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Geology and metallogeny of the Barberton greenstone belt: a survey

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Abstract - The principal episodes of the geotectonic history of the Barberton greenstone belt span some 400 Ma from the initial submarine eruption of the Onverwacht lavas (at about 3500 Ma) to the compression and inversion of the depository as a fold-and-thrust belt between 3230 and 3215 Ma, followed by late-tectonic hydrothermal Au mineralization at about 3100 Ma. The crust beneath the greenstone belt was fortified by sialic underplating due to episodic felsic plutonism at about 3445 Ma, 3227 Ma and 3105 Ma. The main stratigraphic units of the Barberton Supergroup comprise the Onverwacht Group, a thick assortment of ultramafic and mafic volcanics including a number of sill-like layered ultramafic complexes, which is overlain by the Fig Tree Group of turbiditic greywacke sandstones and associated mudstones and banded ferruginous shales. These strata are, in turn, paraconformably overlain by continentally-derived shallow-water arenites of the Moodies Group. The broader aspects of the deformation of the Barberton depository can be considered in terms of one of two models: either the structures represent a single compressional episode of upright folding accompanied by high-angle thrusting between 3230 and 3215 Ma, or they resulted from an early episode of tectonic stacking, followed by a later event involving the rotation of early thrust slices into upright folds and accompanied by high-angle thrusting. Substantial deposits of chrysotile asbestos, magnesite and talc were developed in the ultramafic host rocks of the Onverwacht Group and important mesothermal deposits of Au were formed in the reactive host rocks belonging to all three Groups in splays adjacent to regionally-developed shear zones. There is still uncertainty as to the primary metallogenic setting of the belt during Onverwacht times and what part accretionary or collage tectonics may or may not have played in the evolution of the Barberton greenstone belt.

Résumé - Les épisodes géotectoniques principaux qui ont affecté la ceinture de roches vertes de Barberton s'étalent sur quelque 400 Ma, depuis les éruptions sous-marines initiales vers 3500 Ma, en passant par la compression et l'inversion des dépôts en tant que chaîne charriée et plissée vers 3230-3215 Ma, jusqu'à, enfin, la minéralisation aurifère tardi-tectonique et hydrothermale vers 3100 Ma. La croûte de cette ceinture de roches vertes s'est consolidée par accrétion sialique sous-crustale lors de phases plutoniques felsiques épisodiques à 3445, 3227 et 3105 Ma. Les unités stratigraphiques principales du Supergroupe de Barberton comprennnent le Groupe de l'Onverwacht surmonté par le Groupe de Fig Tree qui est lui-même situé sous le Groupe de Moodies. Le Groupe de l'Onverwacht est un ensemble de roches volcaniques mafiques-ultramafiques des sills de complexes ultramafiques stratiformes. Le Groupe sus-jacent de Fig Tree est composé de grès greywackeux turbiditiques associés à des argilites et des schistes ferrugineux rubanés. Il est surmonté de manière paraconforme par les arénites d'origine continentale et de faible profondeur d'eau du Groupe de Moodies. Les aspects principaux de la déformation qui a affecté les dépôts de Barberton peuvent être intégrés dans deux modèles alternatifs: ou bien ces structures représentent un seul épisode de compression de plis droits lors d'un charriage redressé entre 3230 et 3215 Ma, ou bien elles résultent d'un épisode précoce d'empilement tectonique suivi par une phase plus tardive incluant la rotation des écailles du charriage précoce en plis droits et un charriage redressé. Des gisements importants d'asbeste chrysolite, de magnésite et de talc se sont développés dans l'encaissant des roches ultramafiques du Groupe de l'Onverwacht. Dans les trois Groupes, d'important gisements mésothermaux d'or se sont formés dans des excroissances adjacentes à des mega-shear zones. Il reste une incertitude quant au milieu métallogénique primaire de la ceinture à l'époque Onverwacht, et quant au rôle qu'a pu avoir joué la tectonique de type collage ou d'accrétion dans l'évolution de la ceinture verte de Barberton.

INTRODUCTION

The Barberton greenstone belt is internationally renowned for its antiquity, preservation, exposure, lithologies, early life forms, structure and mineralization (Hall, 1918; Visser *et al.*, 1956; Ramsay, 1963; Anhaeusser, 1965 *et seq.*; Viljoen and Viljoen, 1969 *et seq.*; Lowe and Knauth, 1977 *et seq.*; Knoll and Barghoorn, 1977; de Wit, 1982 *et seq.*; de Ronde *et al.*, 1991 *et seq.*). Accordingly, and as part of the development of SAMINDABA (the South African mineral deposit database), the Council for Geoscience has undertaken a metallogenic assessment of the region. The work has involved compiling mapping and other information on record, and information made available in publications and otherwise by mining houses, research institutes and individual researchers, to produce a metallogenic map of the greenstone belt. The process of compilation has facilitated a synoptical appreciation of the geology and structure of the Barberton fold-and-thrust belt, which has assisted in the understanding of the tectogenesis and mineralization of the region.

Following the outstanding pioneer field work by Hall (1918), the publication by the Geological Survey of South Africa of 1:50 000 scale maps and a description of the geology of the Barberton region (Visser *et al.*, 1956) was a significant advance in the understanding of the Early Archaean greenstone belt. Subsequent work by the mining companies operating in the area, and by research groups led by Anhaeusser, Viljoen and Viljoen, Lowe, de Wit and others (see Anhaeusser, 1976, 1986, 1992), has served to refine and elaborate this knowledge, but there is still a basic scientific enigma with regard to the geotectonic setting of the Barberton basin: namely, was the depository episialic, ensialic or ensimatic? There is no ready answer to this question, which provides fertile ground for speculation and continuing research.

Stratigraphically, the Barberton greenstone belt appears to comprise elements of oceanic and continental lithological affinity. It represents an Early Archaean volcano-sedimentary terrain surrounded by invasive sial, which has masked the geotectonic foundations and setting of the deformed depository. Submarine extrusion of high temperature komatiitic and then tholeiitic lavas, with subordinate felsic sills and lavas, was succeeded, after a prominent hiatus, by further ultramafic to mafic volcanicity. Thereafter, early flysch-like and later coarse-clastic sedimentation may have been initiated by post-magmatic thermal relaxation of the overheated lithosphere beneath the Barberton basin. In terms of a speculative model, this response to an incipient or primitive Wilson Cycle probably included the ductile thinning and hot-rifting of the sialic crust, with attendant mid-continental rift, continental margin, back-arc or small ocean basin magmatism, followed by sedimentation and then closure and inversion of the basin by regional compression of indeterminate cause. The brunt of the compression was buffered by the yielding character of the lower plate, including the effects of early-tectonic sialic diapirism, and the crumpling of the interposed greenstone belt. It is probable that evidence of the depository would not have survived for long in a hyperactive oceanic crustal setting, responding to the higher heat-flow in Archaean times, had the oceanic ultramafic to mafic volcanics not been buoyed-up and preserved by sialic underplating, possibly the result of obduction, during or soon after eruption (e.g. Anhaeusser *et al.*, 1967; de Wit and Stern, 1980; Heinrichs, 1980; Jackson *et al.*, 1987; de Wit *et al.*, 1987a, b).

Among other factors, this paper will examine stratigraphical, structural, metamorphic, geochronological and mineralization aspects of the Barberton greenstone belt in an endeavour to substantiate elements of a geotectonic framework to be incorporated in a description to accompany the future publication of a metallogenic map of the region.

STRATIGRAPHY

In arriving at a lithostratigraphical classification of the Barberton Supergroup, the South African Committee for Stratigraphy (Kent, 1980) considered the proposals of Hall (1918), van Eeden (1941), Visser *et al.* (1956), Steyn (1965), Viljoen and Viljoen (1969) and Anhaeusser (1975), among other opinions.

The rocks of the Barberton greenstone belt have been classified into three main groups on the basis of lithostratigraphical associations (Kent, 1980). In short, the principal units of the Barberton Supergroup comprise the Onverwacht Group, the Fig Tree Group and the Moodies Group. The Onverwacht Group is chiefly an assortment of ultramafic and mafic submarine volcanics, including a number of sill-like, layered, ultramafic complexes. This group is overlain by turbiditic greywacke sandstones, associated mudstones and banded ferruginous shales of the Fig Tree Group and shallow-water clastics of the continentally-derived Moodies Group (Visser *et al.*, 1956). The general distribution of the regional geology of the Barberton area is depicted in Fig. 1.

Available radiometric ages (Kamo and Davis, 1991; Kröner *et al.*, 1991; Lopez-Martinez, 1992; de Ronde and de Wit, 1994) indicate that the Barberton cycle began between 3490 and 3470 Ma when voluminous komatiites, komatiitic basalts and tholeiites were extruded subaqueously. However, recent dating of felsic tuffs of the Theespruit Formation in the Steynsdorp anticline has provided an age of between 3547 and 3530 Ma (Kröner *et al.*, 1992) for the lowest formation of the Onverwacht Group (Viljoen and Viljoen, 1969a, b, c, d, e, f). Onverwacht volcanism continued, with a hiatus at the Middle Marker, until about 3445 Ma when the Hooggenoeg Formation felsic volcanic rocks were extruded and comagmatic tonalite-trondjhemite-granodiorite (TTG) plutons intruded the lowermost formations. Then followed a major hiatus in Onverwacht magmatism during which the Buck Reef chert and associated minor ferruginous shales were deposited at the top of the Hooggenoeg Formation. Subsequently, effusion of mafic tuffs and extrusion of ultramafic to felsic lavas of the Kromberg and Mendon Formations (Viljoen and Viljoen, 1969a, b, c, d, e, f; Lowe, 1991) terminated at about 3298 Ma (Kröner, 1993). There followed another period of quiescence during which the Msauli oölite horizon was deposited. Subsequent deposition of debris flow conglomerates, greywacke sandstones, mudstones, tuffs, carbonaceous shales and banded iron formations of the Fig Tree Group (*ca* 3260-3225 Ma) culminated in the eruption of arc-like felsic agglomerates, tuffs and lavas of the Schoon-

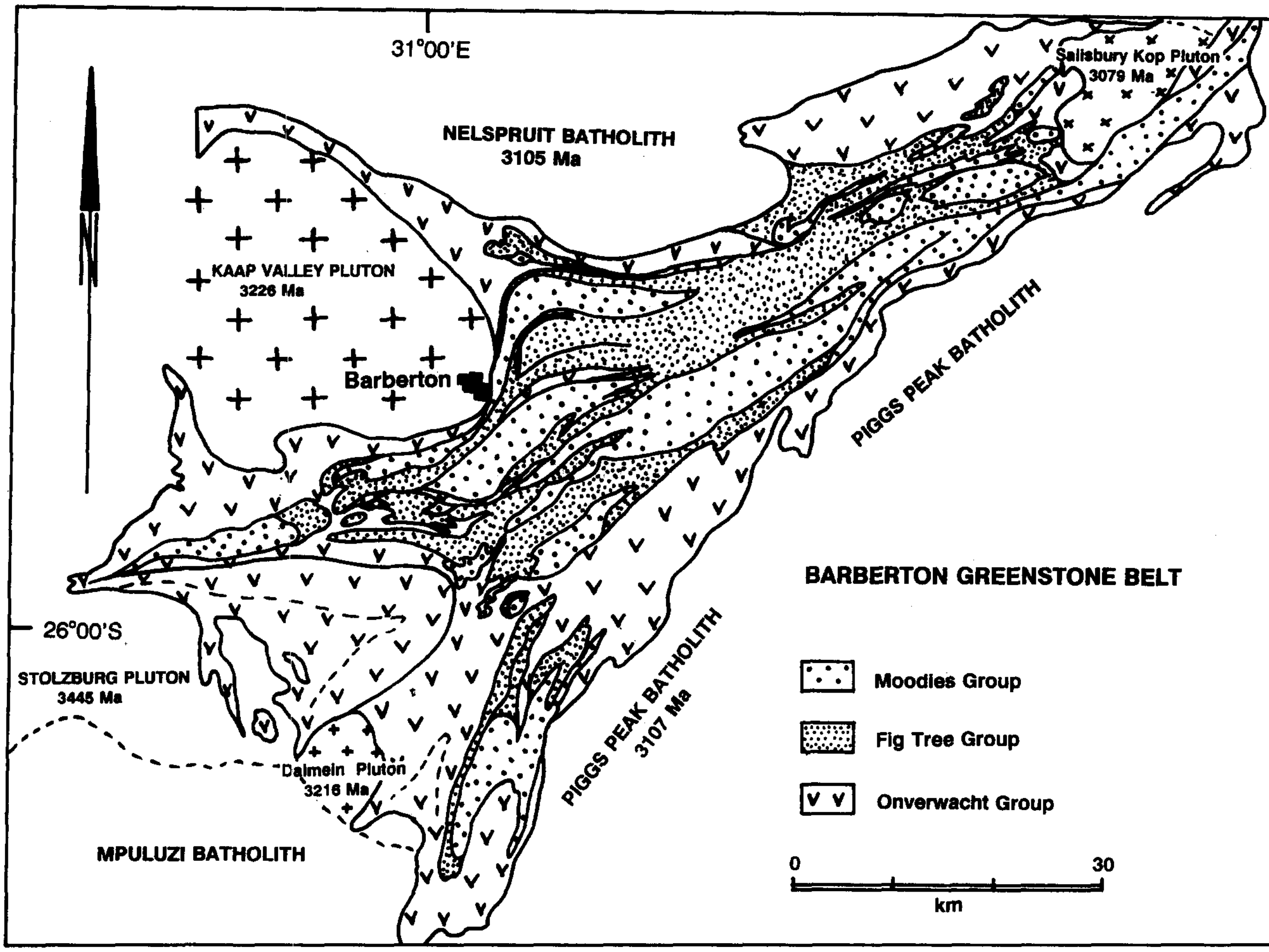

Figure 1. Generalized geology of the Barberton greenstone belt (after Kent, 1980).

gezicht Formation at the top of the group at about 3225 Ma (Armstrong *et al.*, 1990; Kröner *et al.*, 1991). Arenites of the Moodies Group were then deposited paraconformably on the underlying Fig Tree Group. Field evidence suggests that part of the continentally-derived Moodies sedimentation was syntectonic with respect to the main episode of deformation, which created the regionally-developed upright folds (Lamb, 1984, 1987; Heubeck and Lowe, 1994a, b). Figure 2 is a schematic representation of the Barberton stratigraphy showing the principal lithostratigraphic divisions, prominent marker horizons, and some key radiogenic ages.

Prior to deformation, the Barberton basin may have accommodated as much as 20 km in thickness of lavas and sediments. If this was the case, the widespread occurrence of pillow structures in the Onverwacht mafic lavas indicates comparatively rapid basin subsidence under subaqueous conditions in order to permit the accumulation of up to 15 km of ultramafic and mafic volcanics. Following extensive mapping of the Onverwacht Group in the type area (Viljoen and Viljoen, 1969a, b, c, d, e, f), it was proposed that the rocks be classified into a lower ultramafic unit and an upper mafic to felsic unit, each comprising a number of formations. During Onverwacht times there was an eruptive hiatus between the Komati and Hooggenoeg volcanicity, which is recorded by the Middle Marker sedimentary horizon, and, in geochronological terms, there was a major apparent hiatus at the top of the 3445 Ma Hooggenoeg Formation, as evidenced by the Buck Reef Chert Marker. Lowe and Byerly (1991) have identified serpentinized komatiites and other volcanics of the Mendon Formation (3298 Ma) as the uppermost unit of the Onverwacht Group along the western limb of the Onverwacht anticline and above the Buck Reef chert. In spite of some criticism (Williams and Furnell, 1979; de Wit, 1982; de Wit *et al.*, 1987), the broad outline of the field classification appears to be sound and it is evident that a great thickness of volcanics accumulated under subaqueous conditions in the lower part of the Barberton basin. The extent to which the Onverwacht stratigraphy was tectonically thickened as a result of thrust-related structural duplication or tectonic stacking has yet to be demonstrated and the postulation that the stratigraphy comprises juxtaposed accreted terranes has yet to be verified (de Ronde and de Wit, 1994; Lowe, 1994).

The geochemistry of the metavolcanics comprising

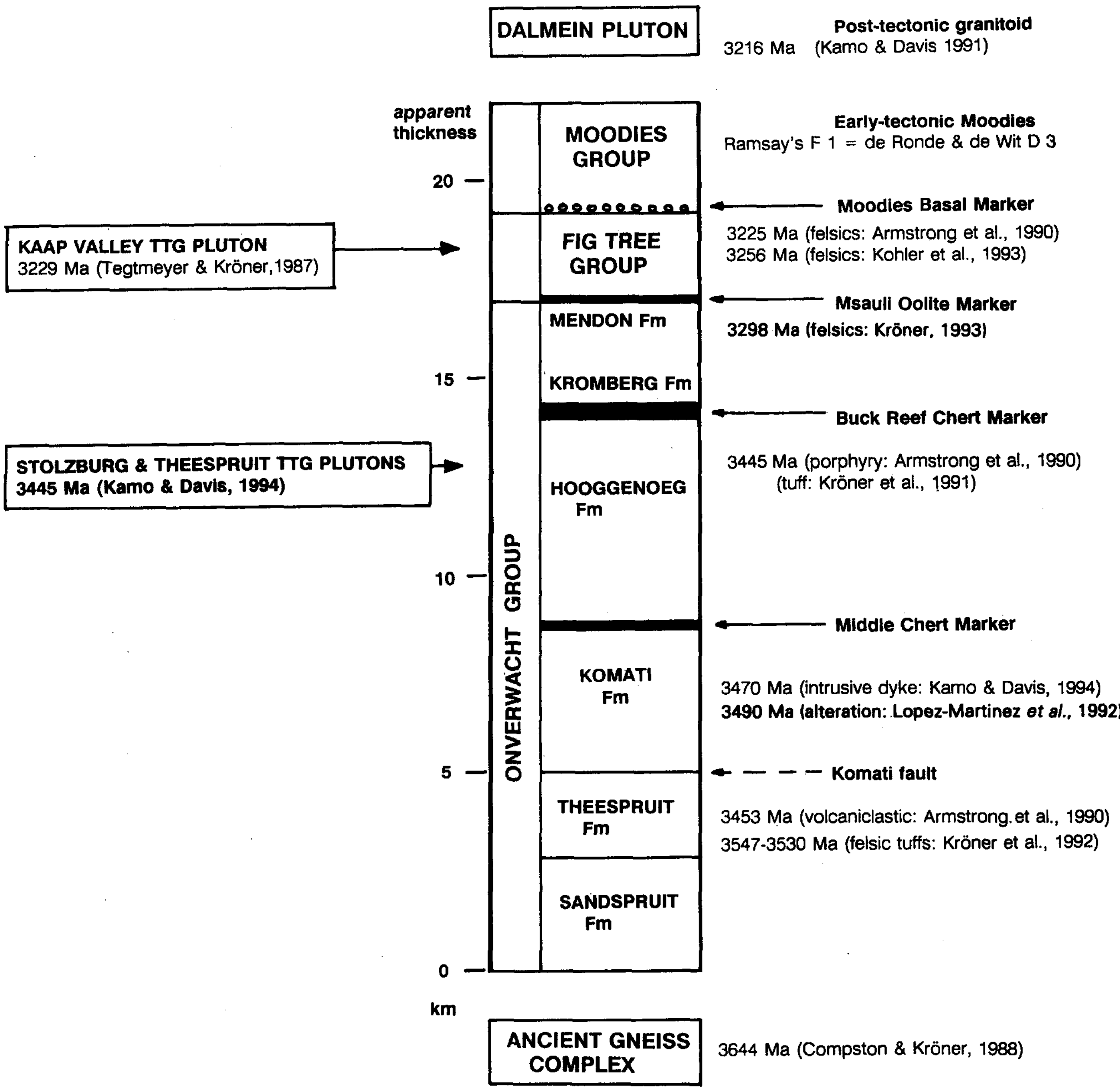

Figure 2. Schematic representation of the apparent stratigraphic column for the Barberton greenstone belt (modified after Lowe, 1982, 1991).

the formations of the Onverwacht Group has received considerable attention since the recognition of primitive ultramafic komatiite lavas (Viljoen and Viljoen, 1969a, b, c, d, e, f; Smith, 1980; Smith and Erlank, 1982; Smith *et al.*, 1980, 1984; Viljoen *et al.*, 1983). Anhaeusser (1981) provided a convenient excursion guide and general summary of the geology and geochemistry of the Onverwacht Group in the Komati valley type-sections. Alteration of Onverwacht lavas has been extensive and includes serpentinization, spilitization, carbonation and silicification. In the case of silicification, komatiites, komatiitic basalts and tholeiites have been so altered as to resemble felsic volcanics (de Wit *et al.*, 1982), to the extent that some silicified zones have been interpreted as the upper members of mafic to felsic volcanic cycles (Viljoen and Viljoen, 1969a, b, c, d, e, f). Identification of palimpsest pillow structure and spinifex textures characteristic of less-viscous mafic volcanics within these zones of silicification led to the recognition of the original nature of the flow-top alteration zones (Heubeck and Lowe, 1986a, b). However, similar zones of silicification have also been interpreted (de Wit, 1982) as fuchsitic décollement zones marking early thrust planes.

The structural and metamorphic complexity of the Onverwacht Group along the northern (Anhaeusser, 1963, 1972; Viljoen, 1963; Wuth, 1980) and southern or Swaziland margins of the greenstone belt (Barton, 1982) is such that the lithologies comprise amphibolites and mafic schists, serpentinites, talc-carbonate rocks, tuffs and agglomerates, felsic schists and porphyries together with minor intercalated sediments. Large, differentiated ultramafic bodies, which were either subvolcanic sills (Viljoen and Viljoen, 1969a, b, c, d, e, f; Anhaeusser, 1976a, b, c) or tectonic peridotites (Barton, 1982; de Wit *et al.*, 1987a, b), are prevalent in the greenstones of the northern and southern margin. Although the Onverwacht Group in the type area may be 15,000 m thick, in the Jamestown Hills area it is probably much less and in Swaziland the group is only 1500-2000 m thick (Barton, 1982). The Onverwacht Group underlying the northern margin of the greenstone belt has been correlated, in the main, with the Lower Onverwacht Theespruit and Komati Formations (Viljoen and Viljoen, 1969a Fig. 6). This correlation has yet to be verified by radiometric dating and it remains a possibility that the greenstones underlying the northern and southern margins of the greenstone belt may correlate with the Upper Onverwacht Mendon Formation (3298 Ma) rather than the Lower Theespruit and Komati Formations (3547-3460 Ma). Lowe (1991, 1994) has suggested that the proposed Weltevreden Formation to the north of the Inyoka fault may be equivalent to the upper part of the Mendon Formation.

The Msauli (Umsoli) oölite horizon in the type locality has been used to define the basal contact of the overlying Fig Tree Group (Heinrichs, 1980) and Jackson *et al.* (1987) have drawn attention to the fact that the "upper contact of the Onverwacht Group marks the extremely significant transition from exclusively oceanic sedimentation in a volcanic environment to sedimentation dominated by the massive influx of terrigenous debris". The Fig Tree Group has been divided into a deeper-water northern facies, in which turbiditic greywackes predominate, and, south of the regionally-extensive Inyoka fault, a shallower-water southern facies comprising shales and jaspilites with subordinate greywackes (Heinrichs and Reimer, 1977; Heinrichs, 1980). While the simple categorization of the Barberton Supergroup into three groups based on gross lithostratigraphical differences remains valid, at formation level there have been some additions, noted below, that modify the original Kent (1980) classification.

The lower and middle strata of the northern facies of the Fig Tree Group may be as much as 1800 m in thickness and comprise the Sheba and Belvue Road Formations (Reimer, 1967; Condie *et al.*, 1970; Anhaeusser, 1976a, b, c). The strata consist of massive volcaniclastic greywackes in the lower formation, grading upwards into finer-textured greywackes exhibiting a progressive increase in granitic detritus higher in the group, together with more abundant shale, chert and banded ferruginous shale horizons. The sedimentary textures have been interpreted as turbiditic in origin, but whether the depositional environment was deep or shallow-water is debatable (Kuenen, 1963; Eriksson, 1980). Condie *et al.* (1970) concluded that up to 50% of the Fig Tree greywacke is composed of chert and quartz and, in a diagram illustrating the secular variation in the content of rock fragments in greywackes from the Sheba Formation, Condie *et al.* (1970) emphasised the predominance of siliceous volcanic fragments over mafic fragments. Danchin (1967) noted the unusually high content of Cr and Ni in the Fig Tree shales and concluded that the sediments were derived from ultramafic source rocks.

The southern facies of the Fig Tree Group, which was subdivided after the Kent (1980) classification, comprises the ferruginous Ngwenya Formation, probably deposited under relatively deep-water conditions, overlain by the coarsening-upwards fan-delta sediments of the Mapepe Formation (Heinrichs, 1980). In the Mapepe Formation, Nocita and Lowe (1990) have recognized the onset of tectonism in the sedimentary record, or alternatively concurrent volcanic activity and concluded that the "depositional model is that of a fan delta building into a relatively shallow body of water in response to a developing fold and thrust belt". Following the lead of Jackson *et al.* (1987) they also concluded that the depositional environment was that of an evolving foredeep. The ferruginous strata of the underlying Ngwenya Formation indicate formation within a relatively stable basin (Heinrichs, 1980) prior to deposition of the coarsening-upwards fan-delta sediments of the Mapepe Formation, when the basin began to close gradually in response to regional compression. As a result of field work in the southeastern portion of the greenstone belt, Fig Tree Group equivalents underlying the Ngwenya range of Swaziland have been called the Diepgezet Group and comprise 1800 m of sediments, which have been interpreted as deep water, submarine fan deposits (Lamb and Paris, 1988).

Overlying the two lower formations of the Fig Tree Group, in both the northern and southern facies, is the volcanogenic Schoongezicht Formation. In the southern facies of this formation, Heinrichs (1980) recorded the felsic and calc-alkaline character of the metavolcanics, whereas in the Stolzburg syncline of the northern facies Reimer (1967) noted that the Schoongezicht Formation comprises 500 m of coarse-grained tuffs grading upwards into finer-grained tuffs capped by lava and agglomerates. Dacitic tuffs and agglomerate in the Fig Tree Group have been dated at 3259-3225 Ma (Armstrong *et al.*, 1990; Kamo and Davis, 1991; Kröner *et al.*, 1991) with the younger age (3225 Ma) being characteristic of the Upper Fig Tree Schoongezicht Group whereas the older age (3259 Ma) has been assigned to basal Fig Tree Group felsics (Kröner *et al.*, 1991). In the Ulundi-Eureka area, the Schoongezicht Formation is less well-developed and comprises tuffs and agglomerates (Anhaeusser, 1976a, b, c) together with shales, grits, cherts and banded iron formations having a total thickness of about 100 m. To the northeast, similar rocks, known as the Bien Venue Formation, have been dynamothermally altered to silicic schists and were probably originally rhyodacitic to rhyolitic in composition. They have been dated at 3256 Ma (Kröner *et al.*, 1991; Kohler *et al.*, 1993).

The arenaceous Moodies Group includes more distal shales, shaley sandstones, jaspilites and a thin horizon of amygdaloidal andesite north of the Inyoka fault, which are absent from the more proximal arenites in the southern facies (Visser *et al.*, 1956). Anhaeusser (1976) has pointed out that the Moodies sediments are continental in character and, north of the Inyoka fault, he defined three upward-fining cycles as the Clutha, Joe's Luck and Baviaanskop Formations. Jackson *et al.* (1987) have deduced that upward-coarsening sequences in the Moodies Group, and the recycled oceanic volcanic detritus from the Onverwacht in the Fig Tree Group imply an active tectonic setting, such as a foredeep or foreland basin, rather than the passive continental margin setting originally proposed by Eriksson (1980). Widespread evidence for desiccation in the form of mudcracks in the Moodies sediments attests to a shallow-water depositional environment. North of the Inyoka fault, the Moodies strata are up to 3700 m thick and Heubeck and Lowe (1994), adopting Anhaeusser's (1976) unit codes and classification, confirmed that the Lower Moodies Group north of the fault comprises a fining-upward sequence, which was derived from a provenance to the north.

Eriksson *et al.* (1988) observed that, whereas the Moodies Group conformably overlies the Fig Tree Group in the northern parts of the Barberton greenstone belt, in the southern section there is a marked unconformity between the groups. In the southern section, the Moodies Group is made up of a considerable thickness of conglomerate, grading upwards into sandstone while north of the Inyoka fault, in the Stolzburg, Saddleback and Eureka synclines, the Moodies Group is composed of one upward-fining and two upward-coarsening sequences. More recently, Heubeck and Lowe (1994) have concluded, on the basis of extensive field work, that the Lower Moodies was "deposited in one or more intramontane basins in an extensional setting." In the Upper Moodies, a southward-thinning fan-delta conglomerate, which was derived from the north and unconformably overlies lower strata, suggests that the Upper Moodies was deposited in a foreland basin in the initial stages of southward shortening and basin closure during a deformational episode that eventually incorporated most of the Barberton depository into a major fold-and-thrust belt (Heubeck and Lowe, 1994a, b).

In southern Swaziland, the Moodies Group has been called the Malolotsha Group and comprises about 1800 m of coarse siliciclastics with minor finer-grained sediments exhibiting desiccation cracks, which overlie the Diepgezet or Fig Tree Group with an angular unconformity and contain internal unconformities. The relationship of these progressive unconformities to regional fold structures has been interpreted to indicate the syndepositional timing of the deformation (Lamb, 1987).

STRUCTURE

The broader aspects of the deformation of the Barberton depository can be considered in terms of one of two models - either the structures represent a combination of early upright folding accompanied by high-angle thrusting (Anhaeusser, 1975), or they resulted from early tectonic stacking followed by a later event involving the rotation of early thrust slices into upright folds, accompanied by high-angle thrusting (de Wit *et al.*, 1987a, b). In other words, and in elementary terms, crustal shortening across the Barberton basin has been interpreted as being accommodated either by a relatively simple response to regional stress, whether dominantly vertical or lateral or a combination of both, involving the development of early upright folds during which later steep thrusting was initiated (Visser *et al.*, 1956; Anhaeusser, 1975, 1984, 1986), or by a more complicated combination of stress reaction involving early sub-horizontal thrusting and tectonic stacking, including the possibility of obduction of oceanic crust, followed, after a period of sedimentation, by later upright folding and steep thrusting (de Wit, 1982; de Wit *et al.*, 1987a, b; Tomkinson and King, 1991; de Ronde and de Wit, 1994).

Anhaeusser (1975, 1984) emphasised the effects of vertical tectonics and noted that the arcuate pattern of relationships between the greenstone belts and the surrounding granitoids resulted from "dominantly gravity-induced deformational styles that accompanied the emplacement of granitic magmas and diapiric plutons." According to this model, sialic diapir-modified gravity-induced structures resulting from the unstable ensimatic geotectonic setting of the developing greenstone belt and such diapirism produced structures that were superimposed onto lithologies that may have experienced a complex structural pre-history (Anhaeusser, 1984). Citing structural evidence from the New Consort and Sheba areas straddling the Eureka syncline, Anhaeusser concluded that almost the entire range of structural features resulted from granitic diapirism, but he conceded that in some granite-greenstone terrains diapir-

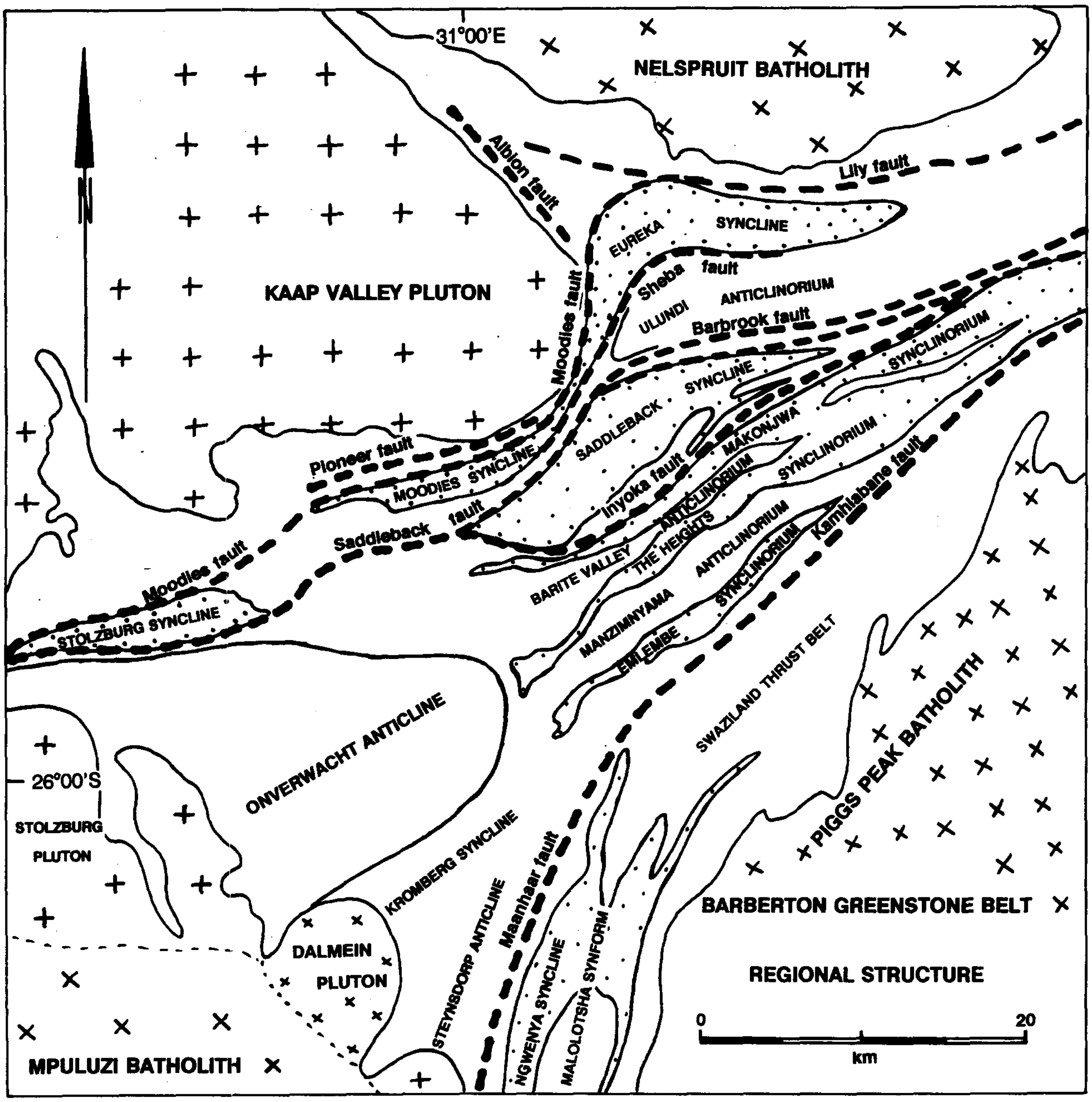

Figure 3. Central portion of the Barberton greenstone belt showing some of the major regional folds and strike faults mentioned in the text (modified after Visser, 1956; Anhaeusser, 1976a, b, c; Heinrichs, 1984; Heubeck and Lowe, 1994a, b). For stratigraphy see Fig. 1.

ism alone is insufficient to explain the structural history and that horizontal shortening, resulting from regional compression, must have been a contributory factor.

An integrated geophysical survey of the Barberton greenstone belt has revealed, as a result of a detailed gravity survey constrained by deep electrical resistivity soundings, that the maximum depth extent of the Barberton Supergroup is "less than 8 km, but always more than 4 km, on all profiles" (de Beer *et al.*, 1988). This evidence for the shallow, physical nature of the greenstone belt, which is shown as being underlain by undifferentiated granitoids, is a practical constraint on the extent of the depth to which vertical tectonics (sagduction) or horizontal tectonics (subduction) can be invoked as processes in the geotectonic evolution of this shallow fold-and-thrust belt. Metamorphic evidence from the Komati and Hooggenoeg Formations in the Onverwacht type area (Cloete, 1994) indicates that these formations were not subjected to pressures of greater than 4 kbar in anything other than a hydrostatic stress field due to burial metamorphism. Evidence for regional dynamic or dynamothermal metamorphism, which would tend to provide support for the interpretation that imbricate thrusting and tectonic stacking has thickened the Onverwacht greenstones, perhaps by as much as an order of magnitude (de Wit *et al.*, 1987a, b), is absent from the Onverwacht type locality.

In a traverse across the Barberton fold belt, from Piggs Peak in the southeast to Consort mine in the northwest, the effects of compression and basin inversion due to regional overthrusting from the southeast can be recognized in the kinematic evidence of the regional variation in structural style. Diminishing finite strain conditions from southeast to northwest may be interpreted from the imbricate thrust-stacking of the rocks underlying the Swaziland sector of the belt (Jones, 1969; Barton, 1982; Lamb, 1987), the regionally developed synclinoria and anticlinoria underlying the Lomati and Manzimnyama basin sector (Visser *et al.*, 1956) and the more open regional folds of the Saddleback and Eureka synclines and the intervening Ulundi anticlinorium (Visser *et al.*, 1956; Ramsay, 1963). In an analysis of strain using deformed pebbles in the Moodies conglomerate, and in apparent contradiction to the generalization that regional strain decreased in a northwesterly direction across the deforming Barberton greenstone belt, Gay (1969) concluded that the regional strain had been much greater in the northern part of the Mountain Land than in the centre. However, he suggested that this was probably because the rocks along the northern margin of the Eureka syncline have experienced two phases of deformation, whereas those in the Makonjwa synclinorium do not show cleavage, thus indicating that Ramsay's "second phase of deformation did not extend this far into the Mountain Land" (Gay, 1969, p. 392). Although he did not specifically exclude the possibility of earlier deformation, Ramsay (1963) referred to the upright Eureka syncline generation of folds, which includes the Makonjwa synclinorium (Visser *et al.*, 1956), as "the first recognised phase of deformation". Heubeck and Lowe (1994) produced a semi-quantitative strain map of the Barberton greenstone belt showing higher strain along the southern as well as the northern margin of the belt, but they did not consider the possibility of distortion of the regional strain markers occasioned by two episodes of deformation affecting the rocks underlying the northern margin.

Following investigation of the mineralized Sheba hills area, van Eeden (1941) concluded that the regional application of major stress from the southeast had resulted in a single period of deformation, during which overfolding of the rocks towards the northwest was distorted by the buttress effect of the Kaap Valley pluton. Publication of mapping by members of the Geological Survey of South Africa (Visser *et al.*, 1956) depicted the structural work of the time, which was most clearly displayed in the mapped stratigraphy of the Moodies Group. Thus, the Eureka and Moodies syncline, the Stolzburg, Saddleback, Lily and Hlambanyati synclines and the Makonjwa, Heights and Emlembe synclinoria were all plainly identified by regional mapping and recorded in the text. The periclinal character of some of the structures was implicit in the descriptive terms 'canoe shaped' and 'boat shaped' as well as 'inverted canoe' folds (Visser *et al.*, 1956, pp165-171). At the same time, major regional 'strike-faults' or 'longitudinal faults' were identified and shown on the 1956 geological map, including the Sheba, Scotsman, Barbrook, Saddleback, Inyoka and Kamhlabane faults. The location of many of these regional structures is shown in Fig. 3. In addition, the Ulundi, Barite Valley and Manzimnyama anticlinoria are depicted. These complex regional structures, affecting relatively incompetent Fig Tree strata, have been incorrectly shown as synclines in the past (Ramsay, 1963; Anhaeusser, 1976a, b, c; Heinrichs, 1984).

Ramsay (1963) published the first detailed structural analysis of the Eureka and Ulundi folds in the general area of the Consort, Sheba and Fairview Au mines. Ramsay concluded that the "first deformation gave rise to many folds whose steeply-inclined axial planes were probably initially oriented in a northeast-southwest direction. The second deformation resulted in the widespread development of slaty cleavage and schistosity which cuts obliquely across the first folds". As de Wit (1982) has noted, Ramsay's (1963) measurement of bedding-cleavage relationships led him to conclude that the cleavage-forming event post-dated a fabric-free fold forming episode. In a commentary on the regional faults in the area he examined, Ramsay (*op cit*, p. 392) concluded that the "great strike-faults seem to be developed in the overturned limbs of the first-fold anticlines" and he suggested that the faults were probably initiated during the first deformation, although they appeared to have been subsequently reactivated.

Ramsay's work was followed by a number of publications presenting the results of research undertaken by staff from the Economic Geology Research Unit at the University of the Witwatersrand (Anhaeusser, 1963, 1969, 1976; Anhaeusser *et al.*, 1967, 1969; Gay, 1969; Poole, 1964; Viljoen, 1963; Roering, 1965). With the exception of Roering, these authors accepted Ramsay's structural episodes and, in particular, his conclusion that the first episode of deformation produced the major regional synclines on which the subsequent development of cleavage and schistosity was superimposed. Roering (1965), on the other hand, suggested that the "systematic spread of fold axes within the axial plane indicates that this folding must have been superimposed on an already deformed strata". Anhaeusser (1976) also noticed the spread of the plunge of minor fold axes in the Eureka syncline and concluded that "some of the folds may represent a superimposed phase of deformation". It is now recognized that the variable plunge of fold axes within the axial plane can be indicative of sheath-like folding (Cobbold and Quinquis, 1980) and the variation in the plunge of fold axes noted by Roering (1965) may be due to the curvilinear character of the folds along the axes resulting from the differential contractive response of the developing folds to regional compression and "elongation of the (folded) beds radially upwards" (Visser *et al.*, 1956, p. 179). In a comment on the variable axial plunge displayed by folds of the first deformation (F_1), Ramsay (1963, pp371-2) noted that "it seems unlikely that the regional compression would be taken up by uniform shortening across the folds" and he went on to conclude that "the plunges of folds developing in adjacent zones of large and small strain might be expected to show considerable variation." In a recent study of structures in the Moodies Group, Heubeck and Lowe (1994) have commented that many small- to medium-scale folds in the underlying Fig Tree and Onverwacht rocks have approximately the same orientation as the post-Moodies folds affecting the overlying rocks and that "it was generally impossible to identify these structures as pre-Moodies or post-Moodies in age." In other cases, such as structures in the Fig Tree rocks within the Manzimnyama syncline, the authors found that "folds trend slightly obliquely to the large, regional folds and are apparently refolded by them." It remains to be seen whether this apparent evidence for pre-Moodies folding is real or whether it is more a function of the competence contrast between the Fig Tree and Moodies rocks in a structurally complex area (Heinrichs, 1980).

Sheath folds seem to result from the kinematic amplification of 'deflections' in folded layers, which produced curvilinear folds that became progressively more elongated during deformation (Lacassin and Mattauer, 1985). The steeply-plunging attitude of the fold axis of the Eureka and, especially, the Stolzburg synclines and the generally eye-shaped outcrop pattern of these folds promote the deduction that the structures are sheath-like, although the possible periclinal closures have been disrupted by transcurrent faulting. The fact that the folds in the Sheba anticlinorium, for example the Birthday anticlines where the Zwartkoppie Formation sediments are deformed into a series of parallel isoclines, also prove to be sheath-like, as demonstrated from exposures in the underground workings, serves to enhance the probability that during Ramsay's first fabric-free fold forming event (F_1), conditions were such that sheath-like folds developed on a local and regional scale in the supracrustal rocks under unusually ductile conditions of regional deformation. In relation to the major folds exposed in a narrow transect across the central Barberton mountain land, Heubeck and Lowe (1994, Fig. 4) demonstrate the variable plunge of fold axes within a general axial plane. Certainly what appears to be large-scale periclinal or sheath-like folds, showing axial culminations and depressions, have been described from the Superior Province of the Canadian Archaean (Dimroth *et al.*, 1983; Daigneault *et al.*, 1990) and this indication of crustal plasticity may prove to have been a feature typical of the Archaean 'permobile regime'.

An extraordinary structural feature of the Barberton greenstone belt is the occurrence of these sheath-like folds in rocks that have only experienced low-grade burial and regional metamorphism. It is concluded that this early phase of folding took place in a high-level geodynamic regime, within which pore fluids and a low strain rate promoted ductile conditions and facilitated the development of fabric-free sheath-like folds on a local and regional scale. The superimposed cleavage transects the fold axes in a clockwise sense in the Eureka syncline (Ramsay, 1963, Figs 18 and 20) and this characteristic, together with the apparent en echelon disposition of the major linear synclines, is interpreted to result from regional sinistral transpression (Blewett and Pickering, 1988; Murphy, 1985; Woodcock *et al.*, 1988; Ward, 1991; Pratt and Fitches, 1993). However, in a critical examination of the association between transected folds and transpression, Treagus and Treagus (1992) concluded that "a clockwise sense of transection should not be used as sole indicator of sinistral transpression (nor counter-clockwise for dextral) without supporting structural data". In the Barberton case, some additional corroboration is provided by the prominent sinistral drag evident in the Moodies sediments, adjacent to the Saddleback fault underlying the south-western corner of Dycedale farm. This distinctive left-lateral movement indicator can be observed clearly in the field and on the aerial photograph of the locality.

In another explanation of the structural evolution of the Barberton greenstone belt, it has been proposed (de Wit, 1982; de Wit *et al.*, 1983) that the field relationships in the southern mountain land reveal "an earlier period of deformation, manifested in the Geluk Subgroup and overlying stratigraphic units by regional recumbent folds, inverted stratigraphy, nappes and original sub-horizontal thrusts or glide planes" (de Wit *et al.*, 1983, p. 23). This has been interpreted by the authors to indicate that the 'upright structures' of Ramsay's main phase or first deformation folds (F_1) are, in fact, folds of a later episode of deformation related to sub-horizontal shortening, which was preceded by an episode of thrusting and nappe formation that resulted in the tectonic stacking of the stratigraphy. Geochronological research on the rocks at the base of the Onverwacht has provided some support for this contention in that an ion microprobe study of zircons, extracted from a volcaniclastic horizon in the Lower Theespruit Formation (Armstrong *et al.*, 1990), indicated that at 3453 Ma this formation could be younger than the overlying vol-

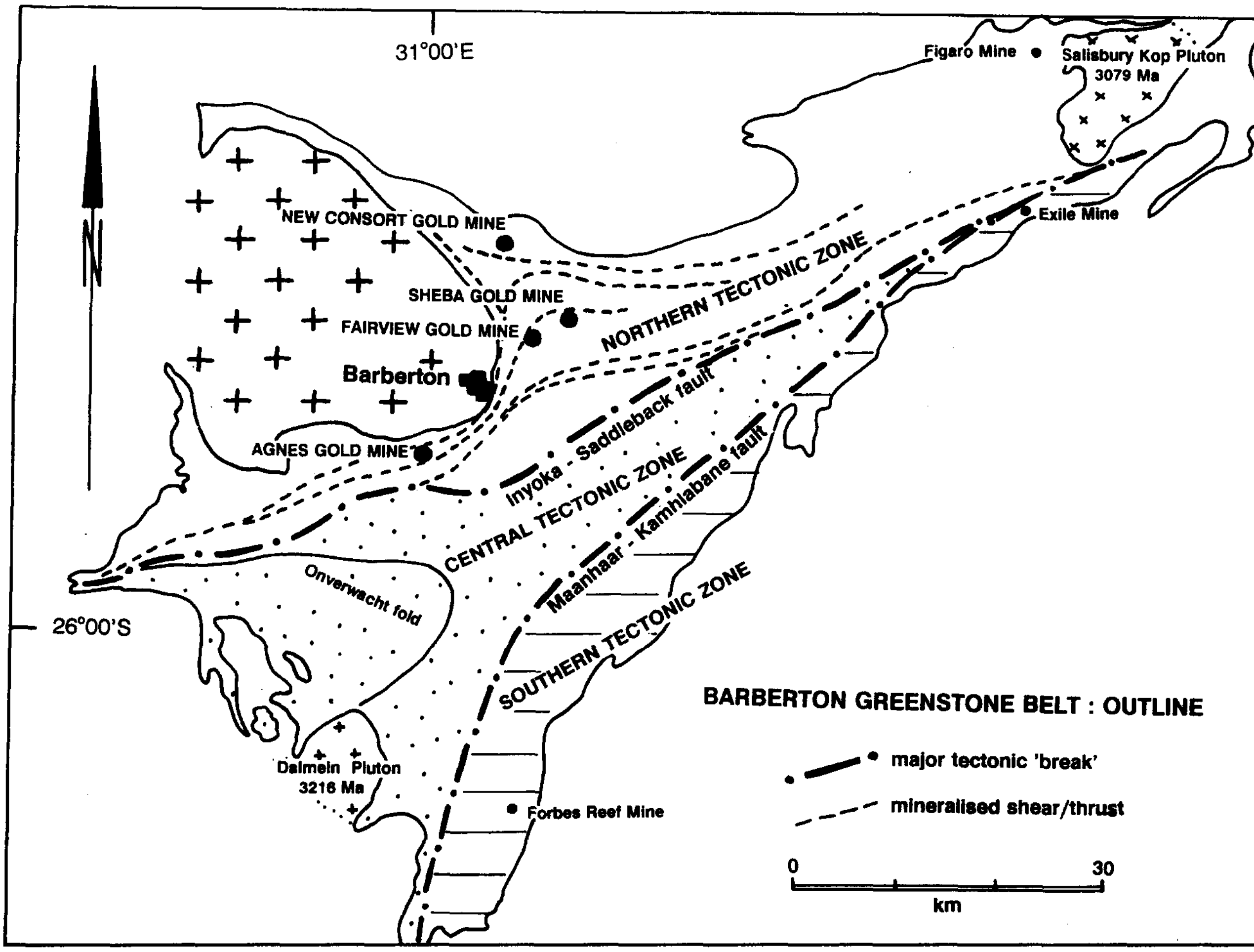

Figure 4. Diagram of the Barberton greenstone belt showing the proposed tectonic zones, as well as the location of four major Au mines and three minor Au mines mentioned in the text.

canics of the Komati Formation and the authors proposed that the two formations may have a major structural discontinuity between them. Subsequently, a maximum age of alteration in the overlying Komati Formation lavas has been measured using $^{40}Ar/^{39}Ar$ at 3490 Ma (Lopez-Martinez, 1992) and a quartz-feldspar porphyry dyke intruding the Komati Formation volcanics has been dated at 3470 Ma (Kamo and Davis, 1994). These data are compatible with the age of zircons from the Middle Marker horizon and a lower volcaniclastic unit in the Komati Formation that yielded ages of 3472 Ma (Armstrong *et al.*, 1990). Accepting the reliability of the 3453 Ma age obtained on zircons from the diamictite at the base of the Theespruit Formation means that the major shear zones in the Komati valley represent planes along which tectonic stacking took place (de Wit *et al.*, 1983). However, detailed mapping at 1:10 000 (V. King, *unpubl.*) clearly demonstrates that the Komati fault is not folded by the Onverwacht bend or anticline and therefore can hardly have acted as the plane along which the Komati Formation was thrust over the Theespruit Formation prior to this deformation. In addition, there is now doubt as to the validity of the younger 3453 Ma age of the Theespruit Formation following work in the Steynsdorp anticline, where samples of felsic tuff have yielded ages of 3547-3530 Ma (Kröner *et al.*, 1992). The fact that the Buck Reef chert at the top of the Hooggenoeg Formation can be followed along the strike for more than 30 km without any indication of the duplication that would result from imbrication along sub-horizontal thrusts also lends little support to the postulation that extensive early tectonic stacking thickened the Onverwacht Group as a whole.

Using a modern geotectonic approach, de Ronde and de Wit (1994) proposed that the Barberton greenstone belt comprises two terrains: Arc and Trench Block I (3445-3416 Ma), which outcrops south of the Inyoka-Saddleback fault system boundary, and Arc and Trench Block II (3259-3222 Ma), which outcrops north of the boundary. As a caveat, the authors note that there is insufficient evidence to define whether the terrains are allochthonous with regard to each other or whether they were part of a larger coherent geologic province which has been subsequently tectonically disrupted. Hence, the writers make use of the term 'block', which does not have the allochthonous connotations often associated with the term 'terrane'. Recently, Lowe (1994) has proposed that the Barberton greenstone belt can be "divided into four tectono-stratigraphic blocks that become younger towards the northwest" and suggested that the Barberton belt grew as a result of magmatic accretion "along the rifted margins of older blocks, and not the tectonic assembly of unrelated terranes along a subduction zone." This postulation readily lends itself to testing by more detailed radiometric dating of the suggested tectono-stratigraphic terranes. Heubeck and Lowe (1994, Table 2) assert that, according to the field evidence from a transect across the central portion of the Barberton greenstone belt, the fold-and-thrust belt can be subdivided into four major structural blocks, according to the orientation and plunge direction of the major fold axes. However, the plunge of the major and minor fold axes varies along strike within the general east-northeasterly striking axial planes, therefore the apparent structural rule from the central transect cannot be regionally extrapolated along strike as a characteristic feature of the proposed four tectonic domains. This is clearly demonstrated by a comparison of the structural evidence from the Heights syncline, underlying the transect in which the fold axes plunge to the northeast (Heubeck and Lowe, 1994a, b), and the Heights synclinorium, underlying the farms Vooruitzicht and Duurstede further along strike to the east, in which the fold axes plunge both to the northeast and the southwest (Visser *et al.*, 1956).

In the present synthesis of Barberton geology, it is convenient to use the term tectonic zone, within which there are distinguishable tectonic domains, in analysing the regional tectonics of the greenstone belt. To this end, Fig. 4 has been prepared to show the situation of the Inyoka fault and the Maanhaar fault which, as regional tectonic breaks, are used to distinguish three tectonic zones as depicted: the Northern Tectonic Zone (NTZ); the Central Tectonic Zone (CTZ) and the Southern Tectonic Zone (STZ). The CTZ and STZ equate to Arc and Trench Block I and the NTZ to Arc and Trench Block II of de Ronde and de Wit (1994). The basis for recognizing the three tectonic zones stems from the obvious difference in the style of regional deformation between the imbricate thrust-stacking characteristic of the STZ (Hunter and Jones, 1969; Jones, 1969; Barton, 1982), the regional anticlinoria and synclinoria as well as the Onverwacht anticline domain underlying the CTZ (Visser *et al.*, 1956; Heinrichs, 1980; Heubeck and Lowe, 1994a, b) and the more open, regional, sheath-like folds typical of the NTZ, which also includes the unique Jamestown hills tectonic domain (Visser *et al.*, 1956; Anhaeusser, 1969; Heubeck and Lowe, 1994a, b).

In post-Onverwacht lithologies, Lamb and Paris (1988) correlated both stratigraphy and structure across the Maanhaar fault and thereby across the boundary between the STZ and the CTZ. Similarly, the Geological Survey (Visser *et al.*, 1956) found no reason to question the correlation of post-Onverwacht lithology and deformation between the CTZ and the NTZ across the Inyoka fault. Subsequently, Heubeck and Lowe (1994) have reported that thinner Moodies sections to the south of the Inyoka Fault may correlate with the basal Moodies Group to the north of the Inyoka Fault, but were probably deposited in separate basins. In the same way that the Saddleback and Inyoka faults are shown to unite, so too the Maanhaar-Kamhlabane fault is shown to join the Saddleback-Inyoka fault on the Singerton farm (Fig. 4). The lithological and structural differences and similarities across the tectonic boundaries, which have been used to distinguish the NTZ, CTZ and STZ, indicate that the Barberton greenstone belt, certainly in post-Onverwacht times, probably comprised a single complex depository rather than a collage of two or three comparatively small allochthonous terranes. Support for a single depository is provided by the similarity in $^{207}Pb/^{206}Pb$ zircon ages of 3225 Ma obtained from a dacitic conglomerate collected from the type-section of the Schoongezicht Formation north of the Inyoka fault and a dacitic tuff collected from the Upper Fig Tree Group in the Barite valley anticlinorium to the south of the Inyoka fault (Kröner *et al.*, 1991).

The late-tectonic Dalmein granodiorite pluton (3216 Ma; Kamo and Davis, 1991) outcrops in the southwestern extremity of the CTZ where the intrusive truncates the Kromberg fold structure and the eastern limb of the Onverwacht anticline (Viljoen and Viljoen, 1969e, Fig. 8, p. 130), otherwise termed the Kromberg antiformal syncline and the Onverwacht bend (de Wit *et al.*, 1987a, b). In the de Ronde and de Wit interpretation, the Kromberg structure is "most likely of D_2 age" and the Onverwacht fold is "almost certainly" D_3 in age. However, the authors suggest that "early D_3 deformation may represent a continuum of D_2 deformation as folds related to both these phases are commonly co-axial". According to the age of the Dalmein pluton, the D_3 deformation was accomplished by 3216 Ma, when structures of this generation were truncated by this discordant granitoid intrusive.

To minimise confusion, it should be stated that in this paper the conventional structural prefix 'F' is used for the deformation episodes recognized by Ramsay (1963) whereas the conventional prefix 'D' is used for the episodes recognized by de Ronde and de Wit (1994). Correlation of these episodes of deformation between the structures in the Southern Tectonic Zone (Jones, 1969; Lamb, 1984), the Central Tectonic Zone (e.g. Visser *et al.*, 1956; Heubeck and Lowe, 1994a, b; de Ronde and de Wit, 1994) and the Northern Tectonic Zone (Ramsay, 1963; Anhaeusser, 1969: Heubeck and Lowe, 1994a, b) is no easy matter and may best be attempted in a semi-quantitative, mechanistic and regional manner rather than on a rigidly statistical basis. In formulating a classification of the episodes of deformation affecting the Barberton greenstone belt (BGB), de Ronde and de Wit (1994) state that "D_3 deformation affects all of the BGB stratigraphy. It is most pronounced along the flanks of the BGB and particularly in the central and northern regions of the belt; this phase of deformation accounts for the NE-SW map pattern of the belt. D_3 deformation is responsible for the approximately NE-SW striking open-syncline/tight-antiform pairs with related thrust- and significant strike-slip components". From this description it is apparent that the de Ronde and de Wit D_3 category of deformation probably includes the earlier-defined main phase or first deformation F_1 (Visser *et al.*, 1956; Ramsay, 1963), which was responsible for the development of the prominent northeasterly-trending regional synclines cored by Moodies arenites. The field evidence for syndepositional deformation during Moodies times (Lamb, 1984; Heubeck and Lowe, 1993, 1994a, b) has been related to sub-horizontal regional compression, possibly resulting from sialic plate collision (Lamb, 1984, 1987; Jackson, 1984; Lamb and Paris, 1988). In agreement with the age of the transgressive Dalmein pluton, Kamo and Davis (1994) have concluded, on the basis of U-Pb dating, that much of the deformation of the Barberton greenstone belt "can probably be accounted for by a single short-lived compressional event within the time span 3230-3215 Ma".

Until recently, comparatively little structural work has been undertaken in the northeastern portion of the Barberton greenstone belt. Ward (1979, 1987) reported from the Figaro mine area near Malelane that the whole sequence has been deformed about northeasterly trending axes, resulting in the development of tight to isoclinal folds (F_1), which can be observed in the field on a scale ranging from that of an outcrop to regional synclines exhibiting a wavelength of 2 km or more. The Minerals Division of Shell, South Africa undertook a thorough prospecting operation for base metals on the farm M'hlati, about 6 km southwest of Figaro mine, including detailed mapping and extensive surface diamond drilling. The structure of the area was evaluated by Vearncombe (1986), who recognized one major and a second minor deformation and noted that both "small and large scale folds of the first deformation are noncylindrical, northward verging and overturned" and that the "mineralised southern synform is probably boat shaped with a flat keel and about 500 m deep". This detailed information, supported by drilling sections, from the Malelane area serves to confirm Ramsay's finding with respect to the primary nature of the main phase or F_1 deformation, which apparently correlates with de Ronde and de Wit's (1994) D_3 in the southern part of the mountain land. Fripp *et al.* (1980) suggested, but did not confirm, that the "early deformations in the northwestern part of the belt involved the formation of fold nappes and thrusts", which, in the case of the Eureka syncline, became steepened and transected during the diapiric rise of the Kaap valley pluton. However, neither the definitive work by Ramsay nor that undertaken by more recent researchers in the field has offered any proof in support of Fripp's structural model.

In an attempt to explain on a regional basis the fact that Ramsay's (1963) first or main phase deformation F_1, that correlates with the D_3 of de Ronde and de Wit (1994), seems to have succeeded two earlier episodes of deformation, D_1 and D_2, it is pertinent to remember that the recognition of the D_2 episode of deformation stems from detailed fieldwork in the Montrose-Mendon area (de Ronde, 1991). In this case "D_2 deformation is responsible for the effects of tight-to-isoclinal folds manifested in Fig Tree Group shales and BIF, and also for the regionally extensive E-W striking shear zones (predominantly thrusts) in the central parts of the belt" (de Ronde and de Wit, 1994). Examination of the field area shows it to be in the hangingwall of the western limb of the Onverwacht 'bend' where there must have been a considerable space problem during the formation of this large regional F_1 (=D_3) fold structure, therefore it is feasible that the D_2 deformation occurred in response to the formation of the Onverwacht fold. This leaves the question of the D_1 deformational episode that apparently involved early sub-horizontal thrusting and tectonic stacking, including the possibility of obduction of oceanic crust (de Wit, 1982; de Wit *et al.*, 1987a), and which affected the type-area of the Onverwacht Group in a sub-domain of the CTZ. The D_1 structural domain has been recognized only below the Buck Reef chert unconformity and may be peculiar to the type-Onverwacht submarine volcanics.

The deformational episode, during which the Eureka syncline and similar structures of Ramsay's F_1 (=D_3) category were formed, took place during and after the emplacement of the Kaap valley tonalite at about 3229-3227 Ma (Tegtmeyer and Kröner, 1987; Kamo and Davis, 1994) and before the intrusion of the discordant Dalmein pluton at 3216 Ma. During regional convergence and the compressional event that formed the upright F_1 folds, the early-tectonic Kaap valley tonalite acted as a diapiric-like structural salient, which indented the deforming greenstone belt. The resulting space problem was accommodated by folding, refolding, faulting and thrusting (Anhaeusser, 1984), including the development of a positive flower structure between the transected Moodies and Saddleback synclines (Ward, 1992), and by prominent secondary inflexion of the Eureka and Moodies synclines. Subsequently, and probably during relaxation of the regional compressive stress, extensive granitic magmatism at about 3105 Ma resulted in the emplacement of remobilized TTG sial in the form of the Piggs Peak-Mpuluzi batholith, along the southern flank of the greenstone belt, and the Nelspruit granite, gneiss and migmatite along the northern flank (Hunter, 1973; Robb *et al.*, 1983; Kamo and Davis, 1991).

Structural investigations at the New Consort Au mine (Tomkinson and Lombard, 1990; Harris *et al.*, 1992) indicate three phases of deformation of which the first is recognized as isoclinal folds and associated tectonic slides, which probably equate with Ramsay's F_1 or first structure and thereby are analogous to the Zwartkoppie anticlines at Sheba Au mine. The possibility of this correlation was considered by Ramsay (1963, p. 371), but metamorphic recrystallization of the New Consort host rocks is such that primary way-up criteria have been obliterated, thereby complicating the recognition of the structural relationship between the Consort bar and the Hangingwall bar. The second episode of deformation is represented by north-south to northwest-southeast striking shear zones, such as the Shires shear zone (Tomkinson, 1991), which are steeply dipping and deform the earlier structures. Later, moderately dipping, normal fault zones, such as the Bluejackets fault, displace the previous structures and constitute a third episode of deformation in the mine.

After the main phase of Au mineralization and the emplacement of the late-tectonic plutons, the Barberton granitoid-greenstone terrain was riven and invaded by a swarm of northwest-striking mafic dykes of pre-Transvaal Supergroup age that follow the 2870 Ma (Hegner *et al.*, 1984) Usushwana mafic complex trend. The swarm is normal to the northeasterly structural trend of the greenstone belt and the dykes probably afford a convenient post-tectonic indication of the maximum principal stress direction during the Barberton deformation episode that developed the northeast-southwest striking regional folds (see Fig. 5).

Finally, in an endeavour to conceptualize the geotectonic setting of the Barberton orogeny, it may be useful to consider the general evidence that is available for reconstructing the evolution of the Barberton depository. It is axiomatic that a knowledge of the floor to any sedimentary basin is a prerequisite for the tectonic classification of the basin (Einsele, 1992; Mitchell and Reading, 1978). In the case of the Barberton depository, it is clear that the siliciclastics of the Fig Tree and Moodies Groups were deposited on ultramafic, mafic and lesser felsic volcanics of the Onverwacht Group. Here one should recall that the dacitic tuff in the Upper Hooggenoeg Formation has been dated at 3445 Ma (Kröner *et al.*, 1992), whereas the upper part of the overlying Mendon Formation has been dated at 3298 Ma (Byerly *et al.*, 1993). In between, the Buck Reef chert horizon at the top of the Hooggenoeg Formation marks an interval in the Onverwacht volcanism that may be as great as 100 Ma. Another hiatus of as much as 50 Ma occurred between the 3298 Ma Mendon Formation and the dacitic tuffs at the base of the Fig Tree Group, which have been dated at 3243 Ma, or the proximal breccias and tuffs in the Fig Tree which have been dated at 3256 Ma (Kröner *et al.*, 1991)

In terms of the interpretation of Japanese accretionary complexes of Cretaceous age, the considerable time interval between the Upper Hooggenoeg Formation and the Mendon Formation, and between the Mendon Formation and the overlying terrigenous sediments of the Fig Tree Group, may be analogous to the case when there is a difference in age between ocean-floor basalts and terrigenous sediment, which is interpreted to indicate the "subduction of an actively spreading mid-ocean ridge", whereas the similarity in age between the ocean-floor basalts and associated terrigenous sediments is taken to indicate proximity to a trench (Osozawa, 1994). This line of reasoning is more in agreement with Cloete's (1994) suggestion that, although the Komati and Hooggenoeg Formations do not exhibit morphological, structural or volcanological characteristics of typical mid-ocean ridge crust, they may represent an oceanic plateau allochthonously thrust onto the back-arc basin stratigraphy of the Theespruit Formation. On the other hand, they may not represent anything of the kind. At an age of 3547-3530 Ma (Kröner *et al.*, 1992; Lowe, 1994) the Theespruit Formation may still constitute the oldest and lowest formation in the Onverwacht Group (Viljoen and Viljoen, 1969a, b, c, d, e, f). If this proves to be the case then the modelling of the basaltic and felsic volcaniclastics of the Theespruit Formation as being consistent with sedimentation in a back-arc basin (Cloete, 1994) could be relevant to the identification of the initial geotectonic setting of the Barberton depository.

The nature of the basement to the Onverwacht eruptives has been obscured by subsequent tectonism and by invasive granitoids which encompass the greenstone belt. Unless the Onverwacht Group itself represents primitive oceanic crust, whether allochthonous or autochthonous, the alternative and reasonable deduction that is left is that the intrusive granitoids represent remobilized tonalite-trondhjemite-granodiorite (TTG) sialic basement. Apart from the evidence for pre-Onverwacht primitive sialic crust in the Ancient Gneiss complex, the older TTG granitoids and the early-tectonic Kaap valley magmatism indicates TTG-style calc-alkaline oceanic or continental arc granitoid plutonism at about 3445 Ma and again at 3227 Ma, if uniformitarian comparisons were valid in the Archaean. This was succeeded by extensive potash-rich magmatism at about 3105 Ma in the intrusion of the Nelspruit, Mpuluzi and Piggs Peak batholiths, which were derived by extensive partial melting of a heterogeneous crust comprising tonalite and trondhjemite gneisses (Robb *et al.*, 1983). The Barberton greenstone belt was preserved by sialic underplating resulting from or reinforced by episodic felsic plutonism at about 3445 Ma, 3227 Ma and 3105 Ma. In this case, the uniformitarian evidence of the remobilized and reconstituted sialic floor to the Sarmiento complex of the Cretaceous back-arc basin in southern Chile (Bruhn *et al.*, 1978; de Wit and Stern, 1981) has considerable modelling relevance to the interpretation of the geotectonic history of the Barberton basin, which may also have been wholly or partially ensialic. Certainly, although not undisputed (Kusky *et al.*, 1994), the evidence from younger Archaean lithologies in Zimbabwe has been interpreted to indicate that the komatiitic greenstones overlie sialic basement above an unconformity and that the Ngezi Group of the Belingwe greenstone belt accumulated on a tonalitic basement (Bickle *et al.*, 1975; Blenkinsop *et al.*, 1993). Therefore, the presence of komatiitic greenstones may not necessarily infer an oceanic environment to the exclusion of proximity to sialic basement.

In the same way that there is no evidence for oceanic crust older than the Onverwacht volcanics, nor is there an obvious sheeted dyke complex (de Wit *et al.*, 1987a, b; Bertrand *et al.*, 1993) to demonstrate beyond doubt that the Onverwacht itself represents a small or large oceanic spreading centre, so too there is little or no evidence for a sialic floor to the Onverwacht Group. However, circumstantial evidence in the form of the contiguous 3644 Ma (Compston and Kröner, 1988) Piggs Peak inlier of the Ancient Gneiss complex of Swaziland, provides a clear indication of sialic crust that is older than the basal Onverwacht by some 150 Ma. In the greenstones of the Theespruit Formation, Lower Onverwacht, a small but significant tectonic wedge of tonalitic gneiss has yielded zircon ages of 3538 Ma (Armstrong *et al.*, 1990) and this has been interpreted as being the basement to the Theespruit. However, felsic rocks in the greenstones near the margin of the 3509 Ma Steynsdorp pluton have provided an age of 3547 Ma (Kröner, 1993) and may indicate that either older Onverwacht or older basement is more prevalent along the southwestern margin of the greenstone belt than formerly recorded (de Ronde and de Wit, 1994). On the other hand, the wedge of tonalitic gneiss in the Theespruit may be a tectonized intrusive, marginally younger than the Theespruit and the 3547 Ma age could well represent the age of the Theespruit Formation (M. Viljoen, *pers. comm.*). In this case, the evidence is dealing with the age of older Onverwacht rather than older basement and the younger apparent age from the Theespruit of 3453 Ma (Armstrong *et al.*, 1990) requires explanation.

Did the Onverwacht lavas pile-up in a mid-continental rift similar to the setting of the Proterozoic Keweenawan flood basalts (Green, 1983), or did the Onverwacht accumulate in a continental margin setting, in an arc-trench environment, in a back-arc basin, or does the Onverwacht represent primitive oceanic crust which was obducted onto continental crust and thereby preserved from mantle engulfment? Everything else in the geotectonic interpretation of the Barberton depository is secondary to the one fundamental uncertainty as to whether the floor to the Onverwacht was originally sialic or simatic and some progress in the testing of these alternatives may be achieved through isotopic studies of the granitoids in the future (Hoffman, 1990). Comparison of such data with isotopic evidence from the remobilized and reconstituted sialic floor to the Chilean Sarmiento complex (de Wit and Stern, 1981) may provide the key to the Barberton enigma with regard to determining the most likely nature of the original floor to the depository. It should then be easier to decide whether the Barberton orogeny was ensialic or ensimatic or a combination of both; whether the orogeny was diachronous or largely synchronous; whether the regional structures developed in response to several separate episodes of deformation between 3445 Ma (D_1) and 3215 Ma (D_3) or whether they resulted from a single compressional episode between 3230 and 3215 Ma (F_1). In either case, the geophysical, structural and metamorphic evidence indicates that the compression and inversion of the Barberton fold-and-thrust belt occurred in a low-pressure high-level geodynamic regime with comparatively little burial or subsequent exhumation of the original depository.

METAMORPHISM

The metamorphic map of South Africa (1992) depicts the Barberton greenstone belt as comprising a central zone of very low-grade zeolite facies terrain that is surrounded by a narrow margin of greenschist facies lithology with a very thin peripheral band of amphibolite facies rocks, which is developed along the contact between the greenstone belt and the surrounding invasive granitoids of different ages.

Following the regional mapping of the geology of the Barberton greenstone belt, Hall (1918) described contact metamorphic rocks which outcrop around almost the entire granitic margin of the belt. Along the southeastern border, from the Figaro and Exile mines in the northeast to the Forbes Reef area in the southwest (Fig. 4), Hall noted the development in appropriate host rocks of andalusite and ottrelite, chiastolite, staurolite, garnet and tourmaline. He reported the occurrence of the same metamorphic index minerals from suitable host rocks underlying the northern margin of the greenstone belt, from the New Consort mine eastwards, with the addition of sillimanite and cordierite.

Visser *et al.* (1956) confirmed Hall's observations with regard to the index minerals developed along the northern margin of the greenstone belt, with the exception of cordierite, but they also noted the prevalence of schists in the Onverwacht. The Komati valley, Jamestown schist belt and the Kaapmuiden-Malelane areas were described as being underlain by a wide variety of chlorite-talc-amphibole schists. Viljoen (1963) mapped the area to the east of the New Consort mine and produced a diagram showing the distribution of greenschist facies, albite-epidote hornfels facies, and hornblende hornfels facies lithologies with increasing grade towards the margin of the Nelspruit batholith. In fact, the tremolite-diopside skarn exposed underground in the New Consort mine (Tomkinson and Lombard, 1990) extends the hornblende hornfels or amphibolite facies metamorphism further south of the Nelspruit granite than originally shown by Viljoen. Anhaeusser (1972) observed that the Jamestown schist belt "has suffered extensive dynamothermal metamorphism that owes its origin mainly to the emplacement of the Kaap valley granite in the south. In addition, there appears to be superimposed on the regional metamorphic event, a thermal or contact metamorphic episode related to the Nelspruit Granite." The regional development of thermal metamorphism along the northern and southern margin of the Barberton greenstone belt may have resulted from the intrusion of the 3105 Ma potash-rich granitoids. This appears to have been superimposed on earlier dynamothermal metamorphism that probably accompanied the deformation of the greenstone belt and the emplacement of the early-tectonic Kaap valley TTG pluton at 3227 Ma. An even earlier episode of metamorphism must have been associated with the intrusion of the 3445 Ma Stolzburg and Theespruit TTG plutons, which are now preserved along the southwestern margin of the greenstone belt (Lopez-Martinez *et al.*, 1992).

De Wit *et al.* (1982, pp1784-87) drew attention to the extensive silicification affecting the cherts and volcanics of the Hooggenoeg Formation and which can be identified as metasomatism due to the preservation of palimpsest sedimentary and igneous textures in the affected rocks. Paris *et al.* (1985, p. 127) concluded that hydrothermal alteration and leaching of pillow lavas accompanied the convection of sea water through the Archaean oceanic crust, thus promoting the silicification of the upper part of the volcanic pile. This metasomatism led earlier workers (Viljoen and Viljoen, 1969a, b, c, d, e, f) to describe as pillowed dacites and rhyodacites cream-coloured silicified lava horizons which have been shown by de Wit *et al.* (1982) to display relict spinifex texture and ocelli, as well as pillow structures, and the horizons can be traced into less-altered serpentinites and mafic lavas. Subsequently, Lowe and Byerly (1986) and Duchac and Hanor (1987) agreed that these bleached flow-top alteration zones which exhibit remnant spinifex texture probably resulted from sub-surface hydrothermal alteration of submarine komatiitic lavas during periods of volcanic quiescence. Exhalative products of this sea floor metasomatism probably included stratiform barytes deposits (Heinrichs and Reimer, 1977) and sea floor hydrothermal vents or ironstone pods (de Wit *et al.*, 1982, 1987). Fluid inclusion work indicates that the formation of the ironstone pods may have been linked to the hydrothermal circulation of evaporitic brines (de Ronde *et al.*, 1994).

Although not proved beyond doubt, it seems probable that pervasive serpentinization, carbonation, spilitization and silicification were metasomatic processes that resulted from sub-surface hydrothermal convection (de Wit and Stern, 1976) in rocks beneath the Barberton sea floor. Initially, percolating aqueous solutions would have been heated by the unusually high temperature of the pile of rapidly-extruded submarine komatiitic volcanics. Therefore, attendant sea floor metasomatism in the Archaean may not necessarily signify proximity to an associated high heatflow Phanerozoic-style oceanic spreading centre marked by gabbro and a sheeted dyke complex. Cloete (1991, 1994) suggested that an early high-temperature episode of hydration and mostly greenschist facies sea floor metamorphism affected both the Lower and Upper Onverwacht Group strata, whereas lower-temperature pervasive silicification was largely confined to rocks in the Upper Onverwacht, which tend to be more siliceous in composition.

The complicated tectonic history of the Barberton greenstone belt has been accompanied by an equally complex metamorphic record that ranges from the early development of sea floor metasomatism, through extensive burial or static metamorphism, to regionally-developed dynamothermal metamorphism, associated with the periodic emplacement of TTG plutons, and superimposed peripheral thermal metamorphism related to the intrusion of the younger potash-rich granitoids.

Following the determination of the composition of minerals in pillow lavas of the Komati Formation, Cloete (1991, 1994) concluded from the results that, at approximately 4 kbar, conditions at the base of the Komati Formation correspond to the apparent overlying stratigraphic thickness of 15-16 km. The common preservation of primary textures in these igneous rocks suggests that prevailing pressure conditions were dominantly hydrostatic and parallel isopleths in the Komati Formation indicate an increasing grade of metamorphism with increasing depth. Cloete deduced that the regional burial metamorphism was superimposed on earlier sea floor metasomatism and he noted the documentation of similar burial-type static metamorphism in the Canadian and Australian Archaean.

However, between the extensive burial or static metamorphism that accompanied the filling of the Barberton basin and the peripheral thermal metamorphism that resulted from the intrusion of the younger granitoids, the Onverwacht was invaded by early TTG granitoids and, some 200 Ma afterwards, the entire Barberton basin was compressed and inverted during folding and thrusting accompanying Ramsay's (1963) first deformation episode (F_1), which was a fabric-free folding event. The widespread development of slaty cleavage and schistosity on a regional scale accompanied the second episode of deformation (F_2), according to Ramsay, and evidence for the imposition of this cleavage on the earlier fabric-free folds can be recognized in the transected fold that is the Eureka syncline. Elsewhere, the cleavage and schistosity is more or less parallel to the northeasterly-trending regional fold axes. The episode of regional dynamothermal metamorphism must have accompanied and succeeded the emplacement of the early tectonic Kaap valley pluton at 3227 Ma. This diapiric-like intrusive (Anhaeusser, 1975) indented the deforming greenstone belt and deflected the Eureka syncline so that the superimposition of transecting cleavage can be identified readily in the Sheba-Fairview area (Ramsay, 1963). Regional cleavage and schistosity appears to have been developed in the rocks of the greenstone belt prior to the invasion of the younger 3105 Ma granitoids and the accompanying thermal metamorphism.

Subsequent dynamic metamorphism, which accompanied the development of the major strike faults (Ramsay, 1963), produced localized cataclastic fabrics associated with shear zones accommodating thrusting, oblique-slip and strike-slip movements. Either during or following the reactivation of these faults, they were silicified and carbonated and can be recognized as linear zones of metasomatized host rocks, which sometimes provide favourable sites for Au mineralization. The Pioneer fault is a good example of one such mineralized, carbonated and silicified shear zone (Wuth, 1980), which traverses greenstones of the Onverwacht Group that had already experienced sea floor, burial, regional dynamothermal and extensive peripheral thermal metamorphism prior to the late-tectonic Au mineralization. Initial age measurements indicate that the deposits formed in suitable structural settings of low mean stress at about 3100 Ma (de Ronde *et al.*, 1991) and therefore just after the intrusion of the younger potash-rich granitoids.

In summary, the complex metamorphic history of the Barberton greenstone belt includes the recognition of the effects of sub-sea floor metasomatism (M_1), followed by extensive burial or static metamorphism (M_2), which was succeeded by peripheral metamorphism accompanying the emplacement of the early 3445 Ma TTG plutons (M_3) and the much younger Kaap valley TTG pluton at about 3227 Ma (M_4). Around 3220-3215 Ma, a cleavage and schistosity-forming episode of regional dynamothermal metamorphism (M_5) affected the entire greenstone belt during the deformational event that resulted in the formation of the NE-trending, regional folds. The regional dynamothermal metamorphism was succeeded by extensive, superimposed, peripheral, thermal metamorphism (M_6) that accompanied the intrusion of the younger potash-rich granitoids at about 3105 Ma. Finally, localized, dynamic metamorphism (M_7) resulted in the development of a number of thrusts and wrench faults, which exhibit a complicated history of localized metasomatism, reactivation and mineralization. Commonly, the regional aspects of this polymetamorphism have been recorded in a general manner and few facets have been researched in detail.

ECONOMIC MINERALIZATION

Introduction

As has been recorded above, the principal units of the Barberton Supergroup comprise of the Onverwacht Group, mainly an assortment of ultramafic and mafic submarine volcanics, including a number of sill-like layered ultramafic complexes, which is overlain by the Fig Tree Group of turbiditic greywacke sandstones and associated mudstones and banded ferruginous shales. These groups are overlain by continentally-derived shallow-water arenites of the Moodies Group.

The metallogeny of the Barberton greenstone belt has been the subject of several publications (Anhaeusser, 1976a, b, c; Anhaeusser and Viljoen, 1986) and this summary can highlight only some of the more recent developments in metallogenic research and understanding of the mineralization of the Barberton greenstone belt. Asbestos, barytes, cinnabar, Au, hematite, magnesite, stibnite, talc and verdite have all been won from orebodies in the greenstone belt at one time or another during the last hundred years (Hall, 1918; Groeneveld, 1975; Anhaeusser, 1986a, b, c). Currently, Au, asbestos, magnesite and talc are being recovered from operating mines in the area. In addition, there has been considerable prospecting for Cu, Ni and Zn without economic success. The Kuroko-like deposit of Ag-Pb-Zn-barytes in metamorphosed felsic volcanics of the Bien

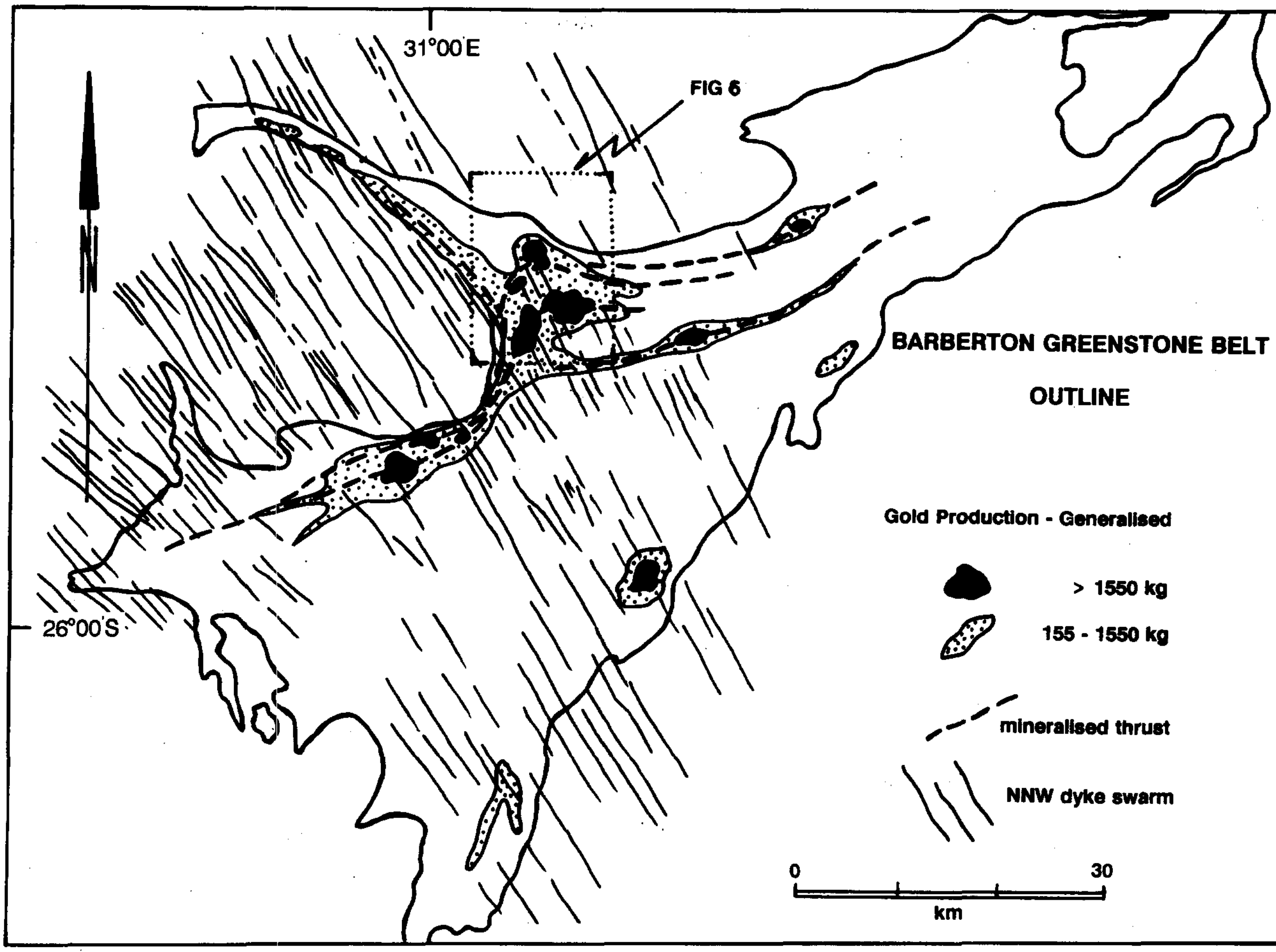

Figure 5. Distribution of Au production from the Barberton greenstone belt in relation to prominent NE-trending mineralized shears/thrusts (modified after Anhaeusser, 1969, 1976a, b, c,), and the NW-striking dyke swarm from Sheet 2530 Barberton, 1:250 000.

Venue Formation, Fig Tree Group, has been extensively explored on the surface and underground within the last decade and is of particular metallogenic interest, although uneconomic at present (Murphy, 1990). The small pod of Ni-spinel in tectonized ultramafite at Bon Accord provides a metallogenic poser, having been interpreted, on geochemical grounds, as either a palaeometeorite or of deep mantle origin (de Waal, 1978, 1986; Tredoux *et al.*, 1989).

Gold

Gold mineralization occurs associated with late-tectonic shears and fractures within the rocks of the Onverwacht, Fig Tree and Moodies Groups (Anhaeusser, 1986a, b, c; Wagener and Wiegand, 1986; Voges, 1986; Wiggett *et al.*, 1986; Wagener, 1986; de Ronde *et al.*, 1992). Host rocks, which have been variably silicified, sericitized, carbonatized and sulphidized along the fractures (Schouwstra and de Villiers, 1988; Tomkinson and Lombard, 1990) vary from greenstones to greywackes, shales, banded ferruginous shales and quartzites or sandstones. The Barberton Au ores are either free milling, moderately refractory or highly refractory depending on the extent to which the precious metal is occluded within the associated sulphides, commonly pyrite and arsenopyrite. Supergene enrichment of Au appears to have been an important factor in the payability of some of the more disseminated deposits, such as those which were worked in the oxidized portion of mineralized banded ferruginous shales.

Figure 5, modified after Anhaeusser (1969), depicts the regional concentration of Au production from the Barberton greenstone belt, including those centres of mineralization from which more than 1.5 tonnes of Au have been mined since the pioneer discoveries in 1884. Also shown on the figure are some of the major regional strike faults or thrusts, with which Au mineralization is spatially associated, as well as the NNW-trending dyke swarm that may serve as a late tectonic marker of the principal stress direction during the main episode of deformation.

The greater part of the Au that has been won from the Barberton greenstone belt was mined from three separate complexes of epigenetic mesothermal ore shoots - the Sheba-Fairview, New Consort, and Agnes-Princeton Au mining centres. The lodes comprise mineralized quartz veins or impregnated and replaced wallrock in systems of both concordant and discordant shears and fractures within the Onverwacht, Fig Tree and Moodies host rocks. All three complexes sustain producing Au mines which have been in operation for over a century. Geochronological work so far indicates that Au mineralization in the Fairview mine occurred between 3126 and 3084 Ma (de Ronde *et al.*, 1991).

More than 120 tonnes of Au have been recovered from the Sheba-Fairview complex, 60 tonnes from the New Consort complex and 30 tonnes from the Agnes-Princeton complex out of a total of some 250 tonnes won from the entire goldfield. For the rest of the workings in the greenstone belt, 14 mines have yielded more than 1 tonne of Au from each working and another 15 mines have produced over 500 kg of Au each. More than 50 kg of Au have been recovered from each of a further 40 smallworkings. The principal focus of the Au mineralization in the greenstone belt is spatially associated with the arc of the refolded Eureka syncline. More than 40 original surface reef workings comprise the whole Sheba-Fairview complex, which occurs along the inner arc of the regional structure. In general terms, the New Consort Au mine complex is developed within the ambit of the outer arc of the refolded Eureka syncline.

The overall geometric relationship between the Fairview and Sheba reefs in the Sheba-Fairview complex is becoming clearer in the lower levels of the Sheba mine where the mineralization is dominated by two converging oreshoots: the MRC or Main Reef complex and the ZK or Zwartkoppie shoot. The MRC is the down-dip extension of the Fairview mine to the west, whereas the ZK shoot has been mined down along a southwesterly plunge from the original lower Sheba surface workings to the south of Golden Quarry. Current underground development has exposed both oreshoots on 35 level, some 1100 m below surface (see Fig. 6).

Nowadays, approximately 2 tonnes of Au are produced each year from the oreshoots in the Sheba workings at a recovery grade of about 16 g t^{-1}. The MRC comprises three sets of shear zones hosted within the greywackes and shales of the Fig Tree Group and economic mineralization haloes up to 5 m wide are a feature of the main shears. In the case of the ZK shoot, ore occurs in greenschist-grade reactive host rocks of the Zwartkoppie bar adjacent to the southern limb of the Zwartkoppie anticline. The bar is developed along the contact between the Onverwacht and Fig Tree Groups and Au concentrations occur where the chert horizon is intersected by mineralized fractures (Robertson, 1989; Robertson *et al.*, 1993).

The same geological contact in the New Consort mine, between the Onverwacht and Fig Tree Groups, is a cherty horizon termed the Consort bar which separates the amphibolite-grade ultramafic and mafic metavolcanics from the metasediments. Two gently converging payshoots in the mine, the PC (Prince Consort) shoot and the Seven Shaft shoot, are associated with this cherty contact, strike northwest and plunge steeply towards the southeast (Fig. 6). Some 800 kg of Au are produced annually from the New Consort oreshoots at a recovery grade of about 13 g t^{-1} (Voges, 1986; Tomkinson, 1991; Tomkinson and Lombard, 1990).

Hall (1918, p. 236) emphasised the strike-parallel linear belts of economic importance in the distribution of Au mineralization in the Barberton mountain land. Anhaeusser (1965, 1969, 1976 *et seq.*) later drew attention to the regional association between Au deposits and wrench faulting in the Barberton greenstone belt. Gold mineralization is commonly concentrated in secondary shear zones and fractures near regional NE-trending reverse, oblique or strike slip faults, depending on the degree of thrusting and transpression. Faults such as the Lily, Scotsman, Pioneer, Moodies, Sheba, Barbrook and Saddleback all have some adjacent Au deposits of varying size at irregular intervals along the strike (Gribnitz, 1964). In terms of the principal centres of Au concentration, the reefs constituting the Sheba-Fairview complex occur within the general proximity of the Sheba fault, but, according to Tomkinson and Lombard (1990), at least some of the New Consort Au mineralization is associated with NW-trending shear zones. From these facts it may be deduced that controls of hydrothermal mineralization or metallotects in the Sheba-Fairview and New Consort case include:

i) a NE-trending thrust or wrench fault;

ii) NW-trending shear zones and tension joints;

iii) regional arcuation of the Eureka syncline;

iv) local shears and fractures that provided a focussed 'plumbing' system; and

v) favourable host rocks, especially near the contact between the Onverwacht metavolcanics and the Fig Tree metasediments.

The northwesterly alignment corresponds to the maximum principal stress direction during the regional deformation which formed the major northeast-southwest trending folds in the greenstone belt and is a prominent joint direction throughout the greenstone belt. Some general aspects of these structural relationships are depicted in Fig. 6.

In terms of the apparent temperature and pressure of formation, major Au deposits in the Barberton greenstone belt are mesothermal in character. According to available geothermometry, the payshoots, in

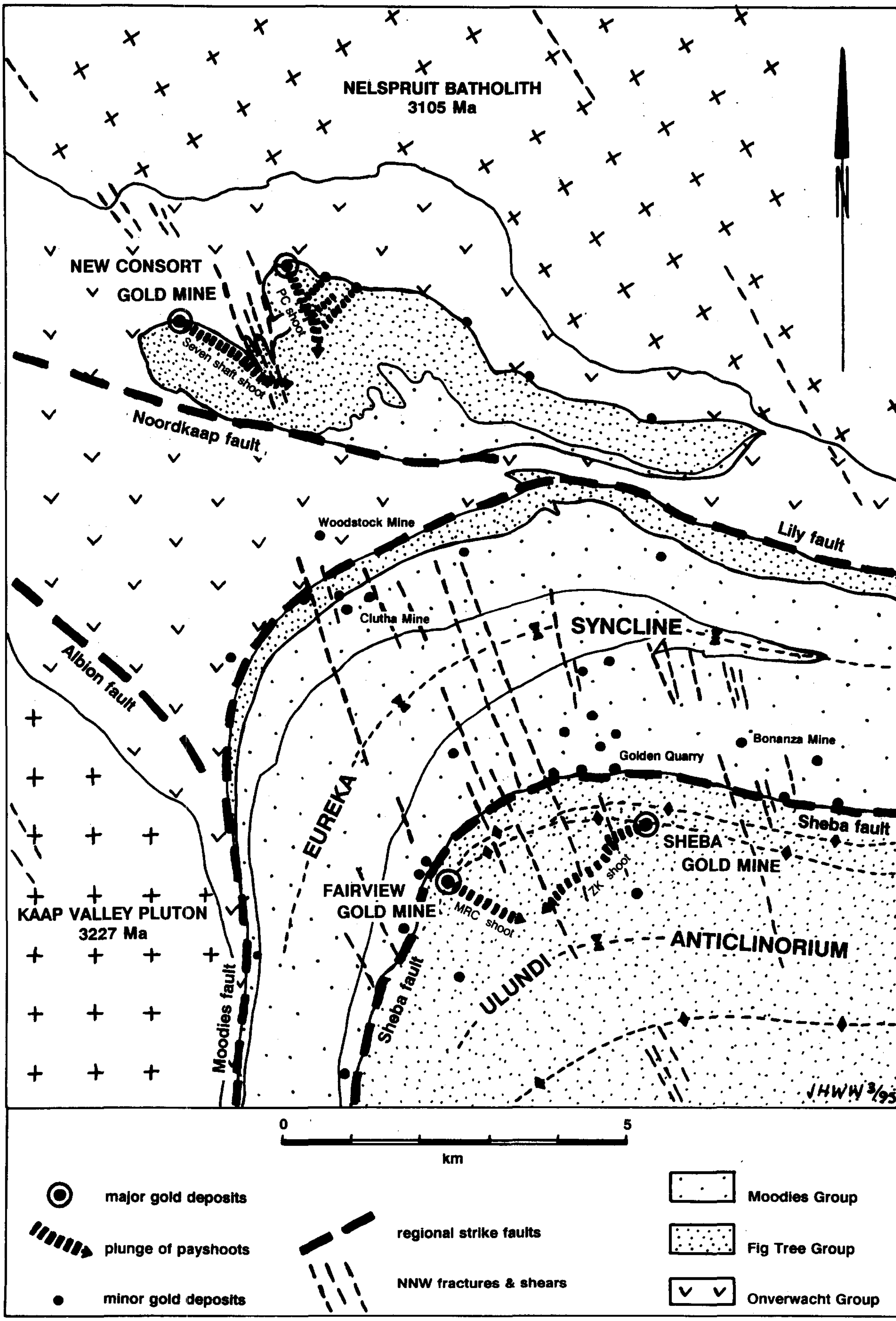

Figure 6. Detailed geology of the New Consort - Fairview - Sheba Au mining area showing the generalized metallotects noted in the text (compiled from Anhaeusser, 1976a, b, c; Gribnitz, 1964; Robertson, 1989; Tomkinson, 1991; Voges, 1986; Wiggett *et al.*, 1986).

which Au is commonly associated with pyrite and arsenopyrite, formed above 500°C in the case of the New Consort mine (Tomkinson and Lombard, 1990) and in the range 300-425°C at 4.2 kbar in the case of the Sheba mine (Schouwstra and de Villiers, 1988). Eastwards from the Sheba-Fairview area, the Au mineralization associated with the regional faults diminishes in tenor from the 17 g t^{-1} Sheba ore reserve to the 6.8 g t^{-1} *in situ* grade of the Barbrook deposit and less beyond. Still further east, and on the extension of the Scotsman fault, occurs the Amo Sb deposit comprising several small bodies grading 5% Sb (Viljoen *et al.*, 1986). Southeast of Amo, underlying the farms Kaalrug and Rusoord, are to be found minor cinnabar occurrences in proximity to the conjoined Saddleback-Inyoka-Kamhlabane fault (Groeneveld, 1975). This gradation in mineralization indicates that there was a variation in the conditions of hydrothermalism channeled along these regional conduits, which is expressed as ores that range from mesothermal Au-sulphide concentrations to epithermal cinnabar deposits. Whether this was a lateral or vertical hydrothermal gradient is uncertain. Some Au lodes underlying the Swaziland section of the greenstone belt contain associated Sb and Hg mineralization, which has been interpreted to represent the distillation of volatile elements along thrusts affecting the low temperature margin of a geothermal field (Barton, 1982).

A study of fluid inclusions within samples of quartz-carbonate vein material collected in the Abbotts, Pioneer, Bellevue and Three Sisters mines established homogenization temperatures of 290 to 310°C as representative for the main stage fluid and it has been suggested that the "phase separation of an H_2O-CO_2-NaCl parental fluid was the principal factor in producing the Barberton gold deposits." (de Ronde *et al.*, 1992).

The determination of light stable isotopes for Au-associated fuchsite- and sericite-rich alteration zones from the same four deposits plus another three, including the Fairview mine, enabled de Ronde *et al.* (1992) to conclude that the source of the mineralizing fluids was likely to have been external to the greenstone belt. The composition of the fluids and the light stable isotope data was found to be remarkably similar to that determined for Archaean Au-quartz veins from Canada and western Australia.

Chrysotile asbestos

Economic deposits of chrysotile asbestos, magnesite and talc in the Barberton greenstone belt are to be found entirely within the host rocks of the Onverwacht Group. The chrysotile and magnesite ores are confined to the sill-like layered ultramafic complexes, whereas deposits of talc are more widespread among the ultramafic greenstones of the Onverwacht Group. Anhaeusser (1976, 1986) has described in detail the metallogeny of chrysotile mineralization in southern Africa, including the Barberton greenstone belt.

In the past, important chrysotile asbestos mining operations have been established in the Kaapsehoop, Mundt's Concession, Koedoe, Stolzburg, Msauli and Havelock layered ultramafic complexes (Groeneveld, 1975; Anhaeusser, 1986a, b, c). Currently there are only two operating chrysotile mines in the area, one at Msauli and the other at Bulembu (Havelock). More than 1.5 million tonnes of fibre have been produced from each of these two mines, which together have accounted for over 85% of the total chrysotile fibre production from the Barberton asbestos field. The orebodies occur within deformed serpentinites along a regional tectonic break known as the Maanhaar fault, which underlies ground adjacent to the Swaziland border in the southern part of the greenstone belt (Fig. 3). It has been suggested that the Msauli-Havelock serpentinite bodies once formed part of a conformable sill (Viljoen and Viljoen, 1969a, b, c, d, e, f), or a differentiated ultramafic layered intrusive (Büttner and Saager, 1982; Rodel, 1993), or that they represent allochthonous serpentinite bodies which were emplaced along a regional thrust (Barton, 1982). In all cases, the regional Maanhaar-Kamhlabane fault zone is a major tectonic break similar to the Inyoka-Saddleback break.

In general, chrysotile orebodies in the layered ultramafic complexes are associated with faulting, fracturing and folding in the host rock and are best developed in altered dunite and peridotite horizons locally known as 'ore zone' serpentinite. There has been an obvious combination of host rock and structural control in the development of economic concentrations of fibre in this asbestos field (Anhaeusser, 1986a, b, c; Barton, 1986; Voigt *et al.*, 1986). Late-stage dilatant stress in the ore zone serpentinite created vein fractures in which cross-fibre seams and stockworks of chrysotile formed from hydrothermal serpentinous solutions by a process of transport, diffusion and crystallization (Laurent, 1975; Moody, 1976).

Magnesite

Magnesite has been recovered profitably from a number of layered ultramafic complexes in the Barberton greenstone belt, especially in the Kaapmuiden area where the Budd, Ship Hill and Magnesite bodies have supported economic operations (Hall, 1918; van Zyl *et al.*, 1942; Viljoen and Viljoen, 1969a, b, c, d, e, f).

Well over 500,000 tonnes of magnesite have been produced from quarries in the Kaapmuiden serpentinite belt, of which the bulk has come from the Strathmore mine. In this operation, low-grade stockworks of magnesite veinlets, known as 'zebra ore', are quarried and up-graded by beneficiation to remove serpentinite and silica impurities. The origin of vein magnesite deposits in ultramafic host rocks is problematical and models range from supergene to hypogene processes and from weathering to epithermal intrusion (Pohl, 1990; Abu-Jaber and Kimberley, 1992). Researchers generally agree that the host rocks were the source of the Mg, but the carbonate source is less obvious, although extensive carbonation is a feature of hydrothermal metasomatism in the Barberton greenstone belt. Preliminary Pb-Pb isotope age dating of magnesite from the Budd orebody indicates that the mineralization occurred at about 2990 Ma (Toulkeridis *et al.*, 1993).

Talc

A number of small, lensoid bodies of talc have been exploited along strike-parallel shear zones within the serpentinites and talc carbonate rocks of the Onverwacht Group underlying the Jamestown and Kaapmuiden areas of the greenstone belt. Total production has been in excess of 150,000 tonnes of industrial-grade talc, of which more than 70% has been won from the Scotia Talc mine underlying the Bon Accord stock farm (Groeneveld, 1975; Keenan, 1986).

The Scotia talc mine occurs within the aureole of the Nelspruit batholith on the northern flank of the greenstone belt. The underground workings expose serpentinite and talc carbonate (magnesite) schist as well as lenses of apple-green and grey talc, which are associated with secondary shear zones. The formation of the orebodies is ascribed to a combination of dynamic and hydrothermal alteration or steatization of serpentinites derived from ultramafic lavas or sills (Anhaeusser, 1972; Groeneveld, 1975). By deduction, it can be inferred that the talc-forming processes involved host rock, structural and hydrothermal controls of mineralization not so very different to those associated with the formation of magnesite or chrysotile asbestos deposits in the Barberton greenstone belt. The mineralogy of the host rock, the nature and intensity of deformation and the composition and temperature of the active hydrothermal solutions are key variables in the genesis of these ores.

General

In addition to the above resources which are currently being exploited, the Barberton greenstone belt has yielded lesser amounts of stibnite, barytes, cinnabar, Ni and verdite, as well as a substantial tonnage of Fe ore. Of these, the stibnite and cinnabar ores were very small epithermal deposits associated with regional faults that constituted hydrothermal conduits (Groeneveld, 1975; Viljoen *et al.*, 1986). Strati-form barytes deposits in the Lower Fig Tree Group siliciclastics have been interpreted as being exhalative-sedimentary in origin and are worked-out in present economic terms (Heinrichs and Reimer, 1977; de Wit *et al.*, 1987a, b). Small bodies of verdite have been exploited in the ultramafic rocks of the Onverwacht Group where they occur as an ornamental variety of serpentinite (Groeneveld, 1975). The Fe deposit at Ngwenya mine in southern Swaziland yielded 28 million tonnes of high-grade hematite ore from the Lower Fig Tree Group banded iron formation, which had been upgraded by supergene enrichment associated with zones of faulting and the leaching of silica from the protore (Barton, 1982).

CONCLUSION

The depositional history of the rocks of the Early Archaean Barberton greenstone belt spans about 300 Ma, from the submarine effusion of the lowermost Onverwacht eruptives at about 3547 Ma to the sedimentation of the continentally-derived Moodies arenites. According to present information, regional compression led to basin inversion and the end of the depositional cycle by about 3220 Ma when major NE-trending regional folds were developed during the main phase of deformation that was completed by 3216 Ma, according to the evidence from the discordant Dalmein pluton.

Episodes of sialic plutonism underplated the greenstone belt and extended from Upper Onverwacht times, at about 3445 Ma, to the extensive peripheral invasion of late-tectonic granitoids at about 3105 Ma. The nature of the basement to the Onverwacht Group has been obscured by the invasive granitoids which encompass the greenstone belt and hypotheses range from support for an episialic or ensialic geotectonic setting to argument in favour of an ensimatic environment. There is no evidence for remnant oceanic crust which is older than the Onverwacht volcanics, but there are increasing indications in the vicinity of the greenstone belt of sialic crust which was in existence prior to the development of the Barberton basin. At present, and despite the currently-fashionable support of an oceanic and accretionary model for the geotectonic development of greenstone belts, there is considerable uncertainty as to whether the floor to the rocks of the Onverwacht Group was sialic or simatic (continental or oceanic) crust.

Thus, there can be no certainty regarding the primary metallogenic setting of the greenstone belt environs during Onverwacht times. Geochemical indications that the Lower Fig Tree Group greywackes and shales were derived from ultramafic source rocks implies that this sedimentation accompanied the unroofing and erosion of the Onverwacht-type volcanics. The Kuroko-like deposit of Ag-Pb-Zn-barytes in deformed and metamorphosed calc-alkaline felsic volcanics of the Fig Tree Group suggests a tenuous uniformitarian analogy to a supra-sialic Japanese island arc-type of geotectonic setting during Fig Tree times. This was followed by syntectonic sedimentation from a northern continental source as indicated by the deposition of the Moodies Group arenites.

The development of the Barberton greenstone belt could have been initiated in response to an incipient or primitive Wilson Cycle involving the extension, ductile thinning and perhaps hot-rifting of sialic crust. Isotopic work on the surrounding granitoids may indicate to what extent the intrusives represent reconstituted and remobilized sial (Hoffman, 1990). This evidence should improve the accuracy of the estimation as to the probable degree of extension from a rudimentary ensialic rift to a primitive back-arc setting or an ancient oceanic spreading centre, which accompanied Onverwacht-style magmatism and the possibility of later stratigraphic thickening due to tectonic stacking. The magmatism was succeeded, after a considerable interval of time, by early flysch-like and later coarse-clastic sedimentation, followed by regional compression of collisional-intensity, which was buffered by the contractive deformation of the greenstone belt.

During and after the widespread response to regional compression in the form of deformation and metamorphism, substantial economic deposits of chrysotile asbestos, magnesite and talc were developed in the ultramafic host rocks of the Onverwacht Group in stress-controlled reactions to the localization and intensity of deformation and facilitated by the activity and composition of percolating hydrothermal solutions that promoted metasomatic alteration. The important mesothermal lode Au deposits in the greenstone belt were formed along suitable late-tectonic fractures within the Onverwacht, Fig Tree and Moodies Group host rocks in association with regionally-developed shear zones and in response to appropriately-focused and mineralized hydrothermal solutions. Supergene enrichment of banded iron formation protore gave rise to the development of one significant high-grade hematite Fe orebody and several lesser occurrences. Exhalative-sedimentary stratiform barytes ores have been developed in the Onverwacht as well as the Lower Fig Tree Group, where several deposits have been exploited. Small epithermal orebodies of stibnite and cinnabar have been worked in the past and there are some minor showings and prospects for Cu and Ni in the Onverwacht rocks.

The Council for Geoscience plans to publish the 1:100 000 metallogenic map of the Barberton greenstone belt on one sheet, following compilation on eleven 1:50 000 sheets. The metallogenic map will depict all the known mineral occurrences and deposits set against the geological and structural background of the greenstone belt. Appropriate metallogenic symbols will indicate the nature and extent of the mineralization in each case. The map, and accompanying explanation, together with data sheets, will provide a useful basic planning tool for environmental management, land use control, mineral resource evaluation, commodity studies and exploration analysis, as well as for tourism, recreation and education.

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Geological Map

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