate precipitation are: increase of temperature and intense evaporation, influx of supersaturated water to a basin where $CaCO_3$ nuclei or catalysers are present, release of CO_2 by decrease of pressure (upwelling or mechanical processes), mixing of water high in CO_3 and low in Ca^{2+} with seawater, organic mineralisation and organically induced processes such as an increase in the pH by bacterial decay, and increasing ammonia and $CaCO_3$ concentration while removing CO_2 by photosynthesis.

Organisms taking part in carbonate production (organic mineralisation) flourish in warm and clear water, and hence carbonate precipitation in the Phanerozoic to Recent, with minor exceptions, is almost restricted to tropical and subtropical, clear and shallow conditions. Present-day carbonates accumulate chiefly north and south of

Fig. 5
Various stromatolites of the Neoarchaean Transvaal Supergroup (**A**, **C** and **E–H**) and their Recent counterparts (**B**, **D**). **A** Oblique view of domical stromatolite columns at Boetsap, Griqualand West (40 cm long hammer in the centre for scale). The stromatolites are closely spaced and internally finely laminated, and originated in a shallow subtidal environment. They are overlain by columnar stromatolites with coarse sediment trapped between the columns, indicating stronger tidal currents in a shallow subtidal to intertidal environment (as in **F**). **B** Living subtidal columnar stromatolites (Shark Bay, Western Australia) in a similar environment to **A**. The stromatolites grow at a depth of about 100–150 cm and are approximately 80 cm high. Their external morphology closely resembles the shape of the stromatolites in **A**, but the internal structure is much coarser and crudely laminated, due to binding of coarse sediment along the laminae. **C** Cross-section of elongated domal stromatolites, with crinkly laminated internal structure, in the Malmani Subgroup (R512 road from Pretoria to Hartebeespoort dam). The stromatolites are about 100 cm high and probably grew in an intertidal to subtidal palaeoenvironment where they were elongated parallel to tidal currents. **D** Modern elongated stromatolite domes, 100–250 cm long and 10–30 cm high, exposed at low tide on the intertidal flats at Shark Bay, Western Australia. Elongation parallels tidal current direction. These living stromatolites have a different internal structure to their Neoarchaean counterparts, being composed of pustular cyanobacterial mats and coarse sediment. **E**, **F** Columnar and bulbous stromatolitic lamination in the Malmani Subgroup (**E**) and at Boetsap (**F**), with well-developed (**E**) lateral linkage between the columns. An intertidal to shallow subtidal palaeoenvironment is inferred; **E** points to low energy, whereas **F** reflects higher energy, as evidenced by their association, respectively, with fine and coarse sediment. The columns in **E** are 5–10 cm high and the columns in **F**, 20–25 cm high. **G** Domical stromatolites in the Malmani Subgroup (R512 road from Pretoria to Hartebeespoort dam). The stromatolites are about 10 cm high and finely laminated. The white bands and lenses below and above the stromatolitic bed are of late chert (silicified laminite beds). The preferential silicification is probably due to a better permeability in flat laminated compared to irregularly domed strata. **H** Giant domal stromatolites at Boetsap, Griqualand West. The elongated domes can attain a length of 30–50 m and a height of 10 m. They represent deep subtidal bioherms of tufted cyanobacterial mats and were elongated parallel to the palaeocurrent (as in **C** and **D**)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the equator below latitudes of 30°, and in water depths where photosynthesis is possible. In the Archaean and early Proterozoic, building of carbonate shells and skeletons was absent. The Archaean and early Proterozoic life record includes exclusively prokaryotic bacteria and cyanobacteria.

The vast majority of Archaean and early Proterozoic carbonates are stromatolitic (organo-sedimentary) deposits (Fig. 5) or the direct result of erosion of stromatolites. Phanerozoic and Recent stromatolite growth is restricted to ecological niches, where the slow growing cyanobacterial mats are not consumed by grazers, and constitute more complicated communities, including algae and other eukaryotes. They are formed by mechanisms involving trapping and binding of sediment (Fig. 5B), inorganic calcification, as well as biologically influenced and skeletal calcification (Burne and Moore 1987). Early Precambrian stromatolites were formed by similar processes in various, deep subtidal to supratidal environments by bacterial and cyanobacterial communities (Fig. 5).

Precambrian stromatolite morphology (Fig. 5) was ruled by many environmental factors, such as water depth, penetration of light, tidal currents or the amount of sediment influx. Thus, with some restrictions, the palaeoenvironment of the stromatolite communities can be reconstructed from the morphology and microstructure of the preserved stromatolites, from their association with sedimentary rocks, and by comparison with modern stromatolitic settings (Fig. 5). The variation in size and morphology of Precambrian stromatolites was greater than in the Recent; however, morphology of Recent and Precambrian stromatolites, and also of cyanobacteria, is astonishingly similar (Schopf 1993).

Schopf (1993) demonstrated that cyanobacteria became abundant after 3500 Ma ago, but there is no evidence of any more advanced form of life, including the simplest eukaryotic algae, before the end of the Palaeoproterozoic (Schopf and Klein 1992). Although there is no direct evidence for it, the Archaean cyanobacteria were most probably photosynthesising (Schopf 1993). Their activity was responsible for the decrease in CO_2 and rise in oxygen level of the Proterozoic hydrosphere and atmosphere, via oxygenic photosynthesis, where CO_2 and H_2O were transformed by light energy and chlorophyll to glucose, sugar and oxygen. Thus, stromatolites in the early Precambrian were most probably restricted to the photic zone, where there was also uptake of CO_2 by cyanobacteria (Schopf 1992; Schopf and Klein 1992).

Lanier (1988) provides evidence that the micro-and macrostromatolites in the Malmani Subgroup dolomites were indeed capable of precipitating large quantities of carbonate sediment in an overall oxygen-poor environmental setting. A biologically diverse microfossil assemblage, including filamentous and coccoidal cyanobacteria and bacteria is described by Klein and others (1987) and Altermann and Schopf (1995). In the superbly preserved cyanobacteria samples of Klein and others (1987), there is evidence of biogenic precipitation of micritic aragonite needles within the cyanobacterial sheaths.

Carbonate precipitation, chemical and, to a lesser extent, biogenic depends on the chemical composition of seawater. The carbonate in modern shallow water lime muds is generally aragonite (orthorhombic $CaCO_3$ in contrast to trigonal calcite), but may contain up to 50% high-Mg calcite (over 10 mol% Mg) and up to 15% low-Mg calcite. The Mg calcite is derived from attrition of shell material. The aragonite is most probably derived from the breakdown of algae or it is inorganically precipitated.

The composition of Archaean and Proterozoic seawater has been widely discussed by many workers. Most of them (Walker 1983; Kasting 1991; Holland 1994) agree that the composition of seawater has not changed greatly from Precambrian to modern times. A higher saturation in $CaCO_3$ of the late Archaean and early Proterozoic seawater was proposed by Grotzinger (1990) to account for abiotic $CaCO_3$ precipitation, which is thought to have been common in the late Archaean by some authors (Grotzinger and Kasting 1993; Sumner and Grotzinger 1996). Grotzinger (1990) argues that direct precipitation of fine-grained $CaCO_3$ (micrite) might have been induced along cyanobacterial mats, as CO_2 was extracted from the seawater by photosynthesis. The higher $CaCO_3$ saturation is inferred to have been responsible for the bloom of microdigitate and conical stromatolites in the early Archaean, and the decrease in this saturation in the middle Proterozoic was probably partially responsible for the decline of stromatolite diversity (Grotzinger 1990).

In a later model, a high concentration of inhibitors to $CaCO_3$ precipitation, like Mn^{2+}, Mg^{2+}, but mainly Fe^{2+}, in seawater is thought to have been responsible for the precipitation of large aragonite crystals and of magnesian calcite beds directly on the sea floor (Sumner and Grotzinger 1996). Subsequently, the metastable aragonite was pseudomorphed by calcite. The interpretation of the large abiotic calcite pseudomorphs as former aragonite is, however, equivocal as they could also be pseudomorphs after giant gypsum ($CaSO_4$*$2H_2O$) crystals. The increase of O_2 partial pressure in the Proterozoic environment around 2200–1900 Ma led to precipitation of these inhibitors as oxides. The high concentration of Ca in seawater was no longer possible, as nucleation of $CaCO_3$ crystals became a common process and thus, the giant aragonite crystals had no chance to form. Calcite was then precipitated mainly as micrite (Sumner and Grotzinger 1996).

Diagenesis

Oxygen-poor palaeoenvironmental and early diagenetic conditions, generally inferred for the late Archaean time period, are supported by the divalent chemical state of both Fe and Mn in the carbonates (Eriksson and others 1975). It is possible that the large surface/volume ratio of a shallow epeiric sea, such as was probably responsible for deposition of the Transvaal carbonates, promoted precipitation of calcite and aragonite, thereby raising Mg:Ca ratios sufficiently to overcome kinetic impediments and enable dolomitisation of the substrate (Eriksson and others 1975). The saturated Fe:Mn ratio of dolomitic limestones supports this proposed model, in which waters close to the limestones were responsible for early dolomitisation, rather than an alternative refluxed brine dolomitisation model (Eriksson and others 1975). Dolomitisation of primary limestones is, however, a diagenetic process, which is in agreement with field evidence from the rocks themselves. Mg-enriched diagenetic pore waters were most likely responsible for early dolomitisation. The subtidal, abiotically precipitated high-Mg calcite beds (Grotzinger and Kasting 1993) would have supplied sufficient amounts of Mg for thorough dolomitisation of the subtidal carbonates, because magnesian calcite increases in solubility with increasing Mg content. On the other hand, it can also be expected from an early dolomitisation model relying mainly on adjacent waters, that dolomitisation was more complete in the inferred subtidal dolomites, which comprise largely recrystallised dolomites. Mixed limestones and dolomitic limestones were more typical of the inferred shallow subtidal to intertidal depth ranges (Eriksson and others 1975) (Fig. 4).

In practice, some limestone beds, particularly stromatolitic marker beds, were more prone to secondary dolomitisation, while other stromatolitic zones and chert bands appear to have acted as barriers to percolating Mg-rich fluids, enriched in Mg through dissolution of high Mg-calcite. Secondary, late diagenetic dolomitisation is widespread in the Transvaal Supergroup carbonates. Such dolomite is usually coarsely crystalline and of equant grain size. The late diagenetic dolomitisation requires low Mg^{2+}/Ca^{2+} ratios. Compaction of intercalated shales can supply the necessary Mg by dewatering of Mg-enriched fluids (Altermann and Wotherspoon 1995).

Silicification is, similar to dolomitisation, a diagenetic and post-diagenetic process. Several generations of silicification in the Griqualand West carbonates, ranging from early diagenetic, predating the degradation of organic matter, to post-depositional (Fig. 5G) are described by Altermann and Wotherspoon (1995), Altermann and Schopf (1995) and Altermann (1997, in review). Most of the silicification is confined to certain stromatolitic formations or horizons and is rather continuous laterally, but silicification along subvertical faults is also present. Although no systematic investigations of the silicification patterns and genesis in the Transvaal Supergroup have been carried out, it is very likely that the most widespread silicification episode occurred during subaerial weathering of the carbonates and overlying sediments in Proterozoic or even Palaeozoic times. Evidence for this can be found in the coarsely recrystallised near-surface silicification of the Kogelbeen Formation in Griqualand West. The silicification of these rocks was probably also related to silicification of the asbestos fibre in the overlying BIF, and to the formation of "tigers eye" deposits in the weathered zone of the BIF in Griqualand West. It must, presumably, have been caused by SiO_2-enriched (weathering of cherts and quartzites) groundwaters percolating in the more permeable stromatolitic horizons.

Early karst weathering of the preserved Transvaal basin carbonates

Karst weathering and the resultant engineering geological problems form a major topic of this issue. Although karst-fill material is commonly of Karoo age and younger (Wilkins and others 1987), there were several much older karst events in the preserved Transvaal basin carbonates. Deposition of the Griqualand West BIFs took place at about 2.5–2.43 Ga (Martini and others 1995). The onset of Pretoria Group sedimentation, which lies unconformably above the Penge Formation (Fig. 2) BIFs, is difficult to assess. An uncertain zircon age of 2350 Ma for the Bushy Bend lavas near the base of the Pretoria Group is, perhaps, supported by a tectono-thermal event which affected Ventersdorp lavas and underlying basement lithologies at about 2320 Ma (Robb and Meyer 1995). This possible event may reflect rifting, minor volcanism and rapid mechanical subsidence accompanying the basal Rooihoogte fan sediments of the Pretoria Group. A major karst event took place during the time interval represented by the unconformity and related hiatus that separates the Chuniespoort and Pretoria Groups, at about 2430 (or less) Ma to 2320–2350 Ma, when erosion had removed much of the Penge BIFs protecting the dolomites from weathering (Martini and others 1995). Large cavities are typically found in association with the Pretoria Group contact, but also occur at several hundred metres below this level (Martini and others 1995).

The resultant cavity systems were filled mainly by mudstones, dolomitic breccias, and, locally, by Pretoria Group sandstones (Martini and others 1995). Mississippi Valley-type (MVT) Pb-Zn-F mineralization is also found in the karstic cavities, particularly in the southwestern portion of the Transvaal region. The absence of syngenetic mineralisation at the base of the Pretoria Group and the lack of reworked mineralisation in basal Pretoria sediments supports strongly that MVT mineralisation occurred after deposition of the Pretoria Group (or at least the lower portion thereof), but while the cavernous dolostones were still porous, with well-linked cavities, i.e. in early Pretoria Group times (Martini and others 1995).

There must also have been a period of karst formation pre-dating or synchronous with deposition of the Duitschland Formation, which lies stratigraphically between the Penge Formation and the Pretoria Group. It is, however, impossible to distinguish this period from any later karst-forming episode in the Transvaal. Most probably, once established, karst systems were reactivated repeatedly and karstification is active along ancient patterns even today. Theoretically, a period of karst-forming carbonate dissolution can be expected at any major period of subaerial erosion (unconformity). Such unconformities are present below and above the Duitschland Formation and below, within and above the Pretoria and Postmasburg Groups (Eriksson and others 1995). An even earlier period of karstification is documented in Griqualand West, where karst systems are filled with BIF breccias and sediments (Plehwe-Leisen and Klemm 1995) and associated ores. This karstification period was active during deposition of the Asbesheuwels BIFs around 2500–2430 Ma ago and was restricted to the rising of the locally developed Maremane dome structure.

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