19
Basin Analysis in Regionally Metamorphosed and Deformed Early Archean Terrains: Examples from Southern Africa and Western Australia

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Abstract

Early Archean (>3.0 Ga) sedimentary rocks are present in high-grade metamorphic terranes and greenstone belts in southern Africa and Western Australia. The viability and quality of basin analysis in these terranes improves with decreasing metamorphic grade and degree of structural complexity. Depending on the state of preservation of the rocks, it is possible to reconstruct some or all of: stratigraphy, source-area composition and age, sediment dispersal patterns and depositional environments, and basin configuration and tectonic setting.

Pre-greenstone (>3.6 Ga) gneisses in the high-grade Limpopo Province and Western Gneiss Terrain are envisaged as remnants of cratonic nucleii. These gneisses represent basement to cover rocks that are exclusively of sedimentary origin and accumulated between ca. 3.6 and 3.2 Ga. Limpopo cover rocks are devoid of primary sedimentary structures and consist of quartzite with detrital zircons (original quartz arenite), marble (limestone), metapelite (mudstone) and aluminous gneiss (possible wacke). This quartz arenite-carbonate association implies a stable tectonic setting and the best analogs may be younger cratonic-shelf deposits. Metapelites in the Limpopo Province have a complex rare earth element (REE) geochemistry indicating a mixed provenance that included differentiated continental crust. A thick sequence (ca. 2.5 km) of conglomerate and crossbedded aluminous gneiss (wacke) in the Western Gneiss Terrain is interpreted as an alluvial deposit and possibly accumulated in a rift setting. Rare earth element geochemistry of metapelites indicates that differentiated continental crust consisting of K-granites was the dominant component of the source terrain.

Predominantly mafic and ultramafic volcanism in the Barberton and Pilbara greenstone belts took place between 3.5 and 3.3 Ga in an oceanic environment distant from any continental influence. Intercalated sedimentary deposits indicate that volcanism took place at relatively shallow-water depths. The volcanic sequences are overlain by predominantly siliciclastic sedimentary intervals. In the Barberton greenstone belt, the distribution of facies, in conjunction with paleocurrent, petrographic, geochemical, and geochronological data indicate that the Fig Tree and overlying Moodies Groups (ca. 5 km thick) were derived by progressive unroofing of a southerly source terrain consisting of the older volcanic rocks with intercalated sedimentary deposits, and a 3.5 to 3.3 Ga gneiss complex. Sedimentation took place initially in a submarine-fan setting. Basin shoaling is indicated by the upward transition into braided-alluvial and shallow-marine sediments. The stratigraphic evolution of the Fig Tree and Moodies Groups is similar to that of Phanerozoic foreland or foredeep basins.

In the Pilbara Block the lower Gorge Creek Group is compositionally similar to the Moodies Group and was derived from a comparable provenance; the tectonic setting of the basin is unclear. The upper Gorge Creek Group unconformably overlies the lower subdivision and contains abundant quartzite clasts recycled from it. This ca. 3 km-thick stratigraphic sequence is exclusively of continental origin; depositional environments include alluvial fans, braided rivers, floodplains, and lakes. Stratigraphic and sedimentological evidence indicates that basin development was controlled by marginal strike-slip faulting.

Available age constraints indicate that sedimentary rocks in the high-grade terranes and greenstone belts developed contemporaneously. Cover rocks in the high-grade terranes accumulated on thick continental crust and reflect a stable cratonic setting. In contrast, volcanic intervals in the greenstone belts are considered to have developed in an oceanic setting, whereas the overlying siliciclastic intervals reflect active tectonic settings associated with crustal shortening.

Introduction

Sedimentary rocks older than 3 billion years are present in southern Africa and Western Australia in the high-grade Limpopo Province and Western Gneiss Terrain and in the greenstone belts of the

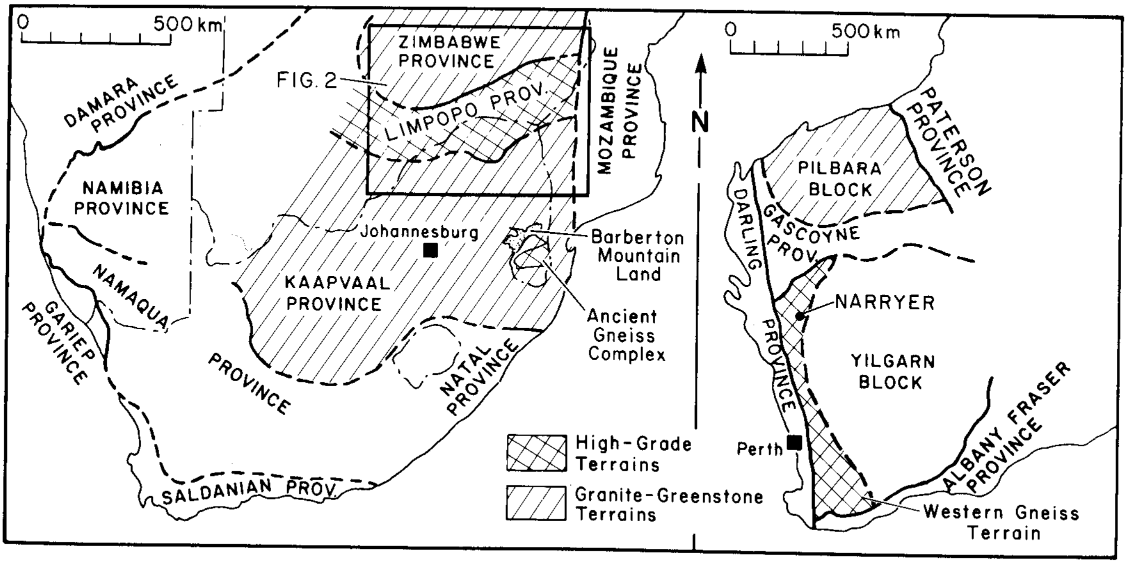


Fig. 19.1. Locality map of the high-grade Limpopo Province and Western Gneiss Terrain and the granite-greenstone Kaapvaal Province and Pilbara Block [adapted from Button (1976), Gee (1979), and Tankard *et al.* (1982)].

Kaapvaal Province and Pilbara Block, respectively (Fig. 19.1). This paper focuses on the predominantly siliciclastic sedimentary sequences in these two different types of crustal terranes. The sediments vary considerably in state of preservation of primary depositional features because of differences in degree of metamorphism and intensity of deformation. Despite the great antiquity of the sediments they are amenable to basin analysis employing many conventional sedimentological techniques (cf. Potter and Pettijohn 1977), as well as some less widely used techniques. The viability and quality of basin analysis improves with decreasing metamorphic grade and structural complexity; depending on the state of preservation of the rocks it is possible to reconstruct some or all of: stratigraphy, source-area composition and age, sediment dispersal patterns and depositional environments, and basin configuration and tectonic setting.

Analysis of these sedimentary rocks also allows a number of fundamental questions to be addressed that are relevant to the early evolution of the Earth's crust. These questions include land-sea proportions, crustal compositions, uniformitarian or nonuniformitarian nature of surface processes, and evidence for exposed land masses prior to 3 billion years ago. Of greater significance are the constraints that basin analysis can place on the temporal and spatial relationship between high-grade and greenstone terranes. High-grade terranes have variably been considered as the metamorphosed roots of greenstone belts, as basement to greenstone belts, or as having developed contemporaneously with greenstone belts but in different tectonic settings (Percival and Card 1983; Windley 1984).

In this chapter we evaluate the application of conventional basin analysis techniques to these metamorphosed sediments before presenting a critical assessment of the problems of basin analysis in metamorphosed and deformed Archean terranes. In the final section of this paper we use the results of the basin analyses to address the broader questions outlined above.

Limpopo Province: Southern Africa

Geologic Setting

The high-grade Limpopo Province in southern Africa is located between the Zimbabwe and Kaapvaal Provinces that consist of arcuate greenstone belts surrounded by granites and gneisses (Fig. 19.2). Within the Limpopo Province a three-fold zonation is recognized; northern and southern marginal zones flank the central zone (Fig. 19.2). The marginal zones were derived by metamorphism and deformation of the adjoining granite-greenstone provinces, whereas the central zone displays a different history.

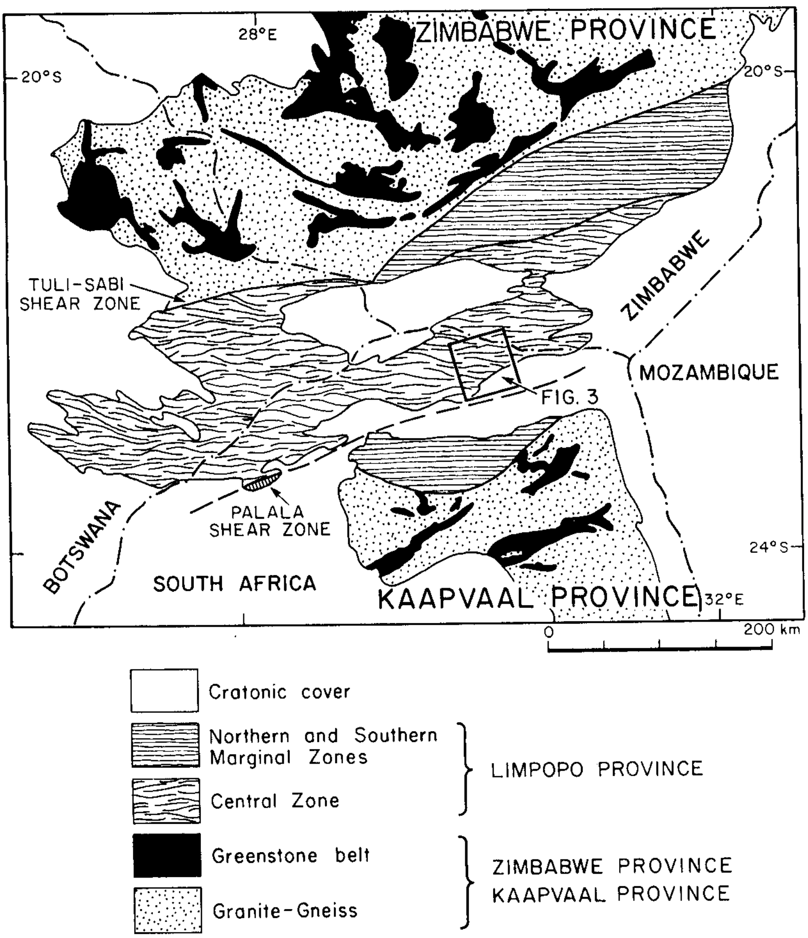


Fig. 19.2. Tectonic zones of the Limpopo Province and their spatial relation to the granite-greenstone terrane of the adjacent Zimbabwe and Kaapvaal Provinces [from Tankard *et al.* (1982)]. Location shown in Figure 19.1.

Within the central zone of the Limpopo Province a complex geological history is recognized dating from 3.8 to 2.4 Ga. The Sand River Gneisses with an age of 3,786 ± 61 Ma (Rb-Sr; Barton *et al.* 1983a) are intruded by 3,560 ± 100 Ma mafic dikes (Rb-Sr; Barton *et al.* 1977) and are recognized as basement to younger cover rocks represented by the Beitbridge Complex (Fig. 19.3; Fripp 1983). While the isotopic ages can be questioned, the basement-cover relationship is based on convincing field evidence that indicates that the oldest recognized structural fabrics are restricted to the Sand River Gneisses (Fripp 1983). In the Beitbridge-Messina-Tshipise region (Fig. 19.3), the Beitbridge Complex consists of four lithological associations, namely, quartzite-pyroxenitic amphibolite (Fig. 19.4); biotite-garnet-cordierite-sillimanite gneiss; magnetite quartzite, calc-silicate gneiss, and marble (Fig. 19.5); and gray gneiss. Intrusive into the cover rocks are the Messina Layered Intrusion (calcic anorthosite suite) and the Singelele Granitoid (Fig. 19.3; Barton 1983a). The Messina Layered Intrusion is dated at 3,153 ± 47 Ma (Rb-Sr; Barton *et al.* 1979a) and $3{,}270^{+105}_{-112}$ Ma (Pb-Pb; Barton 1983b). The latter age is considered to represent the time of emplacement and the lower value considered to record the earliest deformation and metamorphism of the layered intrusion along with the Beitbridge Complex and Singelele Granitoid (Barton 1983a; Watkeys *et al.* 1983) at pressures up to 10 kb and temperatures in excess of 800°C (Horrocks 1983). The ca. 3,150 Ma age of metamorphism is substantiated by mafic dikes of 3,000 to 3,100 Ma age (Rb-Sr; Barton *et al.* 1977, 1983b) that transect a major structural/metamorphic fabric in the basement, Messina Layered Intrusion, and Singelele Granitoid (Barton *et al.* 1983a; Watkeys *et al.* 1983). Subsequent deformation and a second high-grade metamorphism took place at 2.7 Ga (Van Reenen *et al.* 1987), with metamorphism continuing under conditions of

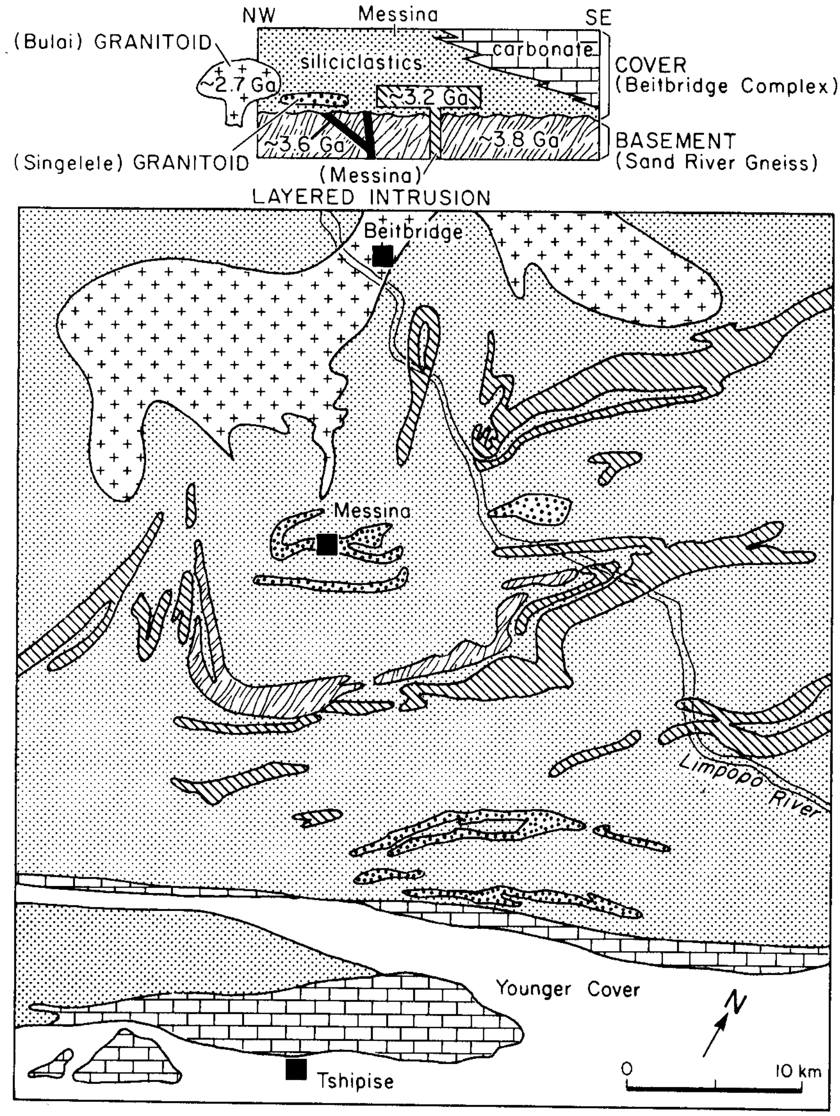


Fig. 19.3. Geological map of the central zone of the Limpopo Province in the Beitbridge-Messina-Tshipise area [modified after Tankard *et al.* (1982)]. Location shown in Figure 19.2.

progressive decompression until 2.2 Ga (Van Reenen 1986). The synkinematic Bulai Granitoid was emplaced early in this structural history (Watkeys *et al.* 1983) and is well dated at ca. 2,650 Ma (Fig. 19.3; Rb-Sr; Van Breemen and Dodson 1972; Rb-Sr and Pb-Pb; Barton *et al.* 1979b).

Stratigraphic Framework of the Beitbridge Complex

In the absence of primary structures such as cross-stratification, and because of great structural complexity, it is not possible to establish a stratigraphic sequence in the Beitbridge Complex. Ductile strain is ubiquitously shown in the central Limpopo Province by a layer-parallel foliation with strongly attenuated limbs of early isoclinal folds and wide-separation boudinage of some less ductile layers (Fig. 19.5). The effect of this strain has been to thin severely the original individual layers, now seen in outcrop at the 1 cm to 100 m scale. On a larger scale (km) it is, of course, likely that large-scale folding and thrust duplication have repeated and therefore thickened the overall original stratigraphic sections.

In this section an attempt is made to interpret primary lithologies and lithologic proportions that can be used to constrain other components of the basin analysis. One of the major problems in the Limpopo Province, as in many high-grade terranes,

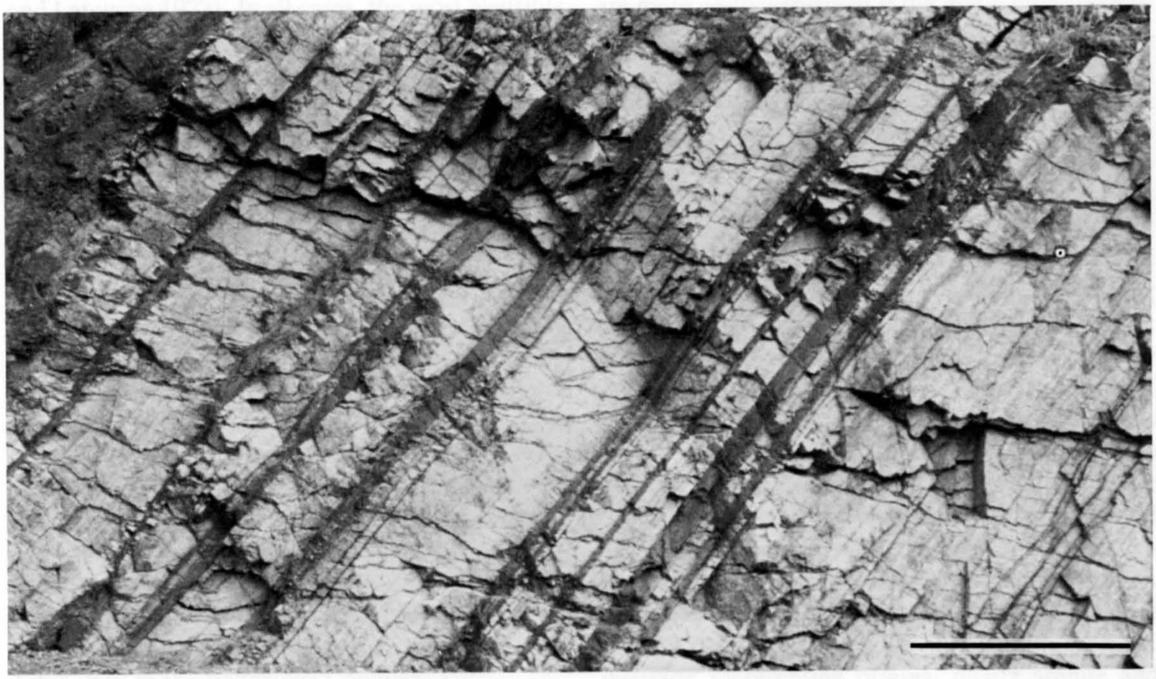


Fig. 19.4. Interlayered quartzites (light) and amphibolites (dark) of the Beitbridge Complex. Outcrop located 10 km southwest of Messina. Scale bar = 1 m.

Fig. 19.5. Quarry outcrop of marble in the Beitbridge Complex. Quarry face is ca. 20 m high. Lower 5 m of outcrop is pink, dolomitic marble; upper 15 m is gray, calcitic marble. Dark attenuated layer with prominent boudins is calc-silicate rock. Outcrop located 20 km northwest of Alldays.

Fig. 19.6. Photomicrograph showing zircons with rounded cores and euhedral metamorphic overgrowths from quartzites in the Beitbridge Complex. Scale bar = .2 mm.

concerns the origin of the quartzite-pyroxenitic amphibolite association in the Beitbridge Complex. This association has conventionally been considered to represent mineralogically mature quartzose sandstone and mafic-ultramafic volcanic rock (Windley 1984). An alternative explanation for this association was proposed by Fripp (1983) who considered the quartzite as original chert and the amphibolite as original volcanic rock. A third possibility is that the amphibolites represent deformed and metamorphosed sills and/or dikes intrusive into quartzose sandstone or chert.

Quartzites in the Beitbridge Complex are presently up to 100 m thick in single occurrences (following severe tectonic thinning) and consist of up to 99.5% quartz. Metapelite partings are rare and no primary sedimentary structures are discernible. Layering defined by magnetite and fuchsite is common in the quartzites, and heavy minerals occur in higher concentrations than in most quartzose sandstones (J. Barton, personal communication 1983). Rutile and zircon are the most common heavy minerals with tourmaline also present; zircons average 0.4 mm in length and typically contain rounded cores with euhedral overgrowths (Fig. 19.6).

A number of the characteristics listed above favor an original detrital sedimentary origin for the quartzites. The high concentration of heavy minerals is incompatible with an interpretation of these rocks as metacherts. Zircon, rutile, and tourmaline are the most stable of the heavy minerals (Hubert 1962); they are typically concentrated in quartz arenite and are preserved despite prolonged source area weathering and reworking within the depositional basin. Other criteria that suggest a detrital sedimentary origin are thickness of the quartzites and the infrequent occurrence of argillaceous partings; thick chert sequences typically contain numerous shale interbeds (Nesbit and Price 1974; Lowe 1976). The presence of fuchsite may be used to suggest an exhalative origin for quartzites (cf. Schreyer *et al.* 1981), but fuchsite is common, and ascribed to the breakdown of detrital chromite, in quartz arenites from the Witwatersrand Supergroup (Antrobus and Whiteside 1964; Pretorius 1964). Although it is probably true that fuchsite is relatively more abundant in Archean sedimentary rocks, examples of prominent fuchsite in quartzose detrital rocks are known through much of the geological record (e.g., Eskola 1933; Padget 1956; Clifford 1957; Fabries and Latouche 1973). In all cases, the most plausible derivation of fuchsite is from detrital chromite of original ultramafic provenance. If a detrital origin for the quartzites is accepted they represent the oldest siliciclastic sedimentary rocks in southern Africa (cf. Tankard *et al.* 1982) and thus would represent first-cycle quartz arenite deposits.

Pyroxenitic amphibolites/granulites (for brevity termed amphibolites below) in the Beitbridge Complex are geochemically indistinguishable from average basaltic tholeiites of intrusive or extrusive origin (Fripp 1983). The amphibolites occur in sharp contact and alternate with quartzite on a scale of tens of metres to centimetres (Fig. 19.4). The cm-scale alternation is frequently developed through stratal thicknesses of up to 10 m. Amphibolites are homogeneous in composition and texture and, from our observations, are most abundant in proximity to the Messina Layered Intrusion (Fig. 19.3). Although most commonly associated with quartzite in the Messina area, amphibolite also occurs with all other lithologies in the Beitbridge Complex.

Although clearly of tholeiitic composition and therefore unlikely to be of sedimentary derivation, the extrusive or intrusive origin of the amphibolite protolith is enigmatic. However, we consider that they are most likely of intrusive origin. The greater abundance of amphibolite that we observed adjacent to the Messina Layered Intrusion suggests a genetic relationship between the two, and the geochemistry of the amphibolites reported by Fripp (1983) is consistent with this suggestion. The amphibolites and metasediments show evidence of significant ductile strain. It is difficult to be confident about the original thickness and scale of interlayering of the amphibolite and quartzite layers because no bulk strain markers are reported, and we observed none. If the strain is high, as we suspect from the qualitative indicators of early folds and boudinage, then the absence of cross-cutting intrusive relationships is understandable, since originally oblique features will have been rotated into parallelism. We observed a few possibly cross-cutting relationships, but they were not sufficiently convincing or oblique to distinguish from original lens-shaped sedimentary or volcanic layer terminations, or from sill edge terminations for that matter, given the likely ductile strain in the rocks. However, the lack of lithologies showing compositions indicative of mixing of quartzite and mafic protoliths, and the sharp contacts everywhere between the two lithologies, are features that are more easily explained if the mafic protolith was largely or entirely intrusive. A tuffaceous protolith for the amphibolites is considered also unlikely. Ash beds do not display the degree of compositional homogeneity apparent in the amphibolite layers. In addition, ash-fall deposits are more commonly siliceous than mafic and typically are interbedded with pelagic sediments or mudstones rather than quartz arenites as in the Beitbridge Complex (cf. Fisher and Schmincke 1984; their Table 7-1). Also, reworking and mixing of the loose ash and arenite would be predicted; appropriate rocks are not seen. If we are incorrect, and the mafic layers are largely flows, then the assemblage of pure quartzites and mafics is one without a Phanerozoic analog. We see no reasons to assume, however, that extrusive mafic rocks were present in any significant quantity, and consider that the evidence at present, while less than completely satisfactory, favors an intrusive origin. We therefore prefer a sedimentary-intrusive origin for the quartzite-amphibolite association with the quartzite representing original quartz arenite and the amphibolite original sills and/or dikes.

The protoliths of two of the other lithological associations comprising the Beitbridge Complex are more easily defined. Marble in the calc-silicate gneiss-marble association attains thicknesses, for single outcrops, of at least 100 m (Fig. 19.5) and, in the Messina-Tshipise region, is developed in a belt parallel to and south of the quartzite occurrences (Fig. 19.3). Marbles display abundant compositional layering and are visualized as original sedimentary carbonate rocks. Calc-silicate gneiss is considered to have originally been calcareous pelite. Associated but subordinate magnetite quartzite may represent either metamorphosed iron-formation or iron-rich quartzite. Overall thicknesses of this association are uncertain. Most workers agree that biotite-garnet-cordierite-sillimanite gneisses represent metapelites and the major element geochemistry of those in the Beitbridge Complex closely resembles average "geosynclinal" shales (Brandl 1983; Fripp 1983). Taylor *et al.* (1986) argue that the mineralogy, and major, trace, and rare earth element (REE) geochemistry, and intimate association with recognizable quartz arenites and carbonates support a pelitic origin for these paragneisses.

The protolith of the fourth lithologic association, the gray gneisses, is highly problematic. On the basis of their geochemistry, Fripp (1983) considers that these gneisses may have been felsic to intermediate volcanic and/or terrigenous sedimentary rocks, whereas Brandl (1983) favors a sedimentary arkosic protolith.

Fripp (1983) estimated the proportions of various lithologies comprising the Beitbridge Complex in the Beitbridge-Messina-Tshipise region (Fig. 19.3). His estimates for rocks considered to be of original volcanic and sedimentary origin are: amphibolite (volcanic) – 30%; biotite-garnet-cordierite-sillimanite gneiss (metapelite) – 30%; quartzite (chert) – 10%; marble-calc-silicate gneiss (limestone-calcareous pelite) – 10%; gray gneiss (sediment or volcanic) – 10%; Singelele gneiss (felsic volcanic) – 10%. Based on these percentages Fripp (1983) considers the Beitbridge Complex to have contained a relatively high ratio of volcanic to sedimentary rocks. Barton (1983a) disputes this ratio arguing that the Singelele Gneiss was intrusive as evidenced by numerous xenoliths. He suggests a somewhat lower proportion of volcanic to nonvolcanic rocks. On the basis of the earlier discussion, it is possible that the Beitbridge Complex contained few, if any, volcanic rocks and our estimate of lithologic proportions of primary surficial deposits, assuming a sedimentary origin for the gray gneisses, is quartz arenite – 17%; limestone and calcareous mudstone – 17%; mudstone – 50%; arkose/feldspathic wacke – 16%.

Source-Area Composition

The Beitbridge Complex represents a mature assemblage of sedimentary rocks and their petrography thus provides little information on source-area composition. Heavy-mineral separates are similarly not informative as zircon, rutile, and rare tourmaline are the only detrital varieties present.

Rare-earth element (REE) patterns preserved in mudstones provide significant constraints on provenance composition because they provide an overall average of REE in upper crustal rocks exposed at the surface to erosion (Taylor and McLennan 1985). Rare-earth element curves for mafic rocks have low slopes, whereas felsic rocks are enriched in light rare-earth elements (LREE), such as La, Ce, Pr, and Nd, to give steep curves. Intermediate curves imply mixing of the two end members. In addition to the steepness of the curve, the presence or absence of a negative Eu anomaly is also informative. Upper crustal rocks lacking such anomalies are considered to represent mantle derivatives. In contrast, upper crustal rocks with negative Eu anomalies such as granite and granodiorite are ascribed to intracrustal melting with retention of Eu in plagioclase in lower- or middle-crustal regions (Taylor and McLennan 1985). Archean mudstones characteristically lack negative Eu anomalies, whereas post-Archean mudstones display such anomalies. This contrast has been interpreted by Taylor and McLennan (1985) to record extensive crustal differentiation and craton development at the Archean-Proterozoic boundary, but their Archean sampling is heavily skewed to greenstone terranes, making the contrast equivocal.

Metapelite samples from the Beitbridge Complex show little evidence of modification by partial melting and hence their REE geochemistry is considered to reflect accurately the composition of the source terrain (Taylor *et al.* 1986). These data (Fig. 19.7) display variable slopes implying a mixed provenance. Samples LP20 and LP28 have typical Archean patterns with no negative Eu anomalies; sample LP20 was derived mainly from tonalitic granitoids or Na-rich felsic volcanic rocks and LP28 from a source with a larger mafic:felsic component. Rare-earth element curves for samples LP4, LP11, and LP30 resemble post-Archean shales. Enrichment of heavy REEs in sample LP30 is attributed to concentration of these elements in garnet as a result of metamorphic differentiation. The steep REE patterns and pronounced negative Eu anomalies displayed by these three samples are compatible with derivation from K-rich granites and granodiorites (Taylor *et al.* 1986).

Tectonic Setting

In the absence of primary sedimentary structures in metasedimentary rocks of the Beitbridge Complex, stratigraphic sequences, sedimentary facies, and sediment dispersal patterns cannot be reconstructed. Detailed paleoenvironmental analysis is thus impractical but the interpreted primary lithologies and lithologic proportions permit some general inferences on depositional environment and tectonic setting. In many respects the Beitbridge Complex is similar lithologically to the widely developed, cratonic-shelf quartz arenite-carbonate association present in younger, unmetamorphosed sequences dating back to 2.5 Ga. Examples include the lower part of the lower Proterozoic Transvaal Supergroup, South Africa (Tankard *et al.* 1982), the 1.9 Ga Epworth Group in Wopmay Orogen, Canada (Hoffman 1973, 1980), and, in particular, Cambro-

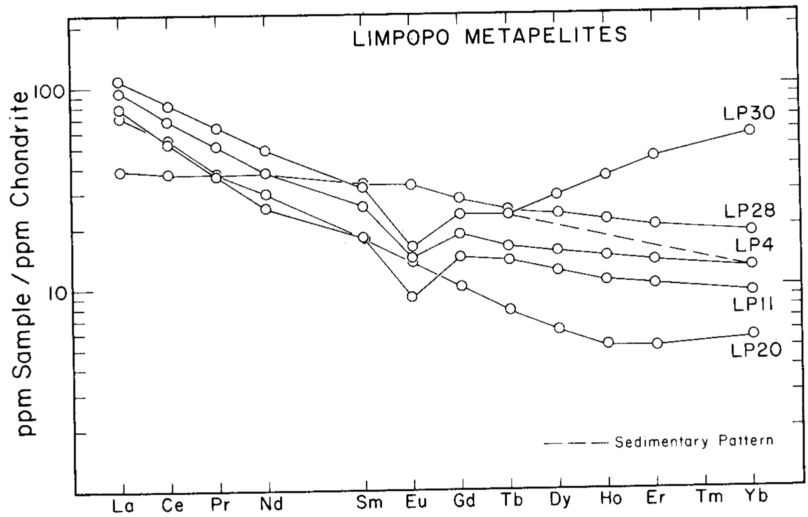


Fig. 19.7. Chondrite-normalized REE plot of metapelite samples in the Beitbridge Complex [compiled from data in Taylor *et al.* (1986)]. Heavy rare-earth element enrichment in sample LP30 is due to metamorphic differentiation and concentration of these elements in garnet. A primary sedimentary pattern would approximate the dashed line.

Ordovician sequences extending around much of North America (Stearn *et al.* 1984). The central Appalachian passive-margin sequence appears to be a particularly appropriate analog; it is notably rich in mudstone because of its near shelf-edge position in contrast to coeval platform sequences inboard to the west, which consist primarily of quartz arenite and carbonate. However, the data base for the Beitbridge Complex is inadequate to draw a direct analogy with a passive margin and the more general term *cratonic shelf* thus is preferred.

The REE geochemistry discussed above provides strong evidence for the exposure to erosion of an at least partly granitic terrain. We suggest that this terrain was continuous with the basement of the stable shelf on which the sedimentary protoliths of the Beitbridge Complex accumulated. Evidence for the erosion of products of intracrustal melting provided by the metapelite samples with steep patterns and negative Eu anomalies implies that 25 to 30 km thick continental crust existed prior to accumulation of the Beitbridge Complex. Taylor *et al.* (1986) consider that this granitic crust constituted small, stable continents or protocratons.

Discussion

The present configuration of the central zone of the Limpopo Province is a result of the following sequence of events: 1) Formation of continental crust greater than 25 to 30 km thick and probably represented by the Sand River Gneisses; 2) Intracrustal melting produced Eu-depleted felsic rocks of granitic and granodioritic composition; deformation of basement rocks took place prior to or after intracrustal melting; 3) Weathering and erosion accompanied/followed by sedimentation of quartz arenite, carbonate and mudstone on a stable continental shelf; 4) Intrusion of the Singelele and Bulai granitoids (Fig. 19.3); 5) Deformation and metamorphism at depths of 25 to 30 km; and 6) Uplift to the present surface.

The first three of these events and particularly the third are relatively well understood. The age of burial and metamorphism is constrained by the 2,650 to 2,700 Ma Bulai granitoid (Fig. 19.3), which was emplaced during or shortly after the peak of granulite metamorphism (Watkeys *et al.* 1983). The metasedimentary rocks of the Beitbridge Complex must therefore have been at mid-crustal depths at approximately 2.7 Ga to achieve the granulite facies assemblages seen at the present surface. Underthrusting of basement and cover rocks is attributed to Himalayan-style collisional tectonics at 2.7 Ga, and subsequent uplift to isostatic readjustment (Kidd 1985; Burke *et al.* 1986; Van Reenen *et al.* 1987).

Western Gneiss Terrain: Western Australia

Geologic Setting

The geologic framework of the Western Gneiss Terrain is similar in many respects to the central zone of the Limpopo Province. At Narryer in the north of

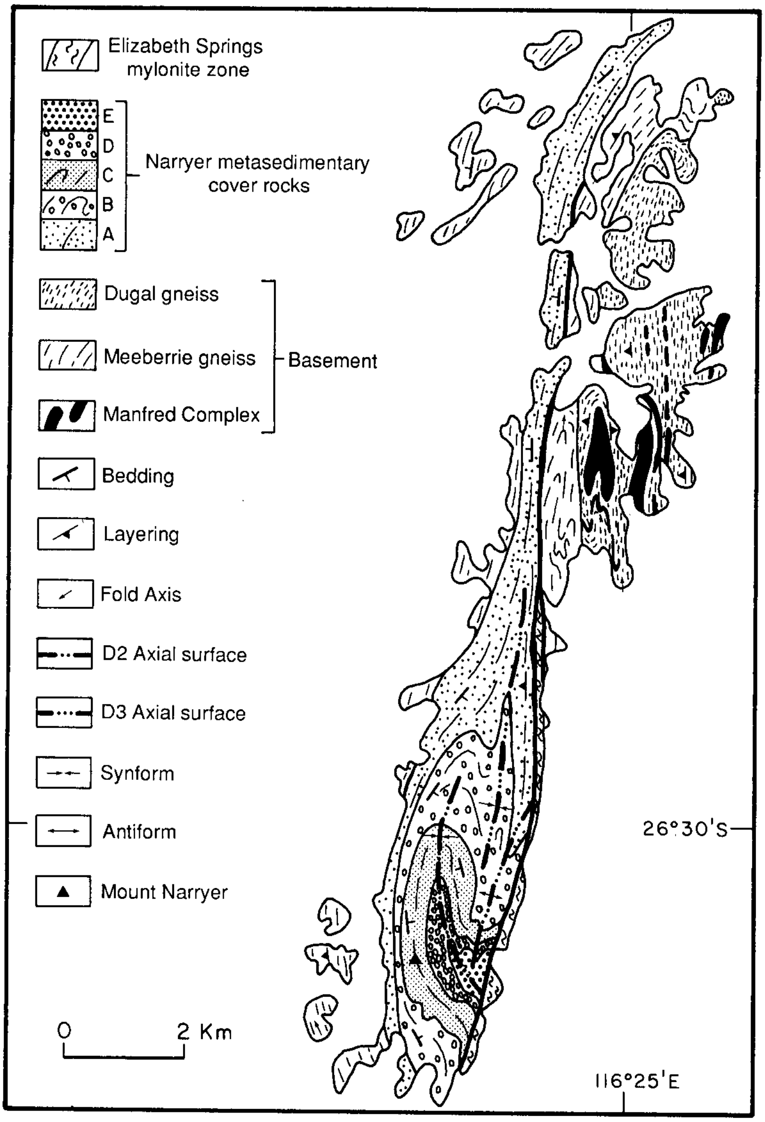


Fig. 19.8. Geological map of the Narryer region in the Western Gneiss Terrain [adapted from Myers and Williams (1985)]. See Figure 19.1 for location and Figure 19.10 for description of cover rocks.

the Western Gneiss Terrain (Fig. 19.1), basement rocks are identified on the basis of a pre-cover structural and metamorphic history (Myers and Williams 1985; Kinny *et al.* in press). Basement rocks (Fig. 19.8) consist primarily of the Meeberrie and Dugel Gneisses dated respectively at 3,630 ± 40 and 3,510 ± 50 Ma (Sm-Nd model; DeLaeter *et al.* 1981) and $3{,}688^{+33}_{-23}$ Ma and $3{,}416^{+82}_{-50}$ Ma (U-Pb zircon; Kinny *et al.* in press); the Dugel Gneiss contains xenoliths of layered gabbroic anorthosite (Manfred Complex; Fig. 19.8) that yield zircon ages of $3{,}750^{+72}_{-40}$ Ma (Kinny *et al.* in press). Zircon rim ages of $3{,}319^{+34}_{-16}$ Ma and a Rb-Sr whole rock isochron age of 3,350 ± 43 Ma are interpreted as metamorphic ages associated with gneiss formation (DeLaeter *et al.* 1981; Kinny *et al.* in press). Metasedimentary rocks, for which no detrital zircon ages are available, are present within the gneisses as either xenoliths or tectonically interleaved bodies of mainly pyroxene and sillimanite-mica quartzite and banded quartz-magnetite-pyroxene rocks (Myers and Williams 1985).

Cover rocks in the Narryer area (Fig. 19.8) consist of a ca. 2.5 km thick sequence of metaconglomerate, quartzite with well preserved crossbeds (Fig. 19.9), and subordinate paragneiss (Gee *et al.* 1981; Myers and Williams 1985). Cores of detrital zircons in the cover rocks are mainly ca. 3,750 Ma and ca. 3,500 Ma but zircons as young as 3,250 Ma and as old as 4,200 to 4,100 Ma are also present (Froude

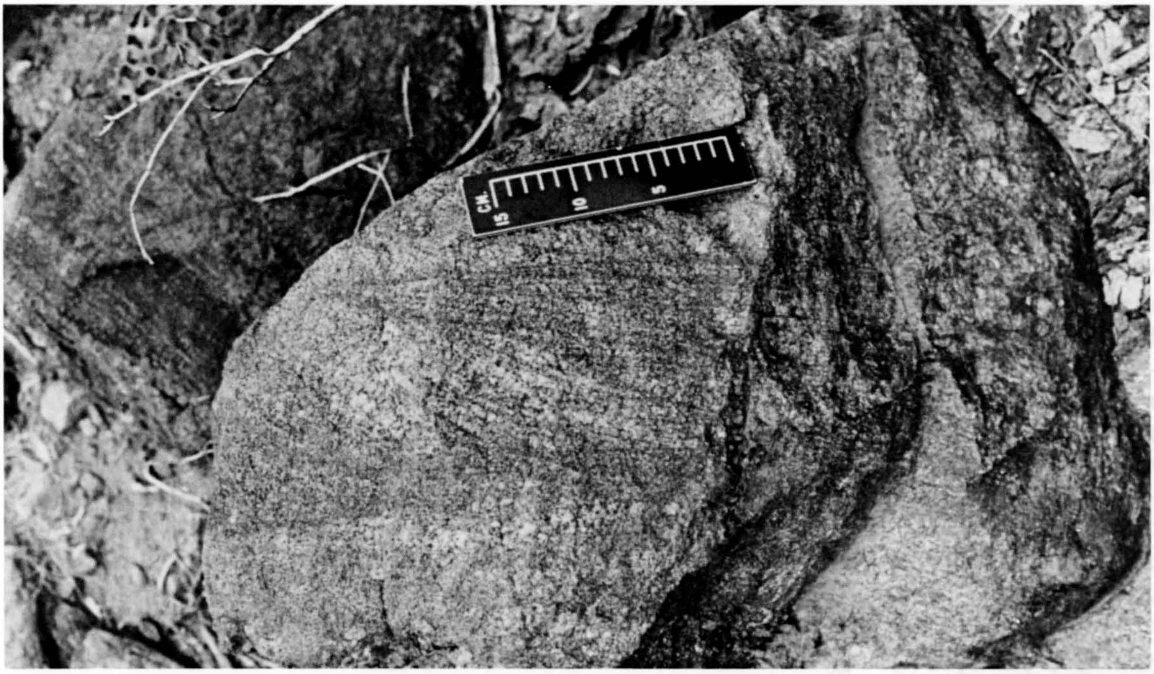


Fig. 19.9. Crossbeds preserved in aluminous gneisses (original wackes) in Unit C in the Narryer cover rocks of the Western Gneiss Terrain. Foresets are defined by metamorphic garnet. Note lack of tectonic modification of primary crossbed shapes.

et al. 1983). Metamorphic rims on detrital zircons cluster around 2,800 Ma (U-Pb zircon; Kinny 1986). The age of the cover rocks is thus constrained at less than 3,250 Ma, the age of the youngest detrital zircons and greater than 2,800 Ma, the age of metamorphic rims on detrital zircons. Sedimentation is considered to have taken place closer to 3,250 Ma than 2,800 Ma (Taylor *et al.* 1986). The 2,800 Ma prograde metamorphism and accompanying deformation (D2) of the basement and cover rocks took place under granulite conditions (Blight and Barley 1981). Subsequent retrograde, amphibolite facies metamorphism was associated with a third phase of deformation (D3, Fig. 19.8; Myers and Williams 1985).

Stratigraphic Framework of the Narryer Cover Rocks

In marked contrast to the Limpopo Province, primary sedimentary structures and clast shapes are remarkably well preserved in the Narryer region despite the deformation and granulite facies metamorphism. Crossbeds display no flattening or distortion of shape (Fig. 19.9) and the conglomerates show essentially no evidence of strain except for local rigid body rotation of some clasts into the plane of weak cleavage. Both early and late mylonites cut across the cover rocks. Primary structures are destroyed in the mylonite zones but are preserved in lensoid bodies between the mylonite zones. Apparently strain partitioning into the mylonites spared most of the primary features from deformation.

The widespread occurrence of conglomerate and crossbedded quartzite provides unequivocal evidence that the Narryer cover rocks are of siliciclastic origin. Unlike the Beitbridge Complex in the Limpopo Province, where no stratigraphic relationships can be deduced, a stratigraphic sequence can be recognized and mapped in the Narryer cover rocks. Primary sedimentary structures indicate that the strata are not overturned and a five-fold subdivision is recognizable (Figs. 19.8, 19.10). Each unit is lithologically distinct and it is possible that the ca. 2.5 km thickness closely approximates the depositional thickness. However, in the absence of marker beds, structural interleaving within units and ductile attenuation cannot be ruled out. Older folded and

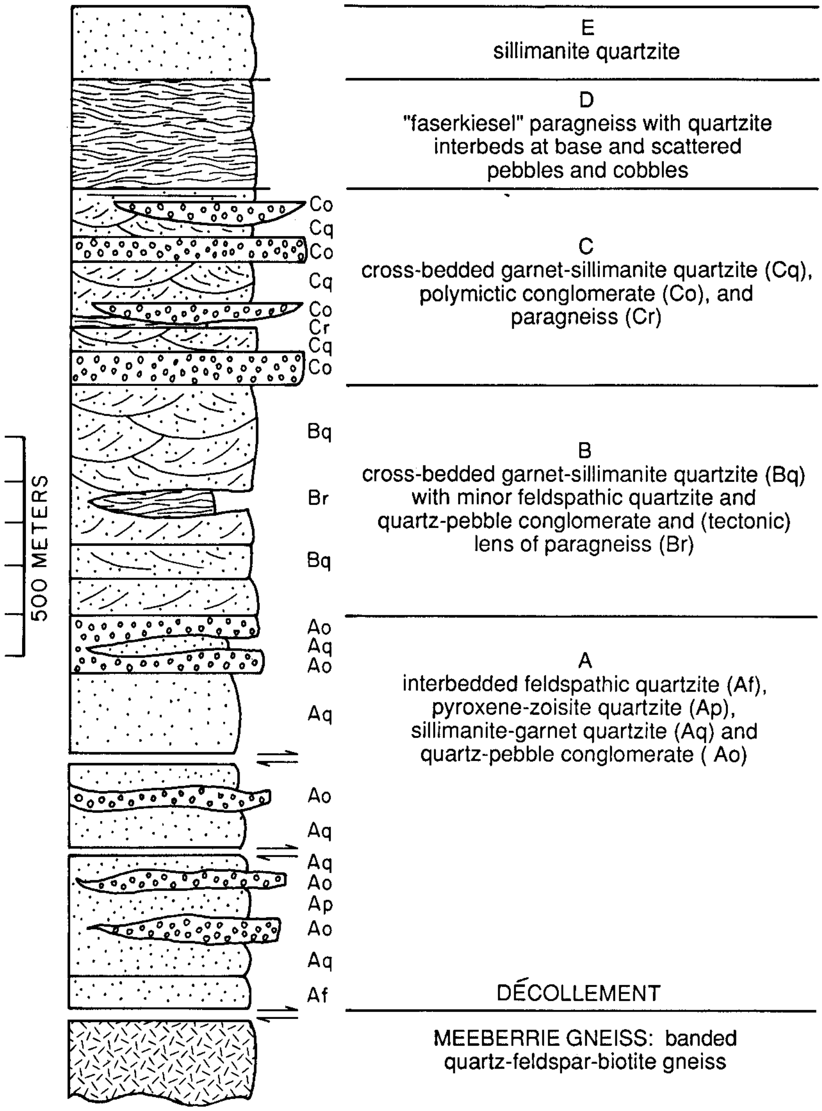


Fig. 19.10. Generalized stratigraphic column through the Narryer cover rocks. Thicknesses from maps of Williams and Myers (in press).

younger unfolded mylonitic shear zones within the cover rocks suggest possible structural duplication.

The base of the cover sequence is a mylonitic shear zone. Unit A at the base of the sequence is a succession of feldspathic, pyroxene-zoisite, and sillimanite-(garnet) quartzites with interbedded sheets of quartz-pebble conglomerate (Fig. 19.10). Unit B is composed of sillimanite-garnet(cordierite) quartzite with well developed crossbedding; also present are subordinate feldspathic quartzite, quartz-pebble conglomerate, and a single lens of paragneiss. Unit C also contains quartz-pebble conglomerate but is defined by the first appearance of polymictic conglomerate interbedded with cross-bedded garnet-sillimanite-(cordierite) quartzite. Cordierite-garnet-biotite-sillimanite gneiss and bimineralic quartz-garnet horizons are locally interbedded with the crossbedded quartzite. Conglomerate occurs as sheets and lenses containing clasts of banded magnetite-quartz-pyroxene-amphibole, vein quartz, quartzite and "garnetite" (garnet-quartz rock). Vein-quartz cobble conglomerate lenses with a sillimanite-rich matrix occur locally. Upper Unit C quartzites interfinger with paragneiss and conglomerate of Unit D that are characterized by the occurrence of distinctive "faserkiesel" of intergrown sillimanite and quartz. Unit D is abruptly overlain by sparsely sillimanitic quartzite of Unit E; primary stratification cannot be identified in either of Units D or E that occur in the core of the Narryer Syncline (Fig. 19.8).

The presence of diverse metamorphic minerals indicates that the original sandstones were texturally immature with variable but significant amounts of clay. Sillimanite is widespread and can constitute up to 20% of the quartzites. A lack of metamorphic K-feldspar in these quartzites indicates that the reaction muscovite + quartz $\rightleftharpoons$ K-feldspar + sillimanite + H_2O did not take place (cf. Winkler 1974); this implies that the protolith was an aluminous, potassium-poor mineral such as kaolinite and not illite. The other widespread metamorphic mineral is the iron-aluminum garnet almandine, which can make up to 25% of the quartzite or as much as 50% of the intercalated garnet-quartz horizons. The protolith of the latter is enigmatic but ferruginous mudstones and siltstones, such as in the Witwatersrand Supergroup, South Africa (Fuller *et al.* 1981), are of similar bulk composition. By analogy, garnet within the crossbedded quartzites could have been derived from ferruginous clay. Pure Fe-Al clays, however, are uncommon, especially in association with kaolinite and the possibility exists that at least some of the iron in the siliciclastic sediments was present as iron oxide. If so, it is tempting to suggest that these metasedimentary rocks originally may have been redbeds in which case they would be the oldest occurrence in the geologic record.

The quartzite, vein-quartz, and banded quartz-magnetite-pyroxene-amphibole clasts are typically rounded; the banded clasts represent original iron formation. "Garnetite" clasts are both rounded and angular and are interpreted as original intraformational ferruginous siltstone-mudstone clasts. The biotite-garnet-sillimanite-cordierite paragneiss horizons represent original pelitic sedimentary rocks.

Source-Area Composition and Age

Information on the composition of the source area from which the Narryer cover rocks were derived can be obtained from clast types, heavy minerals, and REE data from metapelites. Banded quartz-magnetite-pyroxene-amphibole cobbles and boulders are the only diagnostic clasts in the conglomerates; these are identical to banded rocks in the sedimentary outliers within the basement Meeberrie Gneiss and may indicate that these outliers are xenoliths, or older sedimentary rocks tectonically interleaved during gneiss formation at 3,350 Ma prior to deposition of the Narryer cover rocks, instead of tectonically interleaved cover rocks. Quartzite clasts may have been derived from associated lithologies in the outliers or could represent recycled cover rocks; the quartzites from these two locations are indistinguishable in the field.

As in the Limpopo Province, the detrital heavy mineral assemblage in the Narryer cover rocks is dominated by zircon and rutile with sphene and tourmaline comprising less than 1% of the population. Heavy minerals thus provide little information on source-area composition. Rare-earth element data from metapelites are more informative. Three REE curves have steep slopes and prominent negative Eu anomalies (Fig. 19.11; Taylor *et al.* 1986). There is no evidence of a mixed provenance of the type recognized in the Limpopo Province. Rather, these steep curves suggest that the cover rocks were derived from K-rich granitic or granodioritic plutons implying that thick, differentiated crust existed in the source area.

The predominant ages of detrital zircon cores of ca. 3,500 Ma are comparable to the Sm-Nd and U-Pb zircon ages of the basement Meeberrie and Dugel Gneisses. These gneisses are of granitic to granodioritic composition (Myers and Williams 1985) and, together with the sedimentary xenoliths, are a likely source for the cover rocks. The 3,750 Ma detrital zircons were probably derived from the Manfred Complex but the origin of the 4,200 to 4,100 Ma detrital zircons is unknown at present.

Sediment Dispersal Patterns and Depositional Environments

The excellent preservation of primary stratification in original sandstones and primary clast shapes in conglomerates, especially in Units B and C, permits a general facies analysis of the Narryer cover rocks. Three facies associations are present; stratified sandstone, clast-supported conglomerate, and matrix-supported conglomerate. Sandstone facies include trough crossbeds with subordinate tabular-planar crossbeds and horizontal stratification. The crossbeds indicate unimodal flow in a general southeasterly direction but with considerable variability in flow direction. Stratified sandstone is intimately associated with clast-supported conglom-

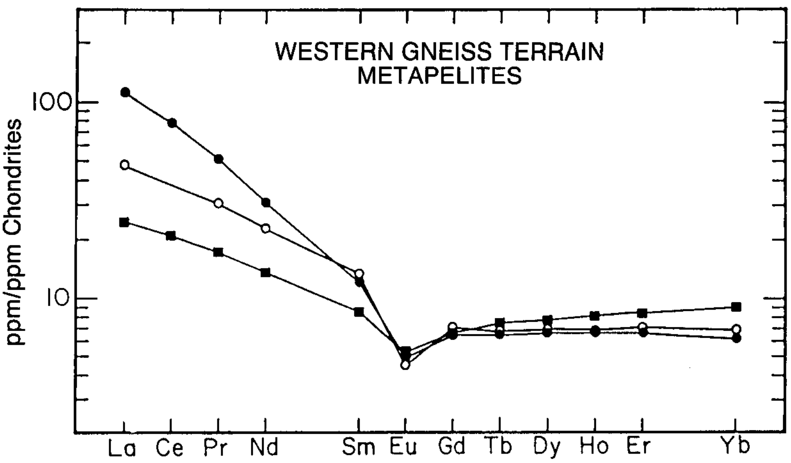


Fig. 19.11. Chondrite-normalized REE plot of metapelite samples from the Narryer cover rocks [from Taylor *et al.* (1986); reproduced with permission from Pergamon Press, Ltd.].

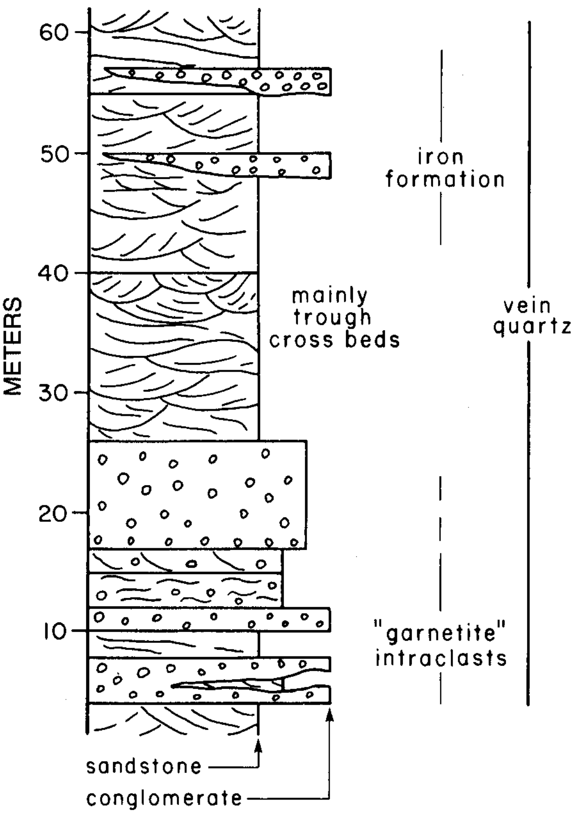


Fig. 19.12. Measured stratigraphic section through part of Unit C in the Narryer cover rocks illustrating the use of conventional sedimentology despite deformation and granulite facies metamorphism. Section located just north of Mt. Narryer.

erate in vertical sequence in Units A and C (Fig. 19.12). Detailed mapping of these two facies associations in Unit C demonstrates a primary lenticularity (Fig. 19.13) reflecting a channel and bar-type geometry typical of a high-energy, braided fluvial system (cf. Rust 1978). Unit A displays a similar interfingering of conglomerate and sandstone. Depositional environments are difficult to reconstruct for the crossbedded sandstones of Units B and E (Fig. 19.10). Possible depositional settings include braided river, tidal, eolian, and shoreface-shelf. Sediment-body geometries and stratification styles characteristic of tidal and eolian systems are not developed (cf. Hunter 1977; Homewood and Allen 1981; Kocurek 1981; Terwindt 1981). The presence of metamorphic sillimanite within the quartzites indicates that kaolinite was a significant constituent of the sandstone protolith. An environment of limited winnowing can thus be inferred; this inference is further evidence against a tidal or eolian origin and probably also against a shoreface-shelf origin for the sandstones. Furthermore, quartz-pebble conglomerates present in Unit B are not characteristic of tidal or eolian settings. Despite the absence of vertical sequences of facies and architectural elements typical of braided-alluvial systems (Miall 1977, 1985), a low-energy braided river is the preferred depositional environment of Units B and E.

The third facies consists of well rounded, vein-quartz cobble, matrix-supported conglomerates with a sillimanite-rich matrix. The matrix may compose up to 90% of the rock and from the earlier discussion

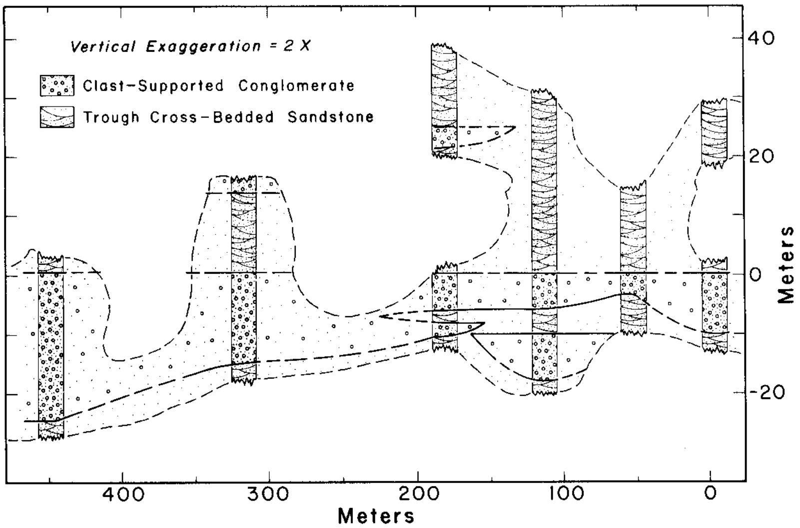


Fig. 19.13. Measured sections through part of Unit C. Top of a conglomerate bed was used as a datum for correlation of sections. Cross section illustrates syn-depositional lenticularity of conglomerate within sandstone.

was probably a sandy kaolinitic mudstone. Matrix-supported conglomerates make up most of Unit D and occur locally in Unit C as broad lenticular bodies hundreds of metres wide and tens of metres thick; these bodies interfinger with clast-supported conglomerates and associated crossbedded sandstones. Based on their close spatial relationship with inferred braided-alluvial deposits, the matrix-supported conglomerates are considered to be of subaerial origin. Possible depositional processes include debris flows and mudflows.

Basin Configuration and Tectonic Setting

Basin configuration and paleogeographic relationships of the facies present in the Narryer cover rocks are unclear. The braided-alluvial sediments may be visualized as a trunk system that flowed in a general southerly direction, whereas the lenticularity, in north-south trending outcrops, of matrix-supported conglomerates in Unit C suggests transverse flow from the east or west. However, paleoflow azimuths are unreliable due to probable rotation of the cover sequence during late stages of deformation.

Rare-earth element geochemistry from the Narryer cover rocks (Fig. 19.11) provides evidence for thick continental crust in that area prior to 3.3 Ga. Other than requiring the existence of continental basement, the tectonic setting in which the cover rocks developed is problematic. The inferred debris-flow or mudflow deposits indicate at least moderate relief, which in intracratonic settings is characteristically associated with faulting. Comparable thicknesses of debris flow or mudflow, and coarse-grained, braided-alluvial deposits have been interpreted as rift sequences (Miall 1981). However, in view of the poor constraints on the configuration of the Narryer basin, it is not possible to differentiate rift from nonmarine foreland basin settings similar to those described by Anderson and Picard (1974), DeCelles (1986), and DeCelles *et al.* (1987).

Discussion

The geological evolution of the Narryer region was similar to the central zone of the Limpopo Province and involved formation of 25 to 30 km thick continental crust, intracrustal melting to produce the K-rich protoliths of the Meeberrie and Dugel Gneisses, deformation and metamorphism of the basement and cover rocks at depths of 16 to 18 km, followed by uplift to the present surface. The age of burial and metamorphism is well constrained at 2.8 Ga but the mechanisms responsible for crustal thickening at this time are not understood. Gee *et al.* (1986) consider the possibility of collision of two unrelated segments of crust to explain the regional relationships between the high-grade Western Gneiss Terrain and the low-grade, granite-greenstone terrane of the Yilgarn Block to the east (Fig. 19.1); this proposed event might have been associated with the necessary crustal thickening in the Western Gneiss Terrain at ca. 2.8 Ga.

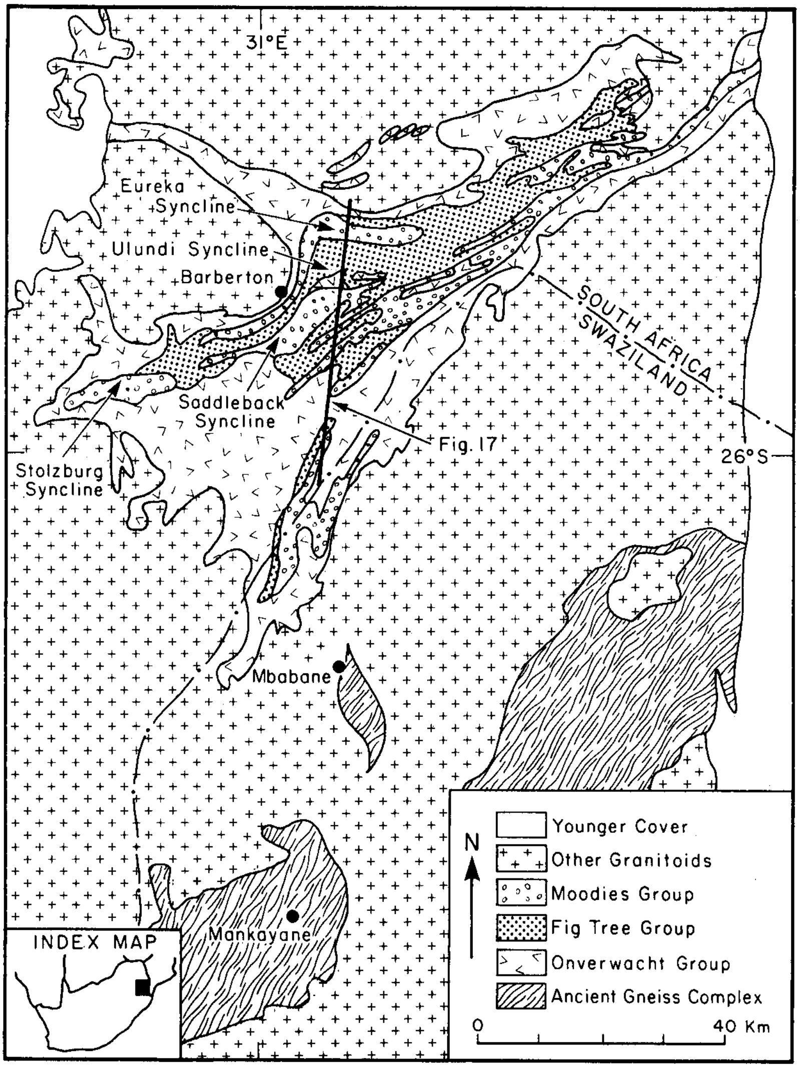


Fig. 19.14. Generalized geological map of the Barberton Greenstone Belt, Ancient Gneiss Complex, and surrounding granitoids [adapted from Tankard *et al.* (1982)]. Index map shows location in southern Africa.

Barberton Mountain Land: South Africa and Pilbara Block: Western Australia

Geologic Setting

The Barberton Mountain Land and Pilbara Block (Fig. 19.1) have long been considered as classical examples of Archean granite-greenstone terranes; in each terrane arcuate greenstone belts are surrounded by granite and gneiss (Figs. 19.14, 19.15). Metamorphism within the greenstone belts is greenschist grade or lower; in proximity to intrusive plutons amphibolite grades are locally attained. The Ancient Gneiss Complex southeast of the Barberton Mountain Land (Fig. 19.14) consists of 3.5 Ga metavolcanic and metasedimentary rocks, a tonalitic-gneiss batholith, metabasite dikes, meta-anorthosites, and ca. 3.3 Ga undeformed granitoids (Jackson *et al.* 1987).

A common two-fold stratigraphic subdivision is recognized within the greenstone belts; namely, lower volcanic-dominated and upper siliciclastic-dominated intervals (Fig. 19.16; Viljoen and Viljoen 1970; Hickman 1981). The lower volcanic-dominated intervals, the Onverwacht and Warrawoona Groups (Fig. 19.16), consist mainly of mafic-ultramafic volcanic rocks that are characteristically pillowed and in the Barberton Mountain Land contain komatiites. Subordinate felsic volcanic rocks are present in both regions; these display calc-

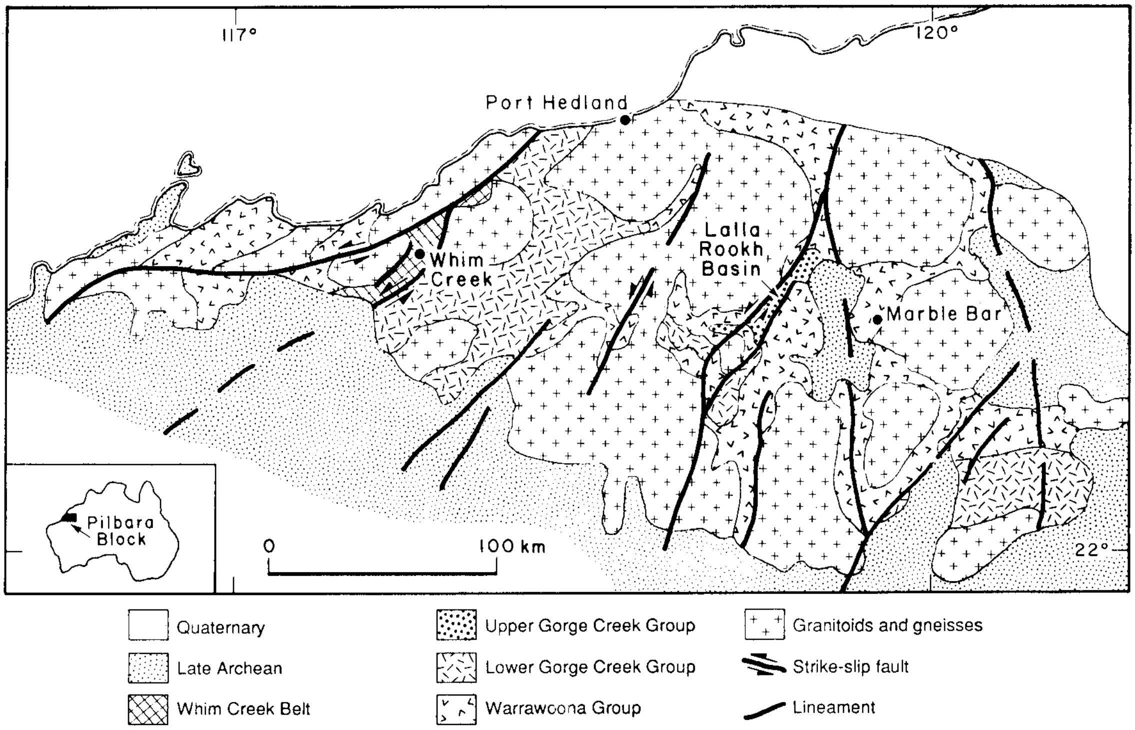


Fig. 19.15. Generalized geological map of a portion of the Pilbara Block [adapted from Hickman (1983)]. Index map shows location in Australia.

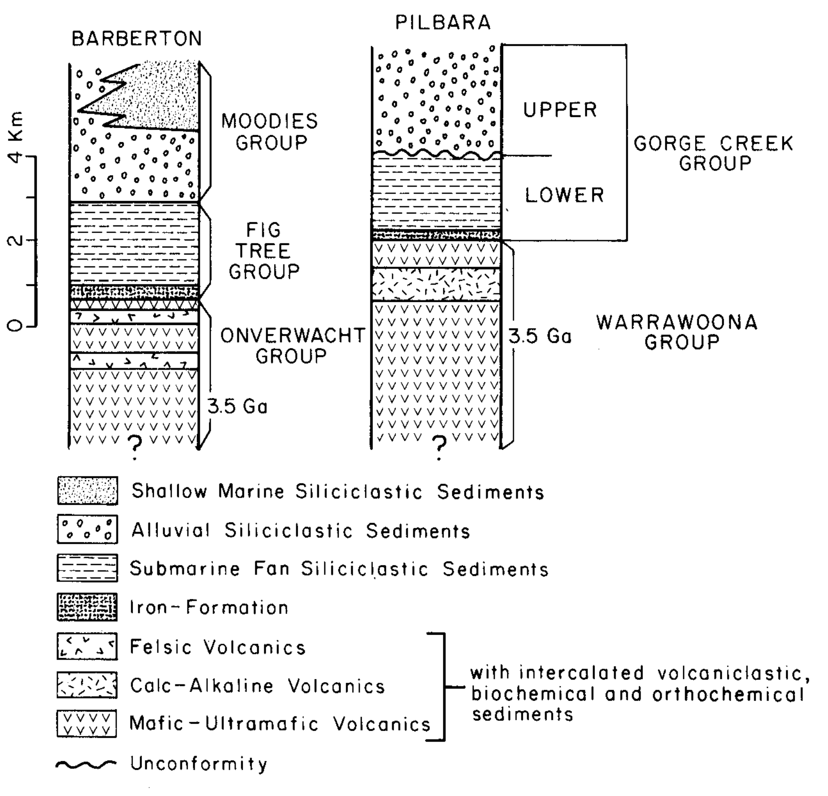


Fig. 19.16. Generalized stratigraphic columns of the Barberton and Pilbara greenstone belts [adapted from Anhaeusser (1973), Hickman (1983), Krapez (1984)].

alkaline affinities in the Pilbara Block (Barley *et al.* 1984). Lower Onverwacht mafic and felsic rocks together give a Sm-Nd isochron age of 3,540 ± 30 Ma (Hamilton *et al.* 1979), later revised to 3,530 ± 50 Ma (Hamilton *et al.* 1983). Warrawoona mafic volcanic rocks give a similar age of 3,560 ±32 Ma (Sm-Nd isochron; Hamilton *et al.* 1981). Minimum ages for the volcanic intervals are ca. 3,300 Ma (U-Pb zircon; Van Niekerk and Burger 1969; Pidgeon 1984).

Sedimentary rocks are minor but important components of the Onverwacht and Warrawoona Groups insofar as providing paleoenvironmental information. Lowe (1982) has recognized three types of sedimentary rocks in the volcanic sequences, which he categorizes as volcaniclastic, biochemical, and orthochemical. The sedimentary detritus is exclusively of intrabasinal derivation indicating an environment removed from continental influence. Furthermore, no continental basement to the volcanic intervals has been recognized in either region. These criteria have been used by Lowe (1982) to suggest that volcanism took place in an oceanic environment. Intercalated sedimentary rocks indicate that the preserved remnants of oceanic crust developed in shallow water (Barley *et al.* 1979; Lowe 1982). The mafic-ultramafic volcanism is considered to have formed relatively flat shield volcanoes with sporadic felsic cones that were often exposed to weathering and erosion (Lowe 1982).

The upper contact of the Onverwacht and Warrawoona Groups marks the base of the siliciclastic-dominated intervals and a dramatic change in tectonics as reflected by the Fig Tree-Moodies and Gorge Creek Groups (Fig. 19.16). Moodies Group sedimentation in the Barberton Mountain Land took place at less than 3,310 ± 10 Ma, the youngest age of granitoid cobbles in the basal conglomerate (U-Pb zircon; Tegtmeyer *et al.* 1981; Tegtmeyer and Kröner 1987), and before 3,200 ± 30 Ma, the age of an intrusive granite in the eastern mountain land (Fig. 19.14; Pb-Pb apatite; Oosthuysen and Burger 1973; recalculated by Cahen *et al.* 1984). The age of the Gorge Creek Group is less well constrained; direct geochronological evidence indicates an age between 3,300 Ma, the end of Warrawoona volcanism, and 2,768 ± 16 Ma, the age of volcanic rocks in an upper Archean sequence in the Pilbara Block (Fig. 19.15; U-Pb zircon; Pidgeon 1984). The upper Gorge Creek Group is considered to have accumulated at ca. 2,950 Ma (Krapez and Barley in press).

In the following sections, detailed basin analyses are presented for the Fig Tree and Moodies Groups in the Barberton Mountain Land (Figs. 19.14, 19.16) and the upper Gorge Creek Group in the Pilbara Block (Figs. 19.15, 19.16). These two sequences are well understood and provide an opportunity to evaluate the role of plate tectonics in the Archean. The lower Gorge Creek Group is not discussed; source-area composition and depositional environments were similar to the Fig Tree and Moodies Groups but sediment dispersal patterns, basin configuration, and tectonic setting are constrained poorly.

Stratigraphic Framework of the Fig Tree and Moodies Groups

Primary depositional textures and structures are well preserved throughout the Fig Tree and Moodies Groups, which crop out in a number of isolated synclines surrounded by Onverwacht volcanic rocks (Fig. 19.14). With the excellent preservation of "right-way-up" indicators in the form of primary sedimentary structures, a coherent stratigraphy can be mapped within each syncline. Correlation between synclines, however, is difficult in the Fig Tree Group because of the absence of marker or time horizons. In the southern half of the Barberton greenstone belt, the Fig Tree strata conformably overlie Onverwacht volcanic rocks and were involved in syndepositional folding and northwesterly directed, thrust-nappe tectonics prior to deposition of the unconformably overlying Moodies Group (Fig. 19.17; Lamb 1984; Lowe *et al.* 1985). The Fig Tree Group in this area commences with a 60 m-thick iron formation displaying evidence of synsedimentary deformation (Eriksson 1980a), overlain by an upward-coarsening sequence of sandstone and conglomerate. The overlying Moodies Group consists of a thick sequence of conglomerate with subordinate sandstone grading upward into sandstone (Fig. 19.17; Eriksson 1980a). In the northern half of the greenstone belt the Fig Tree and Moodies Groups are in conformable contact in the Saddleback, Stolzburg, and Eureka Synclines (Fig. 19.14). The Fig Tree Group consists primarily of ca. 2 km of interbedded wacke and mudstone with subordinate iron-formation (Eriksson 1980b), whereas the Moodies Group comprises one upward-fining sequence overlain by two upward-coarsening

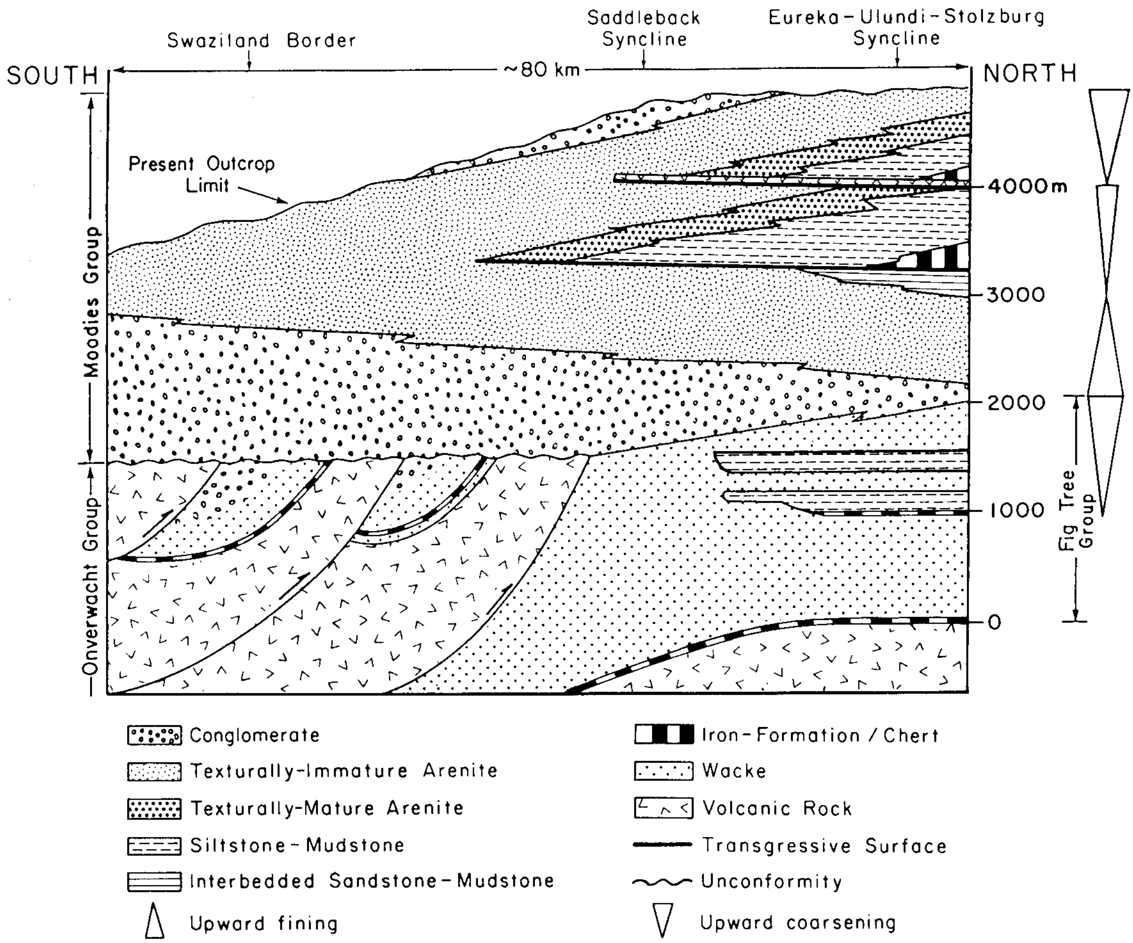


Fig. 19.17. Stratigraphic cross section of the Fig Tree and Moodies Groups and their relationship to the Onverwacht Group [based on data from Condie *et al.* (1970), Eriksson (1979), Lamb (1984), and Lowe *et al.* (1985)]. Location shown on Figure 19.14.

sequences; these sequences are ca. 1 km thick (Fig. 19.17; Eriksson 1979; Jackson *et al.* 1987). The base of the Moodies Group is defined by a conglomerate passing upward into a thick interval of sandstone followed by interbedded sandstone and mudstone. The upward-coarsening sequences are variable; in the Stolzburg and Eureka Synclines they commence with iron-formation grading into mudstone that contains increasing proportions of sandstone and siltstone upward and is capped by a thick sandstone. In the Stolzburg Syncline these sequences are coarser and consist primarily of sandstone and conglomerate (Fig. 19.17). Correlation of the Moodies Group among the Saddleback, Stolzburg, and Eureka Synclines is based on similarity of sequences and on an amygdaloidal lava horizon at the base of the uppermost, upward-coarsening sequence (Fig. 19.17). In the absence of biostratigraphic control, the sequence boundaries and particularly the lava horizon are taken to represent time planes.

Source-Area Composition and Age

Clast types in conglomerates, sandstone petrography, and mudstone geochemistry provide information on the composition of the provenance for the Fig Tree and Moodies Groups. White, black, and banded chert are the predominant clast types; subordinate vein-quartz, iron-formation, and felsic- and mafic-volcanic clasts are also present (Eriksson

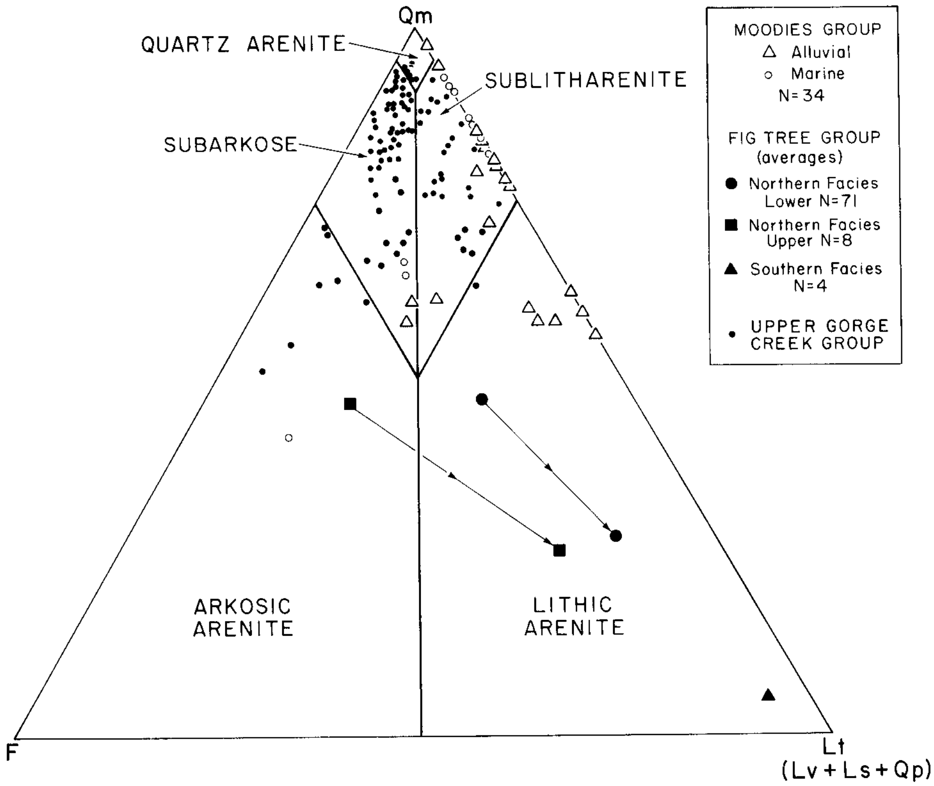


Fig. 19.18. Ternary plot of sandstone compositions in the Fig Tree, Moodies, and upper Gorge Creek Groups [data from Krapez (1984), Jackson *et al.* (1987), Eriksson (1980a), Herget (1966), and Reimer (1975) classification after Pettijohn *et al.* (1972)]. End members are monocrystalline quartz (Qm), total feldspar (F), and total lithics (Lt) incorporating volcanic lithics (Lv), sedimentary lithics (Ls), and polycrystalline quartz, mainly chert (Qp). For northern Fig Tree facies, average matrix component recalculated as Lt with shift towards lithic pole.

1978, 1980a). Granitoid cobbles and boulders occur in conglomerates of the southern Fig Tree outcrop belt and are abundant in the basal Moodies conglomerate in the northern Barberton greenstone belt (Gay 1969; Eriksson 1978; Heinrichs 1980). The granitoid clasts in the Moodies Group are massive or gneissic and vary from alkali granite and granite to granodiorite (Krupicka 1975; Reimer *et al.* 1985). Botryoidal quartz clasts and clasts of polymictic conglomerate are unique to the Moodies Group (Eriksson 1980a).

Sandstones in the Fig Tree and Moodies Groups become progressively more quartzose stratigraphically upward (Fig. 19.18). In the southern Fig Tree outcrop belt, sandstones are highly lithic, consisting of up to 90% metavolcanic rock fragments and angular chert (Eriksson 1980a). Sandstones in the northern outcrop belt are wackes with up to 35% matrix (Herget 1966; Reimer 1975); the matrix is considered to have been derived by diagenetic recrystallization of unstable lithic grains (Reimer 1972). If the matrix is recalculated as rock fragments, these sandstones are lithic arenites with chert and volcanic rock fragments the predominant components but with significant quantities of quartz and feldspar also present, in contrast to southern Fig Tree sandstones (Fig. 19.18). Lithic components include grains with perthitic intergrowth textures. Moodies Group sandstones are mainly sublitharenites with less common lithic, arkosic, and quartz arenites (Fig. 19.18; Jackson *et al.* 1987). The lithic character of most of the sandstones is due to their chert (Qp) component; volcanic rock fragments may constitute up to 10% of the arenites and, where present, orthoclase and microcline are the predominant feldspar varieties.

Clast types and sandstone petrography in the Fig Tree and Moodies Groups indicate a mixed provenance of Onverwacht volcanic rocks and associated chert, and granitoid and gneiss. Recycling of older

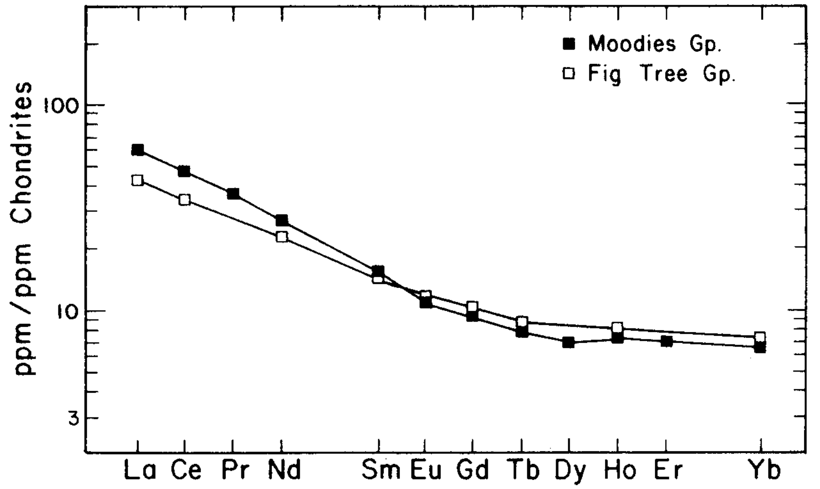


Fig. 19.19. Chondrite-normalized REE plot of averages of mudstone samples from the Fig Tree and Moodies Groups [compiled from data in McLennan *et al.* (1983)].

volcanic and sedimentary components is thus implied. Iron-formation clasts within all three groups suggest recycling within the predominantly siliciclastic intervals as iron formation is absent in the Onverwacht Group (Lowe 1982). Polymictic conglomerate clasts within the Moodies Group provide still stronger evidence of recycling.

Geochemical data from the Fig Tree and Moodies Groups substantiate the above provenance interpretation. Rare-earth element curves are relatively flat as reflected in low La/Yb ratios, and negative Eu anomalies are absent (Fig. 19.19; McLennan *et al.* 1983). Rare-earth element modelling implies a mixed provenance of mafic-ultramafic volcanic rocks and tonalites-felsic volcanic rocks; a significant plutonic contribution is indicated by the granitoid clasts in the Fig Tree and Moodies Groups and the grains of K-feldspar and perthitic intergrowth in several samples. The absence of negative Eu anomalies implies that the granitoid crust exposed to weathering and erosion had not undergone intracrustal melting, although granodiorites and monzonites without substantial negative Eu anomalies cannot be excluded (McLennan *et al.* 1983). In the Barberton Mountain Land, progressive unroofing of granitoids is reflected in the upward increase in quartz and feldspar (Fig. 19.18), Th, U (McLennan *et al.* 1983), and the steeper REE pattern in the Moodies versus the Fig Tree Group (Fig. 19.19). Rare-earth element patterns for the Fig Tree Group require a substantial mafic contribution, whereas light REE enrichment in the Moodies Group requires that felsic igneous rocks constituted up to 80% of the provenance but with less than 20% consisting of granitoids with significant Eu anomalies (McLennan *et al.* 1983). This minor component of the source terrain could have supplied the K-granite clasts and K-feldspar in the Moodies Group. High Ni and Cr contents of the Fig Tree and Moodies Groups (Danchin 1967) support the contention that mafic-ultramafic rocks were a significant component of the source terrain but the high concentrations of these two elements are not compatible with the REE modelling (McLennan *et al.* 1983).

Available data indicate that the age of the provenance of the Fig Tree and Moodies Groups varied between ca. 3.3 and 3.5 Ga. Detrital zircons from the Fig Tree Group have a mean age of 3.52 Ga, whereas granitic and gneissic clasts in the basal Moodies Group vary in age from 3.3 to 3.47 Ga (U-Pb zircon; Tegtmeyer and Kröner 1987). The geochronological data also are consistent with derivation of the Fig Tree and Moodies Groups from the Onverwacht volcanics and Ancient Gneiss Complex.

Sediment Dispersal Patterns and Depositional Environments

A range of paleoenvironments has been recognized in the Fig Tree and Moodies Groups despite the antiquity of the rocks and the consequent lack of fossils. These paleoenvironments are the subjects of a number of papers and will be reviewed only briefly below with the objectives of: 1) illustrating that detailed facies analysis is feasible in Archean strata, and 2) providing the data base for the analysis of basin configuration and tectonic setting.

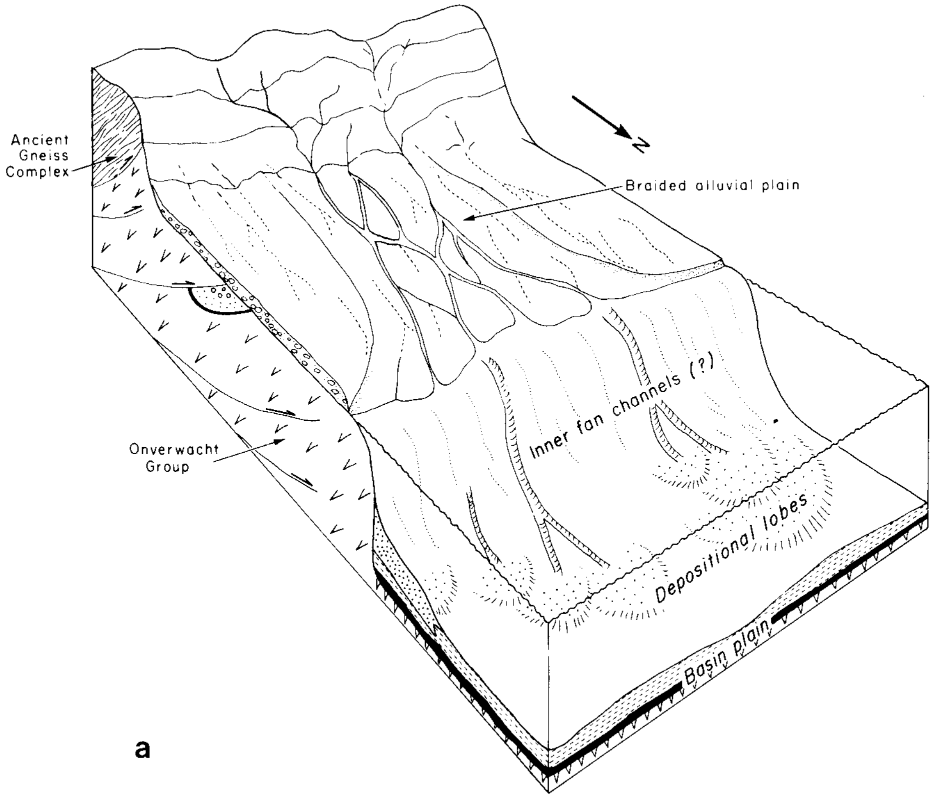
Fig Tree Group strata in the southern outcrop belts (Fig. 19.17) display many affinities with submarine-fan deposits (Eriksson 1980a), but the upward-coarsening sequences preserved in these outcrops are also compatible with a fan delta (Nocita and Lowe 1985). Above a basal chert horizon, the northern Fig Tree Group consists almost entirely of "classical" turbidites consisting of T_{ae}, complete and base-missing Bouma (1962) beds; flute casts indicate flow to the north. Specific subenvironments are not recognized but the overall stratigraphic sequence is interpreted as a lobe to slope transition (Fig. 19.20a; Eriksson 1980b); slope canyons are only inferred. Sandstones of the southern Fig Tree facies are considerably more lithic than those in the northern outcrop belt (Fig. 19.18). This characteristic, coupled with the presence of an unconformity between the Fig Tree and Moodies Groups in the south, has led Jackson *et al.* (1987) to suggest that the southern occurrences of the Fig Tree Group were deposited prior to those in the north and were being involved in northward-directed, fold-thrust tectonics at the time that the northern Fig Tree facies were accumulating (cf. Figs. 19.17, 19.20a). Submarine lobes of the northern Fig Tree Group were fed by a fluvial system represented by coeval braided-alluvial conglomerate and sandstone of the basal Moodies Group to the south (Figs. 19.17, 19.20a); paleocurrents from trough crossbeds indicate flow exclusively to the north.

Overlying conglomerates and texturally immature arenites in the Moodies Group (Fig. 19.17) are also of braided-alluvial origin. Within these sediments representatives are recognized of the Scott and Donjek models (Eriksson 1978; cf. Miall 1977). Northerly paleocurrent vectors, coupled with the northward transition into texturally mature arenite, siltstone, mudstone, and iron formation of inferred shallow-marine origin (Fig. 19.17; Eriksson 1979), indicate that a northerly paleoslope prevailed throughout Fig Tree and Moodies Group sedimentation.

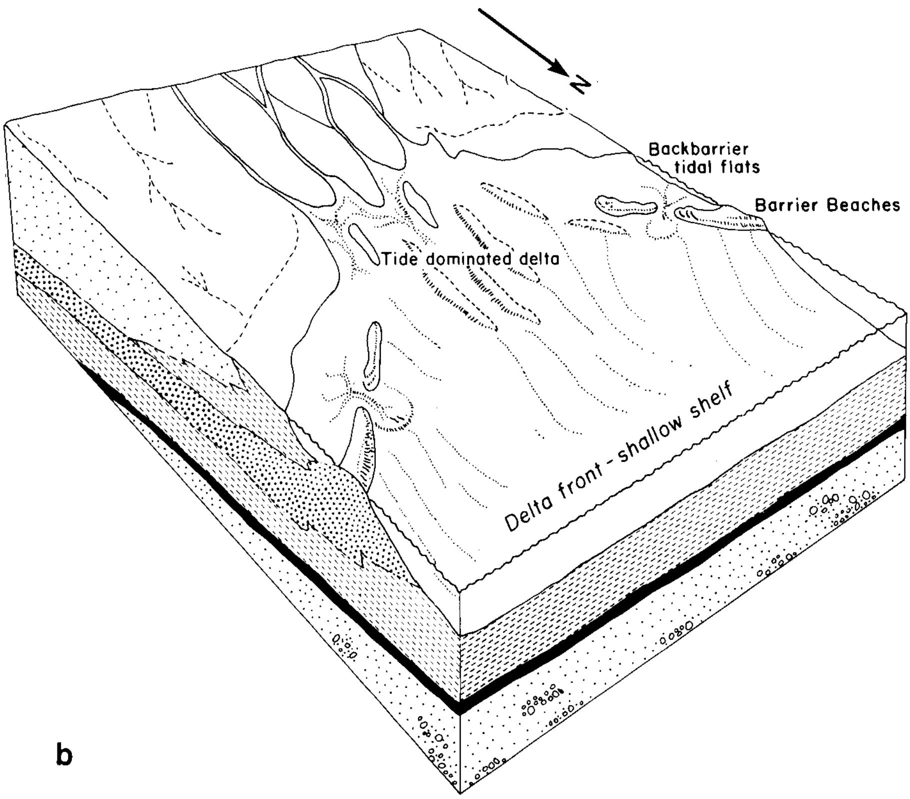
Shallow-marine facies are particularly common in the northern outcrop belt of the Moodies Group where they comprise the two upward-coarsening sequences in the Eureka and Stolzburg Synclines and the lowermost upward-coarsening sequence in the Saddleback Syncline (Fig. 19.17). Both sequences in the Eureka and Stolzburg Synclines are interpreted as progradational barrier-beach deposits in which distal-shelf iron formation passes upward (landward) into outer-shelf siltstone and mudstone followed by texturally mature, shoreface arenites (Eriksson 1979). Foreshore deposits are characterized by swash lamination with low-angle discordances and abundant wave ripples. The lowermost upward-coarsening sequence in the Saddleback Syncline (Fig. 19.17) is interpreted as a tide-dominated delta that developed coevally with barrier-beach sedimentation to the east and west (Fig. 19.20b). Because tide-dominated deltaic deposits are rare in the geologic record, this sequence is discussed in some detail.

Delta-front deposits at the base of the upward-coarsening sequence are composed of coarse-tail graded sandstone laminae capped by mudstone drapes, which contain numerous small-scale flame structures. In addition, dewatering sheets cut across up to 10 m of section. Overlying texturally mature quartz arenites make up large-scale, lenticular bodies oriented perpendicular to the paleoshoreline and are interpreted as linear sandstone shoals (Fig. 19.20b). The interior of the shoals consists of northerly or ebb-oriented crossbeds with reactivation surfaces, whereas bimodal-bipolar crossbeds are present on the shoal margins. Similar asymmetrical flow exists on Holocene tidal sand shoals such as are present along the north coast of Australia. Emergent shoal deposits in the Moodies Group are represented by coarse-grained sandstone with horizontal stratification and small-scale crossbeds, and mudstone laminae and drapes that characteristically display evidence of exposure. An uppermost interval consists of interlaminated sandstone and desiccated mudstone enclosing ca. 2 m thick tidal-channel deposits and is interpreted as a delta-plain sequence. Basal lags in the channel deposits include extrabasinal as well as intrabasinal mudstone clasts and are

Fig. 19.20a. a: Paleogeographic model for the Fig Tree Group in the Barberton Mountain Land illustrating coeval uplift of the Ancient Gneiss Complex and Onverwacht Group [modified after Tankard *et al.* (1982)]. b: Paleogeographic model for the lowermost progradational sequence in the Moodies Group, Barberton Mountain Land [from Tankard *et al.* (1982)].



Ancient Gneiss Complex
N
Braided alluvial plain
Inner fan channels (?)
Onverwacht Group
Depositional lobes
Basin plain
a



N
Backbarrier tidal flats
Barrier Beaches
Tide dominated delta
Delta front - shallow shelf
b

overlain by crossbedded sandstones with reactivation surfaces and mudstone drapes on foresets and between crossbed sets. Horizontally stratified sandstone laminae with numerous desiccated mudstone partings cap the sequences. This deltaic sequence is considered to have accumulated under macrotidal conditions that existed in an embayed reach of the coastline similar to that in the German Bight of the North Sea (Fig. 19.20b; Eriksson 1979).

Basin Configuration and Tectonic Setting

Dispersal patterns coupled with the spatial distribution of facies (Fig. 19.17) indicate that the Fig Tree and Moodies sediments were derived exclusively from the south. Sedimentation took place along a general east to west-trending shoreline (Fig. 19.20). Influx of extrabasinal detritus to the previously oceanic domain represented by the Onverwacht volcanics has been related to uplift of the Ancient Gneiss Complex to the south (Fig. 19.14; Jackson *et al.* 1987). Structures in the complex record progressive crustal shortening beginning at ca. 3.4 Ga. Massive uplift at ca. 3.3 Ga, the maximum age of the Moodies Group, was associated with large-scale, heterogeneous simple shear upwards and to the north. Deepening of the basin at the onset of Fig Tree sedimentation is attributed to crustal loading associated with this thrusting (Fig. 19.20a). The Fig Tree and Moodies Groups thus represent synorogenic deposits and their stratigraphic succession records progressive infilling of the basin under deepwater followed by shallow-water conditions. Jackson *et al.* (1987) have interpreted the above stratigraphic and structural evidence as favoring a foreland or foredeep-basin setting for the Fig Tree and Moodies Groups. The basin was floored by oceanic volcanics that by the time of siliciclastic sedimentation, had been underplated by tonalitic granitoids varying in age from 3.5 to 3.3 Ga (Barton 1983c). Additional evidence in support of a foreland or foredeep-basin setting for the Fig Tree and Moodies Groups includes: 1) the abundance of recycled Onverwacht volcanic and sedimentary components within the Fig Tree and Moodies Groups suggesting that the Onverwacht Group was telescoped onto the Ancient Gneiss Complex during northward folding and shearing of the latter; 2) the presence of older components of the Fig Tree Group within younger units of the siliciclastic interval; and 3) the occurrence of features indicative of syndepositional crustal shortening within the Fig Tree and Moodies Groups such as synsedimentary folding and brecciation of iron-formation, large growth folds, and unconformities above thrust faults verging in the direction of sediment influx (Fig. 19.17).

The wave of crustal shortening that uplifted the Ancient Gneiss Complex advanced progressively northward with time to uplift firstly the Onverwacht Group followed by the Onverwacht Group together with older components of the siliciclastic interval and finally involved the whole greenstone belt in folding and thrusting. Jackson *et al.* (1987) argue that the greenstone belt was sheared off along its base onto tonalities of similar age to the upper Onverwacht Group. The allochthon is considered to have acted as a thermal blanket and dense layer to initiate diapirism, which produced the present upright structural style and arcuate outcrop pattern of the Barberton greenstone belt (Fig. 19.14).

Stratigraphic Framework of the Upper Gorge Creek Group

The upper Gorge Creek Group is confined almost entirely to the Lalla Rookh Basin (Fig. 19.15) where it unconformably overlies shallow-marine quartz arenites, basinal iron formations, and turbiditic conglomerates, sandstones, and mudstones of the lower Gorge Creek Group (Krapez 1984). This upper subdivision of the Gorge Creek Group consists mainly of conglomerate and sandstone with subordinate mudstone. Basin-fill thickness is asymmetric; maximum thicknesses of ca. 3 km are developed adjacent to the northwestern boundary fault and decrease to ca. 1 km towards the southeastern basin margin (Fig. 19.15). Stratigraphic relationships within the basin are complex and are discussed later in this section.

Source-Area Composition

Clasts in the upper Gorge Creek Group consist of black, white and banded chert, iron formation, various quartzites, and quartz-mica schist. Sandstones are predominantly subarkoses but display considerable variability in composition (Fig. 19.18). In general, the sandstones display an upward change from lithic to quartzose to feldspathic arenites (Krapez 1984). Microcline is the dominant feldspar in the sandstones. Clast types and sandstone petrography indicate a mixed provenance consisting of Warrawoona volcanics, lower Gorge Creek sediments, and granitoids. The upward trend in sandstone petrography reflects progressive exposure and erosion of K-rich granites.

Heavy minerals in the upper Gorge Creek Group support the above provenance interpretation. Pyrite and chromite are the predominant heavy minerals; various radioactive placer minerals are also present (Krapez 1984). Rounded pyrite grains are massive, concentrically zoned, laminated, framboidal or porous; euhedral pyrite also occurs. The range of pyrite morphologies is similar to that of syngenetic pyrite in Archean shales and it is probable that the pyrite was derived from sulfide-facies iron formations in the lower Gorge Creek Group. Chromite grains are up to 1 mm in diameter; the most likely source was from layered ultramafic intrusions within the Warrawoona Group as chromite in the extrusives is finer grained. Radioactive heavy minerals reflect the K-granite component of the provenance.

Sediment Dispersal Patterns and Depositional Environments

The upper Gorge Creek Group in the Lalla Rookh Basin consists exclusively of continental deposits; Krapez (1984) recognizes four major depositional environments. Alluvial-fan deposits are represented by breccias of talus-slope origin and clast-supported and matrix-supported conglomerates deposited by debris flows and mudflows. These deposits are confined to the margins of the Lalla Rookh Syncline (Fig. 19.15). Braided-alluvial strata are dominated by two facies associations. Conglomeratic assemblages consist of lenses of clast-supported pebbles and boulders representing longitudinal-bar deposits. Associated trough crossbedded, horizontally stratified and wedge-planar crossbedded sandstone facies are interpreted as adjacent channel, and bar-top and bar-margin deposits, respectively (cf. Miall 1977). Sandstone assemblages consist of trough crossbedded cosets with interbedded sets of tabular-planar crossbeds; this association is comparable to the South Saskatchewan model of Cant and Walker (1978) in which trough crossbeds are produced by aggradation of mid-channel bars and tabular-planar crossbeds develop in response to accretion off bar margins. Paleocurrent data from trough axes suggest that source areas were present on most margins of the basin. The data reveal two dominant fluvial dispersal patterns; one along the basin axis parallel to the boundary faults and the other at a high angle to the faults. Interbedded mudstone and fine-grained sandstone with horizontal lamination, ripple cross-lamination, and climbing ripples interfinger laterally with the braided-alluvial facies and are considered to represent floodplain deposits. Coarsening-upward sequences of mudstone grading into sandstone are visualized as lacustrine fan-delta deposits. The sandstones are trough crossbedded, or wave rippled and horizontally stratified and resemble channel and levee deposits, respectively. Associated lacustrine facies include sandstone with swash lamination and wave ripples, and lenticular bedding with desiccation cracks; these two associations are interpreted as lacustrine beach and mudflat deposits. Also present within the lacustrine association of facies are thinly bedded turbidites which accumulated below wave base.

Basin Configuration and Tectonic Setting

Sediment dispersal patterns and the spatial distribution of depositional environments have been used by Krapez (1984) to argue that the upper Gorge Creek Group accumulated in a small, enclosed, fault-bounded basin with roughly the same configuration as the present outcrop (Fig. 19.15). Alluvial fans along the high-gradient basin margins supplied terrigenous debris to a braided trunk system that was flanked by a floodplain and fed sediment to a lacustrine environment in the northern part of the outcrop belt (Fig. 19.21).

Krapez (1984) has argued that stratigraphic and sedimentological evidence in the upper Gorge Creek Group indicates that basin development was controlled by marginal strike-slip faulting. Criteria cited include: 1) the presence of intraformational low-angle unconformities; 2) the great stratigraphic thickness relative to the small basin area; 3) the elongate shape of the basin parallel to a northwestern boundary fault (Fig. 19.21) – the presence of numerous alluvial fans within the basin imply that this and other faults were active during sedimentation; 4) the asymmetry of basin fill and facies distribution; 5) the vertical stacking and limited lateral migration of depositional environments; and 6) the dominance of longitudinal infilling (cf. Reading 1980; Nilsen and McLaughlin 1985). The Lalla Rookh Basin is comparable in size, geometry and basin fill to small basins associated with strike-slip faulting in southern California (Crowell 1974; Nilsen and McLaughlin 1985), Norway (Steel and Gloppen

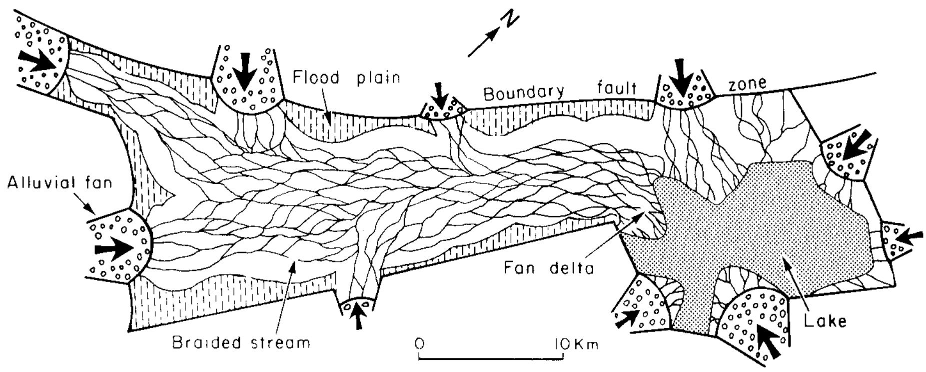


Fig. 19.21. Paleogeographic model for the upper Gorge Creek Group, Lalla Rookh Syncline [from Krapez (1984); reproduced with permission from Extension Division, University of Western Australia].

1980), and Turkey (Hempton *et al.* 1983). The Whim Creek Belt in the western Pilbara Block (Fig. 19.15) is envisaged similarly as a strike-slip basin (Krapez and Barley in press).

Problems and Perspectives

Techniques used in the analysis of the four basins are summarized in Table 19.1. This table clearly illustrates that the degree of sophistication of basin analysis increases with decreasing metamorphic grade and degree of structural complexity.

A stratigraphic framework is the starting point in a basin analysis. As for any Precambrian sequence, the lack of biostratigraphic control in the basins documented in this paper imposes limitations on correlation of stratigraphic units. In the Beitbridge Complex a lack of primary sedimentary structures prevents the recognition of a stratigraphic succession. In contrast, preservation of primary structures in the Narryer cover rocks as well as in the Fig Tree, Moodies, and upper Gorge Creek Groups allow reconstruction of stratigraphic sequences. At Narryer the possibility of structural duplication exists as mylonite zones are pervasive and marker horizons are absent. The stratigraphic succession in the Fig Tree and Moodies Groups is considered to be reliable, especially in the northern half of the Barberton Mountain Land. In this area marker beds are present and an amygdaloidal lava horizon that exists at the same stratigraphic level in a number of synclines can be used for stratigraphic correlation. Furthermore, deformational style is well understood. Despite complex folding associated with transpression, detailed facies mapping in the upper Gorge Creek Group has provided a reliable stratigraphic framework.

Source-area compositions are relatively well constrained using mostly conventional techniques. Slopes and shapes of REE curves are not widely utilized in Phanerozoic sequences to interpret provenance compositions but have received wide acceptance among Precambrian workers as important signatures of crustal compositions through time (Taylor and McLennan 1985). Source-area ages are based on conventional U-Pb (Fig Tree and Moodies Groups), and single-grain ion probe (Narryer) dating techniques. The latter approach to determining source-area ages is relatively new and provides a means of discriminating different ages in a mixed population of detrital zircons.

Determination of sediment dispersal patterns in deformed terrains requires stereonet rotation of paleocurrent data (Potter and Pettijohn 1977). The sequence of deformational events at Narryer can be reconstructed but the amount of rotation of the cover rocks during the late stages of deformation cannot be determined; paleocurrent azimuths thus are unreliable. In the Barberton Mountain Land folds are isoclinal and plunging but no rotation of the siliciclastic intervals accompanied folding; stereonet rotation procedures are thus simple and paleocurrent azimuths are considered to be reliable. Folding in the upper Gorge Creek is complex and noncylindrical; only general, relative flow directions can thus be determined.

The lack of fossils in Precambrian sequences in general imposes certain limitations on paleoenvironmental interpretations, especially in discriminating

Table 19.1. Summary of techniques used in basin analyses.

	Beitbridge Complex Limpopo Province South Africa	Narryer Cover Rocks Western Gneiss Terrain Western Australia	Fig Tree and Moodies Groups Barberton Mountain Land South Africa	Upper Gorge Creek Group Pilbara Block Western Australia
	Granulite facies metamorphism	Amphibolite/granulite facies metamorphism	Greenschist facies metamorphism	Greenschist facies metamorphism
Stratigraphic framework	Protolith identification Protolith proportions	Stratigraphic mapping Measured sections	Stratigraphic mapping Measured sections Correlation using lava horizon and sequence boundaries	Facies mapping Measured sections
Source-area composition	Slope and shape of REE curves Heavy minerals not informative	Clast types Slope and shape of REE curves Heavy minerals not informative	Clast types Sandstone petrography Slope and shape of REE curves	Clast types Sandstone petrography Heavy minerals
Source-area age	Data not available	Detrital zircons	Detrital zircons Granitoid clasts	Data not available
Sediment dispersal patterns	Cannot be reconstructed	Primary sedimentary structures	Primary sedimentary structures	Primary sedimentary structures
Depositional environments	Lithologic proportions	Facies interrelationships Facies modelling	Standard facies modelling	Standard facies modelling
Basin configuration including source-area location	Cannot be determined	Spatial interrelationships of depositional environments	Sediment dispersal patterns Spatial distribution of depositional environments	Sediment dispersal patterns Spatial interrelationships of depositional environments
Tectonic setting	Lithologic proportions REE data	REE data Basin configuration	Stratigraphic and structural relationships Stratigraphic evolution Basin configuration	Stratigraphic and sedimentological evolution Basin configuration

ambiguous marine from nonmarine deposits. Interpretations are based solely on facies modelling of physical criteria. In the absence of primary sedimentary structures in the Beitbridge Complex interpretation of depositional environments is equivocal, being based solely on proportions of interpreted protoliths.

Interpretations of basin configuration and tectonic setting are based largely on, and thus hinge on, the validity of the other components of the basin analyses. For example, the cratonic-shelf interpretation for the Beitbridge Complex may be invalid if the amphibolites were lavas or ash beds instead of sills or dikes as discussed earlier. If the amphibolite protolith were of extrusive origin, the Beitbridge Complex would represent an unique and unusual tectonic environment; no Holocene setting is known in which alternating quartz arenites and lavas or tuffs are accumulating. Furthermore, such an association is not known from less-deformed and metamorphosed sequences in the geological record. An unequivocal tectonic interpretation for the Narryer cover rocks is not possible because of post-depositional tectonic overprint and because basin configuration is not well constrained on the basis of only two depositional environments. In contrast, the interpretations of the Fig Tree and Moodies Groups as a foreland-basin sequence, and the upper Gorge Creek Group as a strike-slip basin deposit, are considered to be much less equivocal.

Discussion and Broader Implications

Available age constraints indicate that the high-grade terranes in southern Africa and Western Australia are of comparable age to the greenstone belts and developed between ca. 3.6 and 3.0 Ga, thus negating the possibility that these high-grade terranes represent basement to the greenstone belts. Primary lithological associations interpreted for cover rocks in the Limpopo Province and Western Gneiss Terrain are clearly different from lithological associations in the Barberton and Pilbara greenstone belts. The high-grade terranes in southern Africa and Western Australia thus do not represent the deep-seated metamorphosed roots of greenstone belts. Rather, the two terrane types discussed in this paper represent coeval but contrasting tectonic settings. Cover rocks in the high-grade terranes accumulated on thick continental crust and reflect a stable cratonic setting. In contrast, volcanic intervals in the greenstone belts are widely considered to have developed in an oceanic setting (Lowe 1982; Hoffman 1984; DeWit 1986), whereas the overlying siliciclastic intervals reflect active tectonic settings associated with crustal shortening.

On a global scale, lower Archean sedimentary rocks of cratonic platform affinity are uncommon. In addition to the examples from the Limpopo Province and Western Gneiss Terrain, pre-3.0 Ga sedimentary associations of this type occur *locally* in Canada (Schau and Henderson 1983), India (Srinivasan and Ojakangas 1986; Argast and Donnelly 1986), the United States (Muller *et al.* 1982), and the Soviet Union (Kazansky and Moralev 1981). In view of the low susceptibility of cratonic platform sedimentary rocks to tectonic recycling (Veizer and Jansen 1985), their rare occurrence in the lower Archean rock record suggests that few deposits of this type developed prior to 3.0 Ga. This was probably due to the limited development of stable continental crust until 3.0 Ga in South Africa (Hunter 1974a; Anhaeusser and Robb 1981), 3.0 to 2.85 Ga in Western Australia (Blake and McNaughton 1984), and at the Archean-Proterozoic boundary at 2.5 Ga in most parts of the world (Eriksson and Donaldson 1986). Veizer and Jansen (1979) and Taylor *et al.* (1986) have proposed that small, stable continents comprised no more than 10% of the pre-3.0 Ga Earth. If continental crust was widespread prior to 3.0 Ga as suggested by Armstrong (1981) and Reymer and Schubert (1984), lower Archean cratonic platform sedimentary rocks should be more common in the rock record than they appear to be.

In contrast to the high-grade cratonic association, remnants of lower Archean oceanic crust and overlying active-margin siliciclastic sediments are relatively common on a global scale; most Archean cratons contain greenstone belts. Such volcanic and sedimentary rocks have a high potential for tectonic recycling (Veizer and Jansen 1985) and their occurrence suggests that greenstone assemblages were too voluminous to be entirely subducted and, in part, were involved in obduction as suggested for the Barberton greenstone belt by Jackson *et al.* (1987).

Data discussed in this paper suggest that between 3.6 and 3.3 Ga the surface of the Earth, at least in southern Africa and Western Australia, consisted of small nucleii of differentiated continental crust surrounded by extensive oceanic environments. The opposing view, that the volcanic intervals in the greenstone belts accumulated on continental crust (e.g., Hunter 1974b; Groves 1982; Kröner 1982), is not supported by field relationships or by geochronologic, isotopic, and sedimentologic evidence. No continental basement to the volcanic intervals in the greenstone belts has been recognized in either the Barberton Mountain Land or Pilbara Rock. Rather, geochronological data indicate that the volcanics are the oldest rocks in both areas and were underplated by tonalitic batholiths from 3.5 Ga (Barton 1983c; Barley *et al.* 1984; Jackson *et al.* 1987). Pre-3.6 Ga gneisses have been recognized locally (e.g., Kröner *et al.* 1986) but not in proximity to the greenstone belt. $^{87}Sr/^{86}Sr$ ratios from carbonates within the volcanic intervals indicate a high mantle:continental flux (Veizer *et al.* 1982) and continental detritus is absent from the volcanic intervals in both greenstone belts (Lowe 1982); these data are incompatible with the presence of extensive continental crust.

Granitoids may have been generated in the granite-greenstone terranes of the Kaapvaal Province and Pilbara Block (Fig. 19.1) as early as 3.5 Ga but were not exposed to erosion until ca. 3.3 Ga. These granitoids were uplifted together with the Onverwacht and Warrawoona volcanics (Fig. 19.16) and subordinate gneisses. The Fig Tree and Moodies Groups as well as the lower Gorge Creek Group were derived from this mixed provenance; the Fig Tree and Moodies Groups accumulated in a foredeep basin prior to 3.2 Ga. The upper Gorge Creek Group and coeval deposits developed in small strike-slip basins following stabilization of the Pilbara Block.

Siliciclastic sedimentary rocks in the Limpopo Province may be of comparable age to the oceanic volcanics in the greenstone belts. If so, this would imply that exposed land masses existed by 3.5 Ga and were subjected to weathering and erosion. In any case, the thick siliciclastic sequences in the greenstone belts provide unequivocal evidence that exposed land masses existed by 3.3 Ga. This evidence is at variance with the proposal of Hargraves (1976) that land masses became emergent only by 2.3 Ga. Geochemical data on mudstones indicate that the continental land masses exposed by 3.3 Ga were mainly tonalitic and that K-granites were a subordinate component of the pre-3.0 Ga continental crust. Early Archean (pre-3.0 Ga) sedimentation took place on the exposed land masses primarily in alluvial-fan and braided-alluvial environments as well as subaqueously in marginal-marine, shallow-marine, and deep-marine environments. Physical sedimentary processes and environments on the early Archean Earth were similar to those existing today.

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Preface

On May 8, 1986, the University of Minnesota awarded an honorary doctorate to Francis J. Pettijohn, Professor Emeritus at the Johns Hopkins University, who received his B.S., M.S., and Ph.D. degrees from Minnesota in 1924, 1925, and 1930, respectively. Present at the ceremony were some 100 researchers from the United States, Canada, and Europe, who had assembled in Minneapolis to honor Professor Pettijohn by celebrating one of his best-known and most far-reaching accomplishments—the development of basin analysis.

The forum for this celebration was a 2-day conference entitled "New Perspectives in Basin Analysis." It was clear from the range of topics discussed at the conference that basin analysis, like any area of vigorous research, has grown a good deal since its inception in the 1940s. It was equally striking that through all this growth and diversification, two of Francis Pettijohn's guiding principles have remained as cornerstones in basin analysis: 1) the importance of an integrated approach to the study of sedimentary basins; and 2) the central importance of careful field work in basin analysis.

This book is a collection of contributions based largely on the Pettijohn symposium. In assembling it, we have adopted what we hope is the spirit of the Pettijohn approach: "Consideration of the sedimentary basin as a whole provides a truly unified approach to the study of sediments" (Potter, P.E. and Pettijohn, F.J., *Paleocurrents and Basin Analysis*. New York: Springer-Verlag, 1977, p. 1). This emphasis on an integrated approach makes our division of the contents of the book into four sections somewhat arbitrary. In general, we have separated reports dealing with the use of basin sediments to infer properties of areas outside the basin (Source-Area Characterization) from those dealing mainly with internal features of basin sediments (Lithostratigraphy and Chronostratigraphy). The broadest category, Tectonics and Sedimentation, includes studies that focus on the relation of tectonic setting, subsidence pattern, and sediment supply to basin sedimentation. Finally, we have included reports on Precambrian Basins as a separate section in recognition of Francis Pettijohn's lifelong interest in the difficult, but fundamentally important, problems of basin analysis in Precambrian rocks.

We offer this book, then, not as a comprehensive survey, but as a testimonial to the breadth, vitality, and innovativeness of research on sedimentary basins today. As a pioneering researcher, Pettijohn has given us not only his own work, but also the foundation for the work that has built upon it. As a teacher, he has given us not only his enormously influential textbooks, but also a set of guiding principles that have remained at the heart of basin analysis through four decades of growth and change. We expect that another such conference 40 years hence would find them there still.

Karen L. Kleinspehn
Chris Paola

Contents